

## Chapter 3

# Numerical Weather Prediction Models

### 3.1 Summary

JMA operates NWP models to meet various kinds of requirements on weather forecasting. The suite of the NWP models covers a wide temporal range of forecast periods from a few hours to two seasons providing a seamless sequence of products for the public. The following is a brief description of the major NWP models.

1. The Global Spectral Model (GSM) produces 132-hour forecasts four times a day (00, 06, 12, 18 UTC) to support short-range forecasting (up to three days ahead) and tropical cyclone forecasts and to provide lateral boundary conditions for the Meso-Scale Model (MSM). The GSM forecast at 12 UTC is extended to 264 hours (11 days) to support one-week forecasting. The specifications of the GSM are shown in Table 3.1.1 and a description is given in Section 3.2.
2. The Global Ensemble Prediction System (GEPS) produces forecasts up to 34 days ahead to support tropical cyclone forecasts, one-week forecasts and one-month forecasts as well as issuance of Early Warning Information on Extreme Weather. In addition, the Seasonal EPS system produces seven-month forecasts to support three-month forecasts, warm- and cold-season outlooks and El Niño outlooks. The specifications and other details of the GEPS and the Seasonal EPS are outlined in Section 3.3, and the coupled atmosphere-ocean general circulation model used in the Seasonal EPS system is described in Section 3.4.
3. The MSM produces 39-hour forecasts eight times a day (00, 03, 06, 09, 12, 15, 18 and 21 UTC) to support disaster prevention, very short-range precipitation forecasts and aviation forecasts, and also provides lateral boundary conditions for the Local Forecast Model (LFM). The specifications of the MSM are shown in Table 3.1.2, and a description is given in Section 3.5.
4. The LFM produces nine-hour forecasts 24 times a day on the hour to support aviation forecasts, disaster prevention and very short-range precipitation forecasts. Its specifications are shown in Table 3.1.3, and a description is given in Section 3.6.

JMA operates a global atmospheric transport model (Section 3.7) to support its RSMC activities for nuclear environmental emergency response. The model is executed on request in coordination with the World Meteorological Organization (WMO).

JMA also operates three kinds of Chemical Transport Models (CTMs) (Section 3.8). The Aerosol CTM produces 96-hour forecast to provide Kosa (Aeolian Dust) information, the stratospheric ozone CTM produces 48-hour forecast to support UV index information, and tropospheric-stratospheric ozone CTM produces 72-hour forecast to support the photochemical oxidant information. These CTMs are operated once a day at 12UTC. The radiative transfer model is also used for UV index information.

The operational verification procedure is outlined in Section 3.9.

Table 3.1.1: Specifications of Global Spectral Model (GSM)

<b>System</b>	
Model (version)	Global Spectral Model (GSM1705)
Date of implementation	December 1987 (Lastest version: 25 May 2017)
<b>Configuration</b>	
Horizontal resolution (Grid spacing)	Spectral triangular 959 (TL959), reduced Gaussian grid system, roughly equivalent to $0.1875 \times 0.1875^\circ$ (20 km) in latitude and longitude
Vertical resolution (model top)	100 stretched sigma pressure hybrid levels (0.01 hPa)
Forecast length (initial time)	132 hours (00, 06, 18 UTC) and 264 hours (12 UTC)
Coupling to ocean / wave / sea ice models	–
Integration time step	400 seconds
<b>Initial conditions</b>	
Data assimilation	Four-dimensional variational (4D-Var) method
<b>Surface boundary conditions</b>	
Treatment of sea surface	Climatological sea surface temperature with daily analysis anomaly Climatological sea ice concentration with daily analysis anomaly
Land surface analysis	Snow depth: two-dimensional optimal interpolation scheme Temperature: first guess Soil moisture: climatology
<b>Other details</b>	
Land surface and soil	GSM land model based on the Simple Biosphere (SiB) scheme
Radiation	Two-stream with delta-Eddington approximation for short wave (hourly) Two-stream absorption approximation method for long wave (hourly)
Numerical techniques	Spectral (spherical harmonics) in horizontal, finite differences in vertical Two-time-level, semi-Lagrangian, semi-implicit time integration scheme Hydrostatic approximation
Planetary boundary layer	Hybrid scheme combining Mellor and Yamada level-2 turbulence closure with local eddy diffusivity model Similarity theory in bulk formulae for surface layer
Convection	Prognostic Arakawa-Schubert cumulus parameterization
Cloud	PDF-based cloud parameterization
Gravity wave drag	Longwave orographic drag scheme (wavelengths $> 100$ km) mainly for stratosphere Shortwave orographic drag scheme (wavelengths $\approx 10$ km) for troposphere only Non-orographic spectral gravity wave forcing scheme

Table 3.1.2: Specifications of Meso-scale Model (MSM)

<b>System</b>	
Model (version)	Meso-scale Model (forecast model: ASUCA)
Date of implementation	1 March 2001 (ASUCA: 28 February 2017)
<b>Configuration</b>	
Domain	Japan, Lambert projection, $817 \times 661$ grid points
Horizontal resolution (Grid spacing)	5 km at 60 and 30°N (standard parallels)
Vertical resolution (model top)	76 stretched height hybrid levels (21.8 km)
Forecast length (initial time)	51 hours (00, 12 UTC) and 39 hours (03, 06, 09, 15, 18, 21 UTC)
Coupling to ocean / wave / sea ice models	–
Integration time step	100/3 seconds (3-stage Runge-Kutta method)
<b>Initial conditions</b>	
Data assimilation	4D-Var analysis with mixing ratios of cloud water, cloud ice, rain, snow and graupel derived from preceding forecasts in consideration of consistency with the analysis field of relative humidity
<b>Surface boundary conditions</b>	
Sea surface temperature	Analyzed SST (fixed during time integration) and sea-ice distribution
Land surface analysis	Climatological values of evaporability, roughness length and albedo Snow cover analysis over Japan using a land surface model
<b>Lateral boundary conditions</b>	
Model providing lateral boundary conditions	GSM
Lateral boundary condition update frequency	4 times/day, 00 – 45-hour GSM forecasts initialized at 00/06/12/18 UTC for (03, 06)/09/(15, 18)/21 UTC forecasts and 00 – 57-hour GSM forecasts initialized at 06/18 UTC for 12/00 UTC forecasts
<b>Other details</b>	
Soil scheme	Ground temperature prediction using an eight-layer ground model Evaporability prediction initialized using climatological values depending on location and season
Radiation	Short wave: two-stream with delta-Eddington approximation (every 15 minutes) Long wave: two-stream absorption approximation method (every 15 minutes)
Large-scale dynamics	Finite volume method with Arakawa-C-type staggered coordinates, a horizontally explicit and vertically implicit time integration scheme, combined third- and first-order upwind finite difference schemes in flux form with a limiter as proposed by Koren (1993) in advection treatment for monotonicity, a time-splitting of vertical advection Fully compressible non-hydrostatic equations
Planetary boundary layer	Mellor-Yamada-Nakanishi-Niino Level-3 scheme Similarity theory in bulk formulae for surface layer
Convection	Kain-Fritsch convection scheme
Cloud/microphysics	Three-ice bulk cloud microphysics Consideration of PDF-based cloud distribution in microphysics Time-split treatment for rain and graupel precipitation Cloud water and cloud cover diagnosed using a partial condensation scheme
Orography	Mean orography smoothed to eliminate shortest-wave components
Horizontal diffusion	–
Gravity wave drag	–

Table 3.1.3: Specifications of Local Forecast Model (LFM)

<b>System</b>	
Model (version)	Local Forecast Model (forecast model: ASUCA)
Date of implementation	30 August 2012 (ASUCA: 29 January 2015)
<b>Configuration</b>	
Domain	Japan, Lambert projection, $1,531 \times 1,301$ grid points
Horizontal resolution (Grid spacing)	2 km at 60 and 30°N (standard parallels)
Vertical resolution (model top)	58 stretched height hybrid levels (20.2 km)
Forecast length (initial time)	10 hours (00, 01, 02, 03, 04, 05, 06, 07, 08, 09, 10, 11, 12, 13, 14, 15, 16, 17, 18, 19, 20, 21, 22, 23 UTC)
Coupling to ocean / wave / sea ice models	–
Integration time step	50/3 seconds (3-stage Runge-Kutta method)
<b>Initial conditions</b>	
Data assimilation	LA produces initial conditions via a three-hour analysis cycle based on hourly assimilation with 3D-Var and one-hour forecasts
<b>Surface boundary conditions</b>	
Sea surface temperature	Analyzed SST (fixed during time integration) and sea-ice distribution
Land surface analysis	Climatological values of evaporability, roughness length and albedo Snow cover analysis from MSM
<b>Lateral boundary conditions</b>	
Model providing lateral boundary conditions	MSM
Lateral boundary condition update frequency	8 times/day, 00 – 13-hour forecasts using the latest MSM information
<b>Other details</b>	
Soil scheme	Ground temperature prediction using an eight-layer ground model Evaporability prediction initialized using climatological values depending on location and season
Radiation	Short wave: two-stream with delta-Eddington approximation (every 15 minutes) Long wave: two-stream absorption approximation method (every 15 minutes)
Large-scale dynamics	Finite volume method with Arakawa-C-type staggered coordinates, a horizontally explicit and vertically implicit time integration scheme, combined third- and first-order upwind finite difference schemes in flux form with a limiter as proposed by Koren (1993) in advection treatment for monotonicity, a time-splitting of vertical advection Fully compressible non-hydrostatic equations
Planetary boundary layer	Mellor-Yamada-Nakanishi-Niino Level-3 scheme Similarity theory in bulk formulae for surface layer
Convection	Convective initiation
Cloud/microphysics	Three-ice bulk cloud microphysics Time-split treatment for rain and graupel precipitation Cloud water and cloud cover diagnosed using a partial condensation scheme
Orography	Mean orography smoothed to eliminate shortest-wave components
Horizontal diffusion	–
Gravity wave drag	–

## 3.2 Global Spectral Model (JMA-GSM1705)

### 3.2.1 Introduction

The Global Spectral Model (GSM) employs primitive equations to express resolvable motions and states of the atmosphere. It also incorporates sophisticated parameterization schemes for physical processes. In the horizontal, prognostic variables are spectrally discretized using triangular truncation at wave number 959 (TL959). The corresponding transform grids cover about  $0.1875^\circ$  in both longitude and latitude. In the vertical, the model has 100 layers up to 0.01 hPa.

JMA has operated GSM since March 1988. The model originally had a horizontal resolution of T63 and 16 vertical layers up to 10 hPa with a sigma coordinate system.

In a model upgrade implemented in November 1989, the truncation wave number and the number of vertical layers were increased to T106 and 21, respectively, and a hybrid  $\eta$  vertical coordinate system was adopted.

In March 1996, the horizontal resolution was doubled to T213 and the number of vertical layers was increased to 30. The cumulus parameterization was changed from a Kuo scheme to a prognostic Arakawa-Schubert scheme.

In December 1999, the parameterization schemes underwent extensive refinement. Treatment of cloud water content as a prognostic variable was introduced, and the moist convection process was improved.

In March 2001, the number of vertical layers was increased to 40 and the vertical domain was extended up to 0.4 hPa. The model was highly parallelized to suit massively distributed-memory parallel computer operation.

In February 2005, the Eulerian advection scheme was replaced with a semi-Lagrangian one, and the spectral resolution was increased from T213 (quadratic grid) to TL319 (linear grid). Incremental non-linear normal mode initialization and vertical mode initialization were also introduced.

In March 2006, operations at 06 and 18 UTC were begun with a forecast range of 36 hours in addition to those conducted at 00 UTC with a forecast range of 90 hours and 12 UTC with a forecast range of 216 hours.

In November 2007, the horizontal resolution of GSM was enhanced to TL959, while the number of vertical layers was increased to 60 and the vertical domain was extended up to 0.1 hPa (Iwamura and Kitagawa 2008; Nakagawa 2009). The numerical integration scheme was upgraded from the three-time-level leap-frog scheme to a two-time-level scheme. The forecasts run at 00, 06 and 18 UTC were altered to each cover a uniform period of 84 hours. At the same time, the 20-km-resolution Regional Spectral Model (RSM) and the 24-km-resolution Typhoon Model (TYM) were retired from operational use.

In August 2008, a reduced Gaussian grid was incorporated into GSM as a new dynamical core. This removed redundant grid points at higher latitudes, thereby saving on computational resources (Miyamoto 2006). Incremental non-linear normal mode initialization and vertical mode initialization were eliminated.

In December 2012, a relative humidity threshold was introduced to the diagnostic stratocumulus scheme (Shimokobe 2012).

In March 2013, the coverage period of the forecast run at 12 UTC was extended from 216 hours to 264 hours.

In April 2013, the radiation scheme was improved by updating the coefficients used for the short-wave parameterization of water vapor.

In March 2014, the number of vertical layers was increased to 100 and the vertical domain was extended up to 0.01 hPa. The parameterization schemes for variables such as the boundary layer, radiation, non-orographic gravity waves and deep convection were also revised (Yonehara *et al.* 2014).

In March 2016, various parameterization schemes such as deep convection, cloud, radiation, land model, and sea surface were substantially revised (Yonehara *et al.* 2017).

In May 2017, the parameterization schemes underwent extensive refinement (Yonehara *et al.* 2018).

In June 2018, the coverage period of forecasts run at 00, 06 and 18 UTC was extended from 84 to 132 hours.

### 3.2.2 Dynamics

The GSM is based on the framework of a semi-implicit semi-Lagrangian global model. In order to reduce the general shortcomings of semi-Lagrangian models (such as the lack of conservation properties and the high computational cost of three-dimensional interpolations), a vertically conservative semi-Lagrangian scheme (Yukimoto *et al.* 2011) is adopted for the GSM.

#### 3.2.2.1 Governing Equations

The GSM is run on an  $\eta$  vertical coordinate system, which is a hybrid between pressure  $p$  and  $\sigma$  ( $\sigma = p/p_S$ , where  $p_S$  is surface pressure), implicitly defined as  $p = A(\eta) + B(\eta)p_S$ .  $\eta$  ranges from 0 to 1;  $\eta = 1$  corresponds to the lower boundary (ground surface) and  $\eta = 0$  corresponds to the upper boundary. The prognostic variables (wind vector  $\mathbf{u} = (u, v)$ , temperature  $T$ , pressure  $p$ , specific humidity  $q$  and cloud water content  $q_c$ ) follow the system of primitive equations in the  $\eta$ -coordinate system as follows:

$$\frac{d\mathbf{u}}{dt} = -f\mathbf{z} \times \mathbf{u} - (\nabla\Phi + R_d T_V \nabla \ln p) + \mathbf{F}_u \quad (3.2.1)$$

$$\frac{dT}{dt} = \frac{\kappa T_V \omega}{[1 + (C_{pv}/C_{pd} - 1)q]p} + F_T \quad (3.2.2)$$

$$\frac{dq}{dt} = F_q \quad (3.2.3)$$

$$\frac{dq_c}{dt} = F_c \quad (3.2.4)$$

$$\frac{\partial}{\partial t} \left( \frac{\partial p}{\partial \eta} \right) + \nabla \cdot \left( \mathbf{u} \frac{\partial p}{\partial \eta} \right) + \frac{\partial}{\partial \eta} \left( \dot{\eta} \frac{\partial p}{\partial \eta} \right) = 0 \quad (3.2.5)$$

Here,  $d/dt$  is a total derivative defined as  $d/dt = \partial/\partial t + \mathbf{u} \cdot \nabla + \dot{\eta}\partial/\partial\eta$ , and  $\nabla$  is a horizontal gradient operator. The other notations used above are conventional:  $\mathbf{z}$  is the unit vertical vector,  $T_V$  is the virtual temperature,  $f$  is the Coriolis parameter,  $R_d$  is the gas constant for dry air, and  $\kappa = R_d/C_{pd}$ .  $C_{pd}$  is the specific heat capacity at the constant pressure of dry air and  $C_{pv}$  is the specific heat capacity at the constant pressure of water vapor.  $\mathbf{F}_u$ ,  $F_T$ ,  $F_q$  and  $F_c$  are tendencies relating to parameterized processes. In addition,  $\mathbf{F}_u$  and  $F_T$  include the effects of horizontal diffusion (to be described later). Integrating Eq. (3.2.5) with respect to  $\eta$  using the boundary conditions of  $\dot{\eta} = 0$  at  $\eta = 0$  and  $\eta = 1$ ,  $\eta$ -velocity and  $\omega$  are found:

$$\dot{\eta} \frac{\partial p}{\partial \eta} = -\frac{\partial p}{\partial t} - \int_0^\eta \nabla \cdot \left( \mathbf{u} \frac{\partial p}{\partial \eta'} \right) d\eta' \quad (3.2.6)$$

$$\omega \equiv \frac{dp}{dt} = - \int_0^\eta \nabla \cdot \left( \mathbf{u} \frac{\partial p}{\partial \eta'} \right) d\eta' + \mathbf{u} \cdot \nabla p \quad (3.2.7)$$

The geopotential  $\Phi$  is given by the following hydrostatic relation:

$$\frac{\partial \Phi}{\partial \eta} = -R_d T_V \frac{\partial \ln p}{\partial \eta} \quad (3.2.8)$$

#### 3.2.2.2 Vertical Finite Difference Scheme

The vertical finite difference scheme is coded by following Simmons and Burridge (1981). The prognostic variables  $\mathbf{u}$ ,  $T$ ,  $q$  and  $q_c$  are defined on the full levels, while  $\eta$  (including vertical fluxes) is defined on half-integer levels. Pressure on half-integer levels are expressed as

$$p_{k-1/2} = A_{k-1/2} + B_{k-1/2} p_S \quad (k = 1, 2, \dots, k_{\max}) \quad (3.2.9)$$

Here, the level index  $k$  increases with height,  $k_{\max}$  is the index of the highest model level,  $A_{k-1/2} = A(\eta_{k-1/2})$  and  $B_{k-1/2} = B(\eta_{k-1/2})$ . The profiles of  $A_{k-1/2}$  and  $B_{k-1/2}$  are determined by following Kawai (2013).  $A_{1/2}$  is set to zero so that the lowest level coincides with the ground surface, and values of  $B_{k-1/2}$  above 60hPa are set to zero so that these levels coincide with constant pressure surfaces. For intermediate levels,  $A_{k-1/2}$  and  $B_{k-1/2}$  vary smoothly with  $k$ .

From the hydrostatic relation given by Eq. (3.2.8) the finite difference form of geopotential on the full level is chosen as

$$\Phi_k = \Phi_S + \sum_{k'=1}^{k-1} R_d T_{V,k'} \ln \left( \frac{p_{k'-1/2}}{p_{k'+1/2}} \right) + \alpha_k R_d T_{V,k} \quad (3.2.10)$$

$$\alpha_k = \begin{cases} 1 - \frac{p_{k+1/2}}{\delta p_k} \ln \left( \frac{p_{k-1/2}}{p_{k+1/2}} \right) & (1 \leq k < k_{\max}) \\ \ln 2 & (k = k_{\max}) \end{cases} \quad (3.2.11)$$

Here,  $\Phi_S$  is the geopotential at the surface, and  $\delta p_k = p_{k-1/2} - p_{k+1/2}$ . The pressure gradient force term in Eq. (3.2.1) and the adiabatic heating rate term in Eq. (3.2.2) can then be written in discretized form as

$$(\nabla \Phi + R_d T_V \nabla \ln p)_k = \nabla \Phi_k + \frac{R_d T_{V,k}}{\delta p_k} \left[ \ln \left( \frac{p_{k-1/2}}{p_{k+1/2}} \right) \nabla p_{k+1/2} + \alpha_k \nabla (\delta p_k) \right] \quad (3.2.12)$$

and

$$\left[ \frac{\kappa T_V}{C_p / C_{pd}} \frac{\omega}{p} \right]_k = \frac{\kappa T_{V,k}}{C_{p,k} / C_{pd}} \frac{1}{\delta p_k} \left[ \left( \ln \frac{p_{k-1/2}}{p_{k+1/2}} \right) \left( B_{k+1/2} \mathbf{u}_k \cdot \nabla p_S - \sum_{l=k+1}^{k_{\max}} \nabla \cdot (\mathbf{u}_l \delta p_l) \right) - \alpha_k (\nabla \cdot \mathbf{u}_k) \delta p_k \right] \quad (3.2.13)$$

respectively, where  $C_p$  is the specific heat capacity at the constant pressure of moist air, that defined as  $C_p = [1 + (C_{pv}/C_{pd} - 1)q]C_{pd}$ . The vertical mass flux in Eq. (3.2.6) is discretized as

$$\left( \dot{\eta} \frac{\partial p}{\partial \eta} \right)_{k-1/2} = -B_{k-1/2} \frac{\partial p_S}{\partial t} - \sum_{l=k}^{k_{\max}} \nabla \cdot (\mathbf{u}_l \delta p_l) = B_{k-1/2} \sum_{l=1}^{k_{\max}} \nabla \cdot (\mathbf{u}_l \delta p_l) - \sum_{l=k}^{k_{\max}} \nabla \cdot (\mathbf{u}_l \delta p_l) \quad (3.2.14)$$

### 3.2.2.3 Horizontal Grid

To mitigate the overconcentration of grid points at high latitudes and lower the computational cost, a reduced Gaussian grid is adopted for the GSM. The number of east-west grid points at each latitude is determined based on the magnitude of associated Legendre functions, which is negligibly small at high latitudes and in high orders. With this method, the computational cost of Legendre transformation can also be reduced (Juang 2004). The number of east-west grid points is in fact restricted by FFT package specifications, the number of east-west decompositions in parallelization (as described in 3.2.10) and the interval of coarser radiation grids (as shown in 3.2.3).

### 3.2.2.4 Semi-implicit Semi-Lagrangian Formulation

Prior to the time integration, the forecast equations (Eq. (3.2.1) - Eq. (3.2.5)) are rewritten in the form of  $d_H X / dt = \partial X / \partial t + \mathbf{u} \cdot \nabla X = R$  with vertical advection terms incorporated into  $R$  on the right-hand side. These equations are integrated with respect to time along the trajectory of the parcel from the departure point  $D$  at time  $t$  to the arrival point  $A$  at time  $t + \delta t$ . The linear term  $L$  separated from the forcing term  $R$  is treated semi-implicitly (i.e. using a trapezoidal rule), and the remaining  $R$ , including vertical advection terms, are treated with spatial averaging (Tanguay *et al.* 1992).

The resulting linear terms are slightly amplified by the factor  $\beta = 1.2$  for computational stability, and the following is obtained:

$$X^{A+} - X^{D0} = \delta t \frac{R^{A0} + R^{D(+)}}{2} + \delta t \beta \left[ \frac{L^{A+} + L^{D-}}{2} - \frac{L^{A0} + L^{D0}}{2} \right] \quad (3.2.15)$$

Superscript  $A$  represents the arrival point  $\mathbf{x}_{ij}$  assumed to be on the Gaussian grid, and  $D$  is the departure point  $\mathbf{x}_{ij} - \boldsymbol{\alpha}$  (the displacement vector  $\boldsymbol{\alpha}$ , whose calculation will be described later). The abbreviations used above are the same as those for  $X^{A+} = X(\mathbf{x}, t + \delta t)$ ,  $X^{D0} = X(\mathbf{x} - \boldsymbol{\alpha}, t)$ ,  $R^{A0} = R(\mathbf{x}, t)$ ,  $R^{D(+)} = R(\mathbf{x} - \boldsymbol{\alpha}, t + \delta t)$  and others.  $R^{D(+)}$  is calculated based on extrapolation with respect to time. Rearranging the terms of the above equations gives a system of linear equations for the unknown values  $X^{A+}$ :

$$X^{A+} - \frac{\beta \delta t}{2} L^{A+} = \left[ X^0 + \frac{\delta t}{2} \left\{ R^{(+)} - \beta (L^0 - L^-) \right\} \right]^D + \frac{\delta t}{2} [R^0 - \beta L^0]^A \quad (3.2.16)$$

### 3.2.2.5 Vertically Conservative Semi-Lagrangian Scheme

Yoshimura and Matsumura (2003, 2004) developed a vertically conservative semi-Lagrangian scheme in which vertical advection is treated separately from horizontal advection so that conserved quantities such as water vapor under non-dissipative conditions are preserved in the vertical direction. Processing advection separately in the horizontal and vertical directions also reduces the model's cost of interpolation.

Eq. (3.2.16) can be reformulated with flux forms appropriate for a scheme in which vertical advection can retain conservative properties. Beginning with Eq. (3.2.5) and Eq. (3.2.1) - Eq. (3.2.4), rewriting is performed as follows:

$$\frac{d_H}{dt} \frac{\partial p}{\partial \eta} = -D \frac{\partial p}{\partial \eta} - \frac{\partial}{\partial \eta} \left( \dot{\eta} \frac{\partial p}{\partial \eta} \right) \quad (3.2.17)$$

$$\frac{d_H}{dt} \left( X \frac{\partial p}{\partial \eta} \right) = -DX \frac{\partial p}{\partial \eta} - \frac{\partial}{\partial \eta} \left( \dot{\eta} X \frac{\partial p}{\partial \eta} \right) + R_X \frac{\partial p}{\partial \eta} \quad (3.2.18)$$

Here,  $X$  represents  $\mathbf{u}$ ,  $T_V$ ,  $q$  and  $q_c$ , and  $R_X = dX/dt$ . The parallel nature of these equations is easily recognizable. The first term on the right hand side of these equations represents the increase caused by horizontal convergence, and the second term is the increase caused by vertical flux convergence. With respect to the latter, where  $q$  and  $q_c$  being conservative when  $R_X = 0$ , devising a vertically integrated quantity that remains unchanged in vertical advection appears to be a promising approach. A simple outline of the procedure is given here for specific humidity  $q$  without  $R_q$ .

Vertical discretization and time integration during the period  $\delta t$  described earlier give the following equations with the omission of terms related to the semi-implicit method for reasons of simplicity:

$$\begin{aligned} (\delta p_k)^{A+} &= \left[ (\delta p_k)^0 - \frac{1}{2} (D_k \delta p_k)^{(+)} \delta t + \frac{1}{2} \left\{ \left( \dot{\eta} \frac{\partial p}{\partial \eta} \right)_{k+1/2} - \left( \dot{\eta} \frac{\partial p}{\partial \eta} \right)_{k-1/2} \right\}^{(+)} \delta t \right]^D \\ &\quad + \left[ -\frac{1}{2} (D_k \delta p_k)^0 \delta t + \frac{1}{2} \left\{ \left( \dot{\eta} \frac{\partial p}{\partial \eta} \right)_{k+1/2} - \left( \dot{\eta} \frac{\partial p}{\partial \eta} \right)_{k-1/2} \right\}^0 \delta t \right]^A \end{aligned} \quad (3.2.19)$$

$$\begin{aligned} (q_k \delta p_k)^{A+} &= \left[ (q_k \delta p_k)^0 - \frac{1}{2} q_k^0 (D_k \delta p_k)^{(+)} \delta t + \frac{1}{2} \left\{ \left( q \dot{\eta} \frac{\partial p}{\partial \eta} \right)_{k+1/2} - \left( q \dot{\eta} \frac{\partial p}{\partial \eta} \right)_{k-1/2} \right\}^{(+)} \delta t \right]^D \\ &\quad + \left[ -\frac{1}{2} q_k^0 (D_k \delta p_k)^0 \delta t + \frac{1}{2} \left\{ \left( q \dot{\eta} \frac{\partial p}{\partial \eta} \right)_{k+1/2} - \left( q \dot{\eta} \frac{\partial p}{\partial \eta} \right)_{k-1/2} \right\}^0 \delta t \right]^A \end{aligned} \quad (3.2.20)$$

$$p_{k-1/2} = \sum_{k'=k}^{\text{kmax}} \delta p_{k'}, \quad (k = 1, 2, \dots, \text{kmax}) \quad (3.2.21)$$

Here, the vertically cumulative quantity  $Q$  is defined as follows:

$$Q_{k-1/2} = \sum_{k'=k}^{\text{kmax}} \delta Q_{k'}, \quad \delta Q_k = q_k \delta p_k, \quad (k = 1, 2, \dots, \text{kmax} + 1) \quad (3.2.22)$$

Eq. (3.2.20) rewritten for  $\delta Q_k$  is found to be similar to Eq. (3.2.19) for  $\delta p_k$ , and there is a clear correspondence between  $Q$  and  $p$ . Computation of  $Q$  can therefore be carried out in the five steps outlined below in a fashion parallel to that of  $p$ . The first two steps concern the operations inside the square brackets  $[\dots]^D$  in the above equations. The third step involves the calculation of variables at departure points based on interpolation. The fourth and the fifth steps are similar to the first two, but for the operations in the square brackets  $[\dots]^A$ .

1. First step: Horizontal divergence is calculated. As the mass of each layer  $\delta p_k$  varies to  $\delta p'_k$ , the half-level pressure values  $p_{k-1/2}$  by which layers are bound also shift to  $p'_{k-1/2}$ , which can be computed using Eq. (3.2.21). The values of  $q_k$  remain constant under the horizontal convergence  $q'_k = q_k$ .
2. Second step: Vertical flux convergence is calculated using Eq. (3.2.14) as in the Eulerian scheme. In the same way as in the first step,  $\delta p'_k$  varies to  $\delta p''_k$ , and the values of  $p'_{k-1/2}$  shift to  $p''_{k-1/2}$  except  $k = 1$  ( $p'_{1/2} = p''_{1/2}$ ). In this step, the shift in  $Q'_{k-1/2}$  caused by the vertical flux convergence is computed based on interpolation from  $Q'_{k-1/2}(p'_{k-1/2})$  using  $Q''_{k-1/2} = Q'_{k-1/2}(p''_{k-1/2})$ . This procedure ensures the conservation of the total mass-weighted integral  $Q'_{1/2} = Q''_{1/2}$ , because  $p'_{1/2} = p''_{1/2}$  holds and the other values of  $p''_{k-1/2}$  ( $k = 2, 3, \dots, \text{kmax}$ ) merely have their intervals changed in the vertical column. New values of  $q''_k$  are computed using  $\delta Q''_k$  and  $\delta p''_k$  with Eq. (3.2.22).
3. Third step: Horizontal advection is incorporated by computing  $(\delta p_k)^D$  and  $q_k^D$  via quasi-cubic interpolation.
4. Forth step: Vertical flux convergence is calculated at the arrival point via the second step.
5. Fifth step: Horizontal divergence is calculated at the arrival point via the first step.

The time-integration of  $q$  and  $q_c$  is completed based on these five steps, and that of  $\mathbf{u}$ ,  $T_V$  and  $p_S$  is followed by the semi-implicit calculation shown in Eq. (3.2.16).

### 3.2.2.6 Departure Point Determination

The displacement vector  $\boldsymbol{\alpha}$  (as yet undetermined) obeys the implicit equation

$$\boldsymbol{\alpha} = \delta t \left\{ \frac{\mathbf{u}_k(\mathbf{x}_{ij} - \boldsymbol{\alpha}, t + \delta t) + \mathbf{u}_k(\mathbf{x}_{ij}, t)}{2} \right\} \quad (3.2.23)$$

which expresses that the horizontal advection during the time interval  $\delta t$  is related to the average of future time-step wind value at the departure point and current time-step wind value at the arrival point (SETTLS; Hortal 2002). To improve stability, a method based on wind integrated in a semi-Lagrangian scheme rather than the time extrapolated wind is adopted (Yoshimura 2002). This implicit equation is solved by successive insertions of  $\boldsymbol{\alpha}$ . For the computation of these vector components, it is considered that the axes of the local coordinates ( $\lambda, \varphi$ ) rotate due to the spherical metric as a parcel advances along a trajectory, as is the case whenever horizontal vector components are interpolated on a sphere. The wind at the departure point is computed from linear interpolation except for the last third of the iteration, for which a quasi-cubic approach is used.

### 3.2.2.7 Spectral Method and Horizontal Diffusion

Spectral variables (i.e. vorticity  $\zeta (= \mathbf{z} \cdot \nabla \times \mathbf{u})$ , divergence  $D (= \nabla \cdot \mathbf{u})$ ,  $T_V$  and  $\ln(p_S)$ ) are expanded in terms of spherical harmonics with triangular truncation. In accordance with the framework of the semi-Lagrangian scheme, a linear Gaussian transformation grid is used. Solutionis of horizontal Helmholtz equations (derived when Eq. (3.2.16) is solved for  $D$ ), horizontal diffusion and variables such as the differentials on the sphere are

calculated using the spectral method (Bourke 1974; Hoskins and Simmons 1975). The remaining variables  $q$  and  $q_c$  are defined only on grid points.

To prevent the accumulation of small scale noise (spectral blocking), fourth-order linear horizontal diffusion is applied to  $\zeta$ ,  $D$  and  $T_V$ :

$$\left( \frac{\partial \zeta}{\partial t} \right)_{\text{hdif},4\text{th}} = -K_{4\text{th}} \left( \nabla^4 - \frac{4}{a^4} \right) \zeta \quad (3.2.24\text{a})$$

$$\left( \frac{\partial D}{\partial t} \right)_{\text{hdif},4\text{th}} = -K_{4\text{th}} \nabla^4 D \quad (3.2.24\text{b})$$

$$\left( \frac{\partial T_V}{\partial t} \right)_{\text{hdif},4\text{th}} = -K_{4\text{th}} \nabla^4 \left[ T_V - \frac{\partial \bar{T}_V}{\partial \bar{p}} p \right] = -K_{4\text{th}} \nabla^4 \left[ T_V - \frac{\partial \bar{T}_V}{\partial \bar{p}} B(\eta) p_S \right] \quad (3.2.24\text{c})$$

Here,  $K_{4\text{th}}$  is the diffusion coefficient for the fourth-order horizontal diffusion and  $a$  is the radius of the earth. Bars over variables indicate the global average on the  $\eta$ -surface. Angular momentum conservation does not allow the horizontal diffusion process to work on vorticity with total wave number 1 as shown by Eq. (3.2.24a). Diffusion for virtual temperature is modified to work on the constant pressure surface; otherwise, diffusion on a declining  $\eta$ -surface may produce spurious mixing along steep mountain slopes.  $K_{4\text{th}}$  is chosen so that the power spectrum of enstrophy coincides with that expected based on the two-dimensional turbulence theory.

To provide a sponge layer that absorbs waves incident on the upper boundary, second-order linear horizontal diffusion is applied to the divergence term  $D$  in layers above 30hPa:

$$\left( \frac{\partial D}{\partial t} \right)_{\text{hdif},2\text{nd}} = -K_{2\text{nd}} \nabla^2 D \quad (3.2.25\text{a})$$

$$K_{2\text{nd}} = K_0 \sin^2 \left( \frac{\pi}{2} \frac{\ln p - \ln p_{\text{btm}}}{\ln p_{\text{top}} - \ln p_{\text{btm}}} \right) \quad (3.2.25\text{b})$$

Here,  $K_0$  is the base diffusion coefficient for second-order horizontal diffusion,  $p_{\text{top}}$  is the pressure at the highest model level (0.01hPa), and  $p_{\text{btm}}$  is the pressure at the altitude where the sponge layer begins (30hPa). To suppress wave reflection at the upper boundary,  $K_{2\text{nd}}$  is gradually enhanced with height as shown by Eq. (3.2.25b).  $K_0$  is determined experimentally so that spurious wave reflections at the upper boundary can be appropriately removed.

These fourth and second-order horizontal diffusion terms are calculated backward and implicitly in spectral forms as an independent step after semi-implicit time integration.

### 3.2.3 Radiation

The radiative heating rate is computed as the divergence of net radiation fluxes:

$$\left( \frac{\partial T}{\partial t} \right)_{\text{rad}} = \frac{g}{C_p} \frac{\partial F}{\partial p} \quad (3.2.26)$$

where  $F (= F^\uparrow - F^\downarrow)$  is the net radiation flux,  $F^\uparrow$  ( $F^\downarrow$ ) is the upwelling (downwelling) radiation flux,  $g$  is the acceleration of gravity and  $C_p$  is the specific heat at the constant pressure of moist air.

Solving the radiative transfer equation is computationally very expensive. To reduce this burden, full radiation computation is performed only once an hour for longwave and shortwave on a coarser (reduced radiation) grid. The radiative heating rates associated with longwave and shortwave radiation are corrected for every time step using the surface temperature and the solar zenith angle, respectively.

#### 3.2.3.1 Longwave Radiation

The two-stream radiation transfer method involving the absorption approximation approach (Yabu 2013) is adopted for longwave flux and cooling rate computation. The spectrum in the longwave region is divided into 11 bands as shown in Table 3.2.1.

Table 3.2.1: Band configuration for the longwave radiation scheme. The calculation approaches for absorption associated with atmospheric molecules are C-k (correlated  $k$ -distribution method) and S-k ( $k$ -distribution method with scaling approximation). Notation in each parenthesis denotes gas overlap assumption (pf: perfectly correlated; no: perfectly uncorrelated; pt: partly correlated; cg: combined gas). The number of sub-bands is also shown in each parenthesis.

Band Number	1	2	3a	3b	3c	4	5	6	7	8	9
Wavenumber (/cm)	25-340	340-540	540-620	620-720	720-800	800-980	980-1100	1100-1215	1215-1380	1380-1900	1900-3000
Major absorption gas											
H <sub>2</sub> O(Line)	C-k(16)			C-k(16,cg)		S-k(6)	S-k(6,pf)	S-k(16,pf)	S-k(4)	C-k(16)	S-k(6)
CO <sub>2</sub>											
O <sub>3</sub>							C-k(16)	C-k(16)			
H <sub>2</sub> O(Continuum)				S-k(16,pt)		S-k(6,pf)		S-k(16,pf)	S-k(4,pf)	S-k(16,pt)	S-k(6,pf)
Minor absorption gas											
CO <sub>2</sub>						S-k(6,pf)					
N <sub>2</sub> O			S-k(6,pf)						S-k(2,no)		
CH <sub>4</sub>							S-k(16,pf)	S-k(2,no)			
CFC-11,CFC-12,HCFC-22						S-k(6,pf)					
Number of sub-bands	16	16	16	16	16	6	16	16	16 (=4x2x2)	16	6

In the two-stream method, downwelling (upwelling) radiation fluxes are calculated sequentially from the model top (surface) as follows:

$$F_{k-1/2}^{\downarrow} = 0 \quad (k = k_{max} + 1) \quad (3.2.27a)$$

$$F_{k-1/2}^{\downarrow} = F_{k+1/2}^{\downarrow} e^{-\tau_k/\mu} + B_k^{\downarrow} (1 - e^{-\tau_k/\mu}) \quad (k = k_{max}, \dots, 1) \quad (3.2.27b)$$

$$F_{k-1/2}^{\uparrow} = B_s \quad (k = 1) \quad (3.2.27c)$$

$$F_{k-1/2}^{\uparrow} = F_{k-3/2}^{\uparrow} e^{-\tau_{k-1}/\mu} + B_{k-1}^{\uparrow} (1 - e^{-\tau_{k-1}/\mu}) \quad (k = 2, \dots, k_{max} + 1) \quad (3.2.27d)$$

where  $F_{k-1/2}^{\downarrow}$  ( $F_{k-1/2}^{\uparrow}$ ) is the downwelling (upwelling) radiation flux at the half-integer level  $k - 1/2$ ,  $\tau_k$  is the optical thickness of the model layer  $k$ ,  $B_k^{\downarrow}$  ( $B_k^{\uparrow}$ ) is the downward (upward) effective Planck flux (Chou *et al.* 2001) at the model layer  $k$ ,  $B_s$  is the Planck flux at the surface and  $1/\mu$  (= 1.66) is the diffusivity factor.

Depending on the absorber and the spectral band, absorption associated with atmospheric molecules is evaluated using one of two  $k$ -distribution methods (see Table 3.2.1). The correlated  $k$ -distribution method (Fu and Liou 1992) is applied to absorption dominant in the middle atmosphere. Absorption coefficients at 51 pressure levels between 1000 and 0.01 hPa are tabulated in advance based on the Line-By-Line calculation, and gas absorption data are derived from HITRAN2000 (Rothman *et al.* 2003). The  $k$ -distribution method with scaling approximation (Chou *et al.* 2001) is applied to absorption with a Lorentzian line shape assumed, and only one absorption coefficient at a specified pressure level (500 hPa) is prepared. This method is also applied to H<sub>2</sub>O continuum absorption based on the MT-CKD model with scaling parameters from Zhong and Haigh (1995).

To handle the overlapping of gas absorption within each band, one of three assumptions (perfectly correlated, perfectly uncorrelated and partly correlated) (Zhang *et al.* 2003) is made other than that for CO<sub>2</sub> and H<sub>2</sub>O in the 540–800 cm<sup>-1</sup> region, which is handled using the direct combined gas mapping approach of Li and Barker (2005) (see Table 3.2.1).

Maximum-random cloud overlapping (Geleyn and Hollingsworth 1979) is assumed in the longwave radiation scheme, and is implemented to the two-stream method as per Li (2002). Cloud optical thickness is parameterized as per Lindner and Li (2000) for liquid droplets and as per Ebert and Curry (1992) for ice particles.

### 3.2.3.2 Shortwave Radiation

Shortwave scattering and absorption are modeled in two-stream formulation using the delta-Eddington approximation (Joseph *et al.* 1976; Coakley *et al.* 1983). The spectrum in the shortwave region is divided into 16 bands

(10 in ultraviolet, 5 in visible and 1 in near-infrared) based on Freidenreich and Ramaswamy (1999), while absorption by water vapor in the near-infrared region is calculated via exponential-sum fitting of transmissions method with seven sub-bands based on Collins *et al.* (2006).

Assuming a plane-parallel atmosphere, diffuse radiation fluxes are derived from the following simultaneous equations (Meador and Weaver 1980):

$$\frac{dF^\uparrow}{d\tau} = \gamma_1 F^\uparrow - \gamma_2 F^\downarrow - \gamma_3 \omega_0 S_0 e^{-\tau/\mu_0} \quad (3.2.28a)$$

$$\frac{dF^\downarrow}{d\tau} = \gamma_2 F^\uparrow - \gamma_1 F^\downarrow + (1 - \gamma_3) \omega_0 S_0 e^{-\tau/\mu_0} \quad (3.2.28b)$$

where  $\tau$  is the optical thickness,  $\omega_0$  is the single scattering albedo and  $S_0$  is the incident solar irradiance in the direction  $\mu_0$  (the cosine of the solar zenith angle). The coefficients  $\gamma_i$  ( $i = 1, \dots, 3$ ) are given by

$$\gamma_1 = \frac{1}{4}[7 - \omega_0(4 + 3g)], \quad \gamma_2 = -\frac{1}{4}[1 - \omega_0(4 - 3g)], \quad \gamma_3 = \frac{1}{4}(2 - 3g\mu_0) \quad (3.2.29)$$

where  $g$  is the asymmetry factor. In the delta-Eddington method, solar optical properties such as  $\tau$ ,  $\omega_0$  and  $g$  are adjusted using the fraction of forward-scattering peak  $f$ :

$$\tau' = (1 - \omega_0 f)\tau, \quad \omega'_0 = \frac{(1 - f)\omega_0}{1 - \omega_0 f}, \quad g' = \frac{g - f}{1 - f} \quad (3.2.30)$$

The shortwave radiation flux in each column is calculated using the Practical Independent Column Approximation (PICA; Nagasawa 2012) method, which is a simplified ICA approach based on Collins (2001). The total shortwave radiation flux  $F$  is given as a weighted average of the fluxes in each subcolumn as follows:

$$F = \frac{1}{A_{tot}} \sum_{i=1}^{N_{max}} A_i F_i \quad (3.2.31)$$

where  $A_{tot}$  is the total area of the relevant subcolumns,  $A_i$  is the fractional area of each subcolumn,  $F_i$  is the flux in each subcolumn and  $N_{max}$  is the maximum number of subcolumns. The binary cloud configuration in the column is given by cloud cover assuming maximum-random cloud overlapping.

The reflectance and transmittance of the layer are calculated as functions of the total optical thickness  $\tau_{total}$ , the total single scattering albedo  $\omega_{0\ total}$  and the total asymmetry factor  $g_{total}$  of the layer:

$$\tau_{total} = \tau_R + \tau_g + \tau_a + \tau_c \quad (3.2.32a)$$

$$\omega_{0\ total} = \frac{\tau_R + \omega_{0a}\tau_a + \omega_{0c}\tau_c}{\tau_R + \tau_g + \tau_a + \tau_c} \quad (3.2.32b)$$

$$g_{total} = \frac{g_a\omega_{0a}\tau_a + g_c\omega_{0c}\tau_c}{\tau_R + \omega_{0a}\tau_a + \omega_{0c}\tau_c} \quad (3.2.32c)$$

where the subscripts  $R$ ,  $g$ ,  $a$  and  $c$  denote molecular Rayleigh scattering, gaseous absorption, and Mie scattering/absorption caused by aerosols and clouds, respectively.

The cloud optical properties are parameterized following Dobbie *et al.* (1999) for liquid droplets and Ebert and Curry (1992) for ice particles.

### 3.2.3.3 Gas Concentrations and Aerosol Climatology

The radiatively active gases considered in the radiation scheme are water vapor, ozone, carbon dioxide, oxygen, methane, nitrous oxide, CFC-11, CFC-12 and HCFC-22. Prognostic water vapor concentrations are used for the troposphere, while climatological distribution based on Randel *et al.* (1998) is used for areas above it. The three-dimensional monthly mean climatology of ozone concentration is derived from stratospheric ozone Chemical Transport Model (CTM) calculation (see Subsection 3.8.4). Other radiatively active gases have globally uniform concentrations as shown in Table 3.2.2.

Table 3.2.2: Values of the globally uniform gas concentration (unit is ppmv).

CO <sub>2</sub>	O <sub>2</sub>	CH <sub>4</sub>	N <sub>2</sub> O	CFC-11	CFC-12	HCFC-22
396	209490	1.824	0.3259	0.0003	0.0005	0.0002

In the radiation scheme, five aerosol types (sulfate, black carbon, organic carbon, sea salt and mineral dust) are considered (Yabu *et al.* 2017). The three-dimensional monthly mean climatology of aerosol mass concentration is derived from aerosol CTM calculation (see Subsection 3.8.3), and the optical properties of these aerosols are pre-computed via Mie scattering calculation. The climatological distribution of total-column values for aerosol optical depth (ATOD) is used in combination. Monthly mean ATOD climatology is derived from Moderate Resolution Imaging Spectroradiometer (MODIS), Multi-angle Imaging Spectro-Radiometer (MISR) and Ozone Monitoring Instrument (OMI) observations.

### 3.2.3.4 Cloud Properties

Two types of cloud are considered in the radiation scheme. One is stratiform cloud, whose properties (such as cloud cover, cloud water content and cloud ice content) are provided by the cloud scheme. The other is convective cloud, whose properties are diagnosed using the upward convective mass flux calculated in the convection scheme. The effective radius of cloud liquid droplets  $r_{e,liq}$  [ $\mu\text{m}$ ] is parameterized based on Martin *et al.* (1994) as follows:

$$r_{e,liq} = 10^4 \left( \frac{3CWC}{4\pi\rho_w k N_{TOT}} \right)^{1/3} \quad (3.2.33)$$

where CWC is the cloud water content [ $\text{gm}^{-3}$ ],  $N_{TOT}$  is the number concentration of water cloud droplets [ $\text{cm}^{-3}$ ],  $\rho_w$  is the density of water [ $\text{gm}^{-3}$ ] and  $k$  is a constant.  $N_{TOT}$  is set as 100 and  $300 \text{ cm}^{-3}$  over ocean and land areas, respectively. The effective radius of cloud ice particles  $r_{e,ice}$  [ $\mu\text{m}$ ] depends on temperature  $T$  [K] and cloud ice content IWC [ $\text{gm}^{-3}$ ] as follows (Wyser 1998):

$$\begin{aligned} B &= -2 + 10^{-3}(273 - T)^{1.5} \log_{10} \frac{\text{IWC}}{\text{IWC}_0} \\ r_{e,ice} &= 377.4 + 203.3B + 37.91B^2 + 2.3696B^3 \end{aligned} \quad (3.2.34)$$

where  $\text{IWC}_0$  ( $= 50 \text{ gm}^{-3}$ ) is a constant.

### 3.2.4 Cumulus Convection

The GSM employs a spectral mass-flux convective parametrization scheme based on Arakawa and Schubert (1974) and Moorthi and Suarez (1992). Prognostic closure based on Randall and Pan (1993) is used, although many modifications are made to the original. In addition, a triggering mechanism based on the DCAPE concept (Xie and Zhang 2000) is adopted to suppress excessive convective activity. Convective downdraft, convective momentum transport and mid-level convection are also included in the scheme.

#### 3.2.4.1 Convective Effect on Large-scale Variables

The convective effect on large-scale variables is estimated using the following equations:

$$\left( \rho \frac{\partial \bar{s}}{\partial t} \right)_{\text{conv}} = \sum_n D_n^u (s_n^u - \bar{s}) + D^d (s^d - \bar{s}) + \left( \sum_n M_n^u - M^d \right) \frac{\partial \bar{s}}{\partial z} - L_i c - L_v e - \delta E_s \quad (3.2.35)$$

$$\left( \rho \frac{\partial \bar{h}}{\partial t} \right)_{\text{conv}} = \sum_n D_n^u (h_n^u - \bar{h}) + D^d (h^d - \bar{h}) + \left( \sum_n M_n^u - M^d \right) \frac{\partial \bar{h}}{\partial z} - L_i c - \delta E_h \quad (3.2.36)$$

where  $\rho$  is the density of air,  $s$  and  $h$  are dry and moist static energy,  $D$  is detrainment to the environment,  $L_v$  and  $L_i$  are the latent heat of vaporization and melting,  $e$  is the evaporation amount under the convective cloud

base, and  $c$  is the melt amount. Details of  $\delta E_s$  and  $\delta E_h$  are provided in Subsection 3.2.4.2. The over-bar denotes an environmental value, the superscripts  $u$  and  $d$  indicate updraft and downdraft respectively, and the subscript  $n$  indicates each plume of an ensemble of convection. Multiple plumes are considered for updraft, and a single plume is calculated for downdraft.

The first terms on the right of Eqs. (3.2.35) and (3.2.36) represent detrainment from updraft to the environment, the second terms indicate detrainment from downdraft to the environment, the third represent compensating subsidence, and the fourth represent the melting effect below freezing level. The fifth term in Eq. (3.2.35) denotes the evaporation effect under the cloud base.

### 3.2.4.2 Cloud Model

Based on Arakawa and Schubert (1974), the ensemble effect of multiple cumulus types is considered.

The mass flux of each plume is expressed as

$$M_n = M_B(t)\eta_n(z) \quad (3.2.37)$$

where  $M_B$  is the mass flux at the cloud base and  $\eta$  is the normalized cloud mass flux, which is 1 at the cloud base. The details of  $M_B$  calculation are provided in Subsection 3.2.4.3.

The cloud base of each plume is fixed near 900 hPa in the model. Each type of cumulus is defined by the level of the cloud top, where the updraft cloud mass loses buoyancy and detrainment occurs. The vertical profile of the upward mass flux  $\eta$  is assumed to be a linear function of height  $z$ , as proposed by Moorthi and Suarez (1992), and can be expressed as

$$\eta_n = 1 + \lambda_n(z - z_b) \quad (3.2.38)$$

where  $\lambda$  is the entrainment rate and  $z_b$  is the cloud base height.  $\lambda$  is diagnosed using a condition in which each plume loses buoyancy at its cloud top.

Cloud water content in the updraft is converted to precipitation, and the conversion is formulated as an autoconversion scheme as proposed by Kessler (1969).

The mass flux below the cloud base is calculated based on Jakob and Siebesma (2003):

$$\frac{\partial \eta}{\partial z} = \frac{C}{z}\eta \quad (3.2.39)$$

where  $C$  is a constant set to 0.5.

The plume ascends with the entrainment rate obtained from Eq. (3.2.39) below the cloud base. Assuming the occurrence of convection to be associated with positive subgrid scale fluctuations of temperature and moisture, air with higher dry and moist static energy than the grid-mean environment entrains into the plume.  $\delta E_s$  and  $\delta E_h$  in Eqs. (3.2.35) and (3.2.36) represent the entrainment of excess energy to the updraft below the cloud base.

### 3.2.4.3 Closure

Closure is based on Randall and Pan (1993), and numerous modifications are made to the original scheme. For deep convection, the following prognostic equation is used to calculate the upward mass flux at the cloud base  $M_B$  for each plume (the subscript  $n$  is omitted for simplicity):

$$\frac{dM_B}{dt} = \max\left(\frac{A - fA_0}{2\alpha}, 0\right) \min\left(\frac{\lambda}{\lambda_{\min}}, 1\right) \max(\lambda_{\max}, 0) \left(\frac{\Delta p}{\Delta p_{\text{eff}}}\right) - \frac{M_B}{2\tau_d} \quad (3.2.40)$$

where  $A$  denotes the cloud work function,  $A_0$  is the average of observed cloud work functions as given by Lord and Arakawa (1980),  $\Delta p$  is the depth of model cloud top layer,  $\Delta p_{\text{eff}}$  is the effective depth of the cloud top, and  $\tau_d$  is the time constant of cumulus kinetic energy decay. The parameter  $f$  is introduced to incorporate the effects of grid-scale vertical wind and convective inhibition. This is given by

$$f = \frac{\omega}{\omega_0} + \frac{A_i}{A_{i0}} + c \quad (3.2.41)$$

where  $\omega$  denotes the vertical pressure velocity at the lowest level,  $A_i$  represents the work involved in lifting the parcel to the level of free convection, and  $\omega_0$ ,  $A_{i0}$  and  $c$  are empirically determined constants. The constraint  $0 \leq f \leq 2$  is imposed to ensure realistic tendency of  $M_B$ . In order to suppress tall cumuli in dry conditions and incorporate the effects of turbulence in the planetary boundary layer, the parameter  $\lambda_{\min}$  is defined as follows:

$$\lambda_{\min} = \max \left( \frac{0.9 - \text{RH}}{0.2}, 10^{-3} \right) \frac{0.3}{5l_0} \quad (3.2.42)$$

where RH denotes the vertical mean of relative humidity between the cloud base and the cloud top, and  $l_0$  represents the mixing length of the planetary boundary layer. The parameter  $\lambda_{\max}$  is introduced to suppress tall cumuli with unnaturally large entrainment rates, and is defined as

$$\lambda_{\max} = \min \left( \frac{\lambda - \lambda_2}{\lambda_1 - \lambda_2}, 1 \right) \quad (3.2.43)$$

where  $\lambda_1 = a_1 / (z_t - z_b)$ ,  $\lambda_2 = a_2 / (z_t - z_b)$ ,  $z_t$  is the cloud top height, and  $a_1$  and  $a_2$  are empirically determined constants.

For shallow convection, a simplified version of Eq. (3.2.40) is adopted for closure.

#### 3.2.4.4 Triggering Mechanism

The convective triggering mechanism proposed by Xie and Zhang (2000) known as the dynamic CAPE generation rate (DCAPE) is used in the cumulus parameterization. DCAPE is defined as follows:

$$\text{DCAPE} = (\text{CAPE}(T^*, q^*) - \text{CAPE}(T, q)) / \Delta t \quad (3.2.44)$$

where  $T$  is the temperature,  $q$  is the specific humidity, and  $(T^*, q^*)$  are  $(T, q)$  plus the change caused by overall large-scale advection over a certain time period  $\Delta t$  (the integration time step used in the model). These values are equivalent to  $(T, q)$  just after dynamics calculation. CAPE is defined as

$$\text{CAPE} = \int_{z_{\text{LFC}}}^{z_{\text{LNB}}} g \frac{T_v^u - T_v}{T_v} dz \quad (3.2.45)$$

where  $z_{\text{LFC}}$  and  $z_{\text{LNB}}$  are the height of the level of free convection and that of neutral buoyancy, respectively,  $g$  is the acceleration of gravity, and  $T_v$  is the virtual temperature. The superscript  $u$  denotes a lifted air parcel. Deep convection is prohibited when DCAPE does not exceed an empirically determined critical value.

#### 3.2.4.5 Convective Downdraft

For reasons of economy, only one type of downdraft is assumed, while many types are considered in the updraft scheme.

Downdraft is initiated at the level where the net upward mass flux is reduced to half that at the cloud base. The downdraft mass flux  $M^d$  at the cloud base is given by,

$$M^d = 0.4M_B \quad (3.2.46)$$

where  $M_B$  is the net mass flux at the cloud base of updraft as calculated using Eq. (3.2.40).

Entrainment from the environment is assumed to occur above the cloud base, while detrainment is assumed to occur both above and below it. The entrainment and detrainment rates are set to the same constant value above the cloud base, so that the mass flux of downdraft is constant with height.

#### 3.2.4.6 Mid-level Convection

A mid-level convection scheme is incorporated to represent cumulus convection with a cloud base on a frontal system in the extratropics. The height of the cloud base is given by the maximum moist static energy level in the vertical column, and the cloud top is defined as the level where an air mass rising from the cloud base with a constant entrainment rate loses buoyancy. Closure employs a simpler form of Eq. (3.2.40).

### 3.2.4.7 Convective Momentum Transport

Convective momentum transport is parameterized in a different way from heat and moisture transport. A multiple plume model is adopted both for updraft and downdraft. The entrainment and detrainment rates are set to the same value both for updraft and downdraft between the cloud base and the cloud top, making each mass flux constant with altitude. The magnitude of each updraft  $M_{cn}^u$  is set to the mass flux at the cloud base in the heat and moisture transport scheme as calculated using Eq. (3.2.40), and the magnitude of the downdraft is set to  $0.4 \times M_{cn}^u$ .

### 3.2.4.8 Melting and Re-evaporation of Precipitation

Melting of snow is calculated below freezing level, with formulation similar to that of the cloud scheme (Eq. (3.2.57)). Re-evaporation of precipitation is considered below the cloud base. Related calculation is based on the equation used in the cloud scheme (Eq. (3.2.59)) with a minor modification.

## 3.2.5 Clouds and Large-scale Precipitation

Clouds are prognostically determined in a fashion similar to that proposed by Smith (1990). The simple statistical approach proposed by Sommeria and Deardorff (1977) is adopted for the calculation of cloud amounts and their water content.

### 3.2.5.1 Cloud Scheme

Representing conserved quantities in phase transition between water vapor and cloud water, the total water content (water vapor and cloud water)  $q_w$  and the liquid water temperature  $T_L$  are defined as follows:

$$q_w = q_v + q_c \quad (3.2.47)$$

$$T_L = T - \frac{L}{C_p} q_c \quad (3.2.48)$$

where  $q_v$  is specific humidity,  $q_c$  is cloud water content,  $T$  is temperature,  $L$  is the latent heat of condensation and  $C_p$  is specific heat at a constant pressure. In each grid box,  $q_w$  is assumed to vary due to unresolved atmospheric fluctuations having a probability distribution function with a top-hat shape. The cloud fraction  $C$  is given by the part of the grid box where  $q_w$  exceeds the saturation specific humidity  $q_s$ , and cloud water content is given as the condensation amount in the grid box:

$$C = \frac{a_L (\bar{q}_w - q_s(T_L)) + \Delta q_w}{2\Delta q_w} \quad (3.2.49a)$$

$$q_c = C^2 \Delta q_w \quad (3.2.49b)$$

$$a_L = \frac{1}{1 + \frac{L}{C_p} \left( \frac{\partial q_s}{\partial T} \right)_{T=T_L}} \quad (3.2.49c)$$

where  $\Delta q_w$  is the maximum local deviation from the grid-box mean total water content  $\bar{q}_w$ , the overbar denotes an average over the grid box, and  $C$  is under the constraint  $0 \leq C \leq 1$ .  $\Delta q_w$  is calculated as follows:

$$\Delta q_w = \frac{a_L}{2} \left( \bar{q}_w'^2 - 2b \bar{q}_w' s_l' + b^2 \bar{s}_l'^2 \right)^{\frac{1}{2}} \quad (3.2.50)$$

where  $s_l = C_p T_L + gz$  is dry static energy,  $g$  is acceleration under gravity,  $z$  is height above the surface,  $b = \left( \frac{\partial q_s}{\partial T} \right)_{T=T_L} / C_p$ , and the prime denotes a deviation from the grid average.  $\bar{q}_w'^2$ ,  $\bar{q}_w' s_l'$ , and  $\bar{s}_l'^2$  are calculated using the level 2 turbulence closure model of Mellor and Yamada (see Subsection 3.2.7) in the boundary layer scheme.  $\Delta q_w$  is limited by  $q_s$ :

$$0.2A a_L q_s(T_L) \leq \Delta q_w \leq 0.5A a_L q_s(T_L) \quad (3.2.51)$$

$$A = \min\left(\frac{p_s - p}{p_s - 850}, 1\right) \quad (3.2.52)$$

where  $p_s$  is surface pressure and  $p$  is pressure.

### 3.2.5.2 Stratocumulus Scheme

For the representation of subtropical marine stratocumulus clouds, the stratocumulus scheme proposed by Kawai and Inoue (2006) is adopted to diagnose the cloud fraction  $C$  and the cloud water content  $q_c$  instead of the cloud scheme:

$$C = 12.0 \left( -\frac{\partial \theta}{\partial p} - 0.07 \right) \quad (3.2.53a)$$

$$q_c = 0.05 a_L C q_s \quad (3.2.53b)$$

where  $\theta$  is the potential temperature. This scheme works when the following three conditions in each model layer and one condition near the surface are satisfied:

[in each model layer]

- $-\frac{\partial \theta}{\partial p} > 0.07$  [K/hPa] (just above the model layer)
- relative humidity  $\geq 80$  [%]
- model layer height below the 924 hPa level

[near the surface]

- $-\frac{\partial \theta}{\partial p} < 0.01$  [K/hPa]

The first condition in each model layer represents the formation of marine stratocumulus clouds under a strong inversion layer, and the second and third prevent the false representation of stratocumulus clouds over dry area and shallow convection area, respectively. The condition near the surface prevents the false representation of stratocumulus clouds over land or sea ice during the night.

In the cloud scheme and the stratocumulus scheme, liquid (ice) cloud is assumed to be present when the temperature is above 0°C (below -15°C). Between -15°C and 0°C, mixed-phase cloud is assumed to be present and the mixing ratio changes in a linear fashion with temperature.

### 3.2.5.3 Cloud Ice Fall and Conversion to Precipitation

The prognostic equation for cloud water content is as follows:

$$\frac{\partial q_c}{\partial t} = C_g + I - O \quad (3.2.54)$$

where  $C_g$  represents the generation rate of clouds in the grid box,  $I$  is the inflow from the upper layer, and  $O$  is the downward outflow.  $C_g$  is calculated using the results described in Subsection 3.2.5.1 or Subsection 3.2.5.2.

For ice-only cloud,  $I$  and  $O$  are calculated based on Kawai (2005). Small cloud particles ( $\leq 100 \mu\text{m}$ ) fall into the lower layer, while large ones ( $> 100 \mu\text{m}$ ) fall to the surface immediately as snow:

$$O = \frac{v_{cice} q_c}{\Delta z} + D_{l2S} q_c \quad (3.2.55)$$

where  $\Delta z$  is the layer thickness,  $v_{cice}$  is the terminal velocity of a small ice particle in cloud, and  $D_{12S}$  is the conversion rate from ice cloud to snow. The first term on the right becomes  $I$  in the next layer down.

For mixed-phase cloud or water-only cloud, parameterization for the rate of conversion from cloud water to precipitation  $P$  follows the scheme proposed by Sundqvist (1978):

$$\begin{aligned} P &= \frac{1}{\tau_p} q_c \left[ 1 - \exp \left\{ - \left( \frac{q_c}{C q_c^{\text{crit}}} \right)^2 \right\} \right] \\ &= O \end{aligned} \quad (3.2.56)$$

where  $\tau_p$  represents a characteristic time scale for the conversion of cloud droplets into raindrops and snowflakes, and  $q_c^{\text{crit}}$  is the critical cloud water content at which the release of precipitation becomes efficient. In this case,  $I$  is not considered because raindrops and snowflakes are assumed to fall to the surface immediately. The coalescence process (collection of cloud droplets by raindrops falling through a cloud) and the Bergeron-Findeisen effect (enhancement of precipitation release in clouds containing a mixture of droplets and ice crystals) are modeled following Sundqvist *et al.* (1989).

### 3.2.5.4 Melting and Evaporation

The snow melting rate  $M$  is parameterized using the same method as ECMWF (2014):

$$M = 0.5 \frac{C_p}{L_m} \frac{T_w - T_0}{\tau_m} \quad (3.2.57)$$

$$\tau_m = \frac{7200}{1 + 0.5(T_w - T_0)} \quad (3.2.58)$$

where  $T_w$  is the wet-bulb temperature,  $T_0$  is the melting temperature,  $L_m$  is the latent heat of fusion, and  $\tau_m$  is the relaxation time of melting. Based on Kessler (1969) and Tiedtke (1993), the evaporation rate  $E$  for large-scale precipitation is parameterized as

$$E = b \frac{1}{\tau_e} (q_s - q_v) \left\{ \left( \frac{p}{p_s} \right)^{1/2} \frac{1}{5.09 \times 10^{-3}} \frac{P_l}{b} \right\}^{0.577} \quad (3.2.59)$$

$$\frac{1}{\tau_e} = 5.44 \times 10^{-4} \quad (3.2.60)$$

where  $b$  is the clear-sky precipitation fraction (set to 0.5),  $\tau_e$  is the relaxation time of evaporation, and  $P_l$  is the local precipitation rate.

## 3.2.6 Surface Turbulent Fluxes

Surface fluxes are represented with bulk formula based on Monin–Obukhov (M–O) similarity theory, with momentum flux  $F_m$ , heat flux  $F_h$  and specific humidity flux  $F_q$  expressed as follows:

$$F_m = \overline{(w' \mathbf{v}')}_s = -C_m |\mathbf{v}_1| \mathbf{v}_1, \quad (3.2.61)$$

$$F_h = \overline{(w' \theta')}_s = -C_h |\mathbf{v}_1| (\theta_1 - \theta_s), \quad (3.2.62)$$

$$F_q = \overline{(w' q')}_s = -C_h |\mathbf{v}_1| (q_1 - q_s). \quad (3.2.63)$$

Here  $\mathbf{v} = (u, v)$  represents horizontal wind,  $\theta$  is potential temperature and  $q$  is specific humidity, and the subscripts “1” and “s” indicate variables at the lowest level of the model grid and at the ground surface, respectively.  $C_m$  and  $C_h$  are exchange coefficients for momentum and heat.

According to M–O similarity theory, the exchange coefficients are defined as follows:

$$\frac{z_1}{L} \frac{\left[ \log\left(\frac{z_1+z_{0m}}{z_{0h}}\right) - \Psi_h\left(\frac{z_1+z_{0m}}{L}\right) + \Psi_h\left(\frac{z_{0h}}{L}\right) \right]}{\left[ \log\left(\frac{z_1+z_{0m}}{z_{0m}}\right) - \Psi_m\left(\frac{z_1+z_{0m}}{L}\right) + \Psi_m\left(\frac{z_{0m}}{L}\right) \right]^2} = \frac{gz_1}{|\mathbf{v}_1|^2} \frac{2(\theta_{v1} - \theta_{vs})}{(\theta_1 + \theta_s)} , \quad (3.2.64)$$

$$C_m = \frac{\kappa^2}{\left[ \log\left(\frac{z_1+z_{0m}}{z_{0m}}\right) - \Psi_m\left(\frac{z_1+z_{0m}}{L}\right) + \Psi_m\left(\frac{z_{0m}}{L}\right) \right]^2} , \quad (3.2.65)$$

$$C_h = \frac{\kappa^2}{\left[ \log\left(\frac{z_1+z_{0m}}{z_{0m}}\right) - \Psi_m\left(\frac{z_1+z_{0m}}{L}\right) + \Psi_m\left(\frac{z_{0m}}{L}\right) \right] \left[ \log\left(\frac{z_1+z_{0m}}{z_{0h}}\right) - \Psi_h\left(\frac{z_1+z_{0m}}{L}\right) + \Psi_h\left(\frac{z_{0h}}{L}\right) \right]} , \quad (3.2.66)$$

where  $\kappa$  is von Kármán's constant ( $= 0.4$ ),  $g$  is the standard acceleration due to gravity ( $= 9.80665 \text{ [m/s}^2]$ ),  $z_1$  is the height of the lowest level of the model grid above the ground, and  $\theta_v$  is the virtual potential temperature, while  $z_{0m}$  and  $z_{0h}$  are the surface momentum and heat roughness lengths, respectively. Eq. (3.2.64) gives Obukhov length  $L$  from the prognostic variables on right. The stability functions  $\Psi_m$  and  $\Psi_h$  are parameterized by Beljaars and Holtslag (1991) as follows:

$$x \equiv (1 - 16\xi)^{\frac{1}{4}} \quad (3.2.67)$$

$$\Psi_m(\xi) = \begin{cases} \frac{\pi}{2} - 2 \arctan(x) + \log \frac{(1+x)^2(1+x^2)}{8} & \xi < 0 \\ -\frac{2}{3} \left( \xi - \frac{5}{0.35} \right) e^{-0.35\xi} - \xi - \frac{2}{3} \frac{5}{0.35} & \xi \geq 0 \end{cases} , \quad (3.2.68)$$

$$\Psi_h(\xi) = \begin{cases} 2 \log \frac{1+x^2}{2} & \xi < 0 \\ -\frac{2}{3} \left( \xi - \frac{5}{0.35} \right) e^{-0.35\xi} - \left( 1 + \frac{2}{3} \xi \right)^{\frac{3}{2}} - \frac{2}{3} \frac{5}{0.35} + 1 & \xi \geq 0 \end{cases} . \quad (3.2.69)$$

Over land grids, surface parameters are determined using the land model on each grid in consideration of vegetation type, soil conditions, and snow cover (Subsection 3.2.9). Over ocean grids, surface fluxes are calculated separately for the different subgrid surface fractions. Tiling between open water and sea ice is used with the coupling approach suggested by Best *et al.* (2004).

Surface wind stress depends on oceanic waves excited by surface winds. Roughness length and wind-induced stress are iteratively calculated in the model. Following the method of Beljaars (1995), surface roughness lengths over open sea (ice-free ocean) are determined from Charnock's relation (Charnock 1955):

$$\begin{aligned} z_{0m} &= \frac{0.11\nu}{u_*} + \frac{\alpha}{g} u_*^2 , \\ z_{0h} &= \frac{0.62\nu}{u_*} , \end{aligned} \quad (3.2.70)$$

where  $u_* (\equiv \sqrt{|(w'\mathbf{v}')_s|})$  is the friction velocity,  $\nu$  is the kinematic viscosity of air ( $= 1.5 \times 10^{-5} \text{ m}^2/\text{s}$ ) and  $\alpha$  is the Charnock coefficient ( $= 0.020$ ). The surface roughness lengths over sea ice are fixed at 0.001m for momentum and 0.0005m for heat.

### 3.2.7 Boundary layer (turbulent transport)

A hybrid approach between turbulence kinetic energy (TKE) closure and eddy diffusivity (ED) type scheme is employed to parameterize vertical turbulent transports of momentum, heat and moisture in the atmosphere. The TKE scheme used is the level 2 turbulence closure approach of Mellor and Yamada (1974, 1982), and the ED-type scheme involves the use of stability functions based on Han and Pan (2011). Turbulent transports are expressed as

$$\overline{w'\mathbf{v}'} = - \max(K_m^{TKE}, K_m^{ED}) \frac{\partial \mathbf{v}}{\partial z} , \quad (3.2.71)$$

$$\overline{w' s'_L} = - \max(K_h^{TKE}, K_h^{ED}) \frac{\partial s_L}{\partial z}, \quad (3.2.72)$$

$$\overline{w' q'_w} = - \max(K_h^{TKE}, K_h^{ED}) \frac{\partial q_w}{\partial z}, \quad (3.2.73)$$

where  $s_L (\equiv C_p T + gz - Lq_c)$  is the liquid water static energy,  $q_w (\equiv q + q_c)$  is the total water content, the superscripts “TKE” and “ED” indicate the scheme types for calculating diffusion coefficients  $K$ , and the subscripts “ $m$ ” and “ $h$ ” indicate momentum and heat, respectively.

Following mixing-length theory, the diffusion coefficients of both schemes can be written as

$$K_m = l^2 \left| \frac{\partial \mathbf{v}}{\partial z} \right| f_m, \quad (3.2.74)$$

$$K_h = l^2 \left| \frac{\partial \mathbf{v}}{\partial z} \right| f_h, \quad (3.2.75)$$

where  $f_m$  and  $f_h$  are stability functions and the mixing length  $l$  is given according to Blackadar (1962) as

$$l = \frac{\kappa z}{1 + \kappa z/l_0}, \quad (3.2.76)$$

where  $\kappa$  is von Kármán’s constant ( $= 0.4$ ).

In the TKE scheme, the mixing length  $l_0$  is calculated from sub-grid scale orographic variances and the planetary boundary layer depth.  $l_0$  is a constant equal to 50 m in the ED-type scheme.

The stability functions in the ED type scheme are functions of the gradient Richardson number  $R_i$ , given as

$$f_m^{ED} = \begin{cases} \frac{1 + 2.1R_i}{(1 + 5R_i)^{1.5}} & R_i \geq 0 \\ 1 - \frac{8R_i}{1 + 1.746\sqrt{-R_i}} & R_i < 0 \end{cases}, \quad (3.2.77)$$

$$f_h^{ED} = \begin{cases} \frac{1}{(1 + 5R_i)^{1.5}} & R_i \geq 0 \\ 1 - \frac{8R_i}{1 + 1.286\sqrt{-R_i}} & R_i < 0 \end{cases}. \quad (3.2.78)$$

Those of the TKE scheme can be written as follows:

$$f_m^{TKE} = S_M \sqrt{B_1 S_M (1 - R_f)}, \quad (3.2.79)$$

$$f_h^{TKE} = S_H \sqrt{B_1 S_M (1 - R_f)}, \quad (3.2.80)$$

$$S_M = A_1 F_1 \frac{RF_1 - R_f}{A_2 F_2 (RF_2 - R_f)} S_H,$$

$$S_H = 3A_2 \frac{(\gamma_1 + \gamma_2)(RF_c - R_f)}{1 - R_f},$$

$$R_f = RI_1 \left( \hat{R}_i + RI_2 - \sqrt{\hat{R}_i(\hat{R}_i - RI_3) + RI_2^2} \right),$$

$$RF_1 = B_1 \frac{\gamma_1 - C_1}{F_1}, \quad RF_2 = B_1 \frac{\gamma_1}{F_2}, \quad RF_c = \frac{\gamma_1}{(\gamma_1 + \gamma_2)},$$

$$RI_1 = \frac{1}{2} \frac{A_2 F_2}{A_1 F_1}, \quad RI_2 = B_1 \gamma \frac{(\gamma_1 - C_1)}{F_1} \frac{A_1 F_1}{A_2 F_2}, \quad RI_3 = 4B_1 \frac{\gamma_1}{F_2} \frac{A_1 F_1}{A_2 F_2} - 2RI_2,$$

$$F_1 = B_1(\gamma_1 - C_1) + 2A_1(3 - 2C_2) + 3A_2(1 - C_2), \quad F_2 = B_1(\gamma_1 + \gamma_2) - 3A_1(1 - C_2),$$

$$\gamma_1 = \frac{1}{3} - 2 \frac{A_1}{B_1}, \quad \gamma_2 = \frac{B_2}{B_1}(1 - C_3) + 2 \frac{A_1}{B_1}(3 - 2C_2),$$

where  $A_1(= 1.0)$ ,  $A_2(= 0.58)$ ,  $B_1(= 24.0)$ ,  $B_2(= 11.0)$ ,  $C_1(= 0.13)$ ,  $C_2(= 0.6)$ , and  $C_3(= 0.14)$  are the closure constants of the TKE scheme. The modified gradient Richardson number  $\hat{R}_i$  used instead of  $R_i$  is defined after the method of Smith (1990):

$$\hat{R}_i = g \left\{ \tilde{\beta}_s \frac{\partial s_L}{\partial z} + \tilde{\beta}_Q \frac{\partial q_w}{\partial z} \right\} \left| \left| \frac{\partial \mathbf{v}}{\partial z} \right| \right|^2, \quad (3.2.81)$$

where  $\tilde{\beta}_s$  and  $\tilde{\beta}_Q$  are buoyancy parameters in terms of the cloud-conserved quantities  $s_L$  and  $q_w$ , respectively.

### 3.2.8 Gravity Wave Drag

#### 3.2.8.1 Orographic Drag

The parameterization for the orographic gravity wave drag consists of two components; one for long waves (wavelength  $> 100\text{km}$ ) and the other for short waves (wavelength  $\approx 10\text{km}$ ). The long waves are assumed to propagate upward until reaching wave-breaking levels mainly in the stratosphere and exert drag there (type A scheme), while short waves are always regarded as trapped and dissipated within the lower troposphere (type B scheme). Therefore the fundamental difference between the two schemes appears in the vertical distribution of the momentum deposit. The type A scheme is based on Palmer *et al.* (1986) with some modifications. Details of type A and B schemes are explained in Iwasaki *et al.* (1989).

In both schemes, the momentum flux  $\tau_r$  excited by subgrid-scale variances of topography  $\sigma^2$  is determined by

$$\tau_r = C_{gw} \rho_r N_r v_r \min \left( \sigma^2, \left( \frac{v_r}{2F_c N_r} \right)^2 \right) v_r / v_r \quad (3.2.82)$$

where  $C_{gw}$  is constant ( $1.6 \times 10^{-5}$  for type A and  $6.4 \times 10^{-4}$  for type B),  $\rho$  is air density,  $N$  is Brunt-Väisälä frequency,  $F_c$  is critical Froude number (1.5 for generation and 1.0 for propagation),  $v$  is the intrinsic velocity and  $v = |\mathbf{v}|$ . The subscript  $r$  denotes the reference level where the gravity wave stresses (momentum fluxes) are generated. There is a maximum of the momentum flux due to the valley blocking phenomenon, which is caused by stagnant flow near bottoms of valleys. This phenomenon occurs when the Froude number is below a critical value. The blocking effectively reduces the amplitudes of gravity waves. The topographic variances  $\sigma^2$  are derived from the GTOPO30, which is  $30'' \times 30''$  geographical data developed by the U. S. Geological Survey. First, the mean elevation ( $h_m$ ) and its standard deviation ( $\sigma_m$ ) over a  $5' \times 5'$  grid box are evaluated from GTOPO30. The standard deviation of  $(h_m - h)$  in a Gaussian grid box is regarded as  $\sigma$  in the type A scheme where  $h$  denotes the model topography, while the average of  $\sigma_m$  in the Gaussian grid box is regarded as  $\sigma$  in the type B scheme.

In the type A scheme, the momentum deposit is determined by the amplitude saturation hypothesis. The gravity wave stress at the  $(k + 1/2)$ -th level is given by

$$\tau_{k+1/2} = \min \left( |\tau_{k-1/2}|, |\tau_{sat}| \right) \tau_r / |\tau_r| \quad (3.2.83)$$

where

$$\tau_{sat} = C_{gw} \rho N \left( v \cdot \frac{\tau_r}{|\tau_r|} \right) \left[ \frac{\epsilon}{2F_c N} \left( v \cdot \frac{\tau_r}{|\tau_r|} \right) \right]^2 \frac{\tau_r}{|\tau_r|} \quad (3.2.84)$$

$\epsilon$  is a function of the Richardson number ( $R_i$ )

$$\epsilon = \left( \frac{1}{R_i^{1/2}} + 2 \right) - \left\{ \left( \frac{1}{R_i^{1/2}} + 2 \right)^2 - \left( \frac{1}{R_i} - 4 \right) \right\}^{1/2} \quad (3.2.85)$$

$$R_i = N^2 \left| \left| \frac{\partial}{\partial z} \left( v \cdot \frac{\tau_r}{|\tau_r|} \right) \right| \right|^2 \quad (3.2.86)$$

The wave stress of short gravity waves decreases with altitude due to nonhydrostatic effects (*e.g.*, Wurtele *et al.* 1987). In the type B scheme, the wave stress is simply assumed to be a quadratic function of pressure  $p$  and to vanish around 700hPa as follows:

$$\tau(p) = \begin{cases} \tau_r \cdot \frac{(p/p_s - 0.7)^2}{0.3^2} & p/p_s \geq 0.7 \\ 0 & p/p_s < 0.7 \end{cases} \quad (3.2.87)$$

where  $p_s$  is surface pressure.

### 3.2.8.2 Non-orographic Drag

The parameterization of non-orographic gravity wave drag follows the scheme proposed by Scinocca (2003). Assuming hydrostatic gravity waves in the absence of rotation, the vertical flux of momentum directed into the  $\phi$  azimuth  $\tilde{F}$  is expressed as follows:

$$\rho\tilde{F} = \rho A \frac{\tilde{c} - \tilde{U}}{N} \left( \frac{\tilde{c} - \tilde{U}}{\tilde{c}} \right)^{2-p} \frac{1}{1 + \left[ \frac{m^*(\tilde{c} - \tilde{U})}{N} \right]^{s+3}} \quad (3.2.88)$$

where,

$$A = C m^{*3} \frac{N_0^{2-p} - f^{2-p}}{2 - p} \quad (3.2.89)$$

$\rho$  is density,  $m^*$  ( $= 2\pi/2000$  [1/m]) is the characteristic vertical wavenumber,  $N$  is the Brunt-Väisälä frequency,  $f$  is the Coriolis parameter,  $p$  ( $= 1.5$ ) and  $s$  ( $= 1$ ) are constants,  $\tilde{c} = c - U_0^\phi$  and  $\tilde{U} = U^\phi - U_0^\phi$ , with  $c$  as the ground based phase speed and  $U^\phi$  as the velocity in the direction of the azimuth  $\phi$ , with the subscript 0 referring to the launch level. The launch level and launch momentum flux (corresponding to the constant  $C$ ) are 450 hPa and 3.5 mPa, respectively. The momentum fluxes are discretized in four equally spaced azimuths (north, south, west and east) and 50 phase speed bins in the range of 0.25 to 2000 m/s.

As momentum deposition processes, critical level filtering and nonlinear saturation are considered. In critical level filtering calculation, when  $\tilde{c} - \tilde{U} < 0$  (assuming that waves encounter the critical level), the momentum flux is deposited to the mean flow in this layer. In nonlinear saturation calculation, when the upward propagating momentum flux  $\rho\tilde{F}$  exceeds the saturated momentum flux  $\rho\tilde{F}^{sat}$ , the excess momentum flux ( $\rho\tilde{F} - \rho\tilde{F}^{sat}$ ) is deposited to the mean flow. The saturated momentum flux is expressed as follows:

$$\rho\tilde{F}^{sat} = \rho C^* A \frac{\tilde{c} - \tilde{U}}{N} \left( \frac{\tilde{c} - \tilde{U}}{\tilde{c}} \right)^{2-p} \quad (3.2.90)$$

where  $C^*$  ( $= 10$ ) is the tuning parameter introduced by McLandress and Scinocca (2005). These vertical momentum flux depositions are calculated at each level for each azimuthal direction and phase speed. To reduce the computational cost, parameterization is performed only once an hour.

### 3.2.9 Land Surface Processes

The land surface model in the GSM employs a two-layer energy balance scheme based on the Simple Biosphere scheme (SiB; Sellers *et al.* 1986; Sato *et al.* 1989). It has evolved to a complex representation for snow and soil, with reference to Oleson *et al.* (2010)

The model is composed of vegetation canopy, snow and soil components, each of which has its own variables of temperature, water and ice content. Canopy air space is used to express paths of heat and water transfer. Figure 3.2.1 shows heat and water flows in an analogy of an electric circuit.

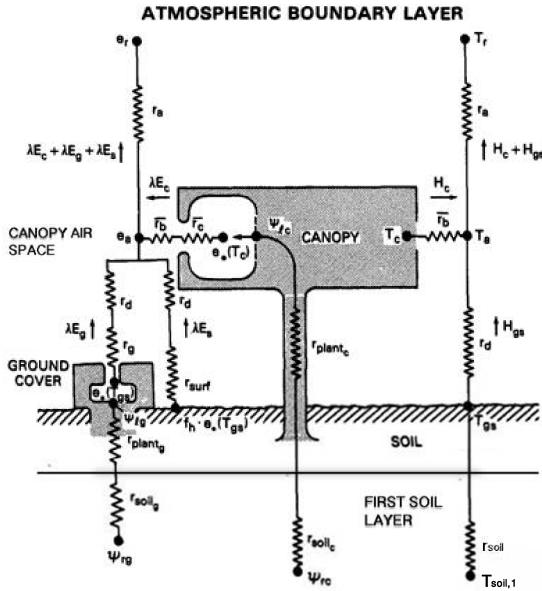


Figure 3.2.1: Schematic illustration of the land surface model (modified from Sellers *et al.* (1986)). Details are provided in the original paper.

### 3.2.9.1 Fluxes

The main role of the land surface model is to provide lower boundary conditions of fluxes to the atmospheric model. The zonal and meridional momentum fluxes ( $\tau_x, \tau_y \equiv \overline{(w'v')}$ , sensible heat flux  $H \equiv \overline{(w'\theta')}$ , and water vapor flux  $E \equiv \overline{(w'q')}$  are based on bulk formulae, as detailed in Subsection 3.2.6. Among these,  $\tau_x$  and  $\tau_y$  can be computed simply using the atmospheric wind velocity  $v_1$ , whereas determination of  $H$  and  $E$  requires the surface temperature and specific humidity values provided by the land surface model. This model in the GSM defines the values as “canopy air space temperature  $T_a (= \theta_s)$  and specific humidity  $q_a$ ”, where sensible heat and water vapor fluxes from vegetation canopy and ground surface are connected.

Sensible heat  $H$  is balanced by the sum of sensible heats from the canopy  $H_c$  and the ground surface  $H_g$ , since canopy air space is assumed to have negligible heat and water vapor capacities,

$$H = H_c + H_g \quad (3.2.91)$$

Similarly, water vapor  $E$  is balanced by the sum of evaporation from the canopy  $E_c^e$ , the ground surface  $E_g^e$  and bare soil  $E_{bs}$ , as well as transpiration from the canopy  $E_c^t$  and the ground surface  $E_g^t$

$$E = E_c^e + E_c^t + E_g^e + E_g^t + E_{bs} \quad (3.2.92)$$

These fluxes are determined with canopy temperature  $T_c$  and ground surface temperature  $T_g$ . Here,  $T_g$  represents grass or bare soil temperature, but is regarded as snow surface temperature when the ground surface is covered with snow.

### 3.2.9.2 Radiation and Albedo

The net radiation fluxes at the canopy  $R_c^n$  and the ground surface  $R_g^n$  are estimated based on the radiation balance equations

$$R_c^n = (1 - \alpha_c) S_{atm}^\downarrow + L_c^\downarrow \quad (3.2.93)$$

$$R_g^n = (1 - \alpha_g) S_{atm}^\downarrow + L_g^\downarrow \quad (3.2.94)$$

where  $\alpha$  is the albedo, and  $S_{atm}^\downarrow$  and  $L^\downarrow$  are downward shortwave and longwave radiation from the atmosphere, respectively. In (3.2.93) and (3.2.94),  $R_c^n$  and  $R_g^n$  can be estimated using the albedos.

The surface albedo  $\alpha_s$  can be determined as an average of the canopy albedo  $\alpha_c$  and the ground albedo  $\alpha_g$ , weighted by the fraction of canopy cover  $f_c$

$$\alpha_s = f_c \alpha_c + (1 - f_c) \alpha_g \quad (3.2.95)$$

Similarly, the ground albedo  $\alpha_g$  is an average of the grass albedo  $\alpha_{grs}$  and the bare soil albedo  $\alpha_{bs}$ , weighted by the fraction of grass cover  $f_{grs}$  in snow-free areas. When the ground is covered with snow,  $f_{grs}$  is set to zero, and  $\alpha_{bs}$  is replaced by the value for snow  $\alpha_{sn}$

$$\alpha_g = \begin{cases} f_{grs} \alpha_{grs} + (1 - f_{grs}) \alpha_{bs} & \text{(snow-free)} \\ \alpha_{sn} & \text{(snow-covered)} \end{cases} \quad (3.2.96)$$

$\alpha_c$  and  $\alpha_{grs}$  are calculated with radiative transfer equations (Sellers *et al.* 1986) for leaf and stem area.  $\alpha_{bs}$  is provided from climatological data of the MODIS albedo product (Schaaf *et al.* 2002) and modified using the solar zenith angle and soil moisture near the soil surface.  $\alpha_{sn}$  evolves with time, and is corrected using the solar zenith angle.

### 3.2.9.3 Energy and Water Balances

The prognostic equations for  $T_c$  and  $T_g$  are given as

$$C_c \frac{\partial T_c}{\partial t} = R_c^n - H_c - L_{vap} E_c \quad (3.2.97)$$

$$C_g \frac{\partial T_g}{\partial t} = R_g^n - H_g - L_{vap} E_g - G_g \quad (3.2.98)$$

where the subscripts  $c$  and  $g$  denote canopy and ground surface, respectively,  $C$  is heat capacity,  $R^n$  net radiation,  $L_{vap}$  latent heat of vaporization,  $G_g$  ground surface heat flux, and  $E_c \equiv E_c^e + E_c^t$ ,  $E_g \equiv E_g^e + E_g^t + E_{bs}$ .

Water storage on canopy leaves  $M_c$  and grass leaves  $M_g$ , which are sources of evaporation, are predicted by

$$\frac{\partial M_c}{\partial t} = P_{i,c} - P_{d,c} - E_c^e \quad (3.2.99)$$

$$\frac{\partial M_g}{\partial t} = P_{i,g} - P_{d,g} - E_g^e \quad (3.2.100)$$

where  $P_i$  is precipitation intercepted by leaves, and  $P_d$  water drip falling from leaves. The difference between these two values,  $I_{cept} = P_{i,c} + P_{i,g} - (P_{d,c} + P_{d,g})$ , represents interception by canopy and grass. When  $T_c$  ( $T_g$ ) is below the freezing point of water,  $M_c$  ( $M_g$ ) represents ice on canopy or grass leaves.

### 3.2.9.4 Snow

Snow temperature  $T_{sn}$  is predicted based on the principle of energy conservation and Fourier's law of heat conduction

$$C_{sn} \frac{\partial T_{sn}}{\partial t} = \frac{\partial G_{sn}}{\partial z} \quad (3.2.101)$$

$$G_{sn} = -\lambda_{sn} \frac{\partial T_{sn}}{\partial z} \quad (3.2.102)$$

where the subscript  $sn$  denotes snow,  $G$  is heat flux (positive downward),  $z$  snow depth from the snow surface, and  $\lambda$  thermal conductivity. Using a multi-layer model, a snow column is discretized into up to four layers (Figure 3.2.2). The top boundary condition the ground surface heat flux, while the bottom boundary condition is conductive heat flux with the first soil layer

$$G_{sn,0} = G_g \quad (3.2.103)$$

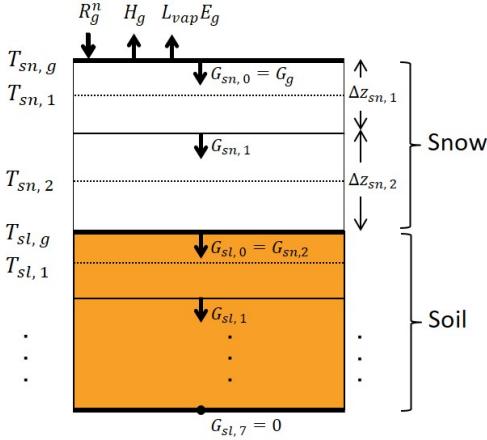


Figure 3.2.2: Schematic diagram of numerical discretization used to solve for snow temperatures  $T_{sn,k}$  with two snow layers on top of soil. The subscripts  $sn$ ,  $sl$  and  $k$  denote snow, soil and the  $k$ -th layer, respectively.

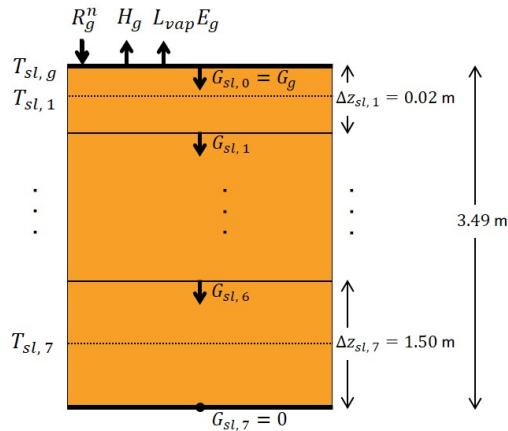


Figure 3.2.3: Schematic diagram of numerical discretization used to solve for soil temperatures  $T_{sl,k}$  in snow-free areas.

$$G_{sn,k_{max}} = \Lambda_{sn,k_{max}} (T_{sn,k_{max}} - T_{sl,1}) \quad (3.2.104)$$

where the subscript  $sl$  denotes soil, and  $\Lambda$  is a thermal conduction coefficient. For integration in time  $t$ , implicit methods are adopted and tri-diagonal matrices are solved.

Snow mass  $M_{sn}$  is predicted based on the snow mass balance equation

$$\frac{\partial M_{sn}}{\partial t} = S_{fall} + (S_{frst} - S_{sub}) + (S_{frz} - S_{melt}) \quad (3.2.105)$$

where  $S_{fall}$  is snowfall reaching the snow surface, including ice drip falling from leaves,  $S_{frst}$  frost,  $S_{frz}$  freezing,  $S_{sub}$  sublimation, and  $S_{melt}$  snowmelt.

Liquid water content in snow  $W_{sn}$  is predicted using

$$\frac{\partial W_{sn}}{\partial t} = (Q_{sn,infl} - Q_{sn,drg}) + (S_{dew} - S_{evap}) - (S_{frz} - S_{melt}) \quad (3.2.106)$$

where  $Q_{sn,infl}$  is infiltration into snow including rainfall and water drip falling from leaves,  $Q_{sn,drg}$  gravitational drainage from the bottom, and  $S_{dew}$  liquid dew,  $S_{evap}$  evaporation. Snow mass and snow water content are predicted using the same  $k_{max}$  layers as those for the snow temperature. When snow water content in a layer exceeds the layer's holding capacity, excess water is moved to the underlying layer. Snow depth data from Snow Depth Analysis (see Section 2.8) are used to set the initial value of snow water equivalent.

### 3.2.9.5 Soil

Soil temperature  $T_{sl}$  is predicted in the same way as snow temperature, and soil is discretized into seven layers (Figure 3.2.3) of soil thicknesses  $\Delta z_{sl,1-7}$  (m) = (0.02, 0.05, 0.12, 0.3, 0.5, 1.0, 1.5). The boundary conditions are

$$G_{sl,0} = \begin{cases} G_g & (\text{snow-free}) \\ \Lambda_{sl,k_{max}} (T_{sn,k_{max}} - T_{sl,1}) & (\text{snow-covered}) \end{cases} \quad (3.2.107)$$

$$G_{sl,7} = 0 \quad (3.2.108)$$

Soil moisture is predicted using Richard's equation, and root extraction for transpiration is also considered. The prognostic equation for the degree of saturation  $W$  is as follows:

$$\frac{\partial W}{\partial t} = \frac{1}{\rho_{wtr} \theta_{sat}} \left( -\frac{\partial Q}{\partial z} - S^t \right) \quad (3.2.109)$$

where  $\rho_{wtr}$  is water density,  $\theta_{sat}$  soil porosity,  $Q$  water flux caused by differences in matric potential and gravitational potential, and  $S^t$  root extraction for transpiration. The top water flux is  $Q_{infl} - E_{bs}$ , where  $Q_{infl}$  is infiltration into soil including precipitation, drip falling from leaves and snowmelt, and the bottom is gravitational drainage  $Q_{drng}$ .

Surface runoff  $R_{off}$  and gravitational drainage  $Q_{drng}$  are counted as total runoff  $R_{total}$

$$R_{total} = R_{off} + Q_{drng} \quad (3.2.110)$$

$$= T_{fall} - (Q_{infl} - Q_{drng}) - (Q_{sn,infl} - Q_{sn,drng}) \quad (3.2.111)$$

where

$$R_{off} = T_{fall} - Q_{infl} - (Q_{sn,infl} - Q_{sn,drng}) \quad (3.2.112)$$

$$T_{fall} = P_g - I_{cept} \quad (3.2.113)$$

$P_g$  is gross rainfall, and  $T_{fall}$  throughfall reaching the ground surface.  $Q_{infl}$  is limited due to the maximum surface infiltration capacity.

### 3.2.9.6 Datasets

The climatological data of base soil albedo is derived from the MODIS albedo product<sup>1</sup> of NASA (Schaaf *et al.* 2002).

Each grid point on land is classified by a specific vegetation type provided from GLC2000<sup>2</sup> (Global Land Cover 2000; Bartholomé and Belward 2005) of the European Commission's Joint Research Center (JRC). The LAI (Leaf Area Index) is based on the MODIS LAI product<sup>3</sup> of NASA (Myneni *et al.* 2002). Fractions of canopy and grass covers are calculated using the 1 km MODIS-based Maximum Green Vegetation Fraction<sup>4</sup> of USGS (Broxton *et al.* 2014), the 1 km Tree Cover Continuous Fields product<sup>5</sup> of GLCF (DeFries *et al.* 2000), and the Cropland and Pasture Area fraction<sup>6</sup> of EarthStat (Ramankutty *et al.* 2008).

The initial condition for soil moisture is given by climatological data calculated using an offline model with the atmospheric forcing dataset of GSWP3 (Global Soil Wetness Project Phase 3; Kim 2017)<sup>7</sup>. The soil property is from HWSD (Harmonized World Soil Database; FAO *et al.* 2012).

## 3.2.10 Parallelization

In the GSM, Open Multiprocessing (OpenMP) is employed for shared memory parallelization, and the Message Passing Interface (MPI) is used for distributed memory parallelization. A two-dimensional decomposition method is adopted for parallelization among processes.

Figure 3.2.4 shows the schematic design of parallelization. There are five computational stages, and appropriate decompositions are selected in each stage. The base is the Grid stage. The Fourier and Legendre stages are used for calculating spherical harmonic transformation, and the Wavenumber stage is used for calculating Helmholtz equations in the semi-implicit scheme. The Horizontal Advection stage is used in the implementation of the semi-Lagrangian advection scheme.

<sup>1</sup><https://search.earthdata.nasa.gov/search>

<sup>2</sup><http://forobs.jrc.ec.europa.eu/products/glcc2000/glcc2000.php>

<sup>3</sup><https://search.earthdata.nasa.gov/search>

<sup>4</sup>[https://archive.usgs.gov/archive/sites/landcover.usgs.gov/green\\_veg.html](https://archive.usgs.gov/archive/sites/landcover.usgs.gov/green_veg.html)

<sup>5</sup><http://glcf.umd.edu/data/treecover/>

<sup>6</sup><http://www.earthstat.org/cropland-pasture-area-2000/>

<sup>7</sup><http://hydro.iis.u-tokyo.ac.jp/GSWP3/index.html>, <https://www.isimip.org/gettingstarted/details/4/>

At the Grid stage, since all vertical levels exist in a same rank for the computation of physical processes and non-linear terms of dynamical processes, variable arrays are decomposed into east-west and north-south directions. North-south decomposition follows a cyclic order, and is applied in such a way that the order of ranking is reversed alternately. This helps to mitigate load imbalances associated with physical parameterization and the number of grid points, since their computational loads depend mainly on latitudinal zones. At the Fourier stage, since all east-west grid points exist in a same rank for the performance of Fourier transformation, variable arrays are decomposed into north-south and vertical directions. At the Legendre stage, since all north-south grid points exist in a same rank for the performance of Legendre transformation, variable arrays are decomposed into vertical and longitudinal wavenumber directions. At the Wavenumber stage, since all vertical levels exist in a same rank for solving Helmholtz equations in the semi-implicit scheme, variable arrays are decomposed into longitudinal and total wavenumber directions. Communication among these four stages can be performed independently within each subset based on the provision of two restrictions for the number of decompositions: 1) the number of decompositions for the east-west direction, the vertical direction and the total wavenumber direction must be the same, and 2) the number of decompositions for the north-south direction and the longitudinal wavenumber direction must be the same.

At the Horizontal Advection stage, variable arrays are decomposed into vertical and north-south directions. To reduce the amount of communication relating to halo regions, the number of decompositions for the north-south direction is made as small as possible. Unlike communication in the stages described above, global communication is required for interaction between the Grid stage and the Horizontal Advection stage.

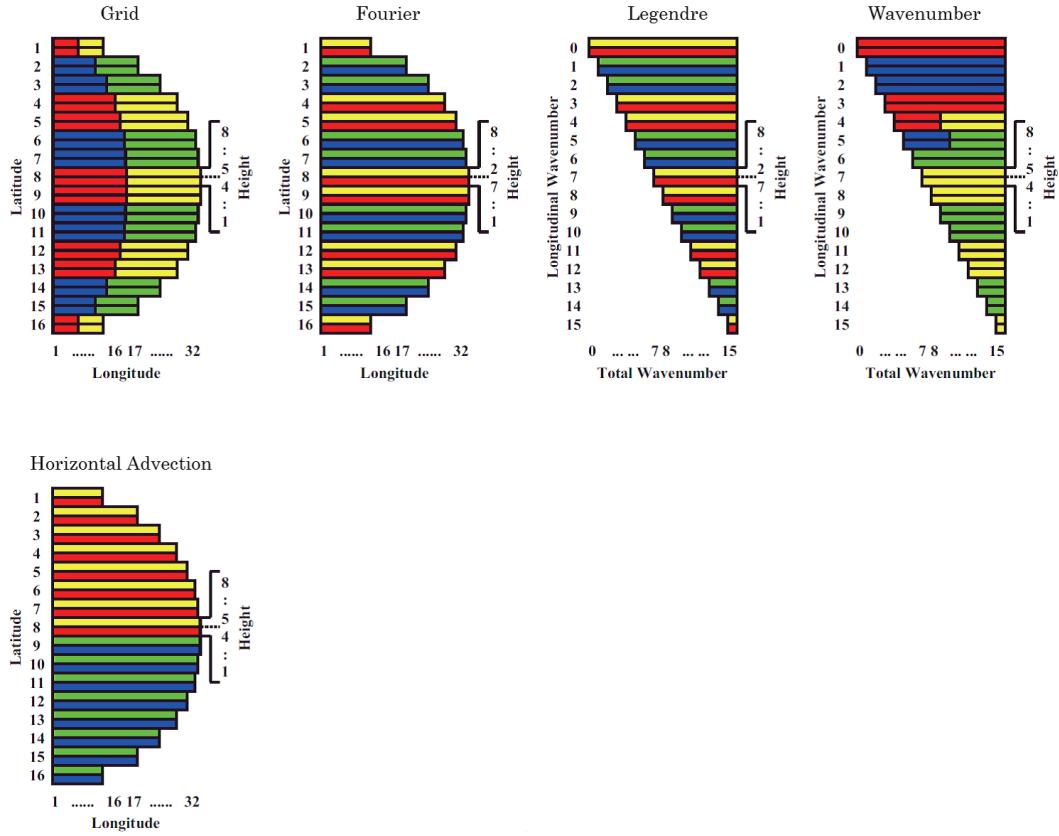


Figure 3.2.4: Schematic design of the parallelization. The number of processes used is assumed to be 4 in this example. Colors in the figure represent the rank for the computation in that area; red is rank 0, yellow is rank 1, blue is rank 2 and green is rank 3.

### 3.2.11 Surface Specifications

#### 3.2.11.1 Orography

The model orography in the GSM is based on Global 30 Arc-Second Elevation (GTOPO30) orographic data developed by the United States Geological Survey (USGS). To derive the orography, the following processing is applied: (i) Elevation data from GTOPO30 on a 30 x 30 arc-second lat-lon grid are averaged on the model's reduced Gaussian grid. (ii) The grid-averaged elevation produced is then spectrally smoothed by multiplying the spectral coefficients by the following smoothing factor

$$f(n) = \exp\left(\log(df)\left(\frac{n(n+1)}{N(N+1)}\right)^2\right) \quad (3.2.114)$$

where  $n$  is the total wavenumber,  $N$  is the truncation total wavenumber, and  $df = 0.1$  is a tunable smoothing parameter.

#### 3.2.11.2 Grid Type

Land-ocean distribution is determined in reference to the Global Land Cover Characteristics (GLCC) database (Loveland *et al.* 2000) compiled by USGS and others. Model grid sections in which the land area ratio is more than 49% in GLCC are regarded as land grids in the GSM to keep the same global ocean area ratio as GLCC. Grids not defined as land are sea (ocean) types can have two tiles fractions (open water and ice). Inland water grids are treated as sea. Meanwhile, each land grid has a particular vegetation type based on the Global Land Cover 2000 database provided by the European Commission's Joint Research Centre in 2003. See Subsection 3.2.9 for the surface properties of land grids.

#### 3.2.11.3 Sea Surface

On sea grids in the GSM, sea surface temperature (SST) and sea ice concentration (SIC) are given as boundary conditions minimally affected by the atmosphere. The amount of change in these variables during the time-integration of the model is equivalent to the time interpolated variation in daily climatological data. The directbeam albedo  $\alpha_B$  of the water surface is derived by the following parameterization (Briegleb *et al.* 1986):

$$\alpha_B = \frac{0.026}{(\mu^{1.7} + 0.065)} + 0.15(\mu - 0.1)(\mu - 0.5)(\mu - 1.0), \quad (3.2.115)$$

where  $\mu$  is the cosine of the solar zenith angle. The diffused albedo  $\alpha_D$  is constant ( $= 0.06$ ).

#### 3.2.11.4 Sea Ice

The ice fraction is modelled as an ice slab, with open water underneath and a skin temperature for thermal contact with the lowest part of the atmosphere. The sea ice parameterization is as follows: (i) The depth of the slab is fixed (the volume remains constant regardless of melting). (ii) Slab material properties are homogeneous and constant. (iii) Snow accumulation on the ice is not considered. Ice heat transfer is assumed to obey the following Fourier law of diffusion:

$$(\rho C) \frac{\partial T_{ice}}{\partial t} = \frac{\partial}{\partial z} \left( \lambda \frac{\partial T_{ice}}{\partial z} \right), \quad (3.2.116)$$

where  $\rho C = 1.93 \times 10^6$  [Jm<sup>-3</sup>K<sup>-1</sup>] is the volumetric ice heat capacity,  $T_{ice}$  is the ice temperature, and  $\lambda = 2.03$  [Wm<sup>-1</sup>K<sup>-1</sup>] is the ice thermal conductivity. As a boundary condition, the temperature at the bottom of the slab is given as  $T_{ice} = 271.51$ K. The temperature at the top of the slab is diagnosed from the net heat flux at the top skin.

The boundary condition at the bottom is the temperature of frozen water, and the top boundary condition is the temperature diagnosed from the net heat flux at the top skin. In the GSM, the ice slab is vertically discretized into four layers. The temperature at each level is solved via implicit time-integration.

The sea ice albedos for near-infrared ( $\alpha^N$ ) and visible ( $\alpha^V$ ) are parameterized as follows:

$$F_{60} = \begin{cases} 0.8 - 0.025(T_{skin} - 263.15) & T_{skin} > 263.15 \\ 0.8 & T_{skin} \leq 263.15 \end{cases},$$

$$F = \max(\min(F_{60} + 0.364(0.5 - \mu), 0.85), 0.07), \quad (3.2.117)$$

$$\alpha^N = 0.86F + 0.01, \quad (3.2.118)$$

$$\alpha^V = 1.14F - 0.01, \quad (3.2.118)$$

where  $T_{skin}$  is the ice surface temperature, and  $\mu$  is the cosine of the solar zenith angle.

### 3.2.12 Initial Conditions

Initial conditions of subsystems such as the atmosphere and land are required for GSM time integration. The specifications of these are detailed in Table 3.2.3, where the term “forecast guess” represents the use of forecast variables with a lead time of six hours. However, land and snow variables are adjusted via snow analysis for consistency.

Table 3.2.3: Initial Conditions of GSM

Subsystem	Variable	Origin
Atmosphere	Zonal wind	4D-Var global objective analysis (see Section 2.5)
	Meridional wind	
	Temperature	
	Specific humidity	
	Surface pressure	
	Cloud water content	Forecast guess
	Diagnosed cloud water content in convection updraft	
	Cloud cover of stratiformis and convective cloud	
	Convective mass flux at cloud base	
Land surface	Grass (or bare soil) temperature	Forecast guess
	Canopy temperature	
	Liquid and ice water content on vegetation	
Land soil	Temperature	Climatological values
	Liquid and ice water content in voids	
Snow	Snow water equivalent	Snow analysis (see Subsection 2.8.1)
	Temperature	
	Density	
	Liquid water content in voids	
	Albedo	
Sea ice	Age	Forecast guess
	Ice concentration	
Open sea	Temperature	Forecast guess
	Sea surface temperature	
	Friction velocity	

### 3.2.13 Forecast Performance

Figure 3.2.5 shows the root mean square error (RMSE) for 24-, 72- and 120-hour forecasts of 500 hPa geopotential height against analysis in the Northern Hemisphere extra-tropics. Dashed lines indicate monthly means, and solid lines represent 12-month running means. There are decreasing trends in RMSEs corresponding to GSM changes (see Subsection 3.2.1), although the impact of the changes in recent years has been relatively small.

Tropical cyclone (TC) track predictions are verified against the best track as analyzed by JMA's RSMC Tokyo - Typhoon Center. The mean position error of GSM TC track predictions in the western North Pacific (Figure 3.2.6) exhibits a gradual reduction from 1996 to 2017 due to GSM improvements, but considerable inter-annual variations are seen in 72-hour forecast errors and elsewhere.

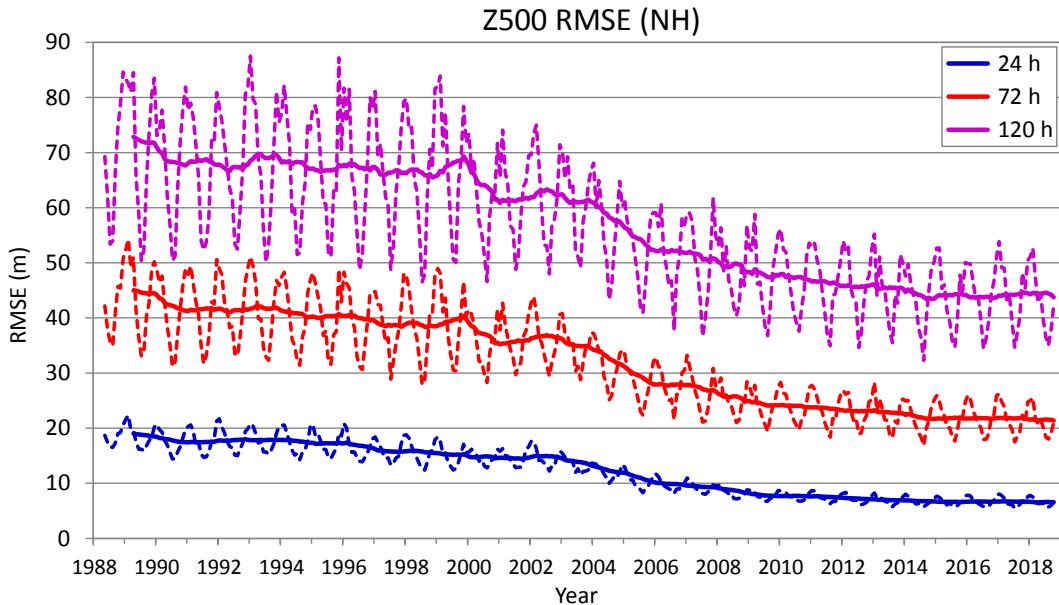


Figure 3.2.5: Root mean square error of GSM 500 hPa geopotential height (Z500) predictions against analysis in the Northern Hemisphere extra-tropics ( $20^{\circ}\text{N}$  –  $90^{\circ}\text{N}$ ). Dashed lines indicate monthly means, and solid lines represent running means calculated for the previous 12 months.

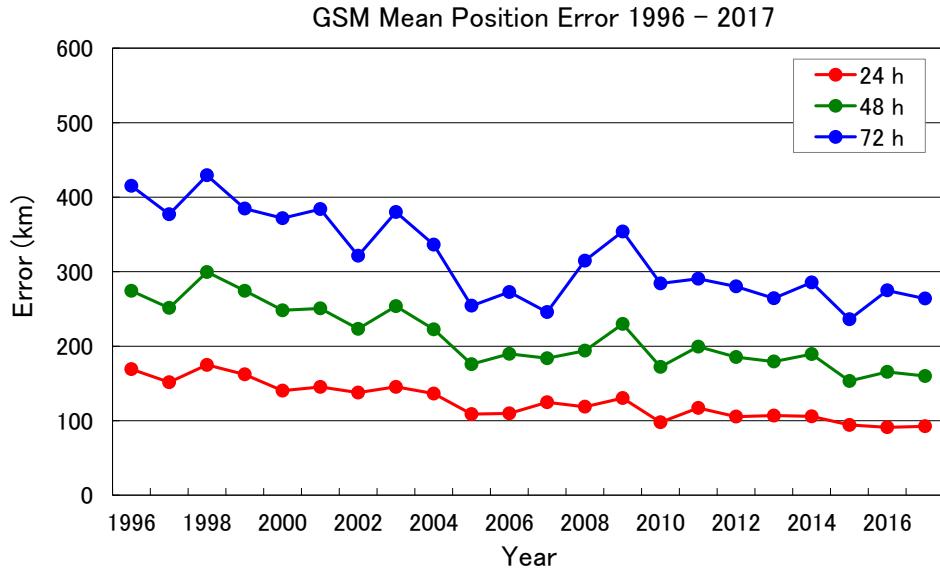


Figure 3.2.6: Mean position error of GSM TC track predictions in the western North Pacific from 1996 to 2017. The lines represent 24- (red), 48- (green) and 72-hour forecasts (blue).

### 3.3 Ensemble Prediction Systems

#### 3.3.1 Introduction

JMA operates its Global EPS and Seasonal EPS ensemble prediction systems to support forecasting work over a wide variety of prediction periods from early medium-range to seasonal.

The Global EPS, which has been operational since January 2017, produces forecasts with lead times of up to 34 days to support the issuance of Five-day Tropical Cyclone (TC) Forecasts, One-week Forecasts, Early Warning Information on Extreme Weather, and One-month Forecasts. The system took over the roles of three previous JMA systems (the Typhoon EPS, the One-week EPS and the One-month EPS; e.g., JMA 2013, 2017; Yamaguchi *et al.* 2014; Hirai *et al.* 2014). The objectives of the integration were to utilize computational resources more effectively and to concentrate efforts into a single EPS system. The Typhoon EPS and the One-week EPS were replaced by the Global EPS in January 2017, and the system inherited the roles of the One-month EPS in March 2017.

The Seasonal EPS, which has been operational since March 2003, produces seven-month forecasts and supports Three-month Forecasts, Warm/Cold Season Forecasts and El Niño Outlooks.

#### 3.3.2 In Operation

##### 3.3.2.1 System Configuration

The specifications of the Global EPS and the Seasonal EPS are shown in Table 3.3.1.

A low-resolution version of JMA's Global Spectral Model (GSM; see Section 3.2) is used in the Global EPS. Accordingly, the dynamical framework and physical processes involved are essentially identical to those of the GSM except for horizontal resolution which varies at the forecast lead time of 18 days. Unperturbed analysis for this EPS is prepared by interpolating the analysis field in global analysis (see Section 2.5). The sea surface temperature (SST) field obtained from JMA's SST analysis (see Subsection 5.2.1) is used as a lower boundary condition. The SST anomaly at the initial time persists during time integration. The sea ice concentration analysis value is also prescribed using the persisting anomaly for forecasts up to 14 days ahead. For forecasts after this time period, sea ice concentration is prescribed by adjusting the previous day's

Table 3.3.1: Specifications of JMA Global EPS and Seasonal EPS

	Global EPS	Seasonal EPS
Start of operation (Latest major implementation)	January 2017 (January 2017)	March 2003 (June 2015)
Initial time	00, 06, 12 and 18 UTC	00 UTC
Forecast range	Initial time 00, 12 UTC: 34 days on Tuesdays and Wednesdays 18 days on Saturdays and Sundays 11 days otherwise Initial time 06, 18 UTC: 132 hours	7 months
Ensemble size	Forecasts up to 11 days: 27 members Forecasts longer than 11 days: 13 members (50-member lagged ensemble with 4 initial times)	13 members (51-member lagged ensemble with 4 initial times)
Model type	GSM (an atmospheric general circulation model)	GSM coupled with the Meteorological Research Institute Community Ocean Model (MRI.COM) (a coupled atmosphere-ocean general circulation model)
Horizontal resolution	Forecasts up to 18 days: Spectral triangular 479 (TL479), reduced Gaussian grid system, roughly equivalent to $0.375^\circ \times 0.375^\circ$ (40 km) in latitude and longitude Forecasts longer than 18 days: Spectral triangular 319 (TL319), reduced Gaussian grid system, roughly equivalent to $0.5625^\circ \times 0.5625^\circ$ (55 km) in latitude and longitude	GSM: Spectral triangular 159 (TL159) reduced Gaussian grid system, roughly equivalent to $1.125^\circ \times 1.125^\circ$ (110 km) MRI.COM: $0.3\text{--}0.5^\circ \times 1.0^\circ$ in latitude and longitude
Vertical resolution (model top)	100 stretched sigma pressure hybrid levels (0.01 hPa)	GSM: 60 stretched sigma pressure hybrid levels (0.1 hPa) MRI.COM: 52 levels and a bottom boundary layer
Initial perturbation generator	SV method, LETKF and LAF method	Atmosphere: BGM method and LAF method Ocean: Perturbed atmospheric forcings and LAF method
Initially perturbed area	Global	Atmosphere: The Northern Hemisphere ( $20^\circ\text{N}$ – $90^\circ\text{N}$ ) and the tropics ( $20^\circ\text{S}$ – $20^\circ\text{N}$ ) Ocean: Global
Model ensemble method	Stochastic physics scheme	
Surface boundary perturbations	SST perturbations	N/A

distribution so that initial sea ice extent anomalies in each hemisphere persist.

JMA's Coupled atmosphere-ocean General Circulation Model (CGCM; Takaya *et al.* 2018) is used in the Seasonal EPS. Atmospheric and land surface initial conditions are obtained from the Japanese 55-year Reanalysis (JRA-55; Kobayashi *et al.* 2015), while oceanic and sea ice initial conditions are taken from MOVE/MRI.COM-G2 (Toyoda *et al.* 2013).

To support the issuance of Early Warning Information on Extreme Weather, One-month Forecasts, Three-month Forecasts, Warm/Cold Season Forecasts and El Niño Outlooks, the models' systematic biases are corrected using the estimated by using the mean forecast error from hindcast experiments.

### 3.3.2.2 Ensemble Size, Forecast Range and Frequency

#### 1. Global EPS

The Global EPS consists of 27 forecast runs implemented up to four times a day from initial times at 00, 06, 12 and 18 UTC. Runs from the initial times of 00 and 12 UTC have a forecast range of 11 days and are used for One-week Forecasts. The forecast range is extended to 18 days on Saturdays and Sundays, and to 34 days on Tuesdays and Wednesdays. The ensemble size of forecasts longer than 11 days is 13 per initial time or 50 in total based on the combination of runs from the four initial times using the lagged averaged forecast (LAF) method. A 50-member lagged ensemble<sup>8</sup> with a forecast range of a month is used for the One-month Forecast issued on Thursdays. A 50-member lagged ensemble<sup>9</sup> with a forecast range of two weeks is used for Early Warning Information on Extreme Weather, which is issued on Mondays and Thursdays when a high probability of seven-day-average very high or very low temperatures or heavy snow is predicted in the week starting five to eight days ahead of the date of announcement. Runs from the initial times of 06 and 18 UTC have a forecast range of 132 hours and are operated when any of the following conditions is satisfied:

- A TC of tropical storm (TS<sup>10</sup>) intensity or higher is present in the RSMC Tokyo - Typhoon Center's area of responsibility ( $0^{\circ}$ – $60^{\circ}$ N,  $100^{\circ}$ E– $180^{\circ}$ ).
- A TC is expected to reach TS intensity or higher in the area within the next 24 hours.
- A TC of TS intensity or higher is expected to move into the area within the next 24 hours.

These 132-hour forecasts and those with initial times at 00 and 12 UTC are used for Five-day Tropical Cyclone Forecasts.

#### 2. Seasonal EPS

The Seasonal EPS consists of 13 forecast runs from an initial time at 00 UTC with a forecast range of seven months. With the initial perturbation method, 13-member ensemble predictions are made every five days from the 1st of January onward. The four latest LAF dates are combined for the provision of a 51-member ensemble to support monthly Three-month Forecasts and for Warm/Cold Season Forecasts issued in February, March, April, September and October. The EPS is also used for the monthly El Niño Outlook.

### 3.3.3 Approach to Ensemble Initial Conditions

In addition to the LAF method, three other approaches are employed in the perturbation of initial conditions for the atmosphere. One is the singular vector (SV) method (Buizza and Palmer 1995), which is used to generate initial perturbation in the Global EPS. Another approach based on the Local Ensemble Transform Kalman Filter (LETKF, Hunt *et al.* 2007) is also used for the Global EPS. The breeding of growing modes (BGM) method (Toth and Kalnay 1993, 1997) is used for the Seasonal EPS. The subsections below describe

<sup>8</sup>Composed of 13 members from 12 UTC on Wednesdays, 13 members from 00 UTC on Wednesdays, 13 members from 12 UTC on Tuesdays and 11 members from 00 UTC on Tuesdays.

<sup>9</sup>Composed in the same way with the One-month Forecast for Thursday's issuance. Likewise, runs from initial times on Saturdays and Sundays are used for Monday's issuance.

<sup>10</sup>A TS is defined as a TC with maximum sustained wind speeds of 34 knots or more and less than 48 knots.

Table 3.3.2: Specifications of the LETKF used in Global EPS

Horizontal resolution	TL319 reduced Gaussian grid system, roughly equivalent to $0.5625^\circ \times 0.5625^\circ$ (55 km) in latitude and longitude
Vertical resolution (model top)	100 stretched sigma pressure hybrid levels (0.01 hPa)
Ensemble size	50 members
Analysis variables	Wind, surface pressure, specific humidity and temperature
Analysis time	00, 06, 12, 18 UTC
First guess	6-hour forecast of its own
Assimilation window	6 hours ( $\pm 3$ hours of analysis time)
Data cut-off time	2 hours and 20 minutes
Observation	Same as global early analysis except for AIRS, IASI and CrIS
Observation operator	RTTOV version 10.2 for satellite radiance ROPP version 8.0 for GNSSRO
Covariance inflation	Adaptive multiplicative covariance inflation
Initialization	Horizontal divergence adjustment based on the analysis of surface pressure tendency (Hamrud <i>et al.</i> 2015)
Model ensemble method	Stochastic physics scheme
Other characteristics	50 analyses are recentered so that the ensemble mean of them become consistent to the analysis of the Global Analysis (GA)

the specifications of these methods and outline how atmospheric ensemble initial conditions are generated for each EPS.

For the Seasonal EPS, initial perturbations for the ocean are also generated, in addition to those for the atmosphere, in two staggered ocean data assimilation streams with a 10-day assimilation window, forced with surface heat and momentum fluxes in the atmospheric initial perturbation fields.

### 3.3.3.1 LETKF Method

The specifications of the LETKF approach used in the Global EPS to represent uncertainties in initial conditions are listed in Table 3.3.2.

Observation datasets assimilated in the LETKF are the same as those of global early analysis (Table 2.2.1) except for those of hyperspectral sounders (AIRS, IASI and CrIS).

Observation localization is applied in the LETKF. Observation errors are multiplied by the inverse of the localization function to give less weight to data collected farther from the analysis grid point. The localization function is given as the Gaussian function of the distance between the analysis grid point and the observation location. The localization scale  $l$  for which the localization function is  $1/\sqrt{e}$  is set to 400 km in the horizontal domain, a 0.4 scale height in the vertical domain (0.8 for surface pressure and ground-based GNSS zenith total-delay observations) and 3 hours in the temporal domain. The tail of the localization function is set to 0 farther than  $2\sqrt{10/3}l$ . For satellite radiance observations, the weighting function divided by its peak value is used as the vertical localization function.

Multiplicative covariance inflation (Anderson 2001) is applied to first-guess (6-hour forecast from the previous analysis) ensembles. Inflation coefficients are estimated for each analysis grid point so that the following relation (Desroziers *et al.* 2005) is observed using locally assimilated observations:

$$tr[\mathbf{d}_{A-B}\mathbf{d}_{O-B}^T] = \alpha tr[\mathbf{H}\mathbf{B}\mathbf{H}^T] \quad (3.3.1)$$

where  $\mathbf{d}_{A-B}$ ,  $\mathbf{d}_{O-B}$ ,  $\mathbf{H}$  and  $\mathbf{B}$  represent the analysis increment projected onto the observation space, innovations, observation operator and background error covariance, respectively.  $tr[]$  represents the trace of the matrix, and  $\alpha$  represents the inflation coefficient. Based on (3.3.1), the raw inflation coefficient is estimated on analysis

grid point  $j$  as

$$\alpha_{j,raw} = \frac{\sum_{m=1}^{N_{obs}} \rho_{m,j} (d_{A-B,m} d_{O-B,m} / \sigma_{o,m}^2)}{\sum_{m=1}^{N_{obs}} [\rho_{m,j}^2 (\mathbf{H}_m \mathbf{X}_b) (\mathbf{H}_m \mathbf{X}_b)^T / (K - 1) / \sigma_{o,m}^2]} \quad (3.3.2)$$

where  $N_{obs}$  is the number of observations assimilated locally,  $K$  is the ensemble size,  $\rho_{m,j}$  is the localization function of the  $m$ th observation and  $\mathbf{H}_m \mathbf{X}_b$  is the first-guess perturbation projected onto the  $m$ th observation. The subscript *raw* represents a raw estimate.  $d_{A-B,m}$  and  $d_{O-B,m}$  are the observational increment and innovation of the  $m$ th observation, where  $d_{A-B,m}$  is computed with a transformation matrix derived from LETKF analysis on grid  $j$ . Temporal smoothing is applied to the estimated inflation coefficient as

$$\alpha_{i,j} = \frac{\alpha_{i-1,j} \sigma_{o,j}^2 + \alpha_{i,j,raw} \sigma_b^2}{\sigma_{o,j}^2 + \sigma_b^2} \quad (3.3.3)$$

where the subscript  $i$  represents the value at the  $i$ th analysis step, and  $\sigma_{o,j}^2$  and  $\sigma_b^2$  are error variances of the estimated and prior coefficients and are set as

$$\sigma_{o,j}^2 = 1 / \sum_{m=1}^{N_{obs}} \rho_{m,j} \quad (3.3.4)$$

$$\sigma_b^2 = 0.005 \quad (3.3.5)$$

Finally, the estimated coefficients are relaxed to the default values as the deviation from these defaults reaches  $1/e$  at 10 days. The defaults are set to 1.21 from the surface to 45 hPa and linearly reduced with the logarithm of pressure to 1.0 at 0.85 hPa. Inflation to specific humidity is reduced by 30% of the estimated inflation coefficients below 230 hPa and linearly decreases with the logarithm of pressure to about 15% at 45 hPa.

Initialization based on analysis of surface pressure tendency (Hamrud *et al.* 2015) is applied after the LETKF analysis update. The surface pressure tendency of the first-guess ensembles  $(\partial p_s / \partial t)_{guess}$  is diagnosed from the continuity equation and the hydrostatic balance as

$$\left( \frac{\partial p_s}{\partial t} \right)_{guess} = - \int_0^1 \nabla \cdot \left( \mathbf{v}_{guess} \frac{\partial p_{guess}}{\partial \eta} \right) d\eta \quad (3.3.6)$$

where  $p_s$ ,  $\mathbf{v}$ ,  $p$  and  $\eta$  are surface pressure, horizontal wind, pressure and the model vertical coordinates (1 at the bottom and 0 at the top), respectively. Surface pressure tendency is analyzed by adding the above variable to the first-guess state variables of the LETKF. The difference in surface pressure tendency diagnosed from horizontal wind and surface pressure analysis and that observed from LETKF analysis is distributed to each model layer so that the value is proportional to the analysis spread of horizontal wind. Horizontal divergence is adjusted using

$$\Delta [\nabla \cdot (\mathbf{v}_k dp_k)] = w_k \left[ \left( \frac{\partial p_s}{\partial t} \right)_{diag} - \left( \frac{\partial p_s}{\partial t} \right)_{anal} \right] \quad (3.3.7)$$

where  $k$  is an index of the vertical model layer,  $w_k$  is the weight on the  $k$ th model layer and  $dp_k$  is the difference in half-level pressure adjacent to the  $k$ th full-level model layer. The horizontal divergence increment is derived from (3.3.7) divided by  $dp_k$ . The increment is multiplied by the square of the cosine of latitude and added to each analysis member.

The analysis ensemble is recentered so that the ensemble mean is consistent with global early analysis.

The initial perturbations for the Global EPS are derived by selecting 26 of the 50 analysis members and subtracting the ensemble mean of the selected members. The perturbations are multiplied by 0.9 and added to the initial perturbations derived using the SV method (Subsection 3.3.3.2).

The stochastic physics scheme (Subsection 3.3.4) is also applied to the forecast ensemble of the LETKF.

Table 3.3.3: Specifications of SV calculation

Resolution	Spectral triangular truncation 63 (TL63), 40 levels		
Norm	Moist total energy		
Target area	Northern Hemisphere (30°N–90°N)	Southern Hemisphere (90°S–30°S)	Tropics (30°S–30°N)
Physical process	Simplified physics		Full physics
Optimization time	48 hours		24 hours
Number of perturbations	13		

### 3.3.3.2 SV Method

Table 3.3.3 summarizes the specifications of SV calculation for the Global EPS. The tangent-linear and adjoint models used for SV computation are lower-resolution versions of those used in the JMA’s 4D-Var system (see Section 2.5) until May 2017. The models involve full dynamical core and physical processes including surface turbulent fluxes, vertical turbulent transports, gravity wave drag, long-wave radiation, clouds and large-scale precipitation, and cumulus convection. SVs based on tangent-linear and adjoint models incorporating full physical processes are called moist SVs, while those based on models incorporating simplified physical processes involving surface fluxes and vertical diffusion are called dry SVs.

#### 1. SV definition

Three SV calculations for different target areas are conducted. The targeted areas are the Northern Hemisphere (30°N–90°N), the tropics (30°S–30°N) and the Southern Hemisphere (90°S–30°S). Dry SVs with a 48-hour optimization time are computed for the Northern Hemisphere and the Southern Hemisphere, while moist SVs with a 24-hour optimization time are computed for the tropics.

#### 2. Norm of SV calculation

The norm for evaluating the growth rate of dry and moist SVs is based on a total energy norm that includes a specific humidity term (Ehrendorfer *et al.* 1999):

$$(x, Ey) = \frac{1}{2} \int_0^1 \int_S \left[ U_x U_y + V_x V_y + \frac{c_p}{T_r} T_x T_y + w_q \frac{L_c^2}{c_p T_r} q_x q_y \right] dS \left( \frac{\partial p}{\partial \eta} \right) d\eta + \frac{1}{2} \int_S \left[ \frac{R_d T_r}{P_r} P_x P_y \right] dS. \quad (3.3.8)$$

Here,  $U_x$ ,  $V_x$ ,  $T_x$ ,  $q_x$  and  $P_x$  are the zonal wind, meridional wind, temperature, specific humidity and surface pressure components of state vector  $x$  respectively, and  $(x, Ey)$  is an inner product of state vectors  $x$  and  $y$  with a norm operator  $E$ .  $c_p$  is the specific heat of dry air at a constant pressure,  $L_c$  is the latent heat of condensation, and  $R_d$  is the gas constant for dry air.  $T_r = 300$  K is a reference temperature,  $P_r = 800$  hPa is a reference pressure, and  $w_q$  is a constant (here 0.04).  $\int dS$  is the horizontal integration for the whole globe, and  $\int \left( \frac{\partial p}{\partial \eta} \right) d\eta$  gives the vertical integration from the surface to the model top. In addition, the norm at the initial time is vertically integrated with a weight that depends on the model level; the kinetic energy term and the available potential energy term are multiplied by a factor of  $10^3$  above the 35th model level, and the specific humidity term is multiplied by a factor of  $10^3$  above the 9th model level. When the surface pressure is 1,000 hPa, the 35th and 9th model levels correspond to about 10 and 750 hPa, respectively. This suppresses initial perturbation around the model top and confines initial specific humidity perturbation in the lower troposphere.

#### 3. Generation of SV-based perturbations

The SV-based perturbations are linear combinations of SVs. A total of 25 SVs are created for each targeted area. In this creation procedure, SVs with extremely high growth rates (which will not grow

sufficiently in a nonlinear model) and SVs with a high level of similarity to others can be eliminated. The 25 combined SVs are transformed in a variance minimum rotation (Yamaguchi *et al.* 2009) to generate 13 SV-based perturbations for each targeted area. The perturbations for the Northern Hemisphere and the Southern Hemisphere are scaled so that their amplitudes of temperature at the 15th model level (or the 6th model level for the tropics) inside the targeted area become 0.23 K (or 0.20 K for the tropics). When the surface pressure is 1,000 hPa, the 15th and 6th model levels correspond to about 500 and 850 hPa, respectively. The perturbations for the three targeted area are linearly combined to create global perturbations.

Finally, the initial conditions of 26 perturbed members are given by adding and subtracting those 13 SV-based perturbations and adding 26 LETKF-based perturbations to unperturbed analysis.

### 3.3.3.3 BGM Method

The processes of the BGM method with separate estimation for the Northern Hemisphere ( $20^{\circ}\text{N}$ – $90^{\circ}\text{N}$ ) and the tropics ( $20^{\circ}\text{S}$ – $20^{\circ}\text{N}$ ), are described here. First, perturbed and unperturbed initial conditions are integrated up to 12 hours for the Northern Hemisphere and 48 hours for the tropics. Then, the difference between the two fields is normalized so that the area-averaged root mean square of the difference for 500-hPa height over the Northern Hemisphere and 200-hPa velocity potential for the tropics (Chikamoto *et al.* 2007) are equal to 14.5 and 20.0 % of the climatological variance, respectively. Third, the normalized perturbations are orthogonalized to each other and added to the analysis to create the next set of initial perturbations. In the Seasonal EPS, the Northern Hemisphere and tropical initial perturbations are combined and added to/subtracted from the analysis.

### 3.3.4 Model Ensemble Approach

The stochastic physics scheme (Buizza *et al.* 1999) is used in the Global EPS and the Seasonal EPS in consideration of model uncertainties associated with physical parameterizations. This scheme represents random errors associated with parameterized physical processes as follows:

$$\frac{\partial \mathbf{x}}{\partial t} = F(\mathbf{x}) + \alpha(\lambda, \phi, t)P(\mathbf{x}). \quad (3.3.9)$$

Here  $t$ ,  $\mathbf{x}$ ,  $F(\mathbf{x})$  and  $P(\mathbf{x})$  are the time, the set of forecast variables, the total tendency of the forecast model and the tendency of the parameterized physical processes, respectively.  $\lambda$  and  $\phi$  show latitude and longitude;  $\alpha(\lambda, \phi, t)$  is a random variable described in a spectral space (Berner *et al.* 2009) featuring spatial correlation with a total wave number of 20 and a time correlation of six hours. The average of  $\alpha$  is set to zero. Its value is limited to a specific range (-0.7 to 0.7 for the Global EPS, and -0.75 to 0.75 for the Seasonal EPS) to avoid excessive perturbations, and its value in the stratosphere is also set to zero.

### 3.3.5 Sea Surface Temperature Perturbations

The same SST and sea ice concentration as in the high-resolution deterministic forecast are used in the unperturbed member of the Global EPS (see Subsection 3.2.11 for details). Perturbations are added to the SST of perturbed members to represent the uncertainty in that of the unperturbed member. Perturbation of SST for member  $i$  ( $\Delta SST_i^f$ ) in the forecast from the initial time  $T_0$  is constructed as

$$\Delta SST_i^f(T_0; t_f) = \alpha [SSTA(T_i + \Delta T + t_f) - SSTA(T_i)] \quad (3.3.10)$$

where  $SSTA$  is the SST anomaly from the daily climatology,  $t_f$  is the forecast lead time and  $T_i$  is a past date randomly sampled from the range within  $\pm 27$  days from that of the current initial time.  $\Delta T$  and  $\alpha$  are tunable parameters, and are set to 1 day and 1.0, respectively.

Perturbation of SST is set to 0 where the maximum sea ice concentration on  $T_0$ ,  $T_i$  and  $T_i + \Delta T + t_f$  is above a certain threshold (0.001) to prevent the use of excessively large perturbations.

### 3.3.6 Performance

The performance of each EPS product is described below. For the sake of completeness, the period before Global EPS operation is also incorporated.

#### 3.3.6.1 Typhoon Forecasting

Typhoon forecasting is supported by the Global EPS, as it was previously by the Typhoon EPS. The results of related verification are provided in the Annual Report on Activities of the RSMC Tokyo - Typhoon Center<sup>11</sup>.

Ensemble TC tracks derived from the EPS enable JMA forecasters to integrate TC track forecast uncertainty into their operational processes. Strike probability data, which indicate the chances of a TC center passing within 120 km of a grid point, are routinely produced as a form of probabilistic guidance. Figure 3.3.1 shows the reliability of typhoon strike probability data for the coming five days<sup>12</sup>. Typhoon track errors were improved by replacement with the Global EPS in 2017. However, the spread of this EPS was greater after a forecast lead time of 72 hours than that of the Typhoon EPS, but lower for forecast lead times of 24 and 48 hours (not shown). Accordingly, the 2017 curve is considered to show relatively high departure from the diagonal, especially in high-probability areas.

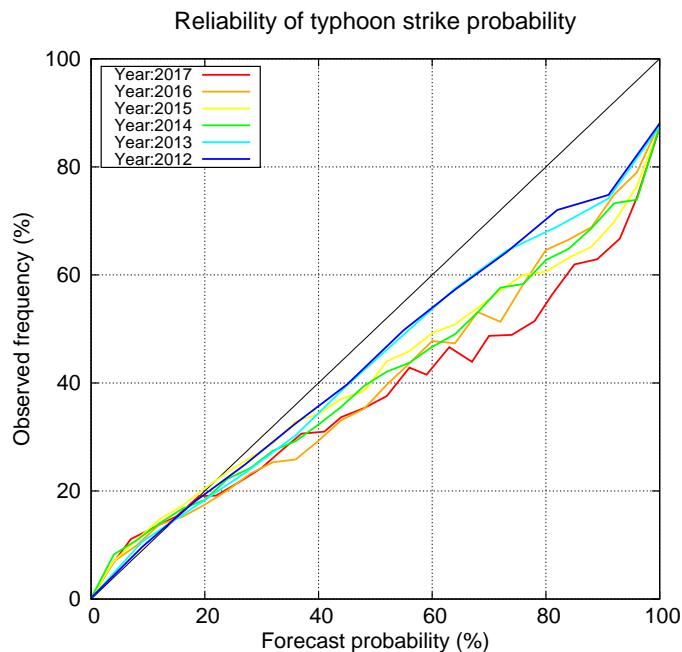


Figure 3.3.1: Reliability diagram for probabilistic verification of typhoon position forecasts as derived from the EPS over a six-year period. The target years for verification are 2012 (blue), 2013 (sky blue), 2014 (light green), 2015 (yellow), 2016 (orange) and 2017 (red). RSMC Tropical Cyclone Best Track information is used as observation data.

#### 3.3.6.2 One-week Forecasting

One-week forecasting is supported by the Global EPS, it was previously by the One-week EPS. The results of related verification are provided in the annual WMO Technical Progress Report on the Global Data-processing

<sup>11</sup><http://www.jma.go.jp/jma/jma-eng/jma-center/rsmc-hp-pub-eg/annualreport.html>

<sup>12</sup>The 2014 result was calculated using forecast data only for the period from April to December to maintain verification data consistency, as a major upgrade to the Typhoon EPS in March 2014 caused a change in score characteristics.

and Forecasting System (GDPFS). Monthly verification data are also provided on the website of the WMO/CBS Lead Centre for EPS Verification<sup>13</sup>.

Figure 3.3.2 shows a time-series representation of monthly-averaged root mean square errors (RMSEs) for the 500-hPa geopotential height ensemble mean forecast against analysis for the Northern Hemisphere (NH; 20°N–90°N). Figure 3.3.3 compares RMSEs of ensemble means, unperturbed members and the spread of the ensemble averaged for the periods of DJF (December/January/February) 2017/2018 and JJA (June/July/August) 2018. A higher level of skill is observed for ensemble means than for unperturbed members, especially for longer lead times. For shorter forecast lead times, the spread is almost the same size as the ensemble means RMSE, but as the forecast lead time increases it tends to become slightly smaller. Figure 3.3.4 shows the Brier skill score (BSS) for 500-hPa geopotential height probabilistic forecasts in the NH. The reference forecast for the skill score is the climatological probability given by the frequency derived from analysis fields for each month. Since the start of operation, performance has improved annually in ensemble mean forecasts and probabilistic forecasts.

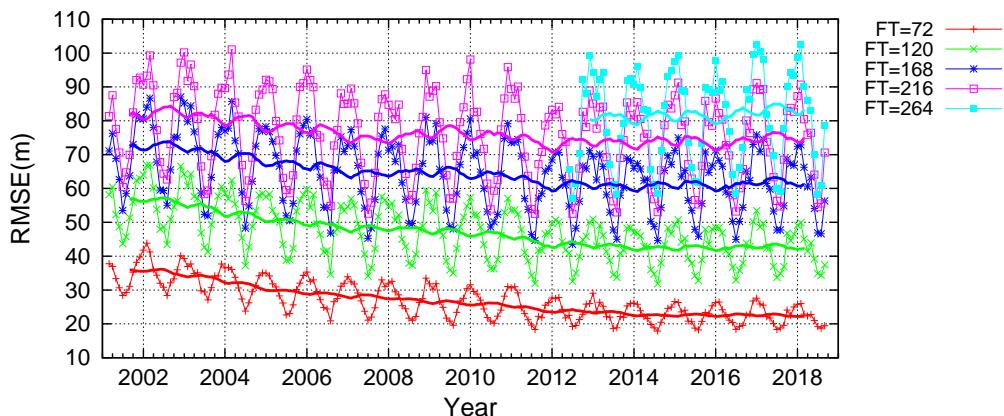


Figure 3.3.2: Time-series representation of ensemble mean scores for the EPS (where the score is the monthly-averaged RMSE of the ensemble mean) for Northern Hemisphere (20°N–90°N) 500-hPa geopotential height forecasts with lead times of 72 (red), 120 (green), 168 (blue), 216 (violet) and 264 (sky blue) hours from March 2001 to September 2018. The thick lines show 13-month running means.

### 3.3.6.3 One-month Forecasting

One-month forecasting is supported by the Global EPS, as it was previously by the One-month EPS. The results of prediction skill evaluation based on hindcast experiments and real-time forecasts are provided on the Tokyo Climate Center website<sup>14</sup>. The hindcast experiments, covering a period of 30 years (1981–2010) and involving five ensemble members, were conducted with atmospheric initial conditions produced from JRA-55. Initial perturbations were created from a combination of initial and evolved SVs, based on the SV method. Perturbations from the LETKF method used in the real-time operational system were not adopted to reduce computational cost.

The skill of ensemble mean forecasts was evaluated using the Anomaly Correlation Coefficient (ACC) and the RMSE for selected areas with respect to several physical variables. Probabilistic forecast skill was also evaluated based on the BSS, the Reliability Skill Score (Brel), the Resolution Skill Score (Bres) and Relative Operating Characteristics (ROC).

<sup>13</sup><http://epsv.kishou.go.jp/EPsv/>

<sup>14</sup><http://ds.data.jma.go.jp/tcc/tcc/index.html>

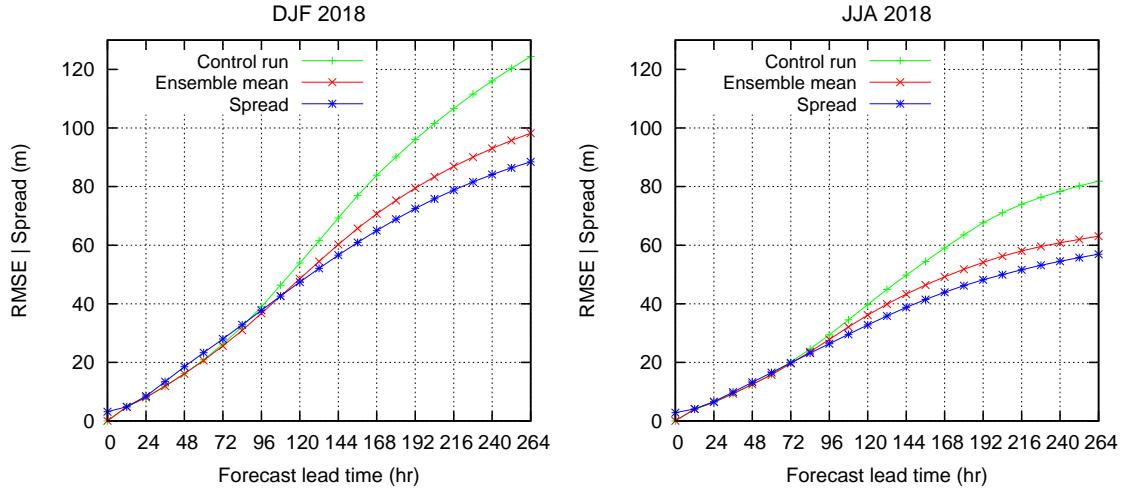


Figure 3.3.3: RMSEs for Northern Hemisphere ( $20^{\circ}\text{N}$ – $90^{\circ}\text{N}$ ) 500-hPa geopotential height forecasts of the ensemble mean (red) and unperturbed members (green) for DJF and JJA 2018 from the EPS. The spread of the ensemble (blue) is also shown.

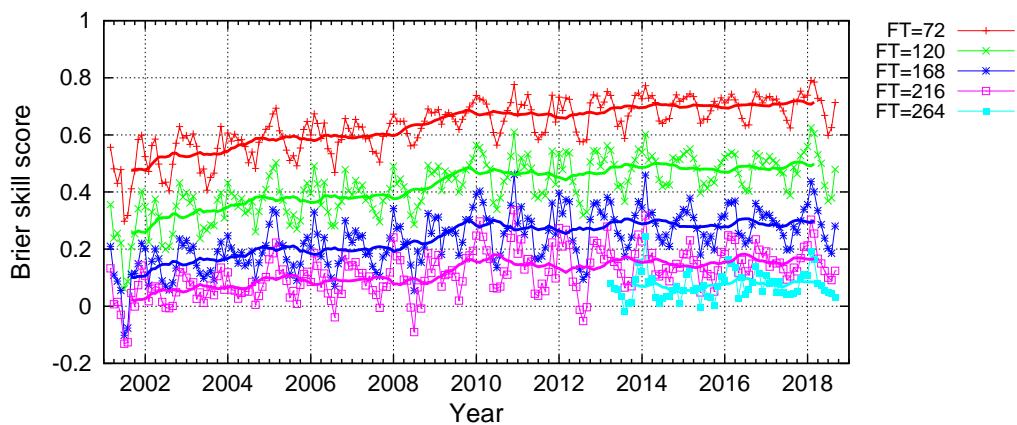


Figure 3.3.4: Same as Figure 3.3.2, but for Brier skill score for probabilistic forecasts of 500-hPa geopotential height negative anomalies with magnitudes less than one climatological standard deviation from JMA's EPS.

Table 3.3.4: ROC areas of 28-day (from day 3 to day 30) mean 2-m temperature (T2m) and 500-hPa geopotential height (Z500) anomaly prediction for positive anomaly events (upper tercile) in the Northern Hemisphere (NH; 20°N–90°N), the tropics (20°S–20°N), and the Southern Hemisphere (SH; 90°S–20°S) based on hindcast experiments covering a period of 30 years (1981–2010). The figures in the table are multiplied by 100. The initial dates are 31 December for January and 30 June for July.

T2m	NH	Tropics	SH	Z500	NH	Tropics	SH
January(Initial:12/31)	74.5	77.3	72.5	January(Initial:12/31)	73.2	91.8	75.4
July(Initial:6/30)	72.6	74.3	68.9	July(Initial:6/30)	72.4	79.3	68.8

Figure 3.3.5 shows a time-series representation of the NH 500-hPa geopotential height ACC for ensemble mean forecasts averaged over 28 days from day 2 to day 29 (the running mean of 52 forecasts) based on operational forecasting conducted from 1997 to 2018. It can be seen that the skill represents a rising trend with fluctuations corresponding to ENSO events. Table 3.3.4 shows ROC areas of 2-m temperature (T2m) and precipitation anomalies based on the hindcast experiments, and indicates that the skill for the tropics is higher than that for the extratropics.

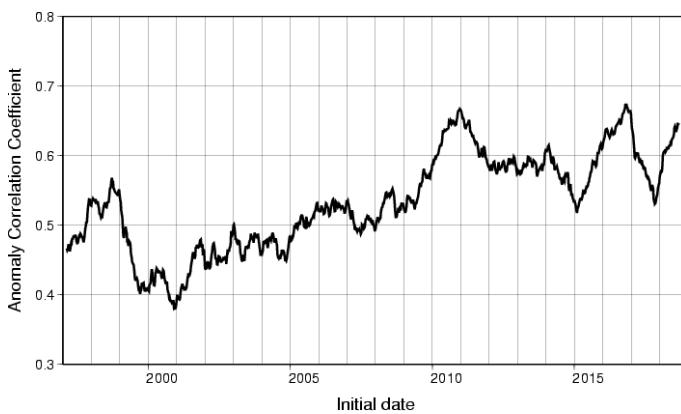


Figure 3.3.5: Time-series representation of the Northern Hemisphere (NH; 20°N–90°N) 500-hPa geopotential height anomaly correlation coefficient in ensemble mean forecasts averaged over 28 days from day 2 to day 29 (the running mean of 52 forecasts) based on operational forecasting conducted from 1997 to 2018

### 3.3.6.4 Seasonal Forecasting

Seasonal forecasting is supported by the Seasonal EPS. The results of prediction skill evaluation based on the WMO Standard Verification System for long-range forecasts (SVS-LRF; WMO 2010) are available on the Tokyo Climate Center website. To verify performance, hindcast experiments covering a period of 30 years (1981–2010) were conducted under conditions identical to those of the operational system, except with an ensemble size of 10 instead of 51.

Figure 3.3.6 shows the ACC between ensemble mean forecasts and observations for SSTs, averaged over the 30 years, in the NINO.3 (5°S–5°N, 150°W–90°W), NINO.WEST (0°–15°N, 130°E–150°E) and IOBW (20°S–20°N, 40°E–100°E) regions. The SST over NINO.3, used as an important indicator in JMA's El Niño outlook, is predicted more accurately than that over NINO.WEST. ROC areas of T2m anomalies and Z500

anomalies are shown in Table 3.3.5. Skill for the tropics is superior to that for the NH and SH. These results are consistent with those obtained from studies on the predictability of seasonal mean fields (e.g., Sugi *et al.* 1997).

Table 3.3.5: ROC areas of three-month means (JJA and DJF) 2-m temperature (T2m) and 500-hPa geopotential height (Z500) anomaly prediction for positive anomaly events (upper tercile) in the Northern Hemisphere (NH; 20°N–90°N), the tropics (20°S–20°N), and the Southern Hemisphere (SH; 90°S–20°S) based on hindcast experiments covering a period of 30 years (1981–2010). The figures in the table are  $\times 100$  values. The initial dates are 26 April for JJA and 28 October for DJF.

T2m	NH	Tropics	SH	Z500	NH	Tropics	SH
JJA(Initial:4/26)	66.3	74.3	62.7	JJA(Initial:4/26)	66.2	83.6	62.9
DJF(Initial:10/28)	65.6	79.0	62.9	DJF(Initial:10/28)	63.0	94.7	69.1

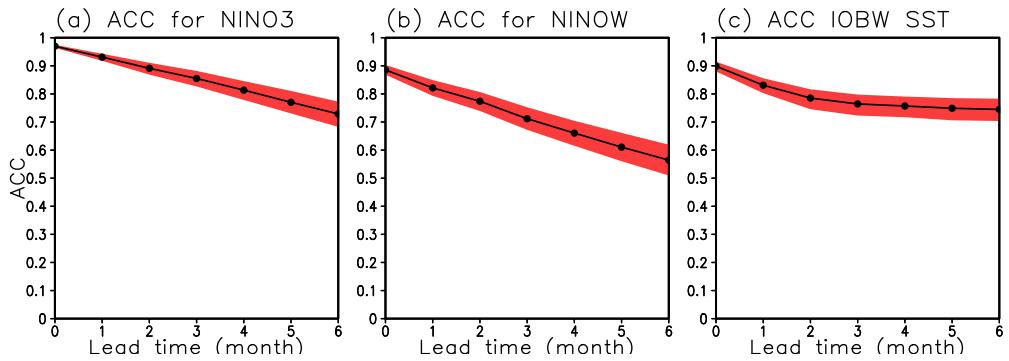


Figure 3.3.6: Anomaly correlations for SSTs, averaged over the 30 years, in (a) NINO.3 ( $5^{\circ}\text{S}$ – $5^{\circ}\text{N}$ ,  $150^{\circ}\text{W}$ – $90^{\circ}\text{W}$ ), (b) NINO.WEST ( $0^{\circ}$ – $15^{\circ}\text{N}$ ,  $130^{\circ}\text{E}$ – $150^{\circ}\text{E}$ ) and (c) IOBW ( $20^{\circ}\text{S}$ – $20^{\circ}\text{N}$ ,  $40^{\circ}\text{E}$ – $100^{\circ}\text{E}$ ). Shading indicates a 90% confidence interval as estimated using the bootstrap method (1,000 samples).

## 3.4 Coupled Atmosphere-Ocean General Circulation Model

### 3.4.1 Model Description

JMA introduced its first coupled atmosphere-ocean general circulation model (CGCM) for the El Niño-Southern Oscillation (ENSO) Outlook in July 1998. Later in February 2010, the CGCM was integrated into the Seasonal Ensemble Prediction System (Seasonal EPS) to provide seasonal climate outlooks, including the ENSO Outlook.

The latest model, jointly developed by the Meteorological Research Institute and the Climate Prediction Division of JMA (JMA/MRI-CGCM2; Takaya *et al.* 2018), was put into operation in June 2015. The model consists of atmospheric and oceanic general circulation models with the SCUP Simple Coupler (Yoshimura and Yukimoto 2008). Table 3.4.1 summarizes the model configurations.

The atmospheric model is based on a low-resolution version of JMA's Global Spectral Model as of 2011 (GSM1011C; JMA 2013), with a horizontal resolution of TL159 (triangular truncation at total wavenumber 159 with a linear grid) corresponding to 110-km grid spacing, and 60 vertical levels with the model top placed at 0.1 hPa. Several parameterization schemes were upgraded to improve representation of atmospheric and oceanic states. These include: entraining sub-cloud plumes (Jakob and Siebesma 2003), independent column

Table 3.4.1: Specifications of the Coupled General Circulation Model

Atmospheric component	Basic equation	Primitive
	Domain	Global
	Resolution	TL159, 60 vertical levels
	Radiation	Two-stream with delta-Eddington approximation for short wave (hourly) Two-stream absorption approximation method for long wave (3 hourly)
	Cumulus convection	Prognostic Arakawa-Schubert scheme
	Land surface process	Simple Biosphere (SiB) scheme
	Planetary boundary layer	Mellor and Yamada Level-2
	Gravity wave drag	Longwave orographic drag scheme (wavelengths > 100km) mainly for stratosphere Shortwave orographic drag scheme (wavelengths $\approx$ 10km) for troposphere only
	Coupling	Frequency
Oceanic component	Basic equation	Primitive, free surface
	Domain	Global
	Resolution	$1^\circ(\text{lon}) \times 0.5^\circ - 0.3^\circ(\text{lat})$ , 52 vertical levels and a bottom boundary layer
	Sea Ice	Mellor and Kantha (1989), Hunke and Lipscomb (2006)
	Vertical diffusion	Noh and Kim (1999)
Coupling	Frequency	Every hour

approximation for cloud overlap (Nagasawa 2012), COARE3.0 sea-surface flux (Fairall *et al.* 2003) with diurnally varying SST (Zeng and Beljaars 2005; Takaya *et al.* 2010), subtropical marine stratocumulus (Kawai 2013) and ocean-current coupling (Luo *et al.* 2005). The minimum entrainment rate in the deep convection scheme (JMA 2013) is modified to follow Tokioka *et al.* (1988). The Simple Biosphere (SiB) land model (Sato *et al.* 1989; Sellers *et al.* 1986) is used in the same configuration as GSM1011C. Model uncertainty is represented with a stochastic physics scheme (Yonehara and Ujiiie 2011; Buizza *et al.* 1999), which is applied to all ensemble members. JMA/MRI-CGCM2 takes into account realistic historical concentrations of six well-mixed GHGs (CO<sub>2</sub>, N<sub>2</sub>O, CH<sub>4</sub>, CFC-11, CFC-12 and HCFC-22) based on historical records from 2005 and on the Coupled Model Intercomparison Project Phase 5 (CMIP5) Representative Concentration Pathway 4.5 (RCP4.5) scenario thereafter (van Vuuren *et al.* 2011).

The oceanic component is the Meteorological Research Institute Community Ocean Model (MRI.COM; Tsujino *et al.* 2010). For JMA/MRI-CGCM2, the model is configured to cover the whole globe with a tripolar grid at a horizontal resolution of  $1^\circ$  longitude and  $0.5^\circ$  latitude in the extratropics with meridional refinement from near the equator to  $0.3^\circ$  (see Figure 5.3.1). The model has 52 vertical layers, with an ocean bottom boundary layer (Nakano and Sugino 2002) in some parts of polar regions. Revised model parameterizations are adopted, including a tracer advection scheme with conservation of second-order moments (Prather 1986), a vertical diffusion scheme incorporating sea surface wave breaking effects (Noh and Kim 1999) and an isopycnal mixing scheme (Gent and McWilliams 1990). The prognostic sea ice scheme of MRI.COM treats formation, accretion, melting, and transfer of sea ice and snow. The sea ice model and ice-ocean coupling framework of Mellor and Kantha (1989) are adopted, with enhancements in areas such as thickness categories, ridging and rheology, following the Los Alamos sea ice model (CICE) version 3.14 (Hunke and Lipscomb 2006).

The atmospheric and oceanic models are integrated at different model time steps. The SCUP exchanges sea surface temperature, sea ice cover and sea surface fluxes (radiation, latent and sensible heat, momentum and fresh water) between the models every hour.

## 3.5 Meso-Scale Model (JMA-MSM1702)

### 3.5.1 Introduction

The meso-scale numerical prediction system has been operated since March 2001 to provide information for disaster prevention and aviation safety. The Meso-Scale Model (MSM) was initially a hydrostatic spectral model, producing 18-hour forecasts every 6 hours at 00, 06, 12 and 18 UTC. The domain covered Japan and its surrounding areas ( $3,600 \times 2,880$  km) at a horizontal resolution of 10 km with 40 vertical layers.

In September 2004, the MSM was replaced with a non-hydrostatic grid model (JMA-NHM; Saito *et al.* 2006, 2007) while retaining similar general configurations in areas such as resolution, forecast time and forecast frequency. In March 2006, the resolutions and operation frequency were enhanced to produce 15-hour forecasts every 3 hours at 00, 03, 06, 09, 12, 15, 18 and 21 UTC with 5-km horizontal grid spacing and 48 vertical layers. After subsequent model updates, the forecast period of the MSM was finally extended to 39 hours for all eight daily operations with an enlarged domain ( $4,080 \times 3,300$  km) in 2013. The extension of the forecast period and the model domain supported improved provision of useful information for disaster prevention and aviation operations with a one-day lead time.

In February 2017, the new-generation nonhydrostatic model ASUCA (Ishida *et al.* 2009, 2010) was incorporated into the operational MSM following its application for LFM usage in January 2015 (Aranami *et al.* 2015). The development of ASUCA was begun in 2007 after the development and widespread adoption of new nonhydrostatic equations allowing conservation of mass as well as sophisticated numerical methods in computational fluid dynamics. Efficient operation of numerical models on scalar multi-core architecture was also required against a background of rapid expansion in the market for massive scalar computers in the supercomputer field (Hara *et al.* 2012).

ASUCA has great potential to meet these demands. In the model, flux-form fully compressible governing equations are adopted and discretized using the finite volume method to guarantee mass conservation. The three-stage Runge-Kutta scheme (Wicker and Skamarock 2002) is employed for time integration, leading to better computational stability, even with a longer time-step interval, than the JMA-NHM. Improvement of parallelization and coding methods yields more effective computation on massive scalar multi-core architecture.

Physical processes equivalent to or better than those of the JMA-NHM are implemented via the use of the Physics Library<sup>15</sup>, in which various subroutines related to physical processes are collected as vertical one-dimensional models with unified coding and interface rules (Hara *et al.* 2012; Hara 2015). This simple one-dimensional realization helps to improve computational efficiency, especially on scalar computers, and facilitates efficient development of physical processes such as evaluation of the straightforward responses of specific processes of interest via idealized single-column model experiments.

As described above, the MSM was significantly upgraded with the introduction of ASUCA in February 2017. This section details the new MSM, with general configurations provided in Subsection 3.5.2. Subsection 3.5.3 describes the design of the dynamical core, and physical processes such as cloud physics, convective parameterization and radiation are detailed in the subsequent subsections. Improvement of the parallelization method in ASUCA is described in Subsection 3.5.10, and forecast performance is evaluated in Subsection 3.5.11.

### 3.5.2 General Configuration

The current ASUCA-based MSM is operated eight times a day, providing 39-hour forecasts every 3 hours at 00, 03, 06, 09, 12, 15, 18 and 21 UTC. Its forecast domain is a rectangular flat area of  $4,080 \times 3,300$  km covering Japan and its surroundings, with a grid spacing of 5 km. The domain configuration is identical to that of 4D-Var Meso-scale Analysis (MA; see Section 2.6) as depicted in Figure 2.6.2, but MA is still based on the JMA-NHM. The rectangular plane is determined via a Lambert conformal conic map projection of the Earth's sphere with a map scale factor applied to correct plane expansion or shrinkage associated with projection from the sphere. Hybrid terrain following the relevant coordinates is adopted for the vertical coordinate to reduce the influences of topography as height increases (Subsection 3.5.3). The lowest atmospheric layer is 10 m above the surface, and the model top is at 21,801 m with 76 layers at intervals increasing from 20 m at the bottom to approximately 650 m at the top.

The prognostic variables are horizontal and vertical momentum, mass-virtual potential temperature, total mass density, density of water vapor and hydrometeors (cloud water, cloud ice, rain, snow and graupel), ground temperature, soil water and four of the second-order moments of turbulent fluctuations (including turbulent kinetic energy). The model is operated with a 100/3-second time step.

Initial conditions for the model are generated via MA. Lateral boundary conditions come from the latest

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<sup>15</sup>The term ASUCA in this section refers to an NWP model incorporating physical processes from the Physics Library. The term sometimes refers only to the related dynamical core in a more narrow sense.

available GSM (Section 3.2) forecast with a 3- or 6-hour time lag. Thus, for example, the MSM at 03 and 06 UTC has lateral boundaries from the GSM initiated at 00 UTC.

The model terrain setting relies on the GTOPO30 data set, which is a global digital elevation model with a horizontal grid spacing of 30 arc seconds developed by the U.S. Geological Survey's EROS Data Center (EDC). To avoid computational instability related to steep slopes on terrain, smoothing is performed so that the valid resolution of the terrain adopted in the model is 1.5 times as coarse as that of the model itself.

The Global Land Cover Characteristics (GLCC) data set, also provided by EDC, is used to determine the land-sea attributes of all grids in the model. To alleviate discontinuities in surface wind and temperature fields around coastlines, a tiling approach is introduced in which land/sea sub-grid effects can be considered in surface flux evaluation. Surface-related parameters such as heat capacity, thermal conductivity, albedo, initial values of soil moisture and roughness are also based on land use as described by the GLCC data set. The National Land Numerical Information data set provided by Japan's Ministry of Land, Infrastructure, Transport and Tourism is also referenced for parameters over Japan.

Grids on land are further classified in terms of snow presence, and sea grids may be covered with ice. This gives a total of four surface categories: land, snow-covered land, sea and ice-covered sea. Snow-covered areas are analyzed using the high-resolution snow depth analysis system (Subsection 2.8.2), and ice-covered areas are identified from sea ice analysis conducted by the Office of Marine Prediction under JMA's Global Environment and Marine Department. As described previously, surface-related parameters are essentially based on land use without assumption of snow- or ice-covered areas. Accordingly, the parameters for these covered grid areas need to be modified with corresponding values.

### 3.5.3 Dynamics

#### 3.5.3.1 Basic Equations

The governing equations used in the MSM consist of non-hydrostatic, fully compressible equations on spherical curvilinear orthogonal and hybrid terrain-following coordinates with the shallow assumption. The equations are described in flux form.

##### 1. Momentum equations

The equations of motion are described as

$$\begin{aligned} & \frac{\partial}{\partial t} \left( \frac{1}{J} \rho u \right) + \frac{\partial}{\partial \xi} \left( \frac{1}{J} \rho u U \right) + \frac{\partial}{\partial \eta} \left( \frac{1}{J} \rho u V \right) + \frac{\partial}{\partial \zeta} \left( \frac{1}{J} \rho u W \right) \\ & + \gamma R_d \pi \left\{ \frac{1}{J} \xi_x \frac{\partial}{\partial \xi} (\rho \theta_m)' + \frac{1}{J} \eta_x \frac{\partial}{\partial \eta} (\rho \theta_m)' + \frac{1}{J} \zeta_x \frac{\partial}{\partial \zeta} (\rho \theta_m)' \right\} \\ & = - \sum_{\alpha} \frac{\partial}{\partial \zeta} \left( \frac{1}{J} \rho u q_{\alpha} W_{t_{\alpha}} \right) - \frac{1}{J} \rho v \Gamma - \frac{1}{J} \rho v f + \frac{1}{J} F_{\rho u}, \end{aligned} \quad (3.5.1)$$

$$\begin{aligned} & \frac{\partial}{\partial t} \left( \frac{1}{J} \rho v \right) + \frac{\partial}{\partial \xi} \left( \frac{1}{J} \rho v U \right) + \frac{\partial}{\partial \eta} \left( \frac{1}{J} \rho v V \right) + \frac{\partial}{\partial \zeta} \left( \frac{1}{J} \rho v W \right) \\ & + \gamma R_d \pi \left\{ \frac{1}{J} \xi_y \frac{\partial}{\partial \xi} (\rho \theta_m)' + \frac{1}{J} \eta_y \frac{\partial}{\partial \eta} (\rho \theta_m)' + \frac{1}{J} \zeta_y \frac{\partial}{\partial \zeta} (\rho \theta_m)' \right\} \\ & = - \sum_{\alpha} \frac{\partial}{\partial \zeta} \left( \frac{1}{J} \rho v q_{\alpha} W_{t_{\alpha}} \right) + \frac{1}{J} \rho u \Gamma + \frac{1}{J} \rho u f + \frac{1}{J} F_{\rho v}, \end{aligned} \quad (3.5.2)$$

$$\begin{aligned}
& \frac{\partial}{\partial t} \left( \frac{1}{J} \rho w \right) + \frac{\partial}{\partial \xi} \left( \frac{1}{J} \rho w U \right) + \frac{\partial}{\partial \eta} \left( \frac{1}{J} \rho w V \right) + \frac{\partial}{\partial \zeta} \left( \frac{1}{J} \rho w W \right) \\
& + \gamma R_d \pi \left\{ \frac{1}{J} \zeta_z \frac{\partial}{\partial \zeta} (\rho \theta_m)' \right\} + \left( \frac{\rho'}{J} - \frac{\pi'}{\bar{\pi}} \frac{\bar{\rho}}{J} \right) g \\
& = - \sum_{\alpha} \frac{\partial}{\partial \zeta} \left( \frac{1}{J} \rho w q_{\alpha} W_{t_{\alpha}} \right) + \frac{1}{J} F_{\rho w},
\end{aligned} \tag{3.5.3}$$

where

$$\Gamma = u \frac{m_2}{m_1} \frac{\partial m_1}{\partial \eta} - v \frac{m_1}{m_2} \frac{\partial m_2}{\partial \xi}. \tag{3.5.4}$$

Here,  $J$  is the Jacobian of coordinate transformation from Cartesian coordinates  $(x, y, z)$  to generalized coordinates  $(\xi, \eta, \zeta)$ , defined as

$$J \equiv \begin{vmatrix} \xi_x & \xi_y & \xi_z \\ \eta_x & \eta_y & \eta_z \\ \zeta_x & \zeta_y & \zeta_z \end{vmatrix}, \tag{3.5.5}$$

where,  $(\partial \xi / \partial x)_{y,z}$  - a metric of coordinate transformation - is described as  $\xi_x$ , and the same description applies to other metrics. A limitation for vertical coordinate to satisfy  $\xi_z = \zeta_z = 0$  is introduced, enabling utilization of the Split-Explicit time integration scheme (see Subsection 3.5.3.3).  $(u, v, w)$  and  $(U, V, W)$  represent velocity components in Cartesian coordinates and generalized coordinates, respectively.  $\gamma = C_p/C_v$ , where  $C_p$  and  $C_v$  are the specific heat of dry air at constant pressure and constant volume, respectively.  $R_d$  is the gas constant for dry air, and  $\rho$  is the total mass density defined as

$$\rho = \rho_d + \rho_v + \rho_c + \rho_r + \rho_i + \rho_s + \rho_g, \tag{3.5.6}$$

where the subscripts  $d, v, c, r, i, s$  and  $g$  represent dry air, water vapor, cloud water, rain, cloud ice, snow and graupel, respectively.

$\pi$  is the Exner function defined by

$$\pi = \left( \frac{p}{p_0} \right)^{\frac{R_d}{C_p}}. \tag{3.5.7}$$

The overlined variables  $\bar{\rho}$ ,  $\overline{\rho \theta_m}$  and  $\bar{\pi}$  represent the hydrostatic state as

$$\gamma R_d \bar{\pi} \frac{1}{J} \zeta_z \frac{\partial}{\partial \zeta} (\overline{\rho \theta_m}) + \frac{\bar{\rho} g}{J} = 0, \tag{3.5.8}$$

and the variables with prime  $\rho'$ ,  $(\rho \theta_m)'$  and  $\pi'$  represent perturbation from the hydrostatic state.  $g$  is gravity acceleration, and  $f$  is the Coriolis parameter.  $q_{\alpha}$  is the ratio of the density of water substances  $\alpha$  to the total mass density ( $\alpha = v, c, r, i, s, g$ ).  $W_{t_{\alpha}}$  is the terminal fall velocity of water substance  $\alpha$ .  $\theta_m$  is defined as

$$\theta_m \equiv \theta \left( 1 + \left( \frac{1 - \epsilon}{\epsilon} \right) q_v - q_c - q_r - q_i - q_s - q_g \right), \tag{3.5.9}$$

where  $\epsilon$  is the ratio of  $R_d$  to the gas constant for water vapor.  $F_{\rho u}$ ,  $F_{\rho v}$  and  $F_{\rho w}$  are terms of the surface friction.

Lambert conformal projection is employed, and the map factors  $m_1$  and  $m_2$  (for the  $x$  and  $y$  directions) are given by

$$m_1 = m_2 = m = \left( \frac{\cos \varphi}{\cos \varphi_1} \right)^{a-1} \left( \frac{1 + \sin \varphi_1}{1 + \sin \varphi} \right)^a, \quad (3.5.10)$$

where  $\varphi$  is the latitude of the relevant point,  $\varphi_1 = 30^\circ$ ,  $\varphi_2 = 60^\circ$  and  $a$  is given by

$$a = \ln \left( \frac{\cos \varphi_1}{\cos \varphi_2} \right) / \ln \left\{ \frac{\tan \left( 45^\circ - \frac{\varphi_1}{2} \right)}{\tan \left( 45^\circ - \frac{\varphi_2}{2} \right)} \right\}. \quad (3.5.11)$$

The hybrid terrain-following vertical coordinate which is based on the same approach as the  $\eta$  coordinate (Simmons and Burridge 1981) is adopted to reduce the influences of topography as height increases (Ishida 2007). The vertical coordinate  $\zeta$  is transformed using the equation:

$$z = \zeta + z_s h(\zeta), \quad (3.5.12)$$

where  $z$  is the height and  $z_s$  is the surface height. The function  $h(\zeta)$  is given by,

$$h(\zeta) = \frac{b \left\{ 1 - \left( \frac{\zeta}{z_T} \right)^n \right\}}{b + \left( \frac{\zeta}{z_T} \right)^n}, \quad b = \frac{\left( \frac{z_l + z_h}{2z_T} \right)^n}{1 - 2 \left( \frac{z_l + z_h}{2z_T} \right)^n}, \quad (3.5.13)$$

where  $z_T$  is the model top,  $z_l = 2000\text{m}$ ,  $z_h = 12000\text{m}$  and  $n = 3$ , respectively.

## 2. Continuity equations

The continuity equation is described as follows:

$$\frac{\partial}{\partial t} \left( \frac{1}{J} \rho' \right) + \frac{\partial}{\partial \xi} \left( \frac{1}{J} \rho U \right) + \frac{\partial}{\partial \eta} \left( \frac{1}{J} \rho V \right) + \frac{\partial}{\partial \zeta} \left( \frac{1}{J} \rho W \right) = - \sum_{\alpha} \frac{\partial}{\partial \zeta} \left( \frac{1}{J} \rho q_{\alpha} W_{t_{\alpha}} \right) + \frac{1}{J} F_{\rho}, \quad (3.5.14)$$

where  $F_{\rho}$  is the tendency by water vapor flux from the surface.

## 3. Prognostic equation of potential temperature

The thermodynamic equation is described as

$$\frac{\partial}{\partial t} \left( \frac{1}{J} (\rho \theta_m)' \right) + \frac{\partial}{\partial \xi} \left( \frac{1}{J} \rho \theta_m U \right) + \frac{\partial}{\partial \eta} \left( \frac{1}{J} \rho \theta_m V \right) + \frac{\partial}{\partial \zeta} \left( \frac{1}{J} \rho \theta_m W \right) = \frac{1}{J} \left( \rho_d + \frac{\rho_v}{\epsilon} \right) Q_{\theta}, \quad (3.5.15)$$

where  $Q_{\theta}$  is the diabatic heating.

## 4. Prognostic equation of water substances

The prognostic equations for the density of water substances are described as

$$\frac{\partial}{\partial t} \left( \frac{1}{J} \rho q_{\alpha} \right) + \frac{\partial}{\partial \xi} \left( \frac{1}{J} \rho q_{\alpha} U \right) + \frac{\partial}{\partial \eta} \left( \frac{1}{J} \rho q_{\alpha} V \right) + \frac{\partial}{\partial \zeta} \left( \frac{1}{J} \rho q_{\alpha} (W + W_{t_{\alpha}}) \right) = \frac{1}{J} F_{\rho\alpha}, \quad (3.5.16)$$

where  $F_{\rho\alpha}$  is source or sink term and tendency by flux from the surface for  $\alpha = v$ .

## 5. State equation

The state equation is

$$p = R_d \pi \rho \theta_m. \quad (3.5.17)$$

### 3.5.3.2 Spatial discretization

The grid structures of the model are the Arakawa C type in the horizontal direction and the Lorenz type in the vertical direction. The equations are spatially discretized using the finite volume method (FVM) to conserve total mass throughout the whole domain in consideration of lateral boundary inflow and outflow. The third-order upwind scheme with the flux limiter function proposed by Koren (1993) is employed to calculate horizontal and vertical advection terms for monotonicity in order to prevent numerical oscillation, and enhance computational efficiency.

### 3.5.3.3 Time integration

The Runge-Kutta (RK3) scheme (Wicker and Skamarock 2002) is adopted for system time integration. The terms responsible for sound waves and gravity waves are treated using a split-explicit time integration scheme with a short time step. Other time-splitting methods are also used to treat vertical advection associated with strong wind and vertical advection of water substances with high terminal velocity such as rain or graupel.

#### 1. Split-Explicit (HE-VI) Scheme

The horizontally explicit and vertically implicit (HE-VI) scheme (Klemp *et al.* 2007) is employed. RK3 scheme is also used for the short time step of HE-VI. Forward time integrations with the short time step  $\Delta\tau$  are used for the horizontal momentum equations:

$$\left(\frac{1}{J}\rho u\right)^{\tau+\Delta\tau} = \left(\frac{1}{J}\rho u\right)^\tau - \gamma R_d \pi^t \left\{ \frac{1}{J} \xi_x \frac{\partial}{\partial \xi} (\rho \theta_m)'^\tau + \frac{1}{J} \eta_x \frac{\partial}{\partial \eta} (\rho \theta_m)'^\tau + \frac{1}{J} \zeta_x \frac{\partial}{\partial \zeta} (\rho \theta_m)'^\tau \right\} \Delta\tau + R_u^t \Delta\tau, \quad (3.5.18)$$

$$\left(\frac{1}{J}\rho v\right)^{\tau+\Delta\tau} = \left(\frac{1}{J}\rho v\right)^\tau - \gamma R_d \pi^t \left\{ \frac{1}{J} \xi_y \frac{\partial}{\partial \xi} (\rho \theta_m)'^\tau + \frac{1}{J} \eta_y \frac{\partial}{\partial \eta} (\rho \theta_m)'^\tau + \frac{1}{J} \zeta_y \frac{\partial}{\partial \zeta} (\rho \theta_m)'^\tau \right\} \Delta\tau + R_v^t \Delta\tau, \quad (3.5.19)$$

where

$$R_u = -\frac{\partial}{\partial \xi} \left( \frac{1}{J} \rho u U \right) - \frac{\partial}{\partial \eta} \left( \frac{1}{J} \rho u V \right) - \frac{\partial}{\partial \zeta} \left( \frac{1}{J} \rho u W \right) + \frac{1}{J} F'_u, \quad (3.5.20)$$

$$R_v = -\frac{\partial}{\partial \xi} \left( \frac{1}{J} \rho v U \right) - \frac{\partial}{\partial \eta} \left( \frac{1}{J} \rho v V \right) - \frac{\partial}{\partial \zeta} \left( \frac{1}{J} \rho v W \right) + \frac{1}{J} F'_v, \quad (3.5.21)$$

and  $\frac{1}{J} F'_u$  and  $\frac{1}{J} F'_v$  are the right hand side of Eq. (3.5.1) and Eq. (3.5.2), respectively. Backward time integrations are used for equations of vertical momentum, potential temperature and density:

$$\left(\frac{1}{J}\rho w\right)^{\tau+\Delta\tau} = \left(\frac{1}{J}\rho w\right)^\tau - \left\{ \gamma R_d \pi^t \frac{1}{J} \zeta_z \frac{\partial}{\partial \zeta} (\rho \theta_m)'^{\tau+\Delta\tau} + \frac{\rho'^{\tau+\Delta\tau}}{J} g - \frac{\pi'^t}{\bar{\pi}} \frac{\bar{\rho}}{J} g \right\} \Delta\tau + R_w^t \Delta\tau, \quad (3.5.22)$$

$$\begin{aligned} \left(\frac{1}{J}(\rho\theta_m)'\right)^{\tau+\Delta\tau} &= \left(\frac{1}{J}(\rho\theta_m)'\right)^\tau - \left\{ \frac{\partial}{\partial\zeta} \left( \frac{1}{J}\zeta\theta_m^\tau(\rho w)^{\tau+\Delta\tau} \right) \right\} \Delta\tau \\ &\quad - \left\{ \frac{\partial}{\partial\xi} \left( \frac{1}{J}\theta_m^\tau(\widetilde{\rho U}) \right) + \frac{\partial}{\partial\eta} \left( \frac{1}{J}\theta_m^\tau(\widetilde{\rho V}) \right) + \frac{\partial}{\partial\zeta} \left( \frac{1}{J}\theta_m^\tau(\widetilde{\rho W}) \right) \right\} \Delta\tau + \frac{1}{J}F_{\rho\theta_m}^t \Delta\tau, \end{aligned} \quad (3.5.23)$$

$$\begin{aligned} \left(\frac{1}{J}\rho'\right)^{\tau+\Delta\tau} &= \left(\frac{1}{J}\rho'\right)^\tau - \left\{ \frac{\partial}{\partial\zeta} \left( \frac{1}{J}\zeta(\rho w)^{\tau+\Delta\tau} \right) \right\} \Delta\tau \\ &\quad - \left\{ \frac{\partial}{\partial\xi} \left( \frac{1}{J}(\widetilde{\rho U}) \right) + \frac{\partial}{\partial\eta} \left( \frac{1}{J}(\widetilde{\rho V}) \right) + \frac{\partial}{\partial\zeta} \left( \frac{1}{J}(\widetilde{\rho W}) \right) \right\} \Delta\tau + \frac{1}{J}F_\rho' \Delta\tau, \end{aligned} \quad (3.5.24)$$

where

$$R_w = -\frac{\partial}{\partial\xi} \left( \frac{1}{J}\rho w U \right) - \frac{\partial}{\partial\eta} \left( \frac{1}{J}\rho w V \right) - \frac{\partial}{\partial\zeta} \left( \frac{1}{J}\rho w W \right) + \frac{1}{J}F'_w, \quad (3.5.25)$$

and  $\frac{1}{J}F'_w$ ,  $\frac{1}{J}F'_\rho$  and  $\frac{1}{J}F_{\rho\theta_m}$  are the right hand side of Eq. (3.5.3), Eq. (3.5.14) and Eq. (3.5.15), respectively, and

$$(\widetilde{\rho U}) = \xi_x(\rho u)^{\tau+\Delta\tau} + \xi_y(\rho v)^{\tau+\Delta\tau}, \quad (3.5.26)$$

$$(\widetilde{\rho V}) = \eta_x(\rho u)^{\tau+\Delta\tau} + \eta_y(\rho v)^{\tau+\Delta\tau}, \quad (3.5.27)$$

$$(\widetilde{\rho W}) = \zeta_x(\rho u)^{\tau+\Delta\tau} + \zeta_y(\rho v)^{\tau+\Delta\tau}. \quad (3.5.28)$$

Here, we can exclude  $(\rho w)^{\tau+\Delta\tau}$  from Eq. (3.5.26) - Eq. (3.5.28) due to the limitation for vertical coordinate to satisfy  $\xi_z = \zeta_z = 0$  as mentioned in Subsection 3.5.3.1, which enables the vertical implicit treatment of Eq. (3.5.22) - Eq. (3.5.24). Eliminating  $\left(\frac{1}{J}(\rho\theta_m)'\right)^{\tau+\Delta\tau}$  and  $\left(\frac{1}{J}\rho'\right)^{\tau+\Delta\tau}$  from Eq. (3.5.22) using Eq. (3.5.23) and Eq. (3.5.24), we obtain the one dimensional Helmholtz type equation of  $\omega \equiv \left(\frac{1}{J}\rho w\right)^{\tau+\Delta\tau}$  as

$$-\Delta\tau^2\gamma R_d\pi^t \frac{1}{J}\zeta_z \frac{\partial}{\partial\zeta} \left( J \frac{\partial}{\partial\zeta} (\zeta_z\theta_m^\tau\omega) \right) - \Delta\tau^2g \frac{\partial}{\partial\zeta} (\zeta_z\omega) + \omega = R, \quad (3.5.29)$$

where

$$R = \left(\frac{1}{J}\rho w\right)^\tau - \gamma R_d\pi^t \Delta\tau \frac{1}{J}\zeta_z \frac{\partial}{\partial\zeta} \left\{ (\rho\theta_m)'^\tau + JR'_{\theta_m} \Delta\tau \right\} - \Delta\tau g \left( \frac{1}{J}\rho'^\tau + R'_\rho \Delta\tau \right) + R'_w \Delta\tau, \quad (3.5.30)$$

and

$$R'_w = \frac{\pi'^t}{\bar{\pi}} \frac{\bar{\rho}}{J} g + R_w = - \left( 1 - \frac{\pi'}{\bar{\pi}} \right) \frac{\bar{\rho}}{J} g + R_w, \quad (3.5.31)$$

$$R'_{\theta_m} = - \left\{ \frac{\partial}{\partial \xi} \left( \frac{1}{J} \theta_m^\tau (\widetilde{\rho U}) \right) + \frac{\partial}{\partial \eta} \left( \frac{1}{J} \theta_m^\tau (\widetilde{\rho V}) \right) + \frac{\partial}{\partial \zeta} \left( \frac{1}{J} \theta_m^\tau (\widetilde{\rho W}) \right) \right\} + \frac{1}{J} F_{\rho \theta_m}^t, \quad (3.5.32)$$

$$R'_\rho = - \left\{ \frac{\partial}{\partial \xi} \left( \frac{1}{J} (\widetilde{\rho U}) \right) + \frac{\partial}{\partial \eta} \left( \frac{1}{J} (\widetilde{\rho V}) \right) + \frac{\partial}{\partial \zeta} \left( \frac{1}{J} (\widetilde{\rho W}) \right) \right\} + \frac{1}{J} F_\rho^t. \quad (3.5.33)$$

Considering  $W = 0$  at the upper and lower boundary and  $u = 0, v = 0$  at the lower boundary, upper and lower boundary conditions are given by  $\omega = 0$ .

## 2. Time splitting of vertical advection

Using RK3 as a time integration scheme and a flux limiter function as an advection scheme, the CFL condition of 3-dimensional advection is given by

$$C_\xi + C_\eta + C_\zeta < 1.25, \quad (3.5.34)$$

where  $C_\xi, C_\eta$  and  $C_\zeta$  are the Courant number in the  $\xi, \eta$  and  $\zeta$  direction, respectively. As this condition can be hard to fulfill with typhoons characterized by stormy horizontal winds and strong updrafts, time splitting of vertical advection is adopted in consideration of computational efficiency and the model's memory alignment with vertical indices placed innermost.

In the time splitting method, each RK3 stage is divided into substeps depending on the relevant Courant numbers. As each RK3 stage can be regarded as a forward time integration with the time steps of  $\Delta t/3$ ,  $\Delta t/2$  and  $\Delta t$ , respectively (as shown in Figure 3.5.1), these time steps are used to evaluate the Courant numbers for each stage. For each column, the number of substeps  $N$  is set to satisfy

$$C_\xi + C_\eta + \frac{C_\zeta}{N} < 1.25, \quad (3.5.35)$$

at each RK3 stage. When time splitting is invoked, the forward form integration at each RK3 stage is replaced with RK3 (i.e., RK3 is nested in the original RK3 time integration) as shown in Figure 3.5.2. This involves greater computational cost, but produces the desired higher stability.

When time-splitting is invoked, fields are updated using the horizontal flux  $F_\xi$  and  $F_\eta$  first, and the vertical flux  $F_\zeta$  is then evaluated with the integrated field as follows.

$$\phi^{H*} = \phi^n - \left( \frac{\partial}{\partial \xi} F_\xi^n + \frac{\partial}{\partial \eta} F_\eta^n \right) \Delta \tau, \quad (3.5.36)$$

$$\phi^{n+1} = \phi^{H*} - \left( \frac{\partial}{\partial \zeta} F_\zeta^{H*} \right) \Delta \tau. \quad (3.5.37)$$

## 3. Time splitting of vertical advection of water substances

To stabilize integration for the vertical advection of water substances with high terminal velocity, a time splitting method is adopted. The short time step  $\Delta \tau_1$  for sedimentation is determined from the Courant number  $C_{t\zeta}$  as follows.

$$\Delta \tau_1 = \begin{cases} \Delta t & (\max(C_{t\zeta}) \leq 1) \\ \beta \frac{\Delta t}{\max(C_{t\zeta})} & (\max(C_{t\zeta}) > 1), \end{cases} \quad (3.5.38)$$

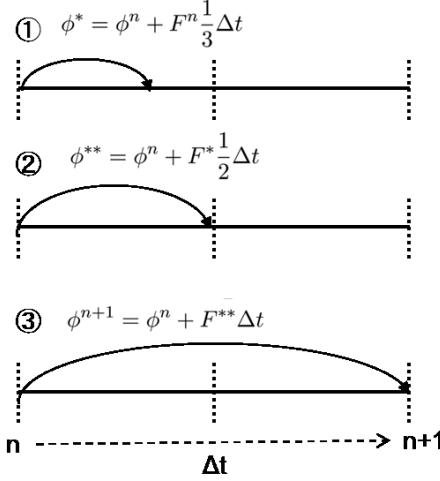


Figure 3.5.1: Schematic of RK3 time integration scheme.

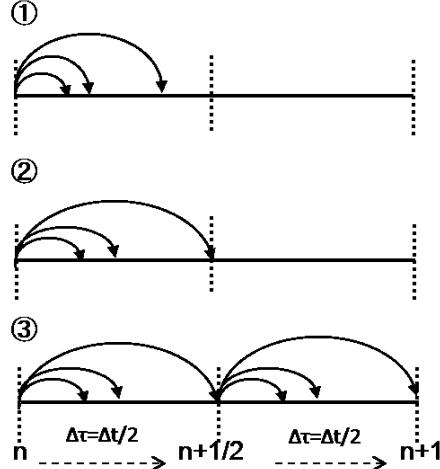


Figure 3.5.2: Schematic of time-splitting of vertical advection. The case needs to split ③ stage into 2 substeps.

where

$$C_{t\zeta} = (W^n + W_{ta}^n)\Delta t / \Delta \zeta. \quad (3.5.39)$$

Here,  $W^n$  is  $W$  at the time level  $n$ ,  $\beta$  is a parameter for determining the short time step, and  $c = 0.9$  is used.

After time integration with  $\Delta\tau_1$ , the residual time step is  $\Delta t' = \Delta t - \Delta\tau_1$ . The next short time step  $\Delta\tau_2$  is decided from the Courant number  $C'_{t\zeta} = (W^n + W_{ta}^{n+\tau_1})\Delta t' / \Delta \zeta$  and the time integration with  $\Delta\tau_2$  is calculated. This procedure is repeated until no residual time step is left.

### 3.5.3.4 Boundary Conditions

Rayleigh damping,

$$D_R = -n(x, y, z) \{\phi - \phi_{EXT}\}, \quad n(x, y, z) = \max\left(\frac{D}{m_L}, \frac{D}{m_U}\right), \quad (3.5.40)$$

is added near the lateral and upper boundaries to the time tendencies of horizontal and vertical momentum, potential temperature and the mixing ratio of water vapor, where  $\phi$  is the prognostic variable and  $\phi_{EXT}$  is the value of the external model.  $m_L$  and  $m_U$  represent coefficients that determine the 1/e-folding time for the lateral and upper boundaries, respectively, while  $m_L = 250$  seconds and  $m_U = 125$  seconds. The location-based function D is unity at the boundary and decreases with subsequent grid point distance.

### 3.5.4 Cloud Microphysics

An explicit three-ice bulk microphysics scheme (Ikawa and Saito 1991) based on Lin *et al.* (1983) is incorporated. The scheme predicts the mixing ratios of water vapor and five hydrometeors designated by  $q_x$  where  $x$  denotes categories defined as  $v$  for water vapor,  $c$  for cloud water,  $r$  for rain,  $i$  for cloud ice,  $s$  for snow, and  $g$  for graupel. The cloud microphysical processes simulated in this scheme are illustrated in Figure 3.5.3 (see Table 3.5.1 for a list of symbols used in the figure). In this scheme, some basic cloud microphysical processes

(e.g., nucleation of cloud particles, conversion from cloud particles to precipitation particles) are parameterized because the related processes occur within a shorter time than the integration time step. However, most of the cloud microphysical processes can be applied directly to calculation related to the size distribution assumed in each hydrometeor category.

The number-weighted mean of the temporal tendency of one cloud microphysical variable  $\phi$  relating to one cloud microphysical process in each particle gives the grid-mean temporal tendency of  $\phi$  as

$$\frac{d\phi}{dt} = \int_0^\infty \frac{d\phi_0(D)}{dt} n(D) dD, \quad (3.5.41)$$

where  $\frac{d\phi_0(D)}{dt}$  is the temporal tendency of  $\phi$  relating to one cloud microphysical process in a particle with diameter  $D$ , and  $n(D) dD$  is the number of particles per unit volume of air with diameters from  $D$  to  $D + dD$ . Hydrometeor size distribution therefore significantly affects time tendency of cloud microphysical variables relating to cloud microphysical processes.

### 3.5.4.1 Mass-size Relationships

The mass-size relationships represent particle mass  $m_x$  as a function of particle diameter  $D_x$  for determination of mixing ratios or mass weighted-mean variables. For example, the mixing ratio  $q_x$  is generally formulated as

$$q_x = \frac{1}{\rho_a} \int_0^\infty m_x(D_x) n_x(D_x) dD_x, \quad (3.5.42)$$

where  $\rho_a$  is the density of air.

Particle sphericity is assumed in the hydrometeor categories of cloud ice, graupel, rain and cloud water, and density is constant in each category ( $\rho_x$ ). Accordingly, the mass-size relationship ( $m_x(D_x)$ ) is given by

$$m_x(D_x) = \frac{\pi}{6} \rho_x D_x^3. \quad (3.5.43)$$

For the category of snow, the mass-size relationship is formulated as

$$m_s(D_s) = a_s D_s^{b_s}, \quad (3.5.44)$$

where  $a_s$  and  $b_s$  are set to  $0.0185 \text{kgm}^{-1.9}$  and  $1.9$  respectively (Brown and Francis 1995).

### 3.5.4.2 Size Distribution Functions

#### 1. Cloud ice and graupel

The size distributions of cloud ice and graupel are assumed to follow an exponential function:

$$n_x(D_x) = N_{0x} \exp(-\lambda_x D_x), \quad (3.5.45)$$

where  $N_{0x}$  is the intercept and  $\lambda_x$  is the slope parameter of the size distribution. Accordingly, the moment formula for cloud ice and graupel is calculated as

$$M_x(p) = \int_0^\infty D_x^p n_x(D_x) dD_x = N_{0x} \frac{\Gamma(1+p)}{\lambda_x^{1+p}}, \quad (3.5.46)$$

where  $M_x(p)$  is the  $p$ -th moment of  $n_x(D_x)$ . The number concentration is the 0-th moment of  $n_x(D_x)$ , and is therefore calculated as

$$N_x = \int_0^\infty n_x(D_x) dD_x = M_x(0) = \frac{N_{0x}}{\lambda_x}. \quad (3.5.47)$$

The mixing ratio  $q_x$  is the third moment of  $n_x(D_x)$ , and is therefore also calculated as

$$q_x = \frac{1}{\rho_a} \int_0^\infty m_x(D_x) n_x(D_x) dD_x = \frac{\rho_x \pi}{\rho_a 6} M_x(3) = \frac{\rho_x \pi}{\rho_a 6} N_{0x} \frac{\Gamma(4)}{\lambda_x^4}. \quad (3.5.48)$$

The intercepts are assumed to be constant. The following formula is therefore used:

$$N_{0x} = \text{const.}, \quad \lambda_x = \left( \frac{\pi \rho_x N_{0x}}{\rho_a q_x} \right)^{\frac{1}{4}}. \quad (3.5.49)$$

## 2. Snow

The size distribution function itself is not directly used for snow, but moments are parameterized based on Field *et al.* (2007) as follows:

$$M_s(p) = \int_0^\infty D_s^p n_s(D_s) dD_s = A(p) \exp[B(p)(T - T_0)] M_s^{c(p)}(2), \quad (3.5.50)$$

$$A(p) = \exp[13.6 - 7.76p + 0.479p^2], \quad (3.5.51)$$

$$B(p) = -0.0361 + 0.0151p + 0.00149p^2, \quad (3.5.52)$$

$$C(p) = 0.807 + 0.00581p + 0.0457p^2, \quad (3.5.53)$$

where  $T_0$  is the freezing temperature ( $= 273.15\text{K}$ ).

The number concentration, the 0-the moment of  $n_s(D_s)$ , is calculated as

$$N_s = M_s(0) = A(0) \exp[B(0)(T - T_0)] M_s^{c(0)}(2). \quad (3.5.54)$$

Based on (3.5.44), the mixing ratio  $q_s$  is

$$q_s = \frac{1}{\rho_a} \int_0^\infty m_s(D_s) n_s(D_s) dD_s = \frac{1}{\rho_a} a_s M_s(b_s) = \frac{1}{\rho_a} a_s A(b_s) \exp[B(b_s)(T - T_0)] M_s^{c(b_s)}(2). \quad (3.5.55)$$

Accordingly, the second moment  $M_s(2)$  is calculated as

$$M_s(2) = \left[ \frac{\rho_a q_s}{a_s} \frac{1}{A(b_s) \exp[B(b_s)(T - T_0)]} \right]^{\frac{1}{c(b_s)}}. \quad (3.5.56)$$

## 3. Rain

Rain size distribution is based on Abel and Boutle (2012), with an intercept as a function of the slope parameter:

$$n_r(D_r) = N_{0r} \exp(-\lambda_r D_r), \quad (3.5.57)$$

$$N_{0r} = N_{00r} \lambda_r^{\beta_r}. \quad (3.5.58)$$

Thus,  $N_r$ ,  $q_r$  and  $\lambda_r$  are calculated as

$$N_r = \int_0^\infty n_r(D_r) dD_r = M_r(0) = N_{00r} \lambda_r^{\beta_r - 1}. \quad (3.5.59)$$

$$q_r = \frac{1}{\rho_a} \int_0^\infty m_r(D_r) n_r(D_r) dD_r = \frac{\rho_r \pi}{\rho_a 6} M_r(3) = \frac{\rho_r \pi}{\rho_a 6} N_{00r} \frac{\Gamma(4)}{\lambda_r^{4-\beta_r}} \quad (3.5.60)$$

$$\lambda_r = \left( \frac{\pi \rho_r N_{00r}}{\rho_a q_r} \right)^{\frac{1}{4-\beta_r}}. \quad (3.5.61)$$

## 4. Cloud water

As cloud water is assumed to be monodisperse, its size distribution follows the  $\delta$ -function:

$$n_c(D_c) = N_c \delta(D_c - \bar{D}_c), \quad (3.5.62)$$

where  $\delta(x)$  satisfies the equation  $\int_{-\infty}^{\infty} \delta(x-a)f(x) dx = f(a)$  and  $\overline{D_c}$  represents the diameter of the monodisperse particle. The moment formula for cloud water is given by

$$M_c(p) = N_c \overline{D_c}^p. \quad (3.5.63)$$

The number concentration of cloud water  $N_c$  is always assumed to be constant and is set to  $1.0 \times 10^8 \text{ kg m}^{-3}$  in this scheme.

The mixing ratio is calculated as

$$q_c = \frac{\rho_c}{\rho_a} \frac{\pi}{6} M_c(3) = \frac{\rho_c}{\rho_a} \frac{\pi}{6} N_c \overline{D_c}^3. \quad (3.5.64)$$

The diameter is therefore determined as

$$\overline{D_c} = \left( \frac{6\rho_a q_c}{\pi \rho_c N_c} \right)^{\frac{1}{3}}. \quad (3.5.65)$$

### 3.5.4.3 Fall Velocity and Sedimentation

#### 1. Cloud ice, snow and graupel

The simple power law is adopted for the fall velocity-size relationship ( $U_x(D_x)$ ) given by

$$U_x(D_x) = \alpha_{ux} D_x^{\beta_{ux}} \left( \frac{\rho_0}{\rho_a} \right)^{\gamma_{ux}}, \quad (3.5.66)$$

where  $\rho_0$  is the density of the reference air, and  $\alpha_{ux}$ ,  $\beta_{ux}$  and  $\gamma_{ux}$  are constants in each hydrometeor category.

#### 2. Rain

Following Abel and Shipway (2007), the fall velocity-size relationship for rain is calculated as

$$U_r(D_r) = \left( \frac{\rho_0}{\rho_a} \right)^{\frac{1}{2}} \sum_{k=1}^2 c_k D_r^{d_k} e^{-f_k D_r}. \quad (3.5.67)$$

where  $c_k$ ,  $d_k$  and  $f_k$  are constants.

The forecast model for the MSM calculates sedimentation processes for hydrometers in its dynamical core. The cloud microphysics scheme diagnoses mass-weighted mean fall velocities as terminal velocities  $W_{tx}$ :

$$W_{tx} = \frac{\int_0^{\infty} U_x(D_x) m_x(D_x) n_x(D_x) dD_x}{\int_0^{\infty} m_x(D_x) n_x(D_x) dD_x}. \quad (3.5.68)$$

The cloud microphysics scheme passes  $W_{tx}$  to the dynamical core.

Table 3.5.2 shows the characteristics of each hydrometeor class. More information on the treatment of each cloud microphysical process in this scheme can be found in the references.

## 3.5.5 Convective Parameterization

A temporal tendency of a grid mean value  $\phi$  associated with subgrid convection under the isobaric coordinate is generally described as

$$\rho \left( \frac{\partial \bar{\phi}}{\partial t} \right)_{\text{convection}} = \rho g \frac{\partial \rho \bar{\phi}' w'}{\partial p} + S_\phi, \quad (3.5.69)$$

where  $\bar{\phi}' w'$  and  $S_\phi$  represent a subgrid transport flux and a source term respectively. To parameterize the subgrid flux and source term for heat and moisture, a mass flux convective parameterization based on the Kain-Fritsch (KF) (Kain and Fritsch 1990; Kain 2004) type scheme is employed in the MSM.

Table 3.5.1: List of symbols in Figure 3.5.3

Notation	Description
Production terms	
p_a_ppp_b	Production of category “a” converted from category “b” via the process “ppp”
p_a_ppp_a_b	Growth of category “a” based on capture of category “b” via the process “ppp”
p_a_ppp_b_c	Generation of category “a” based on category “b” capturing category “c” via the process “ppp”
Categories by hydrometeor	
v	Water vapor
c	Cloud water
r	Rain
i	Cloud ice
s	Snow
g	Graupel
Cloud microphysical processes	
evp	Evaporation
cnd	Condensation
aut, cn	Conversion
ac	Accretion
mlt	Melting
nud	Nucleation
dep	Deposition
sub	Sublimation
frz	Freezing

Table 3.5.2: Assumed hydrometeor parameters and characteristics

Variable	Rain	Snow	Graupel	Cloud ice	Cloud water
	$q_r[\text{kg kg}^{-1}]$	$q_s[\text{kg kg}^{-1}]$	$q_g[\text{kg kg}^{-1}]$	$q_i[\text{kg kg}^{-1}]$	$q_c[\text{kg kg}^{-1}]$
Size distribution [m <sup>-4</sup> ]	$n_r(D_r) = N_{00r} \lambda_r^{\beta_r} \exp(-\lambda_r D_r)$ $N_{00r} = 0.22$ $\beta_r = 2.2$	not used	directly $n_g(D_g) = N_{0g} \exp(-\lambda_g D_g)$ $N_{0g} = 1.1 \times 10^6$	$n_i(D_i) = N_{0i} \exp(-\lambda_i D_i)$ $N_{0i} = 4.0 \times 10^7$	monodisperse, $\overline{D}_c = \left[ \frac{6q_c \rho_a}{\pi N_c \rho_c} \right]^{\frac{1}{3}}$ $N_c = 1.0 \times 10^8$
Mass[kg]-size[m] relationship	$m_r = \frac{\pi}{6} \rho_r D_r^3$	$m_s = 0.0185 D_s^{1.9}$	$m_g = \frac{\pi}{6} \rho_g D_g^3$	$m_i = \frac{\pi}{6} \rho_i D_i^3$	$m_c = \frac{\pi}{6} \rho_c D_c^3$
Density [kg m <sup>-3</sup> ]	$\rho_r = 1.0 \times 10^3$	not used	$\rho_g = 3.0 \times 10^2$	$\rho_i = 1.5 \times 10^2$	$\rho_c = 1.0 \times 10^3$
Fall velocity [m/s]	$U_r(D_r) = \left( \frac{\rho_0}{\rho_a} \right)^{\frac{1}{2}} \sum_{k=1}^2 c_k D_r^{d_k} e^{-f_k D_r}$ $c_1 = 4854.1$ $d_1 = 1.0$ $f_1 = 195.0$ $c_2 = -446.009$ $d_2 = 0.782127$ $f_2 = 4085.35$		$U_x(D_x) = \alpha_{ux} D_x^{\beta_{ux}} \left( \frac{\rho_0}{\rho_a} \right)^{\gamma_{ux}}$ $\alpha_{us} = 17$ $\beta_{us} = 0.5$ $\gamma_{us} = 0.5$	$\alpha_{ug} = 124$ $\beta_{ug} = 0.64$ $\gamma_{ug} = 0.5$	$\alpha_{ui} = 71.34$ $\beta_{ui} = 0.6635$ $\gamma_{ui} = 0.5$

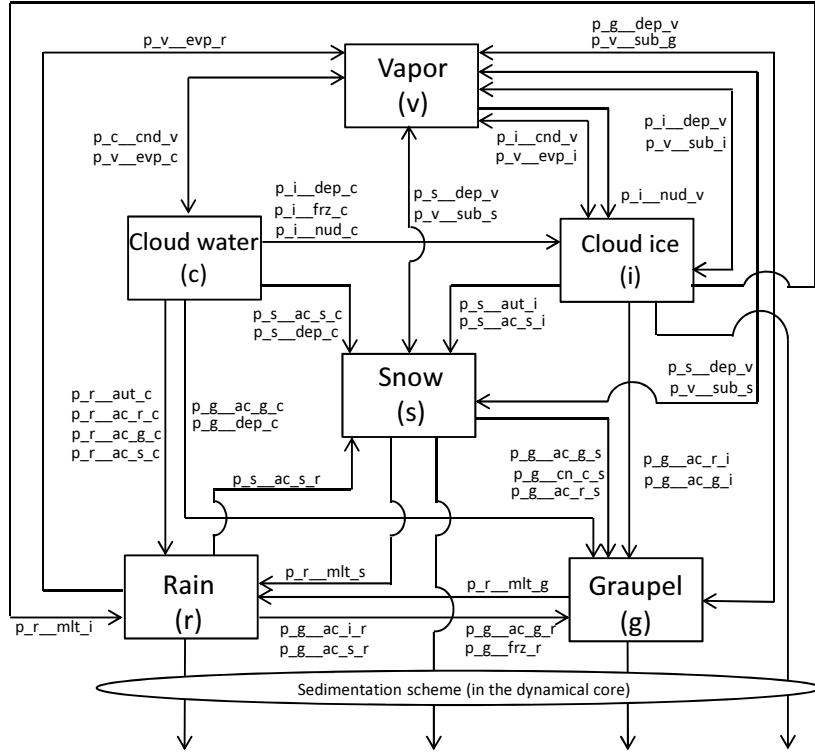


Figure 3.5.3: Cloud microphysical processes in the MSM. For a list of symbols, see Table 3.5.1.

### 3.5.5.1 Cloud Model

The scheme parameterizes convection using a cloud model based on a one-dimensional entraining/detraining plume model incorporating detailed treatment for interactions between convective updraft and the surrounding air. The cloud model consists of an updraft mass flux representing the convective activity within a column. Downdraft is not treated in the scheme. The cloud model involves the assumption that the convective updraft area is in a steady state. Thus, mass conservation is represented as

$$0 = \rho g \frac{\partial M^u}{\partial p} + E^u - D^u, \quad (3.5.70)$$

where  $M$ ,  $E$  and  $D$  represent convective mass flux, and entrainment / detrainment from / to environmental air respectively. The superscript  $u$  shows updraft. For  $\phi$ , the steady state equations can be represented as

$$0 = \rho g \frac{\partial M^u \phi^u}{\partial p} + E^u \bar{\phi} - D^u \phi^u + S_\phi^u. \quad (3.5.71)$$

The scheme also employs the assumption that the area of convection is small enough relative to that of a grid-box. Based on this assumption, the subgrid flux can be represented as

$$\rho \bar{\phi}' w' = M^u (\phi^u - \bar{\phi}). \quad (3.5.72)$$

Substituting Eq. (3.5.70), Eq. (3.5.71) and Eq. (3.5.72) into Eq. (3.5.69), the temporal tendency of  $\phi$  due to subgrid convection can be re-written as a summation of the detrainment and compensating subsidence terms:

$$\rho \left( \frac{\partial \bar{\phi}}{\partial t} \right)_{\text{convection}} = D^u (\phi^u - \bar{\phi}) - \rho g M^u \frac{\partial \bar{\phi}}{\partial p}. \quad (3.5.73)$$

### 3.5.5.2 Determination of Variables in Updraft

Vertical integration with Eq. (3.5.70) and Eq. (3.5.71) from the lifting condensation level (LCL), is applied to determine  $\phi^u$  and vertical profiles of  $M^u$ . During this integration,  $E^u$  and  $D^u$ , representing entrainment and detrainment, are calculated with consideration of the mixing process between the updraft and environmental air.

Following the original KF scheme (Kain and Fritsch 1990), the interaction between updraft and the environment associated with the turbulent mixing is estimated at each vertical model level to determine  $E^u$  and  $D^u$ . It is assumed that the turbulent mixing occurs very near the periphery of the updraft, and that large number of subparcel-like mixtures of the updraft and the environment form at various ratios that can be described by a Gaussian probability distribution function the mean of with a mean of 0.5, representing a scenario in which environmental mass and updraft mass are likely to be equally mixed in subparcels.

The relationship linking  $\delta M_e$ ,  $\delta M_u$  and  $\delta M_t$ , defined as the entrained mass from the environment, the updraft mass mixed with the entrained mass and the total mass respectively, can be expressed as

$$\delta M_u + \delta M_e = \delta M_t = \delta M_t \int_0^1 f(x)dx, \quad (3.5.74)$$

$$\delta M_e = \delta M_t \int_0^1 xf(x)dx, \quad (3.5.75)$$

$$\delta M_u = \delta M_t \int_0^1 (1-x)f(x)dx, \quad (3.5.76)$$

where  $x$  and  $f(x)$  are the fraction of environmental mass in mixed subparcels and the probability distribution function as a function of  $x$  respectively.

$\delta M_e$ , which determines the amounts of entrainment and detrainment, is inversely proportional to the updraft radius,  $R$ :

$$\delta M_e = M_{u0}(a\delta P/R), \quad (3.5.77)$$

where  $\delta P$  is the vertical grid thickness in the pressure coordinate, and the factor  $a$  is set to 0.03 m Pa<sup>-1</sup> as a constant. The radius of the updraft is used only for entrainment rate estimation. The radius  $R$  is set to a constant of 750 m.

Consequently, mixtures with positive buoyancy against the environment entrain into updraft, while those with negative buoyancy detract from it. When  $x_n$  is the fraction with which mixed air is neutral against environmental air,  $E^u$  and  $D^u$  can be formulated as follows:

$$E^u \delta p = \rho g \delta M_t \int_0^{x_n} xf(x)dx, \quad (3.5.78)$$

$$D^u \delta p = \rho g \delta M_t \int_{x_n}^1 (1-x)f(x)dx. \quad (3.5.79)$$

The vertical velocity of the updraft depends on buoyancy and hydrometeor weights. The updraft terminates when the mass flux becomes emaciated through detrainment or when its vertical velocity vanishes.

In the scheme, parameterized convection is divided into deep and shallow convection types. An updraft that does not reach the minimum cloud depth for deep convection is regarded as shallow convection. The minimum cloud depth is a function of temperature at the cloud base (LCL).

### 3.5.5.3 Treatment of convective precipitation

For cloud water content  $q_c$ , Eq. (3.5.71) applies as follows:

$$0 = \rho g \frac{\partial M^u q_c^u}{\partial p} + E^u \bar{q}_c - D^u q_l^u + c_c^u - P_c, \quad (3.5.80)$$

where  $c_c^u [\text{kg m}^{-3}\text{s}^{-1}]$  and  $P_c [\text{kg m}^{-3}\text{s}^{-1}]$  are condensation/glaciation and precipitation generation terms respectively. In updraft evaluation, water condensate /glaciate (depending on updraft temperature at each level) are considered.

The precipitation generation rate is formulated as a Kessler type autoconversion scheme:

$$P_c = A \max(q_c^u - q_{c0}, 0). \quad (3.5.81)$$

The threshold  $q_{c0}$  is set to a constant value of  $2.0 \times 10^{-3} \text{ kg kg}^{-1}$ . The conversion rate  $A$  is set to  $\rho g M^u / \Delta p$  so that excess hydrometeors over the threshold are immediately taken out of the updraft as precipitation. The generated precipitation is added to the tendencies shown as per Eq. (3.5.73) for rain, snow and graupel rather than being represented as falling to the ground. The precipitation fall process is calculated in a sedimentation scheme outside the convective parameterization scheme.

### 3.5.5.4 Closure

The closure process finally determines the magnitude of convective mass flux. Closure is the only difference between deep and shallow convection.

For deep convection, the magnitude of mass flux is determined so that stabilized vertical profiles after convection satisfy the condition that CAPE in the final state should be less than 15% of the initial value. Seeking the stabilized state usually requires iterative adjustment of mass fluxes. The updraft mass flux at the LCL,  $M_{u0} [\text{kg m}^{-2}\text{s}^{-1}]$ , is initially given as follows with the assumption that vertical velocity  $w_{\text{one}}$  is 1 m/s and the initial area occupied by convection is 1 % of a grid:

$$M_{u0} = 0.01 \rho_{u\text{LCL}} w_{\text{one}}, \quad (3.5.82)$$

where  $\rho_{u\text{LCL}}$  is the density of the updraft mass flux at the LCL. Using  $M_{u0}$ , the vertical integration of Eqs. (3.5.70) and (3.5.71) from the LCL are calculated. The value of  $\bar{\phi}_{\text{stabilized}}$ , the grid mean  $\bar{\phi}$  after stabilization, is then calculated based on time integration of Eq. (3.5.73) with the period of the lifetime of convection  $\tau_{\text{lifetime}}$ . If the post-stabilization CAPE is still 15 % more than the initial value, the mass flux at the LCL is increased. Iterating this process, the final value of  $M_{u0}$  is determined.

For shallow convection, the mass flux at the LCL is determined using the maximum turbulent kinetic energy (TKE),

$$M_{u0,\text{shallow}} = \frac{\text{TKE}_{\text{max}}}{k_0} \frac{\Delta p_{\text{parcel}}}{g \tau_{\text{lifetime}}}, \quad (3.5.83)$$

where  $\text{TKE}_{\text{max}}$  is the maximum value of TKE and is set to  $1 \text{ m}^2 \text{ s}^{-2}$  for the MSM.  $k_0$  is set to  $20 \text{ m}^2 \text{ s}^{-2}$ .  $\Delta p_{\text{parcel}}$  [Pa] is the pressure depth from the LCL to the highest model level at which the depth is no larger than 50 hPa.

The temporal tendency of the adjusted physical quantity  $\phi$  can be determined as

$$\left. \frac{d\bar{\phi}}{dt} \right|_{\text{convection}} = \frac{\bar{\phi}_{\text{stabilized}} - \bar{\phi}_{\text{initial}}}{\tau_{\text{lifetime}}}, \quad (3.5.84)$$

where  $\tau_{\text{lifetime}}$  is set to a constant value of 600 s.

### 3.5.5.5 Diagnosis of the Convection as a Triggering Process

Diagnosis is performed to determine whether each column is convectively unstable enough to activate the scheme. This is done for every timestep to identify grids where parameterized convection should occur, and involves two steps.

Firstly, the temperature of a lifted parcel is compared with the environmental temperature  $\bar{T}$ . The parcel temperature  $T_{\text{diag}}$  is defined as the lifted air mass at the LCL  $T_{\text{LCL}}$  with a perturbation:

$$T_{\text{diag}} = T_{\text{LCL}} + \Delta T'. \quad (3.5.85)$$

The second term on the right represents perturbation corresponding to subgrid-scale buoyant flux associated with the planetary boundary layer process.

$$\Delta T' = \max \left[ A_{\text{plume}}, \min \left[ B_{\text{plume}} \sigma_{T_v}, G_{\text{max}} z_h \right] \right] \Pi, \quad (3.5.86)$$

$$\sigma_{T_v} = 1.93 \overline{w' \theta'_{vs}} / w_m, \quad (3.5.87)$$

$$w_m^3 = u_*^3 + 0.25 z_h \overline{w' b'_s}, \quad (3.5.88)$$

where  $\Pi$ ,  $z_h$ ,  $u^*$  and  $\overline{w' b'_s}$  are the Exner function, the height of the planetary boundary layer, friction velocity and turbulent buoyant flux at the surface respectively. The parameters are set to  $A_{\text{plume}} = 0.2$  K,  $B_{\text{plume}} = 3.26$  and  $G_{\max} = 10^{-3}$  Km $^{-1}$ . The diagnosis process lifts the potential updraft source layer (USL) of the lowest 50 hPa depth adiabatically to its LCL for determination of  $T_{\text{LCL}}$ . If  $T_{\text{diag}} > \bar{T}$  is not satisfied, the base of the potential USL is moved up to the next model level and comparison of  $T_{\text{diag}}$  with  $\bar{T}$  is repeated as long as the base of the potential USL is below the lowest 300 hPa of the atmosphere.

As the second step of diagnosis, if  $T_{\text{diag}} > \bar{T}$  is satisfied, the scheme calculates CAPE. To determine this value, the updraft variables are provisionally calculated by vertically integrating Eq. (3.5.70) and Eq. (3.5.71). If the updraft parcel has a positive CAPE value, parameterized convection is activated in the column.

### 3.5.6 Radiation

The radiation process employed in the MSM is almost identical to that in the GSM, as the codes of the GSM radiation process were ported into the MSM. The details are described in Subsection 3.2.3. Some differences are outlined below.

#### 3.5.6.1 Radiatively Active Constituents

Radiatively active gases accounted for in the MSM are identical to those in the GSM, although certain representations of optical properties differ. The absorption coefficients for water vapor used in the shortwave radiation scheme are based on Briegleb (1992), and aerosol optical depth climatology is based on total-column values from MODerate resolution Imaging Spectroradiometer (MODIS) and Total Ozone Mapping Spectrometer (TOMS) observations with seasonal variations. Other optical properties of aerosols are specified as continental and maritime background values without seasonal variation.

#### 3.5.6.2 Cloud Properties

The method of evaluating the effective radius of cloud ice particles is based on Ou and Liou (1995) with modification by McFarquhar *et al.* (2003). Following this method, the effective radius  $r_e$  [μm] is given by

$$r_e = -1.56 + 0.388D_e + 0.00051D_e^2, \quad (3.5.89)$$

$$D_e = 326.3 + 12.42T + 0.197T^2 + 0.0012T^3, \quad (3.5.90)$$

where  $T$  [°C] is the air temperature and  $D_e$  [μm] is the mean effective particle size. The effective radius of cloud water droplets is fixed at 15 μm.

#### 3.5.6.3 Cloud Fraction

The cloud fraction for the radiation scheme is diagnosed using a partial condensation method based on Sommeria and Deardorff (1977) and Mellor (1977), which is also employed to evaluate subgrid scale buoyancy flux in the boundary layer scheme (Subsection 3.5.7). This method involves calculation to determine the variance of the gridbox saturation deficit, associated with fluctuations of liquid water potential temperature ( $\theta'_l$ ) and total water specific humidity ( $q'_w$ ). Assuming unimodal Gaussian distribution for the deficit, the cloud fraction ( $R$ ) and the gridbox mean liquid water content ( $\bar{q}_l$ ) are given by

$$R = \frac{1}{\sqrt{2\pi}(2\sigma_s)} \int_0^\infty \exp\left[-\frac{(q_l - a_l \Delta \bar{q})^2}{2(2\sigma_s)^2}\right] dq_l = \frac{1}{2} \left[ 1 + \operatorname{erf}\left(\frac{Q_1}{\sqrt{2}}\right) \right], \quad (3.5.91)$$

$$\bar{q}_l = \frac{1}{\sqrt{2\pi}(2\sigma_s)} \int_0^\infty q_l \exp\left[-\frac{(q_l - a_l \Delta \bar{q})^2}{2(2\sigma_s)^2}\right] dq_l = 2\sigma_s \left[ R Q_1 + \frac{1}{\sqrt{2\pi}} \exp\left(-\frac{Q_1^2}{2}\right) \right], \quad (3.5.92)$$

where  $\Delta\bar{q} = \bar{q}_w - q_{sat}(\bar{T}_1)$ ,  $T_1$  denotes the liquid water temperature, and  $q_{sat}(T)$  is saturated specific humidity at the temperature  $(T)$ .  $a_l$  and  $Q_1$  are given by

$$a_l = \left[ 1 + \frac{L}{C_p} \left( \frac{\partial q_{sat}}{\partial T} \right)_{T=T_1} \right]^{-1}, \quad (3.5.93)$$

$$Q_1 = \frac{a_l \Delta\bar{q}}{2\sigma_s}, \quad (3.5.94)$$

where  $L$  denotes the latent heat of condensation and  $C_p$  is specific heat at the constant pressure of dry air. The standard deviation of the saturation deficit,  $\sigma_s$ , can be represented using the turbulent prognostic variables ( $\overline{\theta_1^2}$ ,  $\overline{q_w'^2}$ , and  $\overline{\theta'_1 q'_w}$ ) in the boundary layer scheme (Subsection 3.5.7) as follows:

$$\sigma_s^2 = \frac{1}{4} \left( a_l^2 \overline{q_w'^2} - 2a_l b_l \overline{\theta'_1 q'_w} + b_l^2 \overline{\theta_1'^2} \right), \quad (3.5.95)$$

$$b_l = a_l \Pi \left( \frac{\partial q_{sat}}{\partial T} \right)_{T=T_1}, \quad (3.5.96)$$

where  $\Pi$  is the Exner function.

Mixed-phase cloud diagnosis depends on air temperature. The ratio of ice cloud  $R_{ice}$  is given by

$$R_{ice} = r_{ice} R, \quad (3.5.97)$$

$$R_{water} = (1 - r_{ice}) R, \quad (3.5.98)$$

$$r_{ice} = \begin{cases} 0 & T_1 \geq T_{triple} \\ 1 - \left( \frac{T_1 - T_{ice}}{T_{triple} - T_{ice}} \right)^2 & T_{triple} > T_1 \geq T_{ice} \\ 1 & T_{ice} > T_1 \end{cases}, \quad (3.5.99)$$

where  $R_{water}$  is the ratio of the liquid water cloud,  $T_{triple}$  is the triple point temperature of water ( $= 273.16K$ ), and  $T_{ice} = 250.15K$ .

### 3.5.6.4 Radiative Timesteps

Longwave and shortwave radiation schemes are fully calculated every 15 minutes, while heating rates associated with longwave and shortwave radiation are corrected at every time step using the surface temperature and the solar zenith angle, respectively.

## 3.5.7 Boundary Layer

The boundary layer scheme represents vertical turbulent transport of momentum, heat and water. The fluxes exhibit the temporal tendency of the variable  $\phi (= u, v, \theta_1, q_w)$  associated with turbulent transport as follows:

$$\frac{\partial \phi}{\partial t} = - \frac{\partial}{\partial z} \overline{w' \phi'}. \quad (3.5.100)$$

The MSM employs the Mellor-Yamada-Nakanishi-Niino Level 3 model (MYNN3) (Nakanishi and Niino 2009) as a boundary layer scheme. This is a second order turbulent closure model in which it is assumed that the third-order moments of turbulent fluctuation can be depicted by lower-order moments.

### 3.5.7.1 Prognostic Equations and Fluxes

In the MYNN3 with boundary layer approximation, in which horizontal derivatives are ignored, just only four turbulent prognostic variables (including turbulent kinetic energy (TKE)), remain:

$$\frac{\partial q^2}{\partial t} = - 2 \left( \overline{u' w'} \frac{\partial \bar{u}}{\partial z} + \overline{v' w'} \frac{\partial \bar{v}}{\partial z} \right) + 2 \frac{g}{\theta_v} \overline{w' \theta'_v} - 2\varepsilon + \frac{\partial}{\partial z} \left( q \ell S_q \frac{\partial q^2}{\partial z} \right), \quad (3.5.101)$$

$$\frac{\partial \overline{\theta_1'^2}}{\partial t} = -2\overline{w'\theta'_1}\frac{\partial \overline{\theta_1}}{\partial z} - 2\varepsilon_\theta + \frac{\partial}{\partial z}\left(q\ell S_\theta \frac{\partial \overline{\theta_1'^2}}{\partial z}\right), \quad (3.5.102)$$

$$\frac{\partial \overline{q_w'^2}}{\partial t} = -2\overline{w'q'_w}\frac{\partial \overline{q_w}}{\partial z} - 2\varepsilon_q + \frac{\partial}{\partial z}\left(q\ell S_{q_w} \frac{\partial \overline{q_w'^2}}{\partial z}\right), \quad (3.5.103)$$

$$\frac{\partial \overline{\theta'_1 q'_w}}{\partial t} = -\overline{w'\theta'_1}\frac{\partial \overline{q_w}}{\partial z} - \overline{w'q'_w}\frac{\partial \overline{\theta_1}}{\partial z} - 2\varepsilon_{\theta q} + \frac{\partial}{\partial z}\left(q\ell S_{\theta q} \frac{\partial \overline{\theta'_1 q'_w}}{\partial z}\right), \quad (3.5.104)$$

where  $q^2$  is a doubled TKE value,  $\theta_l$  the liquid water potential temperature,  $q_w$  the total water content, and  $\ell$  the mixing length (see Subsection 3.5.7.4).  $\bar{\phi}$  represents an ensemble-averaged  $\phi$ , and  $\phi'$  is the turbulent fluctuation of  $\phi$ . The buoyancy flux,  $(g/\overline{\theta_v})\overline{w'\theta'_v}$ , and the dissipation terms  $\varepsilon_X$  are described in Subsections 3.5.7.2 and 3.5.7.3, respectively. Here,  $q_l$  is the mixing ratio of liquid water (including the ice phase), and

$$q^2 = (\overline{u'^2} + \overline{v'^2} + \overline{w'^2}), \quad (3.5.105)$$

$$\theta_l = \theta - \frac{L}{C_p T} q_l, \quad (3.5.106)$$

$$q_w = q_v + q_l, \quad (3.5.107)$$

where  $L$  denotes the latent heat of condensation and  $C_p$  specific heat at the constant pressure of dry air. The turbulent fluxes are diagnosed as

$$\overline{u'w'} = -q\ell(S_{M2.5} + S'_M)\frac{\partial \overline{u}}{\partial z}, \quad (3.5.108)$$

$$\overline{v'w'} = -q\ell(S_{M2.5} + S'_M)\frac{\partial \overline{v}}{\partial z}, \quad (3.5.109)$$

$$\overline{w'\theta'_1} = -q\ell(S_{H2.5} + S'_H)\frac{\partial \overline{\theta_1}}{\partial z}, \quad (3.5.110)$$

$$\overline{w'q'_w} = -q\ell(S_{H2.5} + S'_H)\frac{\partial \overline{q_w}}{\partial z}, \quad (3.5.111)$$

where  $S_X$  and  $S'_X$  are non-dimensional diffusion coefficients (see Subsection 3.5.7.5).

Once the prognostic equations (3.5.101), (3.5.102), (3.5.103), and (3.5.104) are integrated, the fluxes in Eqs. (3.5.108), (3.5.109), (3.5.110), and (3.5.111) and the tendencies of the turbulent prognostic variables can be calculated.

### 3.5.7.2 Buoyancy Flux

Buoyancy flux  $(g/\overline{\theta_v})\overline{w'\theta'_v}$  is a major origin of TKE production. With consideration of partial condensation effects assuming that the fluctuations of  $\theta_l$  and  $q_w$  from their mean values is expressed by the Gaussian probability density function (PDF) (Sommeria and Deardorff 1977), the width of which depends on  $\overline{\theta_1'^2}$ ,  $\overline{q_w'^2}$  and  $\overline{\theta_1'q'_w}$ , the buoyancy flux can be written as a function of the cloud fraction ( $R$ ) and the gridbox mean liquid water content ( $\overline{q_l}$ ) determined as moments of the PDF (see Subsection 3.5.6). Following Sommeria and Deardorff (1977) and Mellor (1977), the value is given by

$$\frac{g}{\theta_v} \overline{w'\theta'_v} = \frac{g}{\theta_v} \left( \beta_\theta \overline{w'\theta'_1} + \beta_q \overline{w'q'_w} \right), \quad (3.5.112)$$

$$\beta_\theta = 1 + 0.61\overline{q_w} - 1.61\overline{q_l} - \widetilde{R}a_l b_l c_l, \quad (3.5.113)$$

$$\beta_q = 0.61\overline{\theta} + \widetilde{R}a_l c_l, \quad (3.5.114)$$

$$\widetilde{R} = R - \frac{\overline{q_l}}{2\sigma_s} \frac{1}{\sqrt{2\pi}} \exp\left(-\frac{Q_1^2}{2}\right), \quad (3.5.115)$$

$$c_1 = (1 + 0.61\bar{q}_w - 1.61\bar{q}_l)\frac{\bar{\theta}}{T}\frac{L}{C_p} - 1.61\bar{\theta}. \quad (3.5.116)$$

Here,  $a_l$ ,  $b_l$ ,  $\sigma_s$  and  $Q_1$  are given by Eqs. (3.5.93), (3.5.96), (3.5.95), and (3.5.94).

### 3.5.7.3 Dissipation Terms

The dissipation terms  $\varepsilon_X$  appearing in the equations are parameterized on the basis of Kolmogorov's local isotropy assumption as

$$\varepsilon = \frac{q}{B_1\ell}q^2, \quad \varepsilon_\theta = \frac{q}{B_2\ell}\bar{\theta}_l'^2, \quad \varepsilon_q = \frac{q}{B_2\ell}\bar{q}_w'^2, \quad \varepsilon_{\theta q} = \frac{q}{B_2\ell}\bar{\theta}_l'q_w', \quad (3.5.117)$$

with the closure constants  $B_1$  and  $B_2$  (Nakanishi and Niino 2009).

### 3.5.7.4 Mixing Lengths

The mixing length  $\ell$  is given by

$$\frac{1}{\ell} = \frac{1}{L_S} + \frac{1}{L_T} + \frac{1}{L_B}, \quad (3.5.118)$$

where

$$L_S = \begin{cases} kz/3.7 & (\zeta \geq 1) \\ kz(1 + 2.7\zeta)^{-1} & (0 \leq \zeta < 1) \\ kz(1 - 100\zeta)^{0.2} & (\zeta < 0) \end{cases}, \quad (3.5.119)$$

$$L_T = 0.23 \frac{\int_0^\infty qz \, dz}{\int_0^\infty q \, dz}, \quad (3.5.120)$$

$$L_B = \begin{cases} q/N_1 & (\partial\theta/\partial z > 0, \zeta \geq 0) \\ [1 + 5(q_c/L_T N_1)^{1/2}]q/N_1 & (\partial\theta/\partial z > 0, \zeta < 0) \\ \infty & (\partial\theta/\partial z \leq 0) \end{cases}, \quad (3.5.121)$$

with the Brunt-Väisälä frequency  $N_1$ , the von Kármán constant  $k$ ,  $q_c = [(g/\bar{\theta}_v)\bar{w}'\bar{\theta}_v'L_T]^{1/3}$ , and  $\zeta = z/L_{MO}$  with the Monin-Obukhov length  $L_{MO}$ .

### 3.5.7.5 Nondimensional Diffusion Coefficients

$S_{M2.5}$  and  $S_{H2.5}$  are determined using the flux Richardson number and the empirical constants appearing in closure assumptions.  $S'_M$  and  $S'_H$  are correction terms induced by enhancement from the level 2.5 model (in which only TKE is treated as a prognostic variable) to the level 3 model. The correction terms depend on the turbulent prognostic variables ( $q^2$ ,  $\bar{\theta}_l'^2$ ,  $\bar{q}_w'^2$  and  $\bar{\theta}_l'q_w'$ ). Following Nakanishi and Niino (2004),  $S_{\theta_l}$ ,  $S_{\theta_q}$ , and  $S_{q_w}$  are assumed to be the same as  $S_q$ , and  $S_q = S_{\theta_l} = S_{\theta_q} = S_{q_w} = 2(S_{M2.5} + S'_M)$ . For technical details, refer to Nakanishi (2001) and Nakanishi and Niino (2004, 2006, 2009).

## 3.5.8 Surface Fluxes

The main procedures relating to surface processes involve the evaluation of surface fluxes. The surface scheme in the MSM employs a tiled approach in which different subgrid surface types are represented for land and sea. Turbulent fluxes are calculated for all tiles based on properties such as albedo and surface temperature, and are averaged over tiles with land fractions.

Within the surface layer, it is assumed that turbulent fluxes are constant with height and equivalent to surface values. These can be expressed in terms of differences between quantities in the lowest layer of the atmosphere ( $u_1, v_1, \theta_{v1}$ , and  $q_{v1}$ ) and the surface ( $\theta_{vs}$  and  $q_{vs}$ ) as

$$\overline{u'w'} = -C_m U_a u_1 , \quad (3.5.122)$$

$$\overline{v'w'} = -C_m U_a v_1 , \quad (3.5.123)$$

$$\overline{w'\theta_v} = -C_h U_a (\theta_{v1} - \theta_{vs}) , \quad (3.5.124)$$

$$\overline{w'q_v} = -C_q U_a (q_{v1} - q_{vs}) , \quad (3.5.125)$$

where  $u$  and  $v$  are horizontal wind velocity components,  $\theta_v$  virtual potential temperature,  $q_v$  specific humidity, and  $U_a$  wind speed near the surface. Following Kitamura and Ito (2016),  $U_a$  is expressed using turbulent kinetic energy,  $E (= q^2/2)$ , as

$$U_a = \sqrt{u_1^2 + v_1^2 + 2CE} , \quad (3.5.126)$$

where  $C = 2/3$ .  $q^2$  is offered by the boundary layer scheme (see Subsection 3.5.7).  $q_{vs}$  is parameterized with evaporation efficiency,  $\beta$ , and saturated specific humidity at the ground surface temperature,  $q_{sat}(T_s)$ , as follows:

$$q_{vs} = (1 - \beta)q_{v1} + \beta q_{sat}(T_s) \quad (3.5.127)$$

Over land,  $\beta$  is estimated from soil moisture

$$\beta = \begin{cases} w_g/0.3 & (w_g \leq 0.3) \\ 1 & (w_g > 0.3) \end{cases} , \quad (3.5.128)$$

where  $w_g$  is the volumetric water content at the surface, and is predicted using Eq. (3.5.148) (see Subsection 3.5.9). Over the sea, snow, and seaice,  $\beta$  is set to 1.

The transfer coefficients are formulated as

$$C_m(z) = \frac{k^2}{\left[ \ln \frac{z}{z_{0m}} - \psi_m \left( \frac{z}{L_{MO}} \right) + \psi_m \left( \frac{z_{0m}}{L_{MO}} \right) \right]^2} \\ \equiv \frac{k^2}{\Phi_m^2(z, L_{MO})} , \quad (3.5.129)$$

$$C_h(z) = \frac{k^2}{\left[ \ln \frac{z}{z_{0m}} - \psi_m \left( \frac{z}{L_{MO}} \right) + \psi_m \left( \frac{z_{0m}}{L_{MO}} \right) \right] \left[ \ln \frac{z}{z_{0h}} - \psi_h \left( \frac{z}{L_{MO}} \right) + \psi_h \left( \frac{z_{0h}}{L_{MO}} \right) \right]} \\ \equiv \frac{k^2}{\Phi_m(z, L_{MO}) \Phi_h(z, L_{MO})} , \quad (3.5.130)$$

where  $z$  is the height of the lowest model layer,  $L_{MO}$  the Monin-Obukhov length,  $z_{0m}$  and  $z_{0h}$  the roughness length for momentum and heat, and  $k = 0.4$  (von Kármán's constant). The integrated gradient functions for momentum,  $\psi_m$ , and heat,  $\psi_h$ , are given as functions of  $\zeta = z/L_{MO}$  following Beljaars and Holtslag (1991):

$$\psi_m(\zeta) = \begin{cases} -b \left( \zeta - \frac{c}{d} \right) \exp(-d\zeta) - a\zeta - \frac{bc}{d} & (\zeta \geq 0) \\ \frac{\pi}{2} - 2 \tan^{-1} x + \ln \frac{(1+x)^2(1+x^2)}{8} & (\zeta < 0) \end{cases} , \quad (3.5.131)$$

$$\psi_h(\zeta) = \begin{cases} -b \left( \zeta - \frac{c}{d} \right) \exp(-d\zeta) - \left( 1 + \frac{2}{3} a\zeta \right)^{\frac{3}{2}} - \frac{bc}{d} + 1 & (\zeta \geq 0) \\ 2 \ln \frac{1+x^2}{2} & (\zeta < 0) \end{cases} , \quad (3.5.132)$$

with  $a = 1$ ,  $b = 2/3$ ,  $c = 5$ ,  $d = 0.35$  and  $x = (1 - 16\zeta)^{1/4}$ .  $C_q$  over the sea is represented with the same formula as  $C_h$  except with the roughness length for moisture ( $z_{0q}$ ). For values over land, the effect of stomatal resistance is incorporated as follows:

$$C_q = \left[ U_a \left( r_s + \frac{1}{C_h U_a} \right) \right]^{-1}. \quad (3.5.133)$$

Stomatal resistance,  $r_s$ , depends on shortwave radiation flux towards the surface  $S$ :

$$r_s = r_{s, \text{day}} + \frac{r_{s, \text{night}}}{1 + \frac{S}{S_0}}, \quad (3.5.134)$$

where  $S_0 = 1 \text{ W m}^{-2}$ ,  $r_{s, \text{day}}$  is set to  $30 \text{ s m}^{-1}$  from April to October and  $60 \text{ s m}^{-1}$  in other months, and  $r_{s, \text{night}} = 300 \text{ s m}^{-1}$ .

The Monin-Obukhov length is determined from the following relation:

$$Ri_B = \frac{z}{L_{MO}} \frac{\Phi_h(z, L_{MO})}{\Phi_m^2(z, L_{MO})}, \quad (3.5.135)$$

which can be solved by using an iterative approach such as the Newton's method.  $Ri_B$  is the Bulk Richardson Number defined by

$$Ri_B = \frac{gz}{\frac{1}{2}(\theta_{v1} + \theta_{vs})} \frac{(\theta_{v1} - \theta_{vs})}{U_a^2}. \quad (3.5.136)$$

The roughness length on land is set depending on the land use of each grid point. Following Beljaars (1995), values over the sea are expressed as

$$z_{0m} = a_m \frac{\nu}{u_*} + a_{Ch} \frac{u_*^2}{g}, \quad (3.5.137)$$

$$z_{0h} = a_h \frac{\nu}{u_*}, \quad (3.5.138)$$

$$z_{0q} = a_q \frac{\nu}{u_*}, \quad (3.5.139)$$

where  $a_m = 0.11$ ,  $a_{Ch} = 0.018$ ,  $a_h = 0.40$ ,  $a_q = 0.62$ , and  $\nu$  the kinematic viscosity ( $= 1.5 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$ ).  $u_*$  is the friction velocity as defined by

$$u_* = \left( \overline{u'w'}^2 + \overline{v'w'}^2 \right)^{\frac{1}{4}}. \quad (3.5.140)$$

The screen level physical quantities such as temperature and dew point at 1.5 m height and wind at 10 m height are diagnosed by interpolation between the lowest model level and surface assuming the same gradient functions as in the scheme of surface process. Wind velocity at  $z_{10m}$  (10 m height),  $u_{10m}$ , and virtual potential temperature at  $z_{1.5m}$  (1.5 m height),  $\theta_{v1.5m}$  are diagnosed as

$$u_{10m} = \sqrt{\frac{C_m(z)}{C_m(z_{10m})}} u_1, \quad (3.5.141)$$

$$\theta_{v1.5m} = \theta_{vs} + \frac{C_h(z)}{C_h(z_{1.5m})} \sqrt{\frac{C_m(z_{1.5m})}{C_m(z)}} (\theta_{v1} - \theta_{vs}). \quad (3.5.142)$$

### 3.5.9 Ground Temperature and Soil Moisture

Ground surface temperature, which is used in evaluating surface fluxes, is predicted by solving a surface energy balance equation given by

$$c_s \frac{\partial T_s}{\partial t} = (1 - \alpha)S_{w\downarrow} + L_{w\downarrow} - \sigma T_s^4 - H - LE - G_s , \quad (3.5.143)$$

where  $S_{w\downarrow}$  and  $L_{w\downarrow}$  denote the fluxes of shortwave and longwave radiation towards the surface,  $\sigma$  is the Stefan-Boltzmann constant,  $\alpha$  is the surface albedo, and  $G_s$  is heat flux towards the ground.  $H$  and  $LE$  represent fluxes of sensible heat and latent heat from the surface:

$$H = -C_p \rho \overline{w' \theta'_v} , \quad (3.5.144)$$

$$LE = -L \rho \overline{w' q'_v} , \quad (3.5.145)$$

where  $C_p$  is the specific heat of dry air at a constant pressure,  $L$  is the latent heat of vaporization, and  $\rho$  is the density of air near the surface.

Ground temperature ( $T_g$ ), which is used in evaluating  $G_s$ , is predicted using a multi-layer model. The basic formula adopted is the heat conduction equation:

$$c_g \frac{\partial T_g}{\partial t} = -\frac{\partial G}{\partial z}, \quad G = -\lambda \frac{\partial T_g}{\partial z} , \quad (3.5.146)$$

where  $G$  is ground heat flux,  $c_g$  is heat capacity and  $\lambda$  is thermal conductivity. The soil column is discretized into eight layers to solve the above equations, numerically. The soil temperature for the lowest layer is fixed to a climatological value for forecasts. To obtain climatological data for ground temperature, monthly mean temperatures at standard pressure levels were first calculated from objective analysis conducted in 1985 and 1986. Next, these data were interpolated vertically to the model ground surface. Then, only the annual mean and the first harmonic component of annual change in surface temperature were extracted to obtain the climatological underground temperature at the  $k$ -th ground layer with the following equation:

$$T_g = \hat{T} + A \exp\left(-\frac{z}{d}\right) \cos\left\{\frac{2\pi}{365}(D - P) - \frac{z}{d}\right\} , \quad (3.5.147)$$

where  $\hat{T}$  is the mean ground surface temperature,  $A$  and  $P$  are the amplitude and the phase of the annual component of surface temperature, respectively,  $z_k$  is the depth of the  $k$ -th ground layer,  $d$  ( $= 2.65$  m) is the e-folding depth and  $D$  is the number of days since the beginning of the year.

The sea surface temperature is also spatially interpolated from the result of SST analysis (Section 5.2). The value is given as  $T_s$ , and is kept constant during the forecast period.

Soil moisture is predicted using the force-restore method based on Deardorff (1978):

$$\frac{\partial w_g}{\partial t} = -C_2 \frac{w_g - w_2}{\tau} + F_g , \quad (3.5.148)$$

$$\frac{\partial w_2}{\partial t} = F_2 , \quad (3.5.149)$$

where  $w_2$  is the mean volumetric water content under the ground,  $\tau$  a time constant (86400 s), and  $C_2 = 0.9$ .

The forcing terms  $F_g$  and  $F_2$  are given by

$$F_g = -C_g \frac{E - P_r}{\rho_w d_1} , \quad (3.5.150)$$

$$F_2 = -\frac{E - P_r}{\rho_w d_2} , \quad (3.5.151)$$

where  $E$  is the evaporation rate,  $P_r$  is the precipitation rate,  $\rho_w$  is the density of liquid water,  $d_1 = 0.1$  m, and  $d_2 = 0.5$  m.  $C_g$  is given as follows:

$$C_g = \begin{cases} 0.5 & \left( \frac{w_g}{w_{\max}} \geq 0.75 \right) \\ 14 - 22.5 \left( \frac{w_g}{w_{\max}} - 0.15 \right) & \left( 0.15 \leq \frac{w_g}{w_{\max}} < 0.75 \right) \\ 14 & \left( \frac{w_g}{w_{\max}} < 0.15 \right) \end{cases}, \quad (3.5.152)$$

where  $w_{\max}$  is the maximum volumetric water content ( $= 0.4$ ).

### 3.5.10 Parallelization

The Open Multi Processing (OpenMP) interface is employed for shared memory parallelization in the model, and the Message Passing Interface (MPI) is used for distributed memory parallelization. The model domain is split into horizontally two-dimensional sub-domains, and each decomposed sub-domain is assigned to one of the MPI processes (Aranami and Ishida 2004).

The OpenMP interface is used for parallelization inside the sub-domains. OpenMP threads are applied to loops for the y direction, and some horizontal loops (i.e., for the x and y directions) are fused to increase the loop length such that the load imbalance between threads is minimized. The z direction is used as the innermost loop at which vectorization is applied. Thus, kij-ordering is adopted for nested loops.

The sub-domains have halo regions that are exchanged with immediately adjacent MPI processes. As MPI communication and file I/O are time-consuming operations with the current supercomputer architecture, two types of overlapping are used in the model to significantly improve computational efficiency. One is overlapping of halo exchanges with the computation (Cats *et al.* 2008) to minimize the overhead of communication between MPI processes. The OpenMP interface is also used for this operation; while one thread is communicating with another MPI process, the other threads continue independent computation. The other technique involves an I/O server approach (Selwood 2012) to overlap file I/O with computation. In this method, some MPI processes are dedicated to file I/O. While computation continues, dedicated I/O processes read data from files and send them to the relevant computational processes. When output is required, the processes save the data in a dedicated buffer to invoke send operation and immediately continue computation. I/O processes receive the data and output the data to the disk.

For a 39-hour MSM forecast, the domain decomposition and I/O server configuration involve 41 nodes and 656 MPI processes, with 6 threads/MPI used on Cray XC50. The domain is divided into 28 parts in the x direction and 23 in the y direction, and there are 12 I/O servers.

### 3.5.11 Forecast Performance

Forecast verification is an essential process for monitoring the quality of NWP products and improving the model itself. This subsection outlines the performance of MSM precipitation forecasts with evaluation based on comparison with actually observed values.

Figures 3.5.4 and 3.5.5 show time-series representations of threat and bias scores for three-hour cumulative precipitation forecasts produced by the MSM with a 10-mm threshold from January 2011 to December 2017. Verification is performed using Radar/Raingauge-Analyzed Precipitation data (referred to here as R/A; see Subsection Subsection 4.4.1) as reference observations. The verification grid size is 20 km, meaning that forecast and observed precipitation over land or sea within 40 km from the coast is averaged over 20 km meshes. Using all verification grids, contingency tables are created for each initial time by comparing forecasts and observations, and aggregated into monthly or annual tables.

These figures indicate an increasing threat score tendency and a gradual approach of the bias score to unity over the previous seven years. This steady progress is attributable to the ongoing development of the forecast model and its data assimilation system with more extensive use of observation data. Bias and threat scores have increased since the February 2017 introduction of Asuca to the MSM (see Subsection 3.5.1).

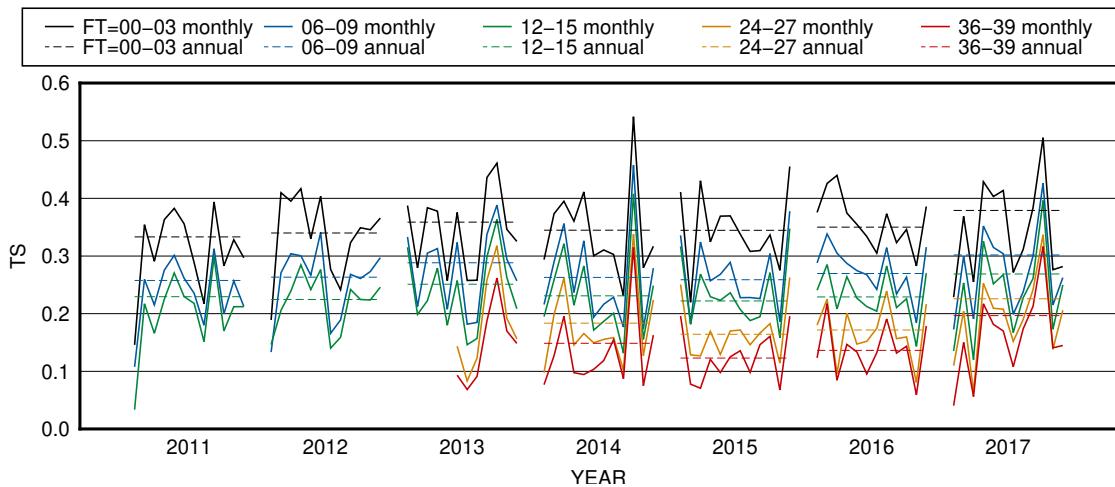


Figure 3.5.4: Monthly and annual threat scores of 3-hour cumulative precipitation at the 10mm threshold, against the R/A within 20km verification grids. The solid and dashed lines represent monthly and annual scores for each, FT represents the forecast range (hours). The verification period is from January 2011 to December 2017, but scores with 27- and 39-hour lead times are available only from June 2013 onward.

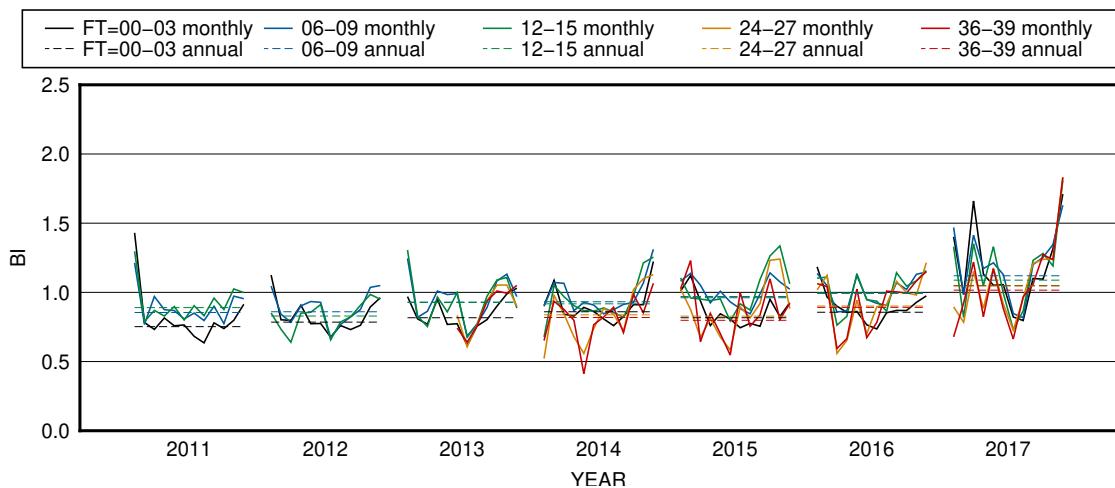


Figure 3.5.5: As per Figure 3.5.4, but for bias scores

## 3.6 Local Forecast Model

### 3.6.1 Introduction

The Local Forecast Model (LFM) was launched in August 2012, along with the Local Analysis (LA) described in Section 2.7, leveraging a supercomputer upgrade implemented in June 2012. The LFM provides weather information for aviation and disaster prevention, running forecasts at an even higher resolution than the 5-km Meso-Scale Model (MSM, Section 3.5). It has 2-km horizontal grid spacing and 58 vertical layers up to a height of approximately 20.2 km above sea level, and is designed to produce forecasts featuring greater detail with emphasis on predicting localized and short-lived extreme events.

The model focuses on providing very-short-range forecasts such as those covering the period nine hours ahead, and allows quick and frequent updating of forecasts using initial conditions with the latest observations assimilated by LA. The operation of the LFM was started with 8 runs per day on a domain covering the eastern part of Japan ( $1,100 \times 1,600$  km), and operation was extended in May 2013 to 24 runs per day on a domain covering Japan and its surrounding areas ( $3,160 \times 2,600$  km).

A new-generation non-hydrostatic model known as ASUCA (Ishida *et al.* 2009, 2010; Hara *et al.* 2012) replaced the previous JMA-NHM (Saito *et al.* 2006, 2007) as the forecast model of the LFM in January 2015 (Aranami *et al.* 2015), ahead of its implementation in the MSM in February 2017 (Subsection 3.5.1). Selected later upgrades of the ASUCA dynamical core and physics library applied to the MSM were also incorporated into the LFM in January 2017.

### 3.6.2 General Configurations

The LFM provides nine-hour forecasts every hour on the hour. The forecast domain covers Japan and its surrounding areas, and has 2-km horizontal grid spacing.

The LFM employs a model identical to the MSM with similar configuration (see Section 3.5). Some differences are described below (see Table 3.1.3).

- The LFM has 58 vertical layers with thicknesses increasing linearly from 40 m at the bottom to 661.5 m at the top. The model top is at a height of 20,189.5 m.
- Boundary conditions are obtained from MSM forecasts.
- The model is operated with a 50/3-second timestep.
- The main part of convection vertical transport is expected to be resolved with grid mean vertical velocity at a horizontal grid spacing of 2 km. However, this does not necessarily mean that all phenomena associated with convection can be resolved. In particular, phenomena on unresolved scales (such as small-scale convergences and topography variances) can induce the forced lifting needed to initiate convection. Accordingly, parameterization to represent convective initiation is used in the LFM (Hara 2015), thereby mitigating delays in the onset of convection.
- The LFM does not incorporate the advanced version of the cloud microphysics scheme used in the MSM with revised particle size distribution functions for snow and rain (see Subsection 3.5.4).
- As described in Subsection 3.5.6, the cloud fraction used in the radiation process is diagnosed in consideration of fluctuations of temperature and water content from their grid mean values over each of the grids. As the fluctuations are expected to be smaller in higher resolution models, the width of the probability density function depicting the characteristics of the fluctuation was made smaller than that used in the MSM.
- The domain decomposition and I/O server configuration in parallelization involve 72-nodes, 864 MPI processes and 8 threads/MPI on Cray XC50. The domain is divided into 34 parts in the x-direction and 25 in the y-direction, and there are 14 I/O servers.

### 3.6.3 Forecast Performance

LFM forecast performance is evaluated in the same way as for MSM forecasts (see Subsection 3.5.11). Figures 3.6.1 and 3.6.2 show time-series representations of threat and bias scores for LFM one-hour cumulative precipitation forecasts at a 10-mm threshold. It can be seen that the LFM tends to underestimate precipitation in the one-hour forecast range and overforecast with lead times from two hours onward, exhibiting a spin-up period during the initial forecast stages.

To verify the skill of LFM convective precipitation forecasts, the Fractions Skill Score (FSS ; see Subsection A.2.12) of one-hour cumulative precipitation from the LFM is compared with that from the MSM. Figure 3.6.3 shows FSS differences between the two models averaged over all initial times in August 2017. The LFM is inferior in the one-hour forecast range when the threshold is less than 10 mm due to the spin-up effect, but is superior in the forecast range of 3 to 9 hours. In particular, in the spatial scale over 80 km, the LFM FSS is better than that of the MSM regardless of the threshold. Figure 3.6.4 shows a Hovmöller diagram of FSS differences between the two models for a spatial scale of 80 km and a forecast range of nine hours averaged over all initial times in each month. The LFM outperforms the MSM at thresholds over 5 mm in summer and under 10 mm in autumn and winter. Contributions are expected from the strength of the LFM in predicting convective rain in summer and weak precipitation caused by the winter monsoon.

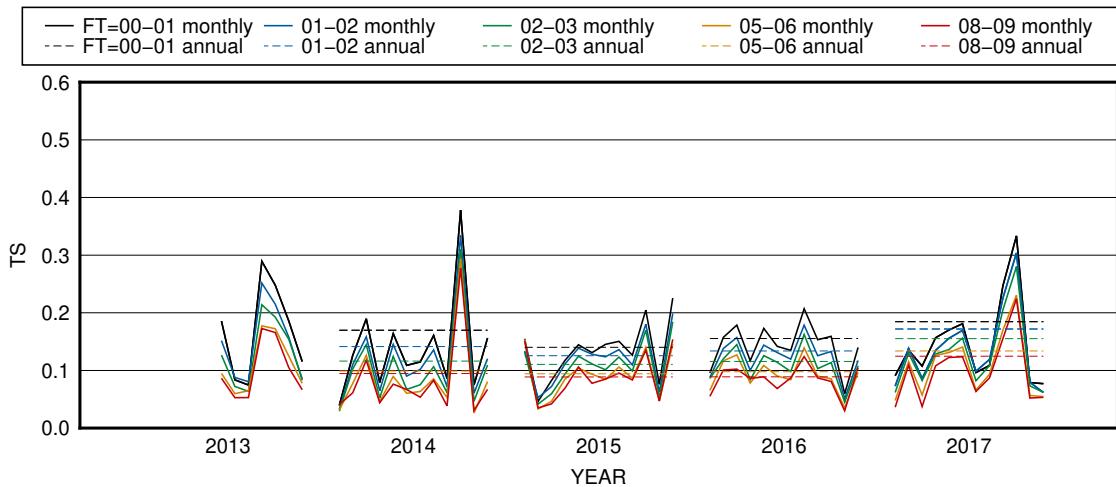


Figure 3.6.1: Monthly and annual threat scores for one-hour cumulative precipitation at the 10mm threshold against the R/A within 20km verification grids. The solid and dashed lines represent monthly and annual scores for each, and FT represents the forecast range (hours). The verification period is from June 2013 to December 2017.

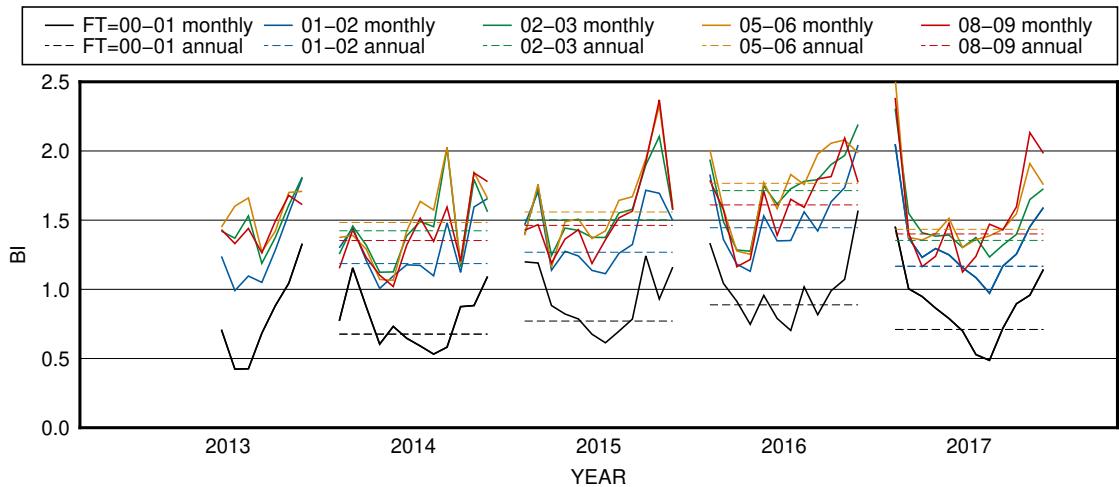


Figure 3.6.2: As Per Figure 3.6.1 but for bias scores

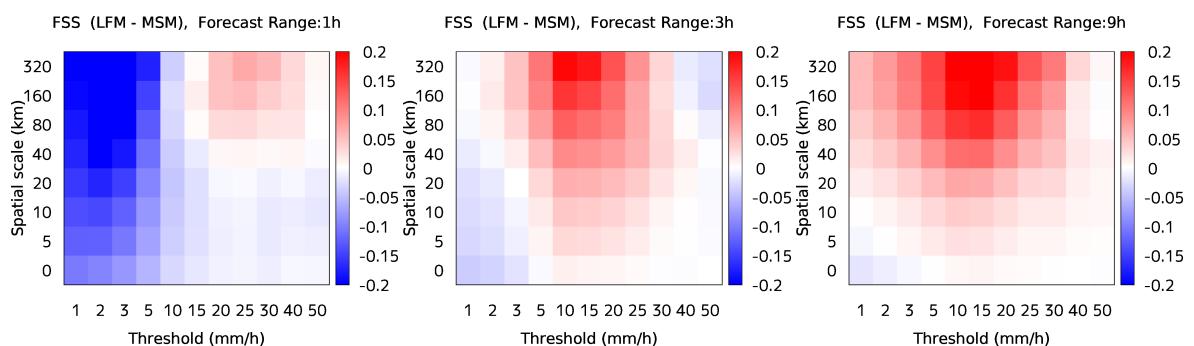


Figure 3.6.3: Monthly averaged subtraction of Fraction Skill Scores for MSM one-hour cumulative precipitation from those of the LFM. The forecast ranges are 1, 3 and 9 hours. The verification period is August 2017.

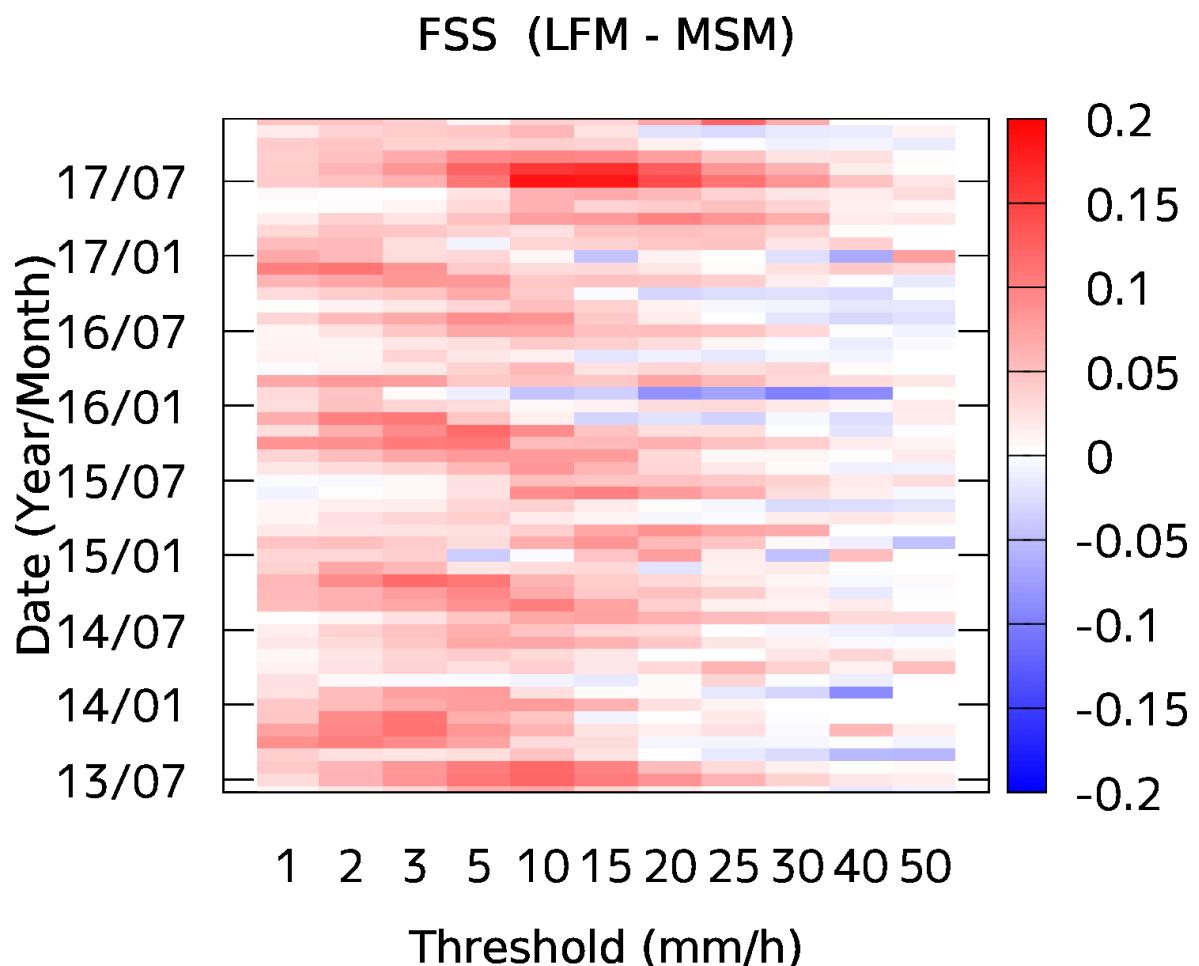


Figure 3.6.4: Monthly averaged subtraction of Fraction Skill Scores for MSM one-hour cumulative precipitation from those of the LFM. The forecast range is nine hours and the spatial scale is 80 km. The verification period is from June 2013 to December 2017.

## 3.7 Atmospheric Transport and Dispersion Model

### 3.7.1 Introduction

In July 1997, JMA was designated as a Regional Specialized Meteorological Centre (RSMC) specializing in the provision of atmospheric transport and dispersion model (ATDM) products for environmental emergency response covering Regional Association II (RA-II) of the World Meteorological Organization (WMO). RSMC Tokyo is required to provide advice on the atmospheric transport of pollutants related to nuclear facility accidents and radiological emergencies. The RSMCs ATDM products are sent to the National Meteorological Services (NMS) of WMO Member States in RA-II and to the secretariats of WMO and of the International Atomic Energy Agency (IAEA). The basic procedure of the service is defined in WMO (2017).

### 3.7.2 Model

#### 3.7.2.1 Basic Model Description

The ATDM used by JMA is based on Iwasaki *et al.* (1998) with modifications developed by Kawai (2002). It involves the use of a Lagrangian approach in which tracer particles released at the temporal and spatial points of pollutant emission are displaced due to horizontal and vertical advection and diffusion and laid down through dry and wet deposition. Computation of advection, dispersion (turbulent diffusion) and deposition is based on the output of the operational numerical weather prediction (NWP) model, involving three-hourly model-level global model (GSM; see Section 3.2) outputs with temporal and spatial interpolation to tracer points. A total of 1,000,000 tracer particles are used in the operational ATDM, and time-integrated concentration and deposition are calculated using 0.5x0.5-degree latitude-longitude grids.

Horizontal velocities of tracers are estimated in accordance with Gifford (1982) as

$$\begin{aligned} u(t) &= u_m(t) + u'(t), \\ u'(t) &= R_h u'(t - \delta t) + \sqrt{1 - R_h^2} \sigma G, \end{aligned} \quad (3.7.1)$$

$$\begin{aligned} v(t) &= v_m(t) + v'(t), \\ v'(t) &= R_h v'(t - \delta t) + \sqrt{1 - R_h^2} \sigma G, \end{aligned} \quad (3.7.2)$$

where  $u$  and  $v$  are zonal and meridional wind speed components, and  $u_m$  and  $v_m$  are those of forecast values from the global NWP system.  $R_h$  is an autocorrelation of Lagrangian velocity as estimated using  $e^{-\delta t/T_L}$ , where  $\delta t$  is the single time step length and  $T_L$  is the Lagrangian time scale.  $\sigma$  is the root mean square of horizontal velocity, which can be estimated as  $(K_h/T_L)^{1/2}$  with reference to the horizontal diffusion coefficient  $K_h$ .  $G$  represents random fluctuation whose statistical distributions have the Gaussian distribution function with a mean of 0 and a standard deviation of 1. The Monte Carlo method is used to determine velocities and displacements of each tracer particle. The horizontal diffusion coefficient  $K_h$  needs to be parameterized in consideration of the model resolution and the temporal and spatial variations of meteorological fields.

An appropriate constant value is set to reduce the computational time burden. Horizontal displacements  $\delta x$  and  $\delta y$  are given by

$$\delta x = u(t)\delta t, \quad (3.7.3)$$

$$\delta y = v(t)\delta t. \quad (3.7.4)$$

The vertical displacement  $\delta z$  for a single time step  $\delta t$  is given as

$$\delta z = w_m(t)\delta t + \Sigma G \sqrt{2K_v \delta t}. \quad (3.7.5)$$

Here,  $w_m$  is the vertical wind speed given by the GSM forecast. The vertical diffusion coefficient depends on atmospheric vertical profiles. The time step for the integration of vertical diffusion  $\delta t'$  is much shorter than

those for the integration of horizontal diffusion and advection. This shorter time step is used so that vertical displacement caused by diffusion does not exceed the thickness of the model layer. The vertical diffusion coefficient  $K_v$  is set with reference to meteorological parameters processed by the NWP model in a way analogous to the molecular diffusion coefficient estimation of Louis *et al.* (1982), and is given as follows:

$$K_v = l^2 \left| \frac{\partial c}{\partial z} \right| F(R_i), \quad (3.7.6)$$

where  $c$  is the horizontal wind speed, and the parameters  $l$  and  $R_i$  are the vertical mixing length of turbulence and the flux Richardson number, respectively. The similarity function of  $F(R_i)$  is defined with reference to Louis *et al.* (1982). The mixing length is expressed as a function of the geometric height  $z$ :

$$l = \frac{\kappa z}{1 + \kappa z/l_0} \quad (3.7.7)$$

where  $\kappa$  is von Kármán's constant and  $l_0$  is the maximum mixing length.

### 3.7.2.2 Dry and Wet Deposition

The surface tracer flux  $F$  associated with dry deposition is presented using deposition velocity  $V(z_r)$  and concentration  $C(z_r)$  at the reference level  $z_r$  as

$$F \equiv V(z_r)C(z_r). \quad (3.7.8)$$

For simplicity, the deposition rate is set to  $F/z_r$  following Kitada *et al.* (1986).

For wet deposition, only wash-out processes are parameterized. The wet deposition rate  $\Lambda[1/h]$  is approximated as a function of precipitation intensity  $P[\text{mm}/\text{h}]$  as predicted by the GSM with the below-cloud scavenging ratio per hour given by Kitada (1994) as

$$\Lambda \approx 0.1P^{0.75}. \quad (3.7.9)$$

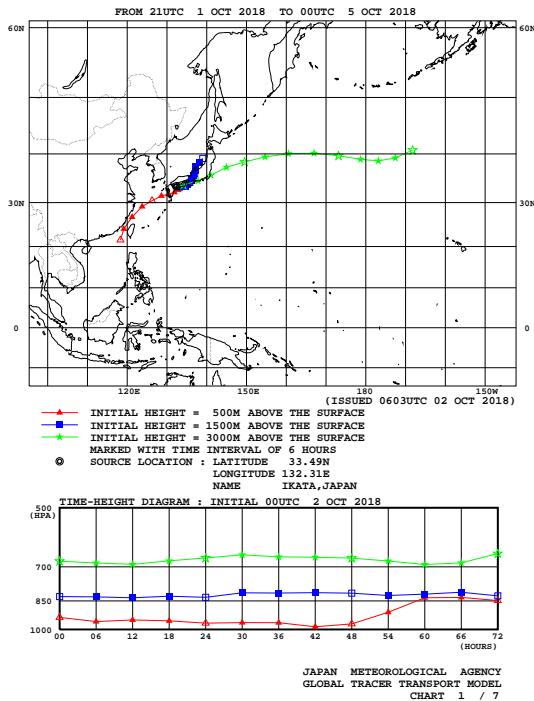
The Monte Carlo method is applied to decide which tracer particles are removed from the atmosphere at the above-mentioned dry and wet deposition rates. Noble gases such as  $^{133}\text{Xe}$  are excluded from these depositing treatments.

### 3.7.3 Products

ATDM products are charts of 3D trajectories, time-integrated pollutant concentrations, total depositions. Sample charts are shown in Figures 3.7.1 - 3.7.7, and information on related interpretation is provided in Appendix 2.2.22 of WMO (2017).

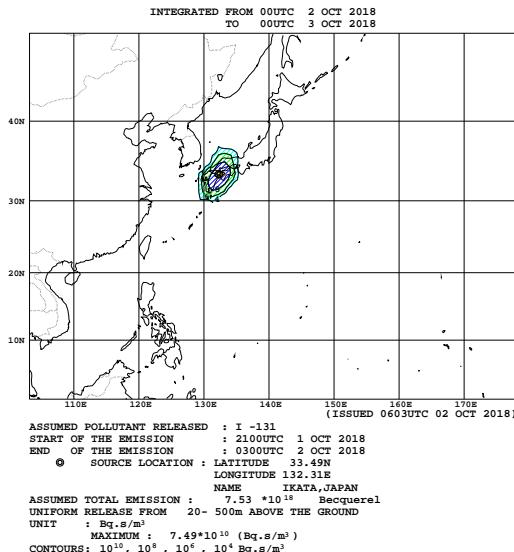
## EXERCISE-EXERCISE-EXERCISE

### 3-D TRAJECTORY

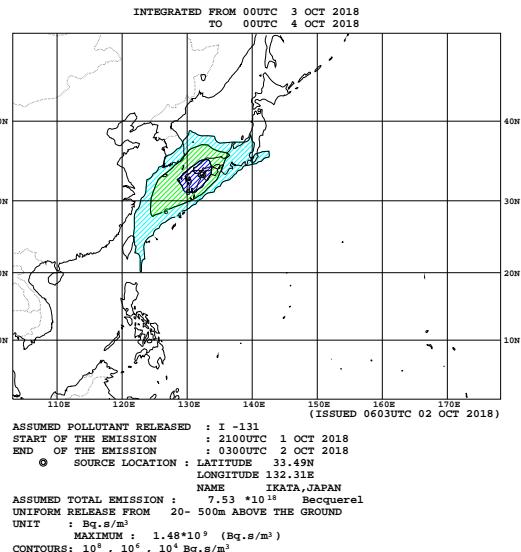


**Figure 3.7.1: Example of the 3D trajectories**  
**EXERCISE-EXERCISE-EXERCISE EXERCISE-EXERCISE-EXERCISE**

### TIME INTEGRATED SURFACE - 500m LAYER CONCENTRATION



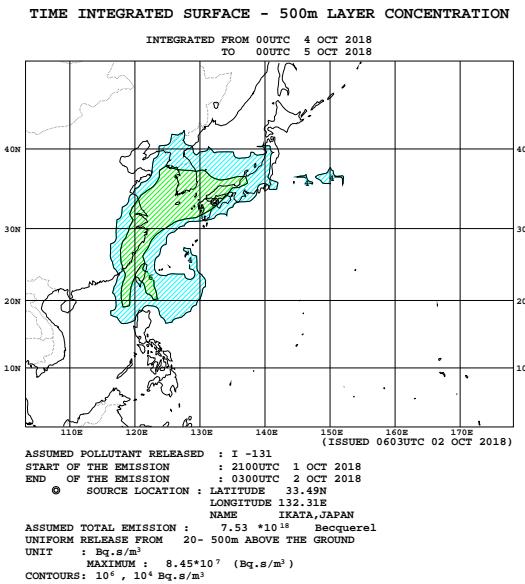
### TIME INTEGRATED SURFACE - 500m LAYER CONCENTRATION



**Figure 3.7.2: Example of time-integrated concentration in forecasts of up to 24 hours**

**Figure 3.7.3: Example of time-integrated concentration in forecasts of up to 48 hours**

### EXERCISE-EXERCISE-EXERCISE



### EXERCISE-EXERCISE-EXERCISE

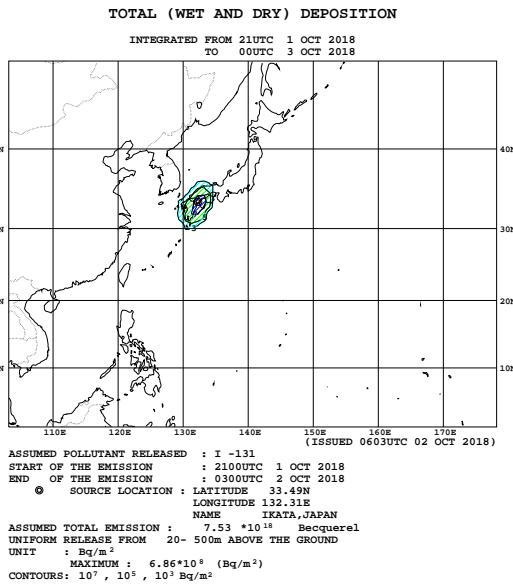


Figure 3.7.4: Example of time-integrated concentration in forecasts of up to 72 hours

### EXERCISE-EXERCISE-EXERCISE

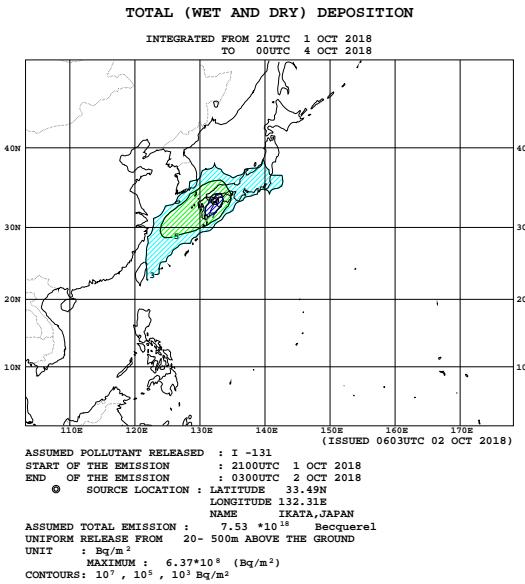


Figure 3.7.5: Example of total deposition in forecasts of up to 24 hours

### EXERCISE-EXERCISE-EXERCISE

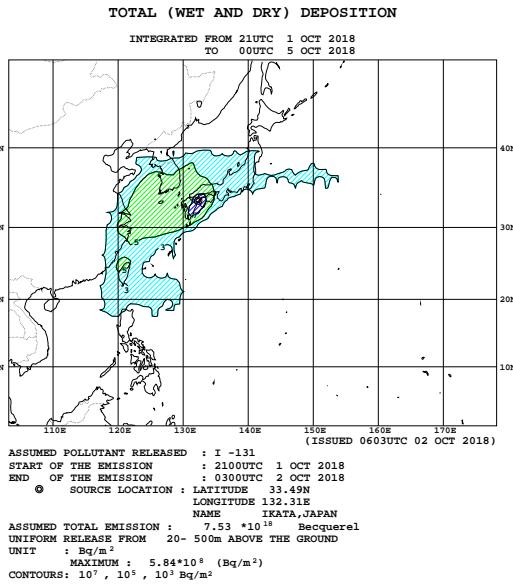


Figure 3.7.6: Example of total deposition in forecasts of up to 48 hours

JAPAN METEOROLOGICAL AGENCY  
GLOBAL TRACER TRANSPORT MODEL  
CHART 6 / 7

Figure 3.7.7: Example of total deposition in forecasts of up to 72 hours

JAPAN METEOROLOGICAL AGENCY  
GLOBAL TRACER TRANSPORT MODEL  
CHART 7 / 7

## 3.8 Chemical Transport Model

### 3.8.1 Introduction

JMA is providing wide-ranging atmospheric environmental information including Kosa (Aeolian dust) forecasts, UV index forecasts and photochemical oxidant information. Information on Kosa (Figure 3.8.1) and UV index (Figure 3.8.2 and Figure 3.8.3) are provided via the official website, and photochemical oxidant information is provided to prefectural governments as a basis for related advisories. This information is based on operational predictions conducted using several chemical transport models (CTMs). JMA has operated a global aerosol CTM for the Kosa forecast since January 2004, and a global ozone CTM for the UV index forecast since May 2005. The organization began to utilize the global ozone CTM for photochemical oxidant information in August 2010, and this was replaced by a regional ozone CTM in March 2015.

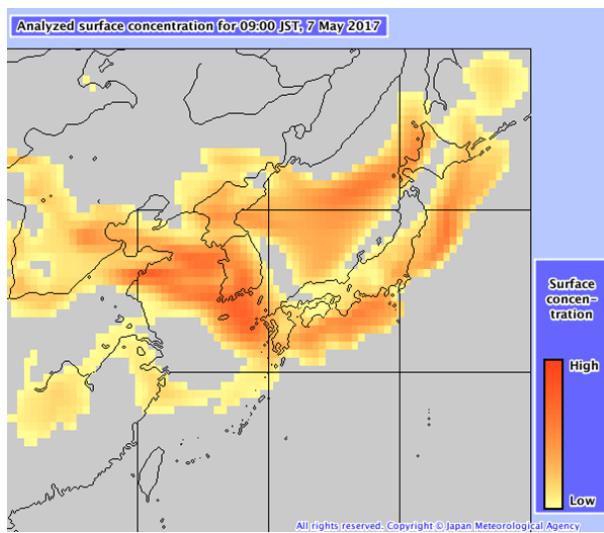


Figure 3.8.1: Kosa (Aeolian dust) prediction web page (<https://www.jma.go.jp/en/kosafcst/>).

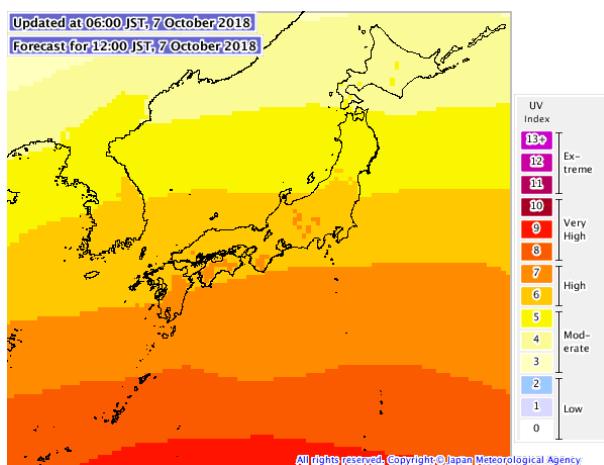


Figure 3.8.2: Clear-sky UV index forecast web page (<https://www.jma.go.jp/en/uv/>).

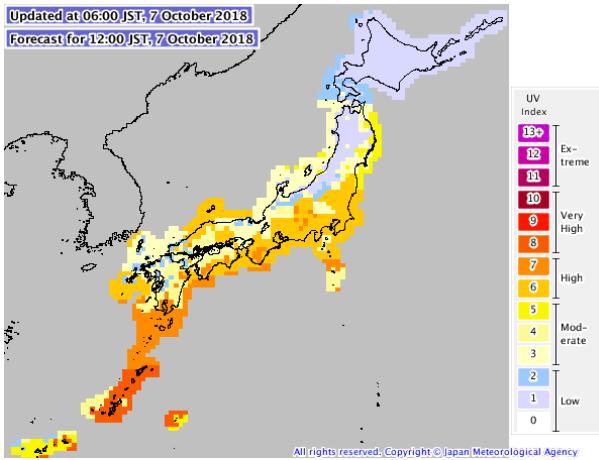


Figure 3.8.3: UV index forecast web page (<https://www.jma.go.jp/en/uv/>).

### 3.8.2 General Description of CTMs

#### 3.8.2.1 Coupling of chemical and meteorological parts

Schematic illustrations of individual CTM structure are shown in Figure 3.8.4 to Figure 3.8.6. The chemical modules are coupled with atmospheric forecast models, which derive meteorological fields such as wind and precipitation as required in chemical computations. In the global CTMs used for aerosol and ozone prediction, chemical modules are directly coupled with the Atmospheric General Circulation Model (MRI-AGCM3; Mizuta *et al.* (2012)) developed by the Meteorological Research Institute (MRI) of JMA as part of the Earth System Model (MRI-ESM; Yukimoto *et al.* (2011)). Online coupling is achieved using Scup coupler software (Yoshimura and Yukimoto (2008)). In the regional CTM utilized for photochemical oxidant information, it is coupled with the regional Non-Hydrostatic atmospheric Model (JMA-NHM; Saito *et al.* (2006)), which is the previous version of the JMA mesoscale weather forecast model. NHM meteorological computation is first conducted to derive meteorological fields, and the chemical modules are then run using them (offline coupling).

#### 3.8.2.2 Relaxation to Analysis/Forecast Field

Due to the need for handling of more chemical variables and processes than atmospheric prediction models such as the GSM, CTMs require more computational resources. Accordingly, lower spatial resolutions of the global CTMs are generally adopted in operational use. As a result, the spatial resolution of the atmospheric model in the global CTM differs from that of the model used for daily meteorological analysis and forecasting. There may also be differences in adopted model processes between the two atmospheric models. Due to these discrepancies, meteorological fields computed in the CTM are not necessarily consistent with those of operational meteorological analysis and forecast. In order to conduct CTM with more accuracy, the nudging techniques shown in Eq. (3.8.1) is often utilized:

$$\left( \frac{\partial x}{\partial t} \right)_{nudging} = - \frac{x - x_{analysis/forecast}}{\tau} \quad (3.8.1)$$

where  $x$  is a meteorological variable of the CTM at a certain time  $t$ ,  $x_{analysis/forecast}$  is the corresponding analysis or forecast value derived from a higher-resolution atmospheric model, and  $\tau$  is a relaxation time of 6-24 hours. This technique enables the CTM to simulate meteorological fields realistically during the prediction period.

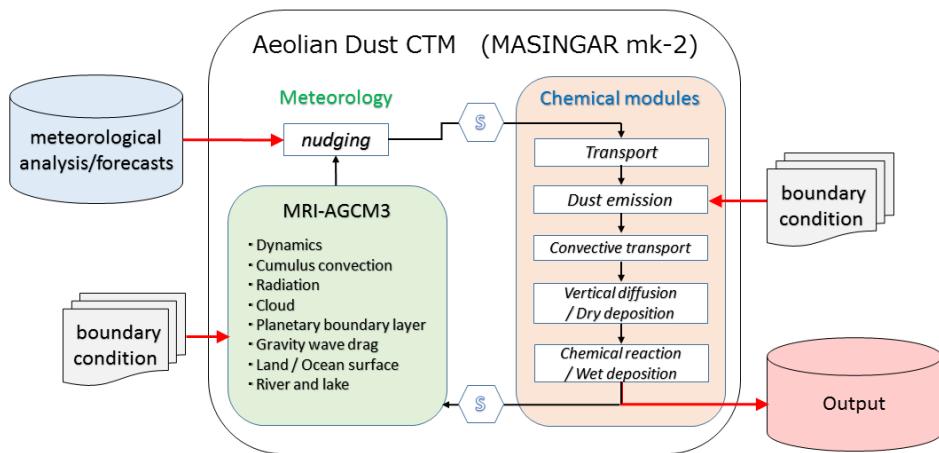


Figure 3.8.4: Schematic illustration of the CTM for Kosa (Aeolian dust) prediction. (The symbol mark S indicates the Scup coupler.)

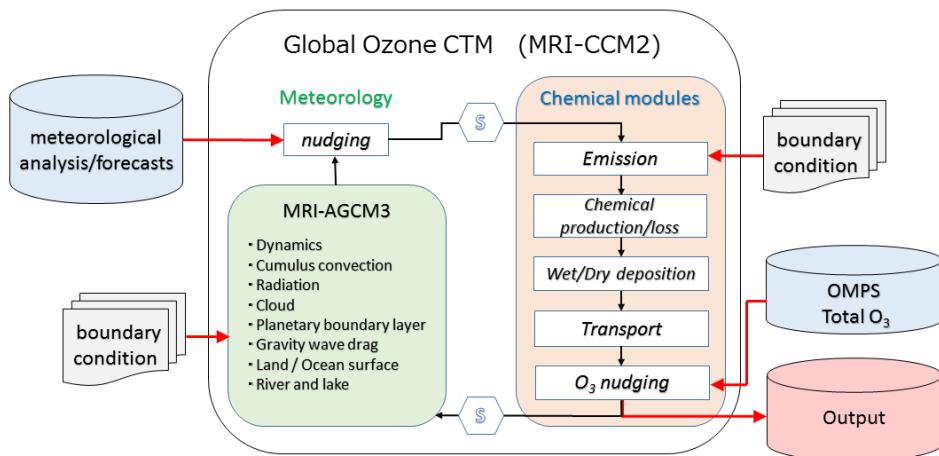


Figure 3.8.5: Same as Figure 3.8.4 except for the CTM for UV index forecast.

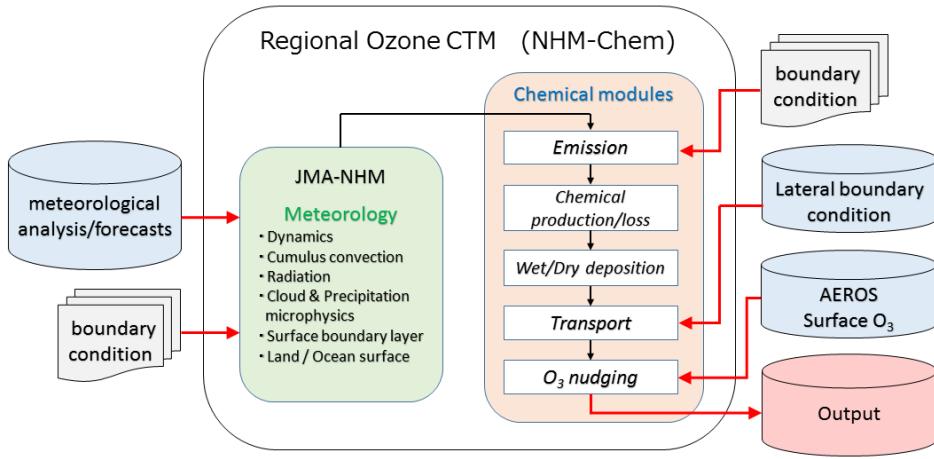


Figure 3.8.6: Same as Figure 3.8.4 except for the CTM for photochemical oxidant information.

### 3.8.3 Aerosol CTM Used for Kosa Prediction

#### 3.8.3.1 Basic Framework

The chemical transport model MASINGAR (Model of Aerosol Species IN Global Atmosphere; Tanaka *et al.* (2003)) used for Kosa (Aeolian dust) prediction consists of transport modules for advective transport (semi-Lagrangian scheme) and sub-grid scale eddy diffusive/convective transport, as well as other modules of surface emission, dry/wet deposition and chemical reactions. It incorporates consideration for sulfate, black and organic carbon, mineral dust and sea salt as aerosol species. The prediction period is 96 hours and the spatial resolution is TL479L40 (horizontal grid interval: approx. 40km; 40 vertical layers up to 0.4 hPa). The emission flux of mineral dust aerosol depends on meteorological, geographical and soil surface conditions such as wind speed, land use, vegetation type, soil moisture and soil types. The emission flux  $F$  of dust (soil particles with diameter  $D$ ) is expressed in proportion to the saltation flux  $Q$ :

$$F(D) = \alpha(D, d_s)Q(d_s) \quad (3.8.2)$$

where  $d_s$  is a diameter of saltation particles and the proportional coefficient  $\alpha$  depends on both  $D$  and  $d_s$ .

The saltation flux  $Q$  is set zero when the friction velocity on a bare surface  $u_*$  is lower than the following threshold velocity:

$$u_{*t}(d_s) = f_w \sqrt{A_N \left( \frac{\rho g d_s}{\rho_a} + \frac{\Gamma}{\rho_a d_s} \right)} \quad (3.8.3)$$

where  $A_N$  and  $\Gamma$  are certain constants,  $\rho$  is soil particle density,  $\rho_a$  is air density,  $g$  is gravitational acceleration and  $f_w$  is a factor depending on soil moisture.

When  $u_*$  is larger than  $u_{*t}$ , the saltation flux is expressed as

$$Q(d_s) = \frac{c(d_s)\rho_a u_*^3}{g} \left( 1 - \frac{u_{*t}(d_s)^2}{u_*^2} \right) \quad (3.8.4)$$

where  $c$  is a coefficient depending on  $d_s$ .

### 3.8.3.2 Relaxation to Analysis/Forecast Field

Although Kosa (mineral dust) aerosol observation data are not assimilated to the MASINGAR, meteorological fields such as wind and the atmospheric temperature are nudged to the operational analysis/forecast data at the initial time and during the prediction period. This enables more reliable simulation of dust emission (described in 3.8.3.1) and transport in the MASINGAR.

### 3.8.3.3 Verification

Predictions by the operational Kosa CTM are verified against surface synoptic observations (reported present weather code (ww)). The model score is calculated using categorical verification as outlined in Appendix A (see Table 3.8.1). The threshold surface dust concentration value for Kosa prediction is currently set to be 90  $\mu\text{g}/\text{m}^3$ . The threat score for Kosa prediction after 24 hours in the Japan area is 0.28 averaged over recent five years (2014-2018) spring season.

Table 3.8.1: Verification indices of categorical prediction for Kosa (Aeolian dust) CTM (ww: present weather code,  $\rho_s^{dust}$ : predicted concentration of surface dust)

	Observed (ww = 06 – 09, 30 – 35, 98)	Not Observed
Forecasted ( $\rho_s^{dust} \geq 90 \mu\text{g}/\text{m}^3$ )	FO	FX
Not Forecasted ( $\rho_s^{dust} < 90 \mu\text{g}/\text{m}^3$ )	XO	XX

## 3.8.4 Global Ozone CTM Used for UV Index Forecast

### 3.8.4.1 Basic Framework

The MRI-CCM2 (Meteorological Research Institute Chemical Climate Model version 2) developed by Deushi and Shibata (2011) is a global ozone CTM to predict distributions of atmospheric ozone and other related trace gases for the UV index forecast. The chemical reaction scheme is based on the chemical families method and the chemical module includes 90 chemical species (64 for the long-lived species and 26 for the short-lived species; see Table 3.8.2.). In the latest version of the chemical module, 259 chemical reactions are considered (184 gas-phase, 59 photolysis and 16 heterogeneous reactions). The prediction period of the ozone CTM is 120 hours and the resolution is TL159L64 (approx. 120 km horizontal grid intervals and 64 vertical layers up to 0.01hPa). The chemical module treats grid scale transport with a semi-Lagrangian scheme, sub-grid scale convective transport and turbulent diffusion, dry and wet deposition and emissions of trace gases from various sources.

The dynamical module in the global ozone CTM also has an assimilation process in the meteorological field. Nudging is applied to operationally conducted global atmospheric analysis and forecast during the integration of the CTM. This starts 72 hours prior to the UV index analysis time and ends 48 hours after the prediction period.

### 3.8.4.2 Data Assimilation

The chemical module in the global ozone CTM has an assimilation process similar to that for the meteorological field (Eq. (3.8.1)). Total column ozone from the Ozone Mapping and Profiler Suite (OMPS/Suomi-NPP) satellite monitoring is assimilated once a day. The weight of the model guess with OMPS data is determined by the ratio of the root mean square error against surface observation with a Dobson and Brewer spectrophotometer. 1/3 is adopted as the ratio of relative contribution of the model guess to OMPS observation for nudging. OMPS data obtained within 72 hours prior to the UV index analysis time are assimilated into the CTM.

Table 3.8.2: List of species used in the atmospheric ozone CTM

<b>Long-lived</b>			
01: N <sub>2</sub> O	02: CH <sub>4</sub>	03: H <sub>2</sub> O	04: NO <sub>y</sub>
05: HNO <sub>3</sub>	06: N <sub>2</sub> O <sub>5</sub>	07: Cl <sub>y</sub>	08: O <sub>x</sub>
09: CO	10: OCIO	11: CO <sub>2</sub>	12: Passive tracer
13: HCl	14: ClONO <sub>2</sub>	15: HOCl	16: Cl <sub>2</sub>
17: H <sub>2</sub> O <sub>2</sub>	18: CINO <sub>2</sub>	19: HBr	20: BrONO <sub>2</sub>
21: NO <sub>x</sub>	22: HO <sub>2</sub> NO <sub>2</sub>	23: ClO <sub>x</sub>	24: BrO <sub>x</sub>
25: Cl <sub>2</sub> O <sub>2</sub>	26: HOBr	27: CCl <sub>4</sub> (CFC-10)	28: CFCl <sub>3</sub> (CFC-11)
29: CF <sub>2</sub> Cl <sub>2</sub> (CFC-12)	30: Br <sub>y</sub>	31: CH <sub>3</sub> Cl	32: CH <sub>3</sub> Br
33: CF <sub>2</sub> ClBr (Halon 1211)	34: Cf <sub>3</sub> Br (Halon 1301)	35: COF <sub>2</sub>	36: HF
37: CH <sub>2</sub> O	38: CH <sub>3</sub> OOH	39: C <sub>2</sub> H <sub>6</sub>	40: CH <sub>3</sub> CHO
41: C <sub>2</sub> H <sub>5</sub> OOH	42: PAN (CH <sub>3</sub> C(O)OONO <sub>2</sub> )	43: CH <sub>3</sub> C(O)OOH	44: C <sub>3</sub> H <sub>8</sub>
45: ACET (CH <sub>3</sub> C(O)CH <sub>3</sub> )	46: C <sub>3</sub> H <sub>7</sub> OOH	47: HACET (CH <sub>3</sub> C(O)CH <sub>2</sub> OH)	48: MGLY (CH <sub>3</sub> C(O)CHO)
49: C <sub>2</sub> H <sub>4</sub>	50: GLY ALD (HOCH <sub>2</sub> CHO)	51: GPAN (HOCH <sub>2</sub> C(O)OONO <sub>2</sub> )	52: GC(O)OOH (HOCH <sub>2</sub> C(O)OOH)
53: C <sub>3</sub> H <sub>6</sub>	54: ONIT (CH <sub>3</sub> C(O)CH <sub>2</sub> ONO <sub>2</sub> )	55: POOH (HOC <sub>3</sub> H <sub>6</sub> OOH)	56: C <sub>4</sub> H <sub>10</sub>
57: C <sub>5</sub> H <sub>8</sub> (isoprene)	58: MACR	59: ISON	60: ISOPOOH
61: NALD	62: MACROOH	63: MPAN	64: C <sub>10</sub> H <sub>16</sub> (terpenes)
<b>Short-lived</b>			
01: O(1D)	02: OH	03: Cl	04: O(3P)
05: O <sub>3</sub>	06: HO <sub>2</sub>	07: NO <sub>2</sub>	08: NO
09: Br	10: N	11: ClO	12: BrO
13: NO <sub>3</sub>	14: BrCl	15: H	16: CH <sub>3</sub> O <sub>2</sub>
17: C <sub>2</sub> H <sub>5</sub> O <sub>2</sub>	18: CH <sub>3</sub> C(O)O <sub>2</sub>	19: C <sub>3</sub> H <sub>7</sub> O <sub>2</sub>	20: ACETO <sub>2</sub> (CH <sub>3</sub> C(O)CH <sub>2</sub> O <sub>2</sub> )
21: EO <sub>2</sub> (HOC <sub>2</sub> H <sub>4</sub> O <sub>2</sub> )	22: EO (HOC <sub>2</sub> H <sub>4</sub> O)	23: GC(O)O <sub>2</sub> (HOCH <sub>2</sub> C(O)O <sub>2</sub> )	24: PO <sub>2</sub> (HOC <sub>3</sub> H <sub>6</sub> O <sub>2</sub> )
25: ISOPO <sub>2</sub>	26: MACRO <sub>2</sub>		
<b>Chemical families</b>			
O <sub>x</sub> = O <sub>3</sub> + O(3P) + O(1D)			
ClO <sub>x</sub> = Cl + ClO			
Cl <sub>y</sub> = ClO <sub>x</sub> + OCIO + 2Cl <sub>2</sub> O <sub>2</sub> + HCl + ClONO <sub>2</sub> + HOCl + 2Cl <sub>2</sub> + CINO <sub>2</sub> + BrCl			
NO <sub>x</sub> = NO + NO <sub>2</sub> + NO <sub>3</sub>			
NO <sub>y</sub> = NO <sub>x</sub> + N + HNO <sub>3</sub> + 2N <sub>2</sub> O <sub>5</sub> + HO <sub>2</sub> NO <sub>2</sub> + ClONO <sub>2</sub> + CINO <sub>2</sub> + BrONO <sub>2</sub> + PAN + GPAN + ONIT + ISON + NALD + MPAN			
BrO <sub>x</sub> = Br + BrO + BrCl			
Br <sub>y</sub> = BrO <sub>x</sub> + HBr + HOBr + BrONO <sub>2</sub>			

### 3.8.4.3 Verification

Figure 3.8.7 compares assimilated and observed ozone profiles for 2015. At all heights, the mean differences are within around 0.5 ppmv and the root mean square errors (RMSEs) of simulation against observation are less than 0.8 ppmv.

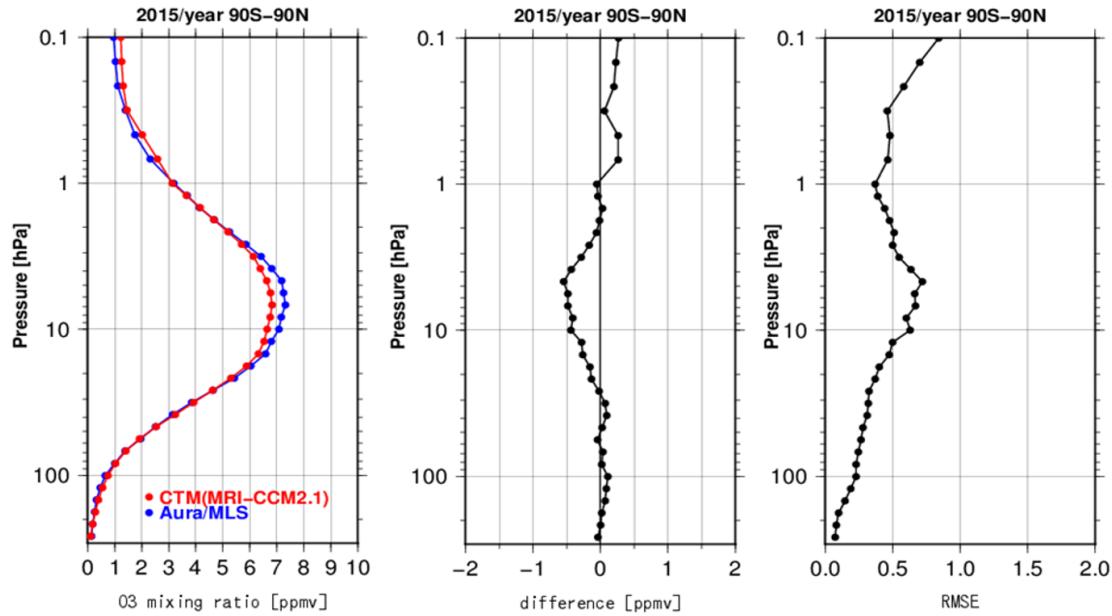


Figure 3.8.7: Annual mean vertical ozone profile in 2015 averaged over the globe. The left panel shows the ozone mixing ratio of the CTM (red) and satellite observation (blue,Aura/MLS). The middle panel shows the difference of the mixing ratios, and the right panel indicates the root mean square difference.

### 3.8.4.4 Radiative Transfer Model for UV Index Forecast

The surface UV dose is calculated under clear-sky conditions by the radiative transfer model (Aoki *et al.* (2002)) in an area from 122°E to 149°E and from 24°N to 46°N with a grid resolution of 0.25° x 0.20°. The look-up table (LUT) method is used to calculate the surface UV dose with reduced computational cost. The basic parameters of the LUT for the clear-sky UV dose are the solar zenith angle and total column ozone predicted by the CTM. The clear-sky UV index is derived from clear-sky UV dose corrected in consideration of climatological aerosol, distance from the sun, altitude and climatological surface albedo. The UV index is derived from correction of the clear-sky UV index with cloud information of the operational weather forecast.

The clear-sky UV index calculated using the LUT is verified against the observed UV index in clear-sky conditions. The results are shown in Figure 3.8.8. The mean error of the calculated clear-sky UV index is 0.1 and the RMSE is 0.4. It is thus verified that modeled UV indices are well simulated to observation values.

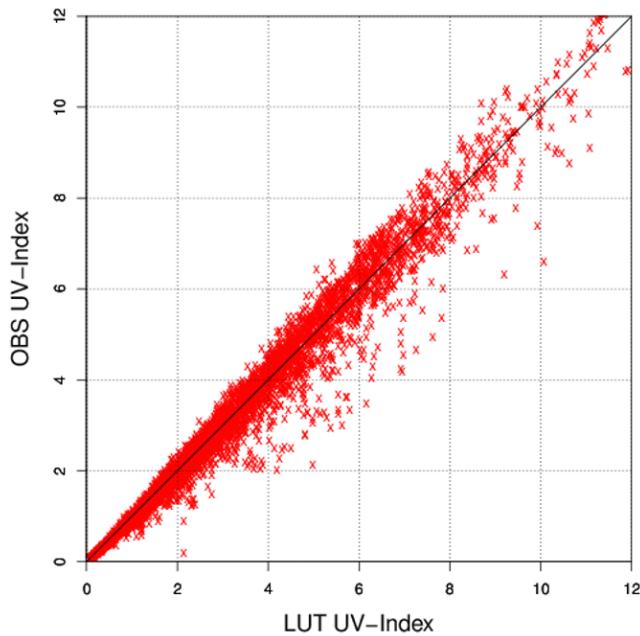


Figure 3.8.8: Relationship between calculated clear-sky UV indices and observed UV indices in clear-sky conditions at three JMA stations from 2015 to 2017. In UV index calculation, climatological aerosol optical depth is applied.

### 3.8.5 Regional Ozone CTM Used for Photochemical Oxidant Information

#### 3.8.5.1 Basic Framework

The regional chemical transport model for photochemical smog bulletins covering the Japan area was developed by Kajino *et al.* (2012). The CTM, known as NHM-Chem, has a horizontal resolution of  $20\text{km} \times 20\text{km}$  in Lambert coordinates and a vertical resolution of 18 layers from the surface to 50 hPa in terrain-following coordinates. The prediction domain is around  $23 - 50^\circ\text{N}$  latitude and  $100 - 145^\circ\text{E}$  longitude. The chemical model, called RAQM2 (Regional Air Quality Model 2), treats 72 chemical species and 214 chemical reactions of SAPRC99 (Carter (2000)), and incorporates major processes for atmospheric trace species such as emissions, advection, turbulent diffusion, sub-grid scale convection, and dry and wet deposition. RAQM2 also implements a fully dynamic aerosol module with a three-moment bulk model using a modal-moment dynamics approach. However, in the operational version of NHM-Chem used for surface photochemical oxidant prediction, aerosol dynamics are not calculated in the interests of computational efficiency.

The lateral and upper boundaries of meteorological fields are given by analysis and forecasts of the global atmospheric model, and the boundaries of the ozone and several related species are nested from the global CTM described in Subsection 3.8.4. The natural and anthropogenic emission inventory dataset listed in Table 3.8.3 are adopted for the regional CTM.

Table 3.8.3: Emission inventories of trace gases used in the regional ozone CTM

	inventory name and reference	emission source	coverage
REAS1.1	(Regional Emission inventory in Asia, version 1.1)	Ohara <i>et al.</i> (2007)	anthropogenic
GFED3	(Global Fire Emission Database, version 3)	Giglio <i>et al.</i> (2010)	natural
MEGAN2	(Model of Emission of Gases and Aerosols from Nature, version 2)	Guenther <i>et al.</i> (2006)	natural

### 3.8.5.2 Relaxation to Observational Data

The regional ozone CTM is operated for a period of 72 hours with an initial time of 12UTC. Running of the model actually starts at 19UTC, and surface ozone concentration data (AEROS: Atmospheric Environmental Regional Observation System in Japan) for 12–18UTC are assimilated using the nudging method like as Eq. (3.8.1). Forcing term for nudging (right side of the equation) is expressed as:

$$\gamma \left( x_{obs}(s) \frac{x_{mdl}(k)}{x_{mdl}(1)} - x_{mdl}(k) \right) \quad (3.8.5)$$

Analysis of ozone concentration in the planetary boundary layer (the k-th model layer) is based on model guess concentration  $x_{mdl}(k)$  and modified by surface observation  $x_{obs}(s)$  with a nudging factor  $\gamma$  set as  $1 \times 10^{-3} [\text{s}^{-1}]$ .

### 3.8.5.3 Verification

Comparison of the simulated surface ozone field for all points with hourly observation data in the Japan area from April to September in 2015 showed that the mean error of surface ozone concentration in the daytime (forecast time: 7 to 23 hours) was 6.6 ppb, the RMSE was 13.7 ppb and the correlation coefficient was 0.69.

## 3.9 Verification

JMA verifies the output of Numerical Weather Prediction (NWP) model forecasts against observation and/or analysis outcomes, and the results are used as reference in research and development regarding NWP models. GSM and Global EPS verification results are exchanged between JMA and other NWP centers via the Lead Centre for Deterministic NWP Verification (LC-DNV) and the Lead Centre for EPS Verification. The standard verification procedures are defined in the Manual on the Global Data-processing and Forecasting System (WMO 2017). This section summarizes operational verification for the GSM and the Global EPS.

The specifications of GSM verification against analysis values are shown in Table 3.9.1, and the scores used for verification are presented in Appendix A.1. GSM forecast performance, including typhoon forecasting, is described in Subsection 3.2.13.

Radiosonde data are used for verification against observation values. The specifications of verification against radiosonde data for the GSM are shown in Table 3.9.2. All radiosonde data passing quality control are used in verification. Stations from which radiosonde data are used in verification are selected on the basis of recommendations from the Commission for Basic Systems.

The specifications of verification regarding Global EPS output for One-week Forecasting are shown in Table 3.9.3. The probabilistic forecast for verification is defined as the ratio of the number of ensemble members in an event to the ensemble size for every grid. The verification results for the Global EPS are described in Subsection 3.3.6.

Table 3.9.1: Operational verification against analysis for the GSM

Verification grid		$1.5^\circ \times 1.5^\circ$ (latitude/longitude)
Elements	Extra-tropics: Tropics:	Geopotential height (Z), temperature (T) and wind at 850, 500 and 250 hPa; Mean sea-level pressure (MSLP) Z, T and wind at 850 and 250 hPa
Scores	Wind: Other elements:	Root mean square vector wind error and mean error of wind speed Mean error, root mean square error, anomaly correlation and S1 score (only for MSLP)
Forecast range (initial time)	Up to 84 hours (00 UTC) and up to 264 hours (12 UTC)	
Forecast steps	Every 12 hours	
Areas	NH extra-tropics ( $90^\circ\text{N} - 20^\circ\text{N}$ ), SH extra-tropics ( $20^\circ\text{S} - 90^\circ\text{S}$ ), Tropics ( $20^\circ\text{N} - 20^\circ\text{S}$ ), North America ( $25^\circ\text{N} - 60^\circ\text{N}, 50^\circ\text{W} - 145^\circ\text{W}$ ), Europe/North Africa ( $25^\circ\text{N} - 70^\circ\text{N}, 10^\circ\text{W} - 28^\circ\text{E}$ ), Asia ( $25^\circ\text{N} - 65^\circ\text{N}, 60^\circ\text{E} - 145^\circ\text{E}$ ), Australia/New Zealand ( $10^\circ\text{S} - 55^\circ\text{S}, 90^\circ\text{E} - 180^\circ\text{E}$ ), Northern polar region ( $90^\circ\text{N} - 60^\circ\text{N}$ ) and Southern polar region ( $90^\circ\text{S} - 60^\circ\text{S}$ )	

Table 3.9.2: Operational verification against radiosonde data for the GSM

Verification grid		Nearest model grid point to the observation location
Elements	Extra-tropics: Tropics:	Z, T and wind at 850, 500 and 250 hPa Z, T and wind at 850 and 250 hPa
Scores	Wind: Other elements:	Root mean square vector wind error and mean error of wind speed Mean error and root mean square error
Forecast range (initial time)	Up to 84 hours (00 UTC) and up to 264 hours (12 UTC)	
Forecast steps	Every 12 hours	
Areas	NH extra-tropics, SH extra-tropics, Tropics, North America, Europe/North Africa, Asia, Australia/New Zealand, Northern polar region and Southern polar region	

Table 3.9.3: Operational verification of the Global EPS for One-week Forecasting

	Deterministic verification	Probabilistic verification	
Analysis	Global analysis on $1.5^\circ \times 1.5^\circ$ grid		
Forecast	Ensemble mean	Probability	
Climatology	Climatological fields and standard deviations are calculated from common climatology provided from LC-DNV. The climatological probability is given by the monthly frequency derived from analysis fields.		
Elements	Z at 500 hPa; T at 850 hPa; u wind component (U) and v wind component (V) at 850 and 250 hPa; MSLP	Anomalies of Z at 500 hPa, T at 2 m, T at 850 hPa and MSLP with thresholds of $\pm 1, \pm 1.5$ and $\pm 2$ climatological standard deviation; Wind speed at 850 hPa with thresholds of 10, 15 and 25 m/s; U and V at 850 and 250 hPa with thresholds of 10th, 25th, 75th and 90th percentile points with respect to the defined climatology; 10 m wind speed with thresholds 10 and 15 m/s	Z at 500 hPa; T at 2 m and 850 hPa; U and V at 850 and 250 hPa; Wind speed at 10 m and 850 hPa; MSLP
Scores	Root mean square error, and anomaly correlation	Reliability table	Continuous ranked probability score
Forecast range (initial time)	Up to 264 hours (00 and 12 UTC)		
Forecast steps	Every 24 hours		
Areas	NH extra-tropics, SH extra-tropics and Tropics		

## 3.10 Global Atmospheric Transport Model for Volcanic Ash

### 3.10.1 Introduction

Since April 1997, JMA has provided information on volcanic ash clouds to airlines, civil aviation authorities and related organizations in its role as the Volcanic Ash Advisory Centre (VAAC) Tokyo. JMA introduced the Global Atmospheric Transport Model (GATM) in December 2013 to create 18-hour predictions for areas where volcanic ash clouds are expected in the Centre's area of responsibility. The forecast is normally updated every six hours (at 00, 06, 12 and 18 UTC) when ash clouds are identified in satellite imagery. If notable changes occur in ash clouds, updates are provided as often as needed.

### 3.10.2 Initial Condition

As the initial condition for the GATM, the boundary of volcanic ash observed by meteorological satellites is adopted. In the model, 40,000 tracer particles are located uniformly in the shape of an inverted pyramidal frustum defined by the observation area and the ash-cloud top height. The specifications of the GATM for VAAC operation are shown in Table 3.10.1.

Table 3.10.1: Specifications of the GATM for VAAC operation

Number of Tracer Particles	40,000
Forecast Time	18 hours
Time Step	10 minutes

For volcanic ash grain size, log-normal distribution is adopted. The probability density function of the diameter  $D$  is given by

$$f(D) = \frac{1}{\sqrt{2\pi\sigma_D^2}} \exp\left[-\frac{\log_2^2(D/D_m)}{2\sigma_D^2}\right] \quad (3.10.1)$$

where  $D_m$  is the mean diameter (set to 0.0316 mm) and  $\sigma_D$  is the standard deviation of distribution (set to 1.0). In addition, the diameter of tracers is limited to between 0.01 mm and 0.1 mm.

### 3.10.3 Model

#### 3.10.3.1 Basic Framework

The GATM adopts a Lagrangian scheme similar to that of the ATDM (see Section 3.7) with gravitational fallout. The location of each tracer after the time step  $\delta t$  (set to 10 minutes) is given by

$$x(t + \delta t) = x(t) + \delta x \quad (3.10.2)$$

$$y(t + \delta t) = y(t) + \delta y \quad (3.10.3)$$

$$z(t + \delta t) = z(t) + \delta z - V_t \delta t \quad (3.10.4)$$

where  $\delta x$ ,  $\delta y$  and  $\delta z$  are given by Eqs. (3.7.3) - (3.7.5). The third term on the right of Eq. (3.10.4) represents gravitational fallout.  $(u_m(t), v_m(t), w_m(t))$  in Eqs. (3.7.1), (3.7.2) and (3.7.5) are the mean wind velocity at  $(x(t), y(t), z(t))$  as derived by interpolating forecast gridded data from the Global Spectral Model (GSM; see Section 3.2) spatially and temporally.

#### 3.10.3.2 Horizontal and Vertical Dispersion

The effect of horizontal dispersion is represented using horizontal wind perturbation  $(u'(t), v'(t))$  from mean wind.  $(u'(t), v'(t))$  are given by Eqs. (3.7.1) and (3.7.2). The parameters are set to  $K_h = 5.864 \times 10^4 \text{ m}^2\text{s}^{-1}$  and

$T_L = 5.0 \times 10^4$  s, and the initial condition of horizontal wind perturbation is  $u'(0) = 0.253G \text{ m s}^{-1}$  following Kawai (2002).

Meanwhile, vertical dispersion is represented as atmospheric vertical turbulence. The diffusion coefficient  $K_v$  is given by Eq. (3.7.6).  $F(Ri)$  in Eq. (3.7.6) represents atmospheric stability as a function of the flux Richardson number  $Ri$  given by the level 2 scheme of Mellor and Yamada (1974, 1982). The mixing length  $l$  is given by Eq. (3.7.7). Von Kármán's constant  $\kappa$  and the maximum mixing length  $l_0$  in Eq. (3.7.7) are set to 0.4 and 30 m, respectively.

In the GATM for VAAC operation, horizontal and vertical dispersion processes are not used in order to avoid excessive diffusion of ash cloud.

### 3.10.3.3 Gravitational Fallout

In the GATM, it is assumed that volcanic ash particles fall at their terminal velocity  $V_t$ , as determined from the equation for balance between gravitational force and air resistance force:

$$\frac{4}{3}\pi\left(\frac{D}{2}\right)^3\rho_p g = \frac{1}{2}C_a\rho_a\pi\left(\frac{D}{2}\right)^2 V_t^2 \quad (3.10.5)$$

where  $\rho_p$  is particle density,  $\rho_a$  is air density and  $C_a$  is the drag coefficient in consideration of the shape parameter  $F = (a_2 + a_3)/a_1$  (where  $a_1$ ,  $a_2$  and  $a_3$  are particle principal axes, with  $a_1$  as the longest) as given by Suzuki (1983):

$$C_a = \frac{24}{Re} \frac{F^{-0.32}}{C_c} + 2\sqrt{1.07 - F} \quad (3.10.6)$$

where  $Re$  is the Reynolds' number represented as  $Re = \rho_a V_t D / \eta_a$  with air viscosity  $\eta_a$ .  $C_c$  is the Cunningham correction factor, which is used to account for the reduction of drag on small particles and is expressed as

$$C_c = 1 + Kn \left[ 1.257 + 0.400 \exp\left(-\frac{1.100}{Kn}\right) \right] \quad (3.10.7)$$

with the Knudsen number  $Kn = 2\lambda_a/D$  based on the mean free path of air  $\lambda_a$ .  $\eta_a$  and  $\lambda_a$  are given as follows:

$$\eta_a(z) = \eta_0 \left[ \frac{T_0 + C_S}{T_a(z) + C_S} \right] \left[ \frac{T_a(z)}{T_0} \right]^{3/2} \quad (3.10.8)$$

$$\lambda_a(z) = \lambda_0 \frac{\eta_a(z)}{\eta_0} \left[ \frac{p_a(z)}{p_0} \right]^{-1} \left[ \frac{T_a(z)}{T_0} \right]^{1/2} \quad (3.10.9)$$

where  $p_a(z)$  is air pressure at height  $z$ ,  $T_a(z)$  is air temperature at  $z$ , and  $C_S$  is the Sutherland constant of air (117 K).  $\eta_0$  ( $18.18 \mu\text{Pa s}$ ) and  $\lambda_0$  ( $0.0662 \mu\text{m}$ ) are the standard values for the reference atmosphere ( $T_0 = 293.15 \text{ K}$  and  $p_0 = 1013.25 \text{ hPa}$ ).

The density of volcanic ash particles  $\rho_p [\text{kg/m}^3]$  is defined as a function of diameter  $D [\text{m}]$ :

$$\rho_p(D) = \frac{0.48 + \rho_{pm}D}{2.0 \times 10^{-4} + D} \quad (3.10.10)$$

where  $\rho_{pm}$  is the density for coarse tephra and is set to  $1 \times 10^3 \text{ kg/m}^3$  as per the density of pumice stone.

### 3.10.3.4 Dry and Wet Deposition

Dry deposition works on tracers within the surface boundary layer, and is simply computed from the depth of the surface boundary layer  $z_r$  and dry deposition velocity  $V(z_r)$  (see Subsection 3.7.2.2). In the GATM,  $z_r$  is set to 100 m and  $V(z_r)$  is fixed as  $0.3 \text{ m s}^{-1}$ .

Wet deposition involves a washout process (below-cloud scavenging) representing the deposition of tracers via rainfall and a rainout process (in-cloud scavenging) representing removal of tracers via their roles as cloud condensation nuclei. In the GATM, only the washout process is considered. The wet deposition rate associated with rain  $\Lambda [\text{h}^{-1}]$  is given by Eq. (3.7.9). In the GATM, tracers below 700 hPa are deposited on the ground in line with the wet deposition rate.

### 3.10.4 Products

VAAC Tokyo issues Volcanic Ash Advisories (VAAs) in text and graphical form as defined in ICAO (2018). VAAs include information on the forecast height and area of ash clouds 6, 12 and 18 hours ahead of observation times based on GATM results. Sample VAAs are shown in Figures 3.10.1 and 3.10.2.

```
FVFE01 RJTD 041926
VA ADVISORY
DTG: 20180404/1926Z
VAAC: TOKYO
VOLCANO: KIRISHIMAYAMA 282090
PSN: N3156 E13052
AREA: JAPAN
SUMMIT ELEV: 1700M
ADVISORY NR: 2018/108
INFO SOURCE: HIMAWARI-8 JMA
AVIATION COLOUR CODE: NIL
ERUPTION DETAILS: EXPLODED AT 20180404/1831Z FL220 EXTD E
OBS VA DTG: 04/1900Z
OBS VA CLD: SFC/FL330 N3155 E13051 - N3200 E13049 - N3200 E13054 -
N3156 E13058 MOV E 30KT
FCST VA CLD +6 HR: 05/0100Z SFC/FL310 N3224 E13215 - N3145 E13222 -
N3056 E13318 - N3058 E13235 - N3221 E13122
FCST VA CLD +12 HR: 05/0700Z SFC/FL290 N3305 E13344 - N3213 E13331 -
N3047 E13526 - N3052 E13437 - N3214 E13246 - N3319 E13159
FCST VA CLD +18 HR: 05/1300Z SFC/FL280 N3407 E13619 - N3227 E13543 -
N3050 E13715 - N3050 E13651 - N3307 E13352 - N3431 E13358
RMK: NIL
NXT ADVISORY: 20180404/2100Z=
```

Figure 3.10.1: Sample VAA in text form

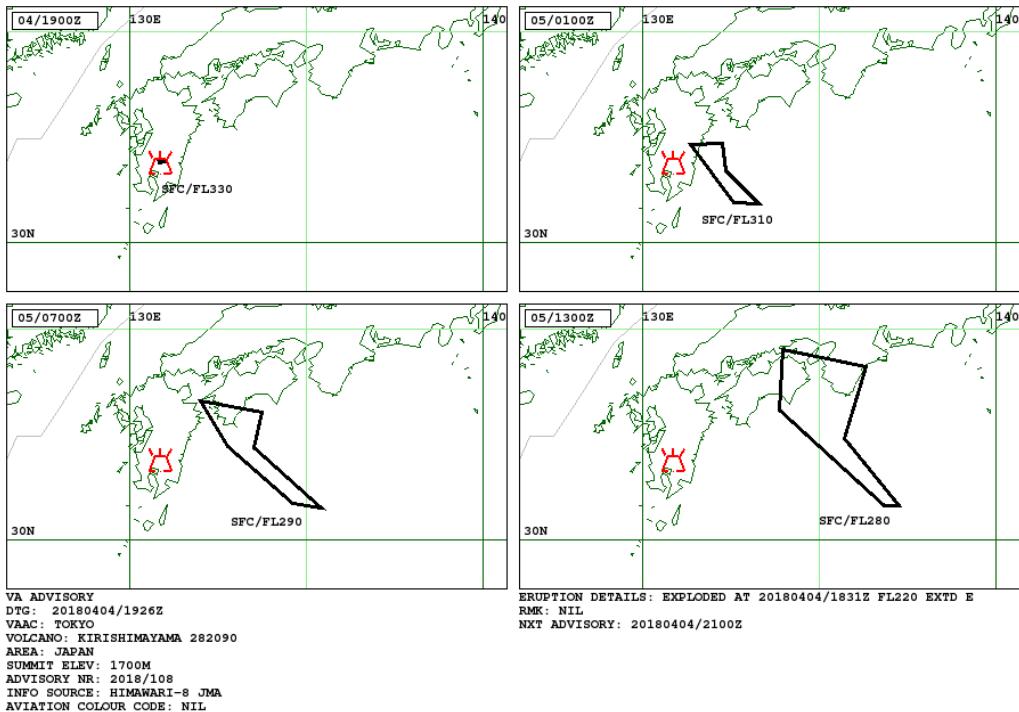


Figure 3.10.2: Sample VAA in graphical form

## 3.11 Regional Atmospheric Transport Model for Volcanic Ash

### 3.11.1 Introduction

JMA launched its Volcanic Ash Fall Forecast (VAFF) product based on the Regional Atmospheric Transport Model (RATM) in March 2008 (Shimbori *et al.* 2009) and updated it in spring 2015 (Hasegawa *et al.* 2015). Three types of forecasts are sequentially provided: VAFFs (Scheduled) are issued periodically based on an assumed eruption for active volcanoes, VAFFs (Preliminary) are brief forecasts issued within 5 - 10 minutes of an actual eruption, and VAFFs (Detailed) are more accurate forecasts issued within 20 - 30 minutes of an actual eruption. The updated VAFFs provide information on expected volcanic ash/lapilli fall areas and/or amounts based on the RATM with Local Forecast Model (LFM; see Section 3.6) or Meso-Scale Model (MSM; see Section 3.5) outputs.

### 3.11.2 Initial Condition

As the initial condition of the RATM, a volcanic plume in the shape of an inverted cone is adopted. The initial plume is based on information from observational reports, including eruption time and plume height, and on the duration of volcanic ash emission. As with the GATM (see Subsection 3.10.2), tracer particle size follows log-normal distribution. The parameters of the probability density function in Eq. (3.10.1) are set to  $D_m = 0.25$  mm and  $\sigma_D = 1.0$ , and the diameter of tracers is limited to between  $0.65 \mu\text{m}$  and 96 mm. The number of RATM tracers is set to be higher than that of the GATM in consideration of the wide range of diameters (see Tables 3.10.1 and 3.11.1).

Table 3.11.1: RATM specifications for VAFFs

	VAFF (Scheduled)	VAFF (Preliminary)	VAFF (Detailed)
Number of Tracer Particles	100,000	100,000	250,000
Forecast Time	18 hours	1 hour	6 hours
Time Step	3 minutes	1 minute	3 minutes

The vertical distribution of volcanic particles in the eruption plume is calculated according to Suzuki (1983). The probability that a tracer with diameter  $D$  is released from a height of  $z$  above vent level is given as follows:

$$P(D, z) = AY(D, z)e^{-Y(D, z)} \quad (3.11.1)$$

where  $A$  is the normalization constant and  $Y(D, z)$  is expressed as

$$Y(D, z) = \beta \frac{W(z) - V_t(D, 0)}{V_t(D, 0)} \quad (3.11.2)$$

with release constant  $\beta$  (set to 0.017).  $V_t(D, 0)$  is the terminal velocity at the height of the volcano summit as calculated using Eq. (3.10.5).  $W(z)$  is the vertical velocity of the eruption column at height  $z$ , which is given by

$$W(z) = W_0 \left(1 - \frac{z}{H}\right) \quad (3.11.3)$$

where  $H$  [km] is the column height and  $W_0$  [ $\text{m s}^{-1}$ ] =  $\sqrt{H/2.2 \times 10^{-4}}$  is the initial velocity. The total amount of volcanic ejecta  $M$  is given in line with Morton *et al.* (1956) as

$$M = K_m H^4 T \quad (3.11.4)$$

where  $T$  is the duration of eruption and  $K_m$  is a constant set to  $6.95 \times 10^5 \text{ kg km}^{-4} \text{ h}^{-1}$ .

### 3.11.3 Model

The basic framework of the RATM is similar to that of the GATM (see Subsection 3.10.3) with the following differences:

- MSM or LFM forecasts are used as meteorological fields instead of data from the GSM forecast. The related high-resolution gridded data are beneficial for predicting volcanic ash fall.
- Since the RATM also predicts volcanic lapilli with high fall velocity, the time step  $\delta t$  of the RATM is set to be shorter than that of the GATM as shown in Tables 3.10.1 and 3.11.1.
- Horizontal and vertical dispersion processes are considered. The maximum mixing length appearing in Eq. (3.7.7) is set to 100 m in the atmospheric boundary layer ( $z' \leq 1 \text{ km a.g.l.}$ ) where vertical dispersion is large, while the value given by Holtslag and Boville (1993) is used for the higher free atmosphere:

$$l_0 = 30 + 70 \exp\left(1 - \frac{z'}{1000}\right) \quad (3.11.5)$$

- Wet deposition (washout) associated with snow and graupel  $\Lambda_s$  [ $\text{h}^{-1}$ ] is considered in addition to that associated with rain, and is given as follows:

$$\Lambda_s \approx 0.1 P^{0.3} \quad (3.11.6)$$

### 3.11.4 Products

The amount of volcanic ash fall and maximum particle size at grid points with spacing of 0.02 degrees are calculated based on tracers falling to the surface, and are used for VAFFs. The amount of ash fall is shown in three qualitative categories combined with recommended action to be taken by the general public (Hasegawa *et al.* 2015). VAFFs include graphical information on ash quantities and lapilli pieces with sizes of 1 cm or more as well as text information on the direction of ash drift, municipalities affected and precautions for disaster prevention. Sample VAFFs are shown in Figures 3.11.1 (Scheduled), 3.11.2 (Preliminary) and 3.11.3 (Detailed).

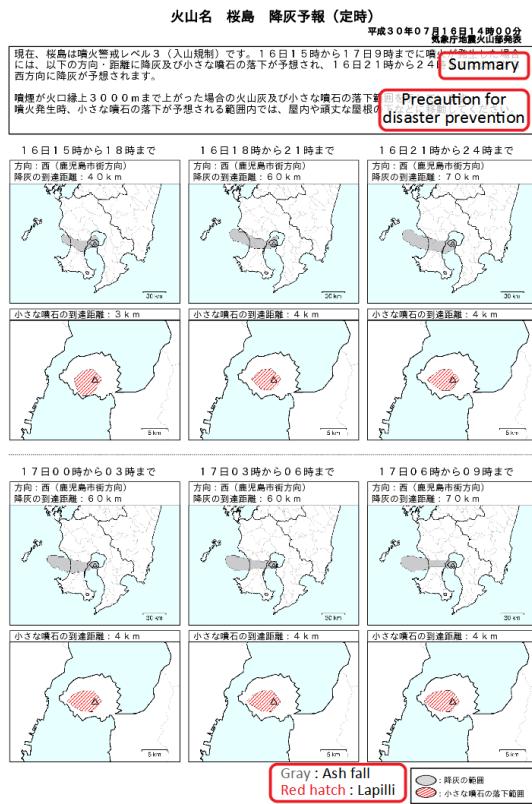


Figure 3.11.1: Sample VAFF (Scheduled)

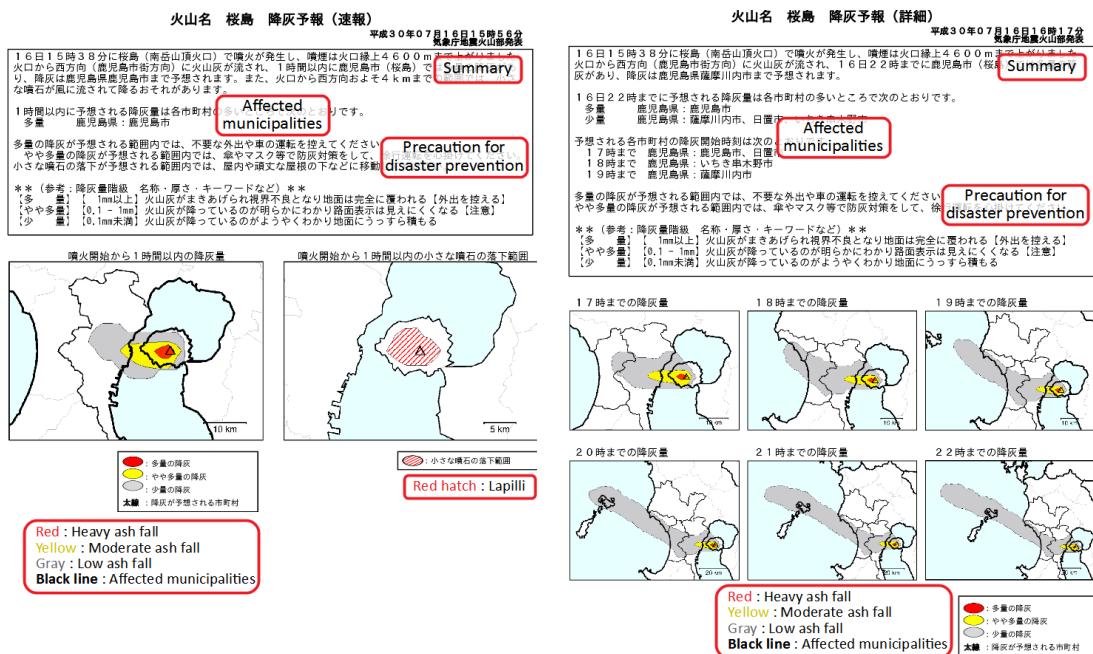


Figure 3.11.2: Sample VAFF (Preliminary)

Figure 3.11.3: Sample VAFF (Detailed)