

OUTLINE

OF

THE OPERATIONAL NUMERICAL WEATHER PREDICTION

AT

THE JAPAN METEOROLOGICAL AGENCY

Appendix to WMO Technical Progress Report on the Global Data-processing and
Forecasting System (GDPFS) and Numerical Weather Prediction (NWP) Research

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Preface

The Japan Meteorological Agency (JMA) started the operation of numerical weather prediction (NWP) in June 1959 after several years of extensive research activities of the Numerical Weather Prediction Group of Japan. That was the third NWP operation in the world, following the Swedish Meteorological and Hydrological Institute (SMHI) in September 1954 and the US Weather Bureau in May 1955. The development of NWP since then has been tremendous. Operational NWP centers, including JMA, have benefited from ever better understanding of meteorological phenomena, improved modeling techniques, increasing computing power, efficient telecommunication systems, and improved observing systems, especially meteorological and earth-observing satellite systems.

This report is published to present technical details of the operational NWP systems of JMA as of June 2018, as an appendix to “WMO Technical Progress Report on the Global Data-processing and Forecasting System (GDPFS) and Numerical Weather Prediction (NWP) Research”.

The first chapter provides an overview of the configurations and specifications of the current computer system at JMA. Thereafter follows a description of the operational suite and the operational job management system on the current computer system. The second chapter deals with three major data assimilation systems for atmospheric fields, namely, Global Analysis, Meso-scale Analysis and Local Analysis including the observation data used in these analyses. A description on the JMA Climate Data Assimilation System as well as more aspects on data assimilation is given in this chapter.

The third chapter describes a suite of NWP models for very short-range prediction of meso-scale disturbances, and for short- and medium-range prediction of synoptic-scale disturbances. The global model is used in ensemble prediction systems for the typhoon forecast, one-week and one-month prediction, and the coupled ocean-atmosphere model is employed for the seasonal forecast and El Niño forecast. An atmospheric transport models are applied to the prediction of transport of trace elements such as radioactive materials and Kosa (Aeolian Dust) for environmental information. The fourth chapter explains various kinds of application products of NWP such as weather charts, gridded data products, very-short-range forecasting of precipitation, hourly analysis in wind and temperature, guidance for short-range forecasting, products for aviation services, products of ensemble prediction systems and atmospheric angular momentum functions. The last chapter is on ocean models, specifically ocean wave models, storm surge models, a sea ice model and an oil spill prediction model as well as sea surface temperature analysis systems and ocean data assimilation systems.

JMA is working forward a further developments to improve the accuracy of NWP models . The reader will find updated information on the NWP systems of JMA on the website of JMA <<https://www.jma.go.jp/jma/en/Activities/nwp.html>> and in the WMO Technical Progress Report on GDPFS and NWP Research that is issued every year.

MUROI Chiashi
Director
Numerical Prediction Division
Japan Meteorological Agency

Chapter 1

Computer System

1.1 Introduction

The Japan Meteorological Agency (JMA) installed its first-generation computer (IBM 704) to run an operational numerical weather prediction model in March 1959. Since then, the computer system at JMA has been repeatedly upgraded, and the current system (Cray XC50) was completed in June 2018 as the tenth-generation computer. Figure 1.1.1 shows the history of computers at JMA, their peak performance, and a change in peak performance calculated using Moore's law¹ from the first computer (IBM 704). The peak performance of the second (HITAC 5020), the third (HITAC 8800), and the eighth (HITACHI SR11000) computers at the beginning of their implementation was almost the same as that projected using Moore's law, while it was lower during the 1980s, 1990s, and the first half of the 2000s. Recent growth is faster and the peak performance of the current computer is higher than the projection.

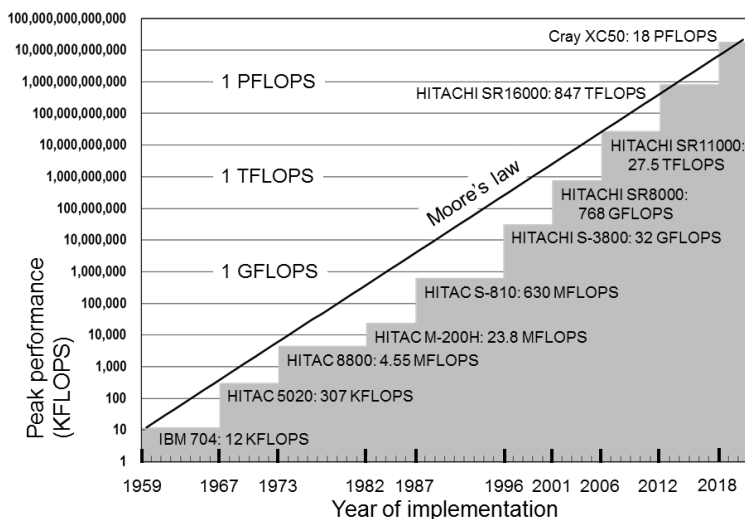


Figure 1.1.1: History of computers used at JMA and their peak performance. The line “Moore’s law” represents the projection of peak performance using Moore’s law from the first computer (IBM 704).

¹The term “Moore’s law” has many formulations. Here we refer to exponential growth of peak performance which doubles every 18 months.

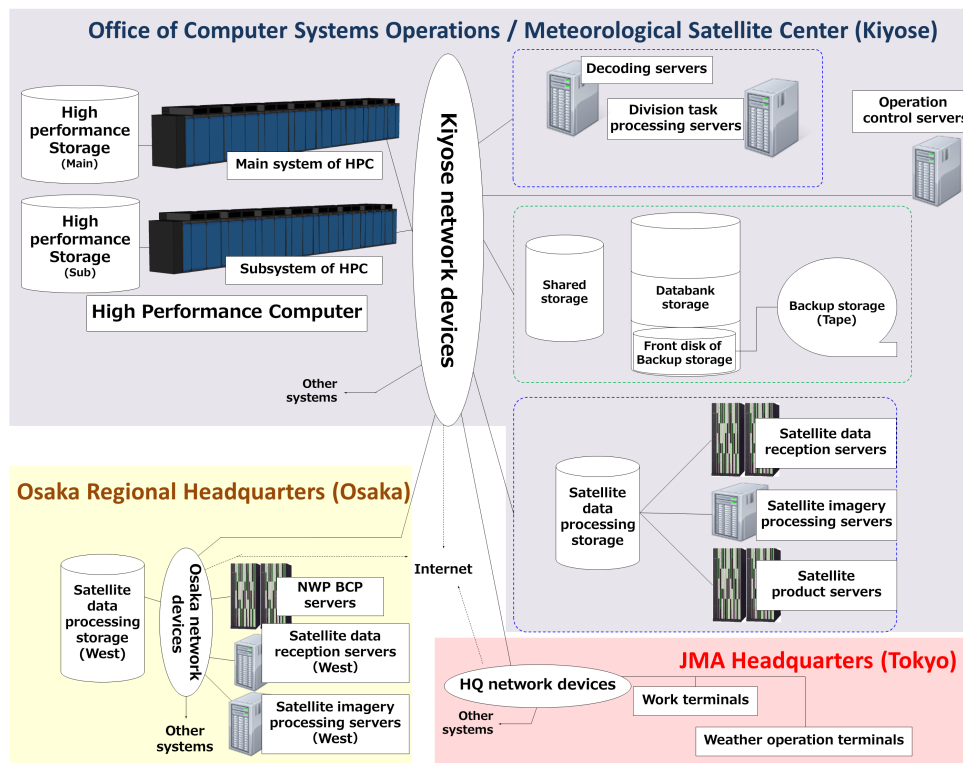


Figure 1.2.1: Schematic illustration of computer system

In this chapter, Section 1.2 briefly describes the configurations and specifications of the current computer system at JMA. Section 1.3 outlines the operational suite and the operational job management system on the current computer system.

1.2 System Configurations and Specifications

1.2.1 Overview

Figure 1.2.1 illustrates major components of the computer system at JMA including Cray XC50 high performance computers, server computers, storages, terminals, and networks. The system has been in operation since 5 June 2018. Most of the computing facilities are located at the Office of Computer Systems Operations and the Meteorological Satellite Center in Kiyose 24 km west of JMA's central-Tokyo HQ, and some servers are located at the Osaka Regional Headquarters for business continuity planning (BCP). A wide area network (WAN) links the Kiyose, HQ and Osaka sites. The specifications of the high-performance computers and server computers are summarized in Table 1.2.1, Table 1.2.2 and Table 1.2.3, respectively.

1.2.2 High Performance Computer

Two independent systems called a main system and a subsystem of a Cray XC50 high performance computer with the same specifications are installed at the Kiyose site. The main system usually runs operational numerical weather prediction jobs, while the subsystem usually runs development jobs. However, in case the main system is under maintenance or out of order, the subsystem runs operational jobs to make the system stable for operational use.

Table 1.2.1: Specifications of high performance computers

Computer	Cray XC50
Number of systems	2
Computational nodes	
Processor, clock frequency	Intel Xeon Platinum 8160, 2.1 GHz
Cores per processor	24
Cores per logical node	48
Logical nodes per system	2741(ESM), 75(MAMU), 8(Tier2), 16(spare)
Peak performance per logical node	3.2256 TFLOPS
Peak performance per system	9,083 TFLOPS
Memory per logical node	96 GiB
Memory per system	264 TiB
Operating system	Cray Linux Environment 6.0(ESM), SUSE 12.2(MAMU, Tier2)
I/O nodes	
Processor, clock frequency	Intel Xeon E5-2699v4, 2.2 GHz
Cores per processor	22
Cores per logical node	22
Logical nodes per system	2(SDB), 4(PBS-MOM), 2(boot), 2(router), 7(network), 15(LNET), 6(data sync), 2(login gateway)
Memory per logical node	128 GiB
Operating system	SUSE 12.2
Login servers	Dell PowerEdge Server
Processor, clock frequency	Intel Xeon Gold 6148, 2.4 GHz
Cores per processor	20
Cores per logical node	40
Number of servers per system	4
Memory per logical node	768 GiB
Operating system	SUSE 12.2

The Cray XC50 consists of computational nodes, I/O nodes, and login servers.

Each of its computational nodes has two sockets for Intel Xeon Platinum 8160 processors with a clock frequency of 2.1 GHz. One socket of the Xeon processor houses a multi-core chip with 24 separate cores, making $2 \times 24 = 48$ cores in each logical node. The theoretical performance per logical node is 3.2256 TFLOPS, and the total memory capacity is 96 GiB per logical node². The computational nodes are ESM³(2741), MAMU⁴(75), Tier2⁵(8), and spare(16) types. The theoretical performance per system is 9,083 TFLOPS for only ESM and MAMU nodes. Each ESM node runs the CLE(Cray Linux Environment) 6.0 operating system, and each MAMU and Tier2 node runs SUSE Linux Enterprise Server 12.2 independently. The inter-node communication rate between each node and the hub processor is 14 GB/s for one-way communication.

The I/O nodes consist of an Intel Xeon E5-2699v4 (2.2 GHz) processor. The types and numbers of these nodes are SDB⁶(2), PBS-MOM⁷(4), boot⁸(2), router⁹(2), network¹⁰(7), LNET¹¹(15), data sync¹²(6), and login gateway¹³(2).

The login system involves four Dell PowerEdge servers with two Intel Xeon Gold 6148 (2.4 GHz) processors. The operating system for both I/O nodes and login servers is SUSE Linux Enterprise Server 12.2.

²The International Electrotechnical Commission approved names and symbols for the power of $2^{10} = 1,024$ instead of 1,000 for prefixes of units. Symbols such as GiB or TiB refer to the former. In contrast, symbols such as GB or TB mean the latter.

³Extreme Scalability Mode nodes. Used for high performance Massively Parallel Processing(MPP) runs.

⁴Multiple Applications Multiple User nodes. Used for smaller applications.

⁵Distribution of computational environments to ESM and MAMU.

⁶Service DataBase node with PBS installation.

⁷PBS Mom daemon applied for ESM node.

⁸Used for boot step.

⁹Used for connection with surveillance network.

¹⁰Used for connection with storage network and servers.

¹¹Lustre NETWORK node. Used for connection with Lustre high performance storage.

¹²Used for connection with main system and subsystem.

¹³Used for connection with login servers.

Table 1.2.2: Specifications of server computers at Kiyose

	Satellite data reception servers	Satellite imagery processing servers	Satellite product servers
Computer	HPE ProLiant DL360 Gen9	HPE ProLiant DL580 Gen9	HPE ProLiant DL380 Gen9
Processor, clock frequency	Intel Xeon E5-2620v3, 2.4 GHz	Intel Xeon E7-8880v3, 2.3 GHz	Intel Xeon E5-2670v3, 2.3 GHz
Cores per processor	6	18	12
Cores per server	12	72	24
Number of servers	5	8	10
Memory per server	64 GiB	256 GiB	192 GiB
Operating system	RHEL 7.3	RHEL 7.3	RHEL 7.3

	Operation control servers	Division task processing servers	Decoding servers
Computer	HITACHI HA8000 RS210AN1	HPE ProLiant DL580 Gen9	HPE ProLiant DL580 Gen9
Processor, clock frequency	Intel Xeon E5-2640v3, 2.6 GHz	Intel Xeon E7-8880v3, 2.3 GHz	Intel Xeon E7-8860v3, 2.2 GHz
Cores per processor	8	18	16
Cores per server	16	72	64
Number of servers	8	12	2
Memory per server	32 GiB	128 GiB	256 GiB
Operating system	RHEL 7.3	RHEL 7.3	RHEL 7.3

The main system and subsystem have high-performance storage configured with a Lustre file system, and have capacities of 1.6x3 PB each. Every time an operational job running on the main system is completed, its output files are copied to the high-performance storage on the subsystem to ensure that the subsystem is ready to run with further operational jobs if operation is switched to it.

1.2.3 Server and Terminal Computers at Kiyose

A number of server computers are used for various tasks, such as processing and decoding of observational data, weather chart analysis and operational suite management.

The satellite data reception servers, satellite imagery processing servers and satellite product servers are used for automatic processing of various kinds of satellite observation data. The five satellite data reception server are HPE ProLiant DL360 Gen9 units with two Intel Xeon E5-2620v3 (2.4 GHz) processors. The eight satellite imagery processing servers are HPE ProLiant DL580 Gen9 units with four Intel Xeon E7-8880v3 (2.3 GHz) processors. The ten satellite product servers are HPE ProLiant DL380 Gen9 units with two Intel Xeon E5-2670v3 (2.3 GHz) processors.

The eight operation control servers used for control of operational suite job groups are HITACHI HA8000 units with two Intel Xeon E5-2640v3 (2.6 GHz) processors.

The division task processing servers are used for weather chart analysis and small operational jobs that are transaction-intensive rather than compute-intensive. The 12 servers of this type are HPE ProLiant DL580 Gen9 units with four Intel Xeon E7-8880v3 (2.3 GHz) processors.

The two servers used for decoding observational data jobs are HPE ProLiant DL580 Gen9 units with four Intel Xeon E7-8860v3 (2.2 GHz) processors.

Other server computers are also used to that manage the operational suite for numerical weather prediction, satellite data processing and other jobs. Server and terminal computers are additionally used to monitor and manage the computer system.

1.2.4 Mass Storage System

Shared, data bank and backup storage systems are used to share data between high-performance computers and server computers.

The shared and databank storage systems are used for jobs running on high-performance computers or server computers. Configuration involves an IBM Spectrum Scale (ISS) file system with RAID 6 magnetic

disks¹⁴ Shared storage comprises three units with a total capacity of 6PB, and databank storage comprises three units with a total capacity of 25PB, one of which is used as a front disk for backup storage.

The backup storage system is used for long-term archiving. It automatically makes backup copies from the front disk of the data bank storage system, and consists of a tape library and four management servers. Its total capacity is about 80 PB.¹⁵

1.2.5 Networks

The Kiyose network connects the high-performance computers, server computers and other network/server elements in the computer system described above.

The storage network connects the high performance computers, server computers, shared storage system, databank storage system, and backup storage system.

Users at HQ remotely log in to computers at the Kiyose site through a WAN consisting of three independent links with transfer speeds of 100 Mbps, 100 Mbps and 1 Gbps (best effort), respectively. The two 100 Mbps links are used for operational jobs, while the 1 Gbps link is used for development jobs. All network equipment is redundantly configured to prevent single failures from causing catastrophic interruption.

The Osaka site is also connected to the Kiyose site through a WAN with two 100 Mbps links.

1.2.6 Server and Terminal Computers at Osaka

Equipment is located in Osaka for NWP BCP operations and redundancy processing of satellite data. There are two HPC ProLiant DL360 Gen9 servers with two Intel Xeon E5-2680v3 (2.5 GHz) processors, which are used for NWP BCP operations.¹⁶ The satellite data reception servers(West) and satellite imagery processing servers(West) are used for processing of satellite observations data in Osaka. The two satellite data reception servers(West) are HPE ProLiant DL360 Gen9 units with two Intel Xeon E5-2620v3 (2.4 GHz) processors. The four satellite imagery processing servers(West) are HPE ProLiant DL360 Gen9 units with two Intel Xeon E5-2698v3 (2.3 GHz) processors.

Table 1.2.3: Specifications of Osaka server computers

	NWP BCP servers	Satellite data reception servers(West)	Satellite imagery processing servers(West)
Computer	HPE ProLiant DL360 Gen9	HPE ProLiant DL360 Gen9	HPE ProLiant DL360 Gen9
Processor, clock frequency	Intel Xeon E5-2680v3, 2.5 GHz	Intel Xeon E5-2620v3, 2.4 GHz	Intel Xeon E5-2698v3, 2.3 GHz
Cores per processor	12	6	16
Cores per server	24	12	32
Number of servers	2	2	4
Memory per server	256 GiB	64 GiB	128 GiB
Operating system	RHEL 7.3	RHEL 7.3	RHEL 7.3

1.3 Operational Aspects

1.3.1 Operational Suite

The JMA operational suite described in later chapters consists of about 80 job groups, including global analysis and global forecasting, with a total of around 20,600 jobs per day. All jobs are submitted from the Routine Operation and Scheduling Environment (ROSE). There are approximately 4,000 and 17,500 constant and variable datasets, respectively.

¹⁴RAID stands for redundant array of independent disks or redundant array of inexpensive disks. In particular, RAID 6 utilizes block-level striping with double distributed parity and provides fault tolerance for two drive failures.

¹⁵The total capacity depends on the volume of the tape cartridge. A capacity of 80 PB is estimated with a 10-TB tape cartridge.

¹⁶Current NWP BCP operations involve online acquisition of gridded data from overseas NWP center sources and processing to create JMA's product format.

1.3.2 ROSE:Job Scheduler

ROSE is a job flow control computer program that automatically controls execution of all operational jobs. Following on from the start of its development in 2008, it has been used by JMA since 2009 in numerical prediction model development environments and other areas. Based on the current system, the program was adopted for operational NWP control.

ROSE is installed on operational control servers, and is used to control all operational jobs based on submission to PBS¹⁷.

1.3.3 RENS:Operational Job Management System

There are complicated dependencies between jobs in a job group and between input and output datasets. To manage a vast number of operational jobs and datasets systematically and assure that jobs run correctly without human error, JMA developed the comprehensive RENS¹⁸ resource using database management systems (DBMSs).

All job information, input/output datasets, and executables are registered in RENS. Dependencies between these elements can be checked using utility programs.

RENS is comprised of four file types, two DBMSs, and several utility programs to allow registration of information, checking of consistency and other tasks as detailed below.

- Files

Registration form: Information about job groups, jobs, datasets, executables, and so on. A registration form is submitted when jobs are added or deleted, datasets or executables are updated, or the configurations of job groups or jobs are modified.

Job definition file: Information about a job group and jobs within the job group such as the job group name, the job name, the schedule (time to run), the order of job groups and jobs (preceding job groups and jobs), and computational resources required (the PBS job class, the number of nodes, the computational time).

Job control language: Information about executables such as a shell script, a ruby script, an awk script and a load module, and input and output datasets used in each job. A job control language file is converted into a shell script using a utility program to be submitted to PBS.

Program build file-format: Information about source files, object modules, libraries, options for compilation, and so on. A program build file-format is converted into a makefile using a utility program to compile load modules.

- DBMSs

DBMS for registration: Information from the above four files is registered using utility programs.

DBMS for job management: Information from the DBMS for registration is stored and this information is used by job schedulers.

When a job control language is converted into a shell script, the following procedures are made:

- Existence test: A shell script tests the existence of all non-optional input datasets at the beginning in order to avoid wasting time if the preceding job failed.
- Quasi-atomic output: Every step of a job calling an executable creates output files with temporary names at first and renames them to final names when the step successfully terminates.

The development of the RENS was started in 2004 on the seventh computer system and installed in the operational system in 2006 when the eighth computer system was implemented. The number of man-made errors after the inclusion of this management system was reduced to about one sixth of that before the adoption.

¹⁷Portable Batch System (the computer program used to perform job scheduling)

¹⁸RENS : Routine Environment for Numerical weather prediction System.

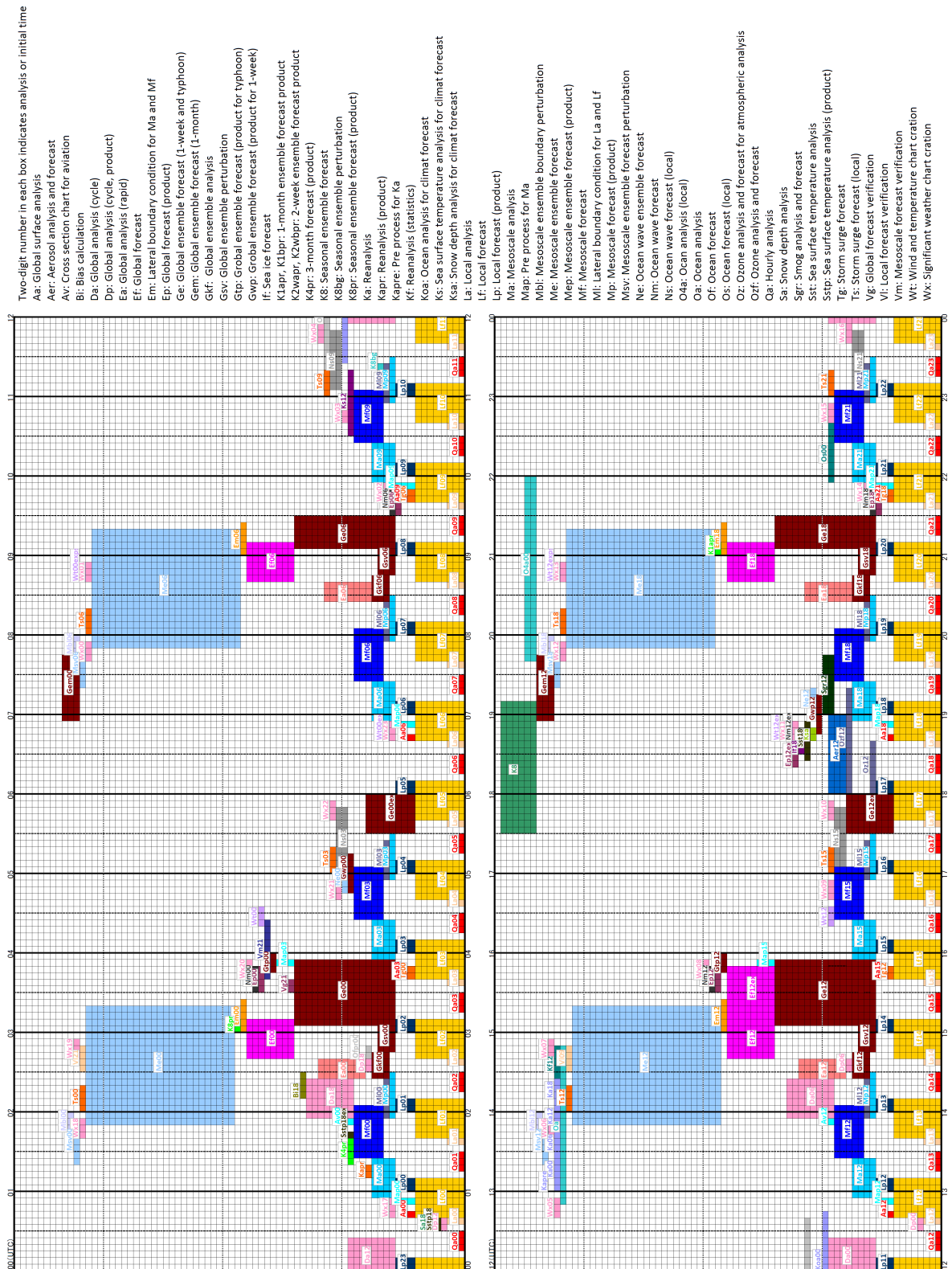


Figure 1.3.1: Daily schedule of the operational suite running on the main system of the high-performance computer as of August 2018. The height and width of each box indicate the approximate number of nodes and the time range, respectively.

Chapter 2

Data Assimilation Systems

2.1 Summary

Three kinds of major data assimilation systems for the analysis for atmospheric fields are operated at JMA: Global Analysis (GA), Meso-scale Analysis (MA) and Local Analysis (LA). Specifications of the JMA data assimilation systems are summarized in Table 2.1.1, Table 2.1.2 and Table 2.1.3. All the analyses are performed by using the procedures shown in Figure 2.1.1.

The following is a brief description of the major components of the analysis systems.

1. Observational data are received from the WMO Information System (WIS) including Global Telecommunication System (GTS), Internet and dedicated networks. The data are decoded according to their code forms. If typhoons exist in the western North Pacific, typhoon bogus profiles are created.
2. Various pre-analysis procedures, such as quality control, data selection and bias correction, are applied to the decoded observational data. In the pre-analysis process, first guess fields retrieved from forecast models are used as a reference of the present atmospheric conditions.
3. The four-dimensional variational method is adopted in Global Analysis and Meso-scale Analysis. To reduce the computational cost, the analysis increment is calculated using a coarser-resolution inner model in the four-dimensional variational method. The resolution of these analysis type is the same as that of the corresponding forecast models.
4. Local Analysis involves a three-hour cycle based on the three-dimensional variational method. Its resolution is coarser than that of the corresponding forecast model.

The atmospheric fields analyzed from the data assimilation systems are used as initial conditions of forecast models. First guess fields and boundary conditions of data assimilation systems are provided from forecast models as shown in Figure 2.1.2.

Sea surface temperature fields (see Section 2.7) and snow depth fields (see Section 2.8) are also analyzed every day.

The Japanese 55-year Reanalysis project has been conducted for the period from 1958 and continues today on a near-real-time basis (see Section 2.10).

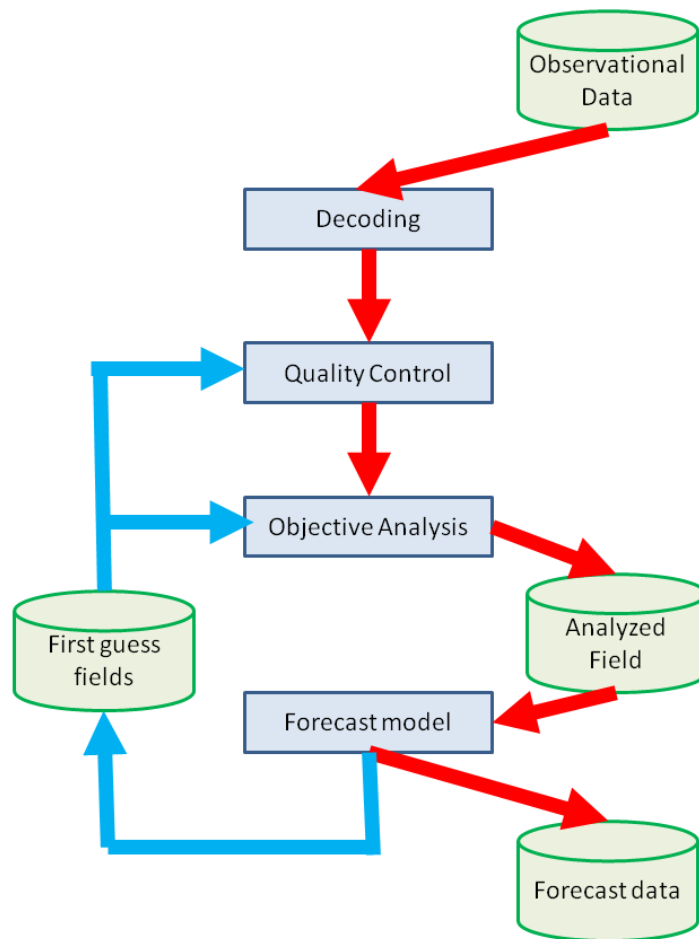


Figure 2.1.1: Major functional components and data flow in the JMA data assimilation system

Table 2.1.1: Specifications of the Global Analysis (GA)

Analysis time	00, 06, 12 and 18 UTC
Analysis scheme	Incremental 4D-Var
Data cut-off time	2 hours and 20 minutes for early run analysis at 00, 06, 12 and 18 UTC 11 hours and 50 minutes for cycle run analysis at 00 and 12 UTC 7 hours and 50 minutes for cycle run analysis at 06 and 18 UTC
First guess	6-hour forecast by the GSM
Domain configuration (Outer step)	Globe TL959,(Reduced Gaussian grid, roughly equivalent to 0.1875 ° (20 km) [1920 (tropic) – 60 (polar)] × 960
(Inner step)	TL319, Reduced Gaussian grid, roughly equivalent to 0.5625 ° (55 km) [640 (tropic) – 60 (polar)] × 320
Vertical coordinate	σ - p hybrid
Vertical levels	100 forecast model levels up to 0.01 hPa + surface
Analysis variables	Wind, surface pressure, specific humidity and temperature
Observation (as of 31 December 2017)	SYNOP, METAR, SHIP, BUOY, TEMP, PILOT, Wind Profiler, AIREP, AMDAR; atmospheric motion vectors (AMVs) from Himawari-8, GOES-[13, 15] and Meteosat-[8, 10]; MODIS polar AMVs from Terra and Aqua satellites; AVHRR polar AMVs from NOAA and Metop satellites; LEO-GEO AMVs; ocean surface wind from Metop-[A, B]/ASCAT; radiances from NOAA-[15, 18, 19]/ATOVS, Metop-[A, B]/ATOVS, Aqua/AMSU-A, DMSP-F[17, 18]/SSMIS, Suomi-NPP/ATMS, GCOM-W/AMSR2, GPM-core/GMI, Megha-Tropiques/SAPHIR, Aqua/AIRS, Metop-[A, B]/IASI and Suomi-NPP/CrIS; clear sky radiances from the water vapor channels (WV-CSRs) of Himawari-8, GOES-[13, 15] and Meteosat-[8, 10]; GNSS RO bending angle data from Metop-[A, B]/GRAS, COSMIC/IGOR, GRACE-[A, B]/Blackjack and TerraSAR-X/IGOR; zenith total delay data from ground-based GNSS
Assimilation window	6 hours

Table 2.1.2: Specifications of the Mesoscale Analysis (MA)

Analysis time	00, 03, 06, 09, 12, 15, 18 and 21 UTC
Analysis scheme	Incremental 4D-Var using a nonlinear forward model in the inner step with low resolution
Data cut-off time	50 minutes for analysis at 00, 03, 06, 09, 12, 15, 18 and 21 UTC
First guess	3-hour forecast produced by the JMA-NHM
Domain configuration (Outer step)	Japan and its surrounding area Lambert projection: 5 km at 60°N and 30°N, 817 × 661 Grid point (1, 1) is at the northwest corner of the domain. Grid point (565, 445) is at 140°E, 30°N
(Inner step)	Lambert projection: 15 km at 60°N and 30°N, 273 × 221 Grid point (1, 1) is at the northwest corner of the domain. Grid point (189, 149) is at 140°E, 30°N
Vertical coordinate	z - z^* hybrid
Vertical levels	(Outer step) 48 levels up to 22 km (Inner step) 38 levels up to 22 km
Analysis variables	Wind, potential temperature, surface pressure and pseudo-relative humidity
Observations (as of 31 December 2017)	SYNOP, SHIP, BUOY, TEMP, PILOT, Wind Profiler, Weather Doppler radar (radial velocity, reflectivity), AIREP, AMDAR; AMVs from Himawari-8; ocean surface wind from Metop-[A, B]/ASCAT; radiances from NOAA-[15, 18, 19]/ATOVS, Metop-[A, B]/ATOVS, Aqua/AMSU-A, DMSP-F[17, 18]/SSMIS, GCOM-W/AMSR2 and GPM-core/GMI; WV-CSRs of Himawari-8; Radar/Raingauge-Analyzed Precipitation; precipitation retrievals from DMSP-F[17, 18]/SSMIS, GCOM-W/AMSR2 and GPM-core/GMI; GPM-core/DPR; GNSS RO refractivity data from Metop-[A, B]/GRAS, COSMIC/IGOR, GRACE-[A, B]/Blackjack, TerraSAR-X/IGOR and TanDEM-X/IGOR; Total Precipitable Water Vapor from ground-based GNSS
Assimilation window	3 hours

Table 2.1.3: Specifications of the Local Analysis (LA)

Analysis time	00, 01, 02, 03, 04, 05, 06, 07, 08, 09, 10, 11, 12, 13, 14, 15, 16, 17, 18, 19, 20, 21, 22 and 23 UTC
Analysis scheme	The three-hour analysis cycle repeats hourly assimilation with 3D-Var and one-hour forecasts
Data cut-off time	30 minutes for analysis at 00, 01, 02, 03, 04, 05, 06, 07, 08, 09, 10, 11, 12, 13, 14, 15, 16, 17, 18, 19, 20, 21, 22 and 23 UTC
First guess	Initial fields produced by the latest MSM
Domain configuration	Japan and its surrounding area Lambert projection: 5 km at 60°N and 30°N, 633 × 521 Grid point (1, 1) is at the northwest corner of the domain. Grid point (449, 361) is at 140°E, 30°N
Vertical coordinate	z - z^* hybrid
Vertical levels	48 levels up to 22 km
Analysis variables	Wind, potential temperature, surface pressure, pseudo-relative humidity, skin temperature, ground temperature and soil moisture
Observations (as of 31 December 2017)	SYNOP, SHIP, BUOY, AMeDAS, TEMP, PILOT, Wind Profiler, Weather Doppler radar (radial velocity, reflectivity), AIREP, AMDAR; AMVs from Himawari-8; radiances from NOAA-[15, 18, 19]/ATOVS, Metop-[A, B]/ATOVS, Aqua/AMSU-A, DMSF-F[17, 18]/SSMIS, GCOM-W/AMSR2 and GPM-core/GMI; WV-CSRs of Himawari-8; soil moisture from GCOM-W/AMSR2 and Metop-[A, B]/ASCAT; Total Precipitable Water Vapor from ground-based GNSS

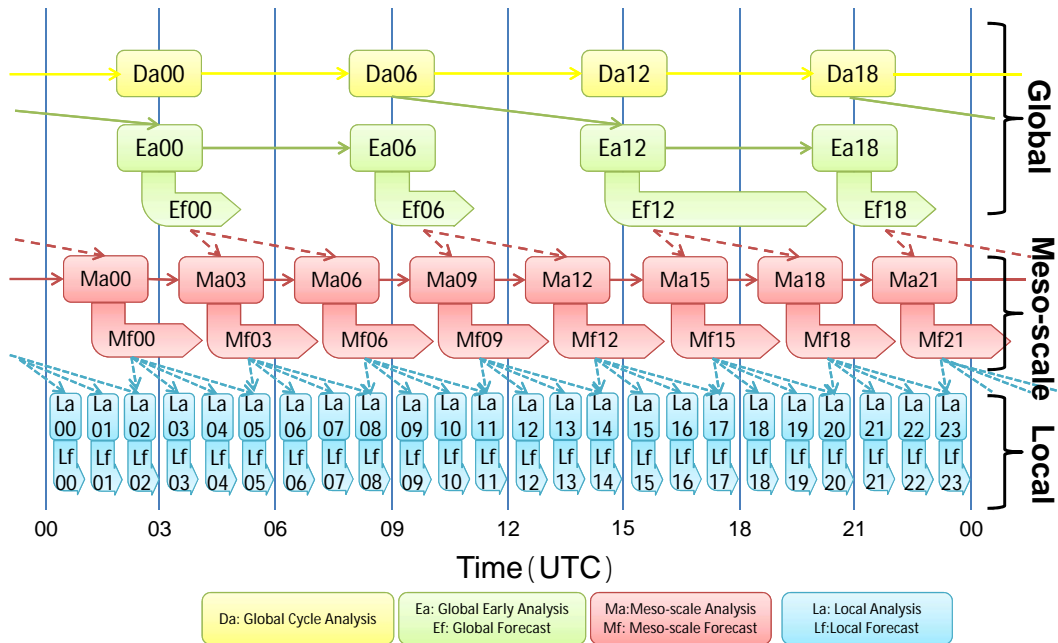


Figure 2.1.2: Main flow of JMA data assimilation systems. The first-guess and boundary conditions for Local Analysis are obtained from the latest MSM output.

2.2 Observation Data

2.2.1 Summary of Observation Data Used in Analysis

A variety of observations are utilized in JMA's current NWP systems. Table 2.2.1 summarizes the types used and the input parameters for the objective analysis systems, as of 1 November, 2018. Additional information on each observation type is provided in the following subsection.

2.2.2 Supplemental Information for Used Observation

2.2.2.1 SYNOP

SYNOP is a numerical code used for reporting surface observations at land stations. About 23,000 reports are produced every six hours.

2.2.2.2 AMeDAS

AMeDAS (the Automated Meteorological Data Acquisition System) is a JMA land surface automated network used to observe near surface temperature and wind at about 930 stations in Japan at 10 minute intervals. To monitor precipitation, around 1,300 raingauges are used in the network.

2.2.2.3 METAR

METAR is a numerical code used for reporting aerodrome weather information. Around 45,000 reports are produced every six hours.

2.2.2.4 SHIP

SHIP is a numerical code used for reporting surface observations performed at sea stations such as ships, oil rigs and moored buoys anchored at fixed locations. Around 6,400 reports are produced every six hours.

2.2.2.5 BUOY

BUOY is a numerical code used for reporting surface observations performed by drifting buoys. Around 10,000 reports are produced every six hours.

2.2.2.6 TEMP

TEMP is a numerical code used for reporting upper-level pressure, temperature, humidity and wind observations performed by radiosondes. Upper air observations are usually taken at the same time each day (00 and/or 12 UTC). Around 650 reports are produced at these times.

2.2.2.7 PILOT

PILOT is a numerical code used for reporting upper-level wind observations performed by rawins or pilot balloons. Around 300, 200 and 100 reports produced at 00, 12, 06 and 18, respectively.

2.2.2.8 Aircraft

Aircraft observations are reported via Aircraft Report (AIREP) and Aircraft Meteorological Data Relay (AMDAR). The numerous reports received from the U.S. are thinned to 1/50th over the continental U.S. in analysis pre-processing. Even after this processing, 80,000-100,000 reports are produced every six hours covering areas around the world. While vertical profile data can be obtained at the vicinity of airports, only flight level data can be collected along the other airways.

Table 2.2.1: Summary of the observation types and parameters used in objective analysis. Third column: P : surface pressure; u : zonal wind; v : meridional wind; T : temperature; Rh : relative humidity; T_B : radiance in brightness temperature; R_1 : precipitation amount; P_{wv} : precipitable water vapor; V_r : radial velocity, S_{mc} : soil moisture content. Fourth column: GA : global analysis; MA : meso-scale analysis; LA : local analysis.

Observation type (or code name used for reporting observation)	Brief description	Parameters used in analysis	Analysis type which observations are used
SYNOP	Land surface observations from world weather stations	P, u, v, T, Rh	GA, MA, LA
AMeDAS	Land surface automated observation network in Japan	u, v, T	LA
METAR	Routine weather report from aerodrome	P	GA
SHIP	Sea surface observations from ships, oil rigs and moored buoys	P, u, v, T, Rh	GA, MA, LA
BUOY	Sea surface observations from drifting buoys	P, u, v, T, Rh	GA, MA, LA
TEMP	Upper-air observations from radiosondes	P, u, v, T, Rh	GA, MA, LA
PILOT	Upper-air wind observations from rawins or pilot balloons	u, v	GA, MA, LA
Aircraft	Upper-air observations from aircraft (mainly commercial)	u, v, T	GA, MA, LA
Wind Profiler	Upper-air wind profile observations from Japan, Hong Kong and Europe	u, v	GA, MA, LA
AMV	Atmospheric motion vector (AMV) wind data from geostationary (GEO) satellites, low earth orbit (LEO) satellites and a combination of LEO and GEO	u, v	GA, MA, LA
Scatterometer	Ocean surface wind vector data from scatterometers on LEO satellites	u, v	GA, MA
MW Sounder	Radiance data from microwave (MW) sounders on LEO satellites	T_B	GA, MA, LA
MW Imager	Radiance data from MW imagers on LEO satellites and precipitation amounts estimated from MW imager radiance data	T_B, R_1 (MA only)	GA, MA, LA
CSR	Clear sky radiance (CSR) data from water vapor channels on GEO satellite infrared imagers	T_B	GA, MA, LA
Hyperspectral IR Sounder	Radiance data from Infrared(IR) sounders on polar orbiting satellites	T_B	GA
GNSS-RO	Bending angle and Refractivity profile data retrieved from radio occultation (RO) measurements of global navigation satellite system (GNSS) receivers on LEO satellites	Bending Angle Refractivity	GA (Bending Angle), MA (Refractivity)
Ground-based GNSS	Zenith total delay (ZTD) data and precipitable water vapor (P_{wv}) data estimated from ground-based GNSS receivers	ZTD, P_{wv}	GA (ZTD), MA, LA (P_{wv})
Radar Reflectivity	Relative humidity data estimated using 3-dimensional reflectivity data from JMA weather (Doppler) radars and Dual-frequency Precipitation Radar (DPR) onboard GPM-core satellite.	Rh	MA
Radial Velocity	Radial velocity data from JMA weather Doppler radars (WDRs) and Doppler radars for airport weather (DRAWs)	V_r	MA, LA
R/A	Radar estimated precipitation amounts calibrated using AMeDAS raingauge network data	R_1	MA
Soil Moisture Contents	Soil moisture data retrieved from a microwave imager radiance and microwave scatterometer observations	S_{mc}	LA
Typhoon Bogus	See Section 2.4.	P, u, v	GA, MA

2.2.2.9 Wind Profiler

Upper air wind speeds and directions are monitored by wind profilers on the ground. A total of 33 wind profilers operated by JMA produce data every 10 minutes. The specifications are detailed in Ishihara *et al.* (2006). Wind profiler data from Europe and Hong Kong are also available.

2.2.2.10 AMVs

Atmospheric motion vector (AMV) wind data are derived by tracing the movement of cloud or water vapor patterns in successive satellite images. AMVs from four geostationary (GEO) satellites (Meteosat-8, -11, GOES-15 and Himawari-8), low earth orbit (LEO) satellites (Terra, Aqua, NOAA and Metop) and LEOGEO-AMVs are used. LEOGEO-AMV data are derived using imagery from a combination of polar-orbiting and geostationary satellites for 60°N and 60°S latitude areas. AMVs from GEO satellites cover 60°N – 60°S and those from polar-orbiting satellites cover polar regions (i.e., latitudes higher than 60°).

2.2.2.11 Scatterometers

Ocean surface wind vectors from scatterometers onboard polar orbiting satellites are used. Data from ASCAT (the advanced scatterometer) onboard Europe's Metop-A, -B polar orbiting satellite are currently utilized.

2.2.2.12 MW Sounders

Clear radiance data from microwave (MW) sounders are used. The data adopted are from AMSU-A (Advance Microwave Sounding Unit A) on NOAA-15, -18, -19, Metop-A, -B and Aqua, MHS (Microwave Humidity Sounder) units on NOAA-18, -19, Metop-A, -B, SAPHIR (Sondeur Atmospherique du Profil d'Humidite Intertropicale par Radiometrie) on Megha-Tropiques, and ATMS (Advanced Technology Microwave Sounder) on Suomi-NPP. AMSU-A is a temperature sounder, and radiance is sensitive to temperature profiles. MHS and SAPHIR are humidity sounders. ATMS has temperature and humidity sounding channels.

2.2.2.13 MW Imagers

Radiances from MW imagers less affected by cloud/rain are used. The data adopted are from AMSR2 (Advanced Microwave Scanning Radiometer 2) on GCOM-W, SSMIS (Special Sensor Microwave Imager Sounder) on DMSP-F17, -F18 and GMI (GPM (Global Precipitation Measurement mission) Microwave Imager) on GPM. Radiance is sensitive to water vapor amounts in the lower troposphere. Precipitation amounts estimated from radiances using the MSC method (Takeuchi and Kurino 1997) are also used in meso-scale analysis.

2.2.2.14 CSR

Clear sky radiance (CSR) is a product providing averaged radiance over cloud-free pixels in GEO satellite imagers. CSR data from four GEO satellites (Meteosat-8, -11, GOES-15 and Himawari-8) are used. CSR data sensitive to water vapor amounts in the upper and middle troposphere are used.

2.2.2.15 Hyperspectral IR Sounders

Clear radiance data from hyperspectral IR sounders are used. The data adopted are from AIRS on Aqua, IASI on Metop and CrIS on Suomi-NPP. Channels located within a CO₂ absorption band sensitive to temperature are used.

2.2.2.16 GNSS-RO

GNSS-RO (Global Navigation Satellite Systems - Radio Occultation) is a technique for measuring atmospheric profiles. With this approach, a set of atmospheric time delay data of GNSS radio signals received by a low earth orbit (LEO) satellite is obtained during each radio occultation event. Since the delay is a result of atmospheric radio refraction along the propagation path of the signal, the vertical profiles of refractivity (or the bending

angle) of the atmosphere at a tangent point can be estimated from the delay data set. As refractivity is a function of temperature, humidity and pressure, it can be used to determine the profiles of these properties. The currently used LEO satellites and their GNSS receivers are IGOR (Integrated GPS Occultation Receiver) onboard COSMIC (Constellation Observing System for Meteorology, Ionosphere and Climate) satellites and TerraSAR-X, GRAS (GNSS Receiver for Atmospheric Sounding) onboard Metop-A and -B.

2.2.2.17 Ground-based GNSS

Ground-based GNSS data are provided from atmospheric time delays of GNSS radio signals collected by ground-based GNSS receivers. JMA uses ground-based GNSS data collected from the global network along with GEONET GNSS receiver data. GEONET is a ground-based GNSS receiver network operated by the Geospatial Information Authority in Japan using around 1,200 receivers located throughout the country.

GNSS-ZTD (GNSS - zenith total delay) data are estimated by averaging the delays of multiple GNSS satellite signals monitored using a single receiver, and are used in global analysis.

GNSS-PWV (GNSS - Precipitable Water Vapor) data based on analysis of GEONET GNSS-ZTD data are used in meso-scale and local analysis.

2.2.2.18 Radar Reflectivity

A total of 20 C-band weather radars with Doppler functionality are operated by JMA. Three-dimensional reflectivity data are obtained every five minutes, and relative humidity profiles are estimated from reflectivity data and NWP grid point values using a technique based on Bayes' theorem (Caumont *et al.* 2010). Relative humidity data are produced for areas within a 200 km radius of each radar site below freezing level. Relative humidity profiles estimated from reflectivity data collected using the Dual-frequency Precipitation Radar (DPR) on the GPM-core satellite are used in meso-scale analysis.

2.2.2.19 Radial Velocity

A total of 20 C-band weather Doppler Radars (WDRs) and 9 Doppler Radars for Airport Weather (DRAWs) are operated by JMA. Three-dimensional radial velocity data are produced every five minutes within a 150 km radius for WDRs and every six minutes within a 120 km radius for DRAWs. The range resolution is 250 m and the azimuthal resolution is 0.703° .

2.2.2.20 R/A

Radar/Raingauge Analyzed Precipitation (R/A) is a product providing composite precipitation data produced by JMA. These data are cumulative precipitation estimations based on weather radar data with a Z-R relationship ($Z = 200R^{1.6}$) calibrated using AMeDAS raingauge data in real time. The details are found in Subsection 4.4.1.

2.2.2.21 Soil Moisture Content

Soil moisture content data estimated from microwave imager (AMSR2/GCOM-W) radiances and microwave scatterometer (ASCAT/Metop-A, -B) observations are used for areas over land in local analysis.

2.3 Quality Control and Related Procedures

Quality control (QC) is a series of procedures by which “bad” observations are screened out. It is a vital component of the objective analysis system because observations sometimes include large errors and erroneous data can significantly impair the quality of atmospheric analysis, leading to low levels of forecast skill. QC procedures in JMA's objective analysis systems are described in the following subsections.

2.3.1 SYNOP, AMeDAS, METAR, SHIP, BUOY, TEMP, PILOT, Aircraft and Wind Profilers

Direct observations (i.e. SYNOP, AMeDAS, METAR, SHIP, BUOY, TEMP, PILOT and aircraft) and wind profilers measures prognostic variables in NWP such as pressure, temperature, wind and humidity. The QC system for these observations consists of internal QC and external QC.

2.3.1.1 Internal QC

Internal QC involves procedures to check and correct observation values using collocated data in reports and several external lists or tables. The checks are outlined below.

1. Blacklist check: The blacklist is a list of problematic stations or data, and is prepared in advance via non-real-time QC (see Section 2.9). Blacklisted observations are rejected in this step.
2. Climatological check: Climatological reasonability is checked in this step. The criteria are defined in advance based on WMO (1993).
3. Trajectory check: Consistency at consecutive locations is checked for reports from moving stations such as SHIP, BUOY and aircraft. The movement velocity and direction are checked in this step and checking is also performed to ensure that SHIP and BUOY locations are in the ocean.
4. Inter-element consistency check: The temporal continuity of consecutive reports from surface stations is checked along with consistency among observation elements within the report.
5. Vertical consistency check: Vertical consistency is checked in TEMP and PILOT data. The check items are (1) instrument icing, (2) temperature lapse rate, (3) hydrostatic relationship, (4) consistency among data at standard pressure levels and those at significant levels and (5) vertical wind shear.
6. Bias correction: Bias correction is applied to TEMP data reported without radiative heating correction or with apparent systematic biases. The correction constants are determined from one-month statistics for the previous month. The same bias correction approach is applied to aircraft temperature data for global analysis but not for meso-scale and local analysis.

2.3.1.2 External QC

External QC involves procedures to check observation values with comparison to (external) first guess and neighboring observations. The checks are outlined below.

1. Gross error check: The departure ($D \equiv O - B$) of the observed value (O) from the first guess (B) is calculated for all observations. The absolute value of D is compared with the tolerance limits C_P (the pass criterion) and C_R (the failure criterion). Data satisfying $|D| \leq C_P$ pass the QC, and those characterized by $|D| > C_R$ are rejected. Data characterized by $C_P < |D| \leq C_R$ are regarded as suspect and sent for spatial consistency checking.
2. Spatial consistency check: The departure D in suspect observation data is compared with departures interpolated using the optimum interpolation method (D_{OI}) with neighboring observations. The absolute difference of D and D_{OI} is compared with the tolerance limit C_S (the criterion for suspect) for final judgment and the data satisfying $|D - D_{OI}| \leq C_S$ are accepted.
Here, the tolerance limits C_P , C_R , and C_S vary with local atmospheric conditions in first guess fields. The limits are made small if the time tendency and horizontal gradient are small in the fields, and *vice versa*. This scheme is called Dynamic QC (Onogi 1998).
3. Duplication check: Duplication is often found in observation reports with data obtained through different communication lines. The most appropriate report is picked out from among duplicates after performance of the above checks in consideration of status.

Table 2.3.1: Summary of blacklist areas for AMVs. IR: infrared; WV: water vapor; CSWV: clear sky water vapor; NH: Northern Hemisphere; SH: Southern Hemisphere; Polar AMV: AMVs from polar-orbiting satellites; GEO AMV: AMVs from geostationary satellites; LEOGEO AMV: AMVs from a combination of low earth orbiting and geostationary satellites.

Kind	Blacklisting area
Polar AMV (IR) at NH	above 300 hPa or below 900 hPa
Polar AMV (WV/CSWV) at NH	above 300 hPa or below 550 hPa
Polar AMV (IR/WV) at SH	above 300 hPa or below 550 hPa
Polar AMV (CSWV) at SH	above 350 hPa or below 550 hPa
Polar AMV (All)	poleward of 88°N or 88°S
GEO AMV (All)	above 175 hPa or below 975 hPa
GEO AMV (IR)	above 275 hPa at poleward of 20°N or 20°S
GEO AMV (WV)	above 225 hPa at poleward of 20°N or 20°S
LEOGEO AMV	above 300 hPa or below 900 hPa
LEOGEO AMV	between 600 hPa and 640 hPa
LEOGEO AMV	at poleward of 70°N or 70°S
LEOGEO AMV	between 640 hPa and 900 hPa at poleward of 60°S

2.3.2 AMV

Blacklisted AMVs (Table 2.3.1) are rejected in the first step, as are those with low quality indicators (QI, Holmlund 1998). QI thresholds are defined for each satellite, domain, vertical level and image type. And thinning is then performed based on a distance of 200 km. Climatological checking (see Subsection 2.3.1.1) and external QC (see Subsection 2.3.1.2) are then performed. The details of the QC and its detailed settings for AMVs are given on the NWP SAF AMV monitoring page¹.

2.3.3 Scatterometers

Level 2 ocean surface wind products are used in global analysis and meso-scale analysis. Low quality data from areas over land or sea ice are rejected at the first step. The most likely wind directions are then selected from the inherent ambiguity wind directions in scatterometer measurements using both the NWP nudging technique and the median filter technique. The next step is gross error checking (see Subsection 2.3.1.2). In this step, correct wind data are occasionally rejected in and around severe weather systems such as cyclones and fronts where the wind direction and speed vary sharply. To avoid such undesirable rejection, specialized quality control named Group-QC is applied. In this step, spatial consistency among wind vectors is checked in terms of smooth transition in wind direction and wind speed. Data that pass Group-QC are excluded from rejection in gross error checking. The details of scatterometer QC are given on the NWP SAF scatterometer monitoring page².

2.3.4 Satellite Radiance

Satellite radiance data are used in global and meso-scale analysis as a form of brightness temperature. The RTTOV-10 fast radiative transfer model (Saunders *et al.* 2012) is employed for radiance assimilation. The common QC procedures for radiance data are blacklist checking, thinning and external QC. The blacklist specifies problematic instruments, and is made in advance based on non-real-time QC (see Section 2.9). Blacklisted data are rejected in the first step. In the next step, data are thinned spatially in each time slot of the assimilation window (approximately one hour) to reduce computational costs. The subsequent external QC includes reduction of instrumental scan biases (except for CSR), cloud/rain contamination checking, location checking,

¹<https://nwpsaf.eu/monitoring/amv/amvusage/jmamodel.html>

²<https://nwpsaf.eu/monitoring/scatter/scatusage/jmamodel.html>

Table 2.3.2: Summary of microwave sounder channel sets used for each condition

	AMSU-A	MHS	ATMS	SAPHIR	SSMIS (Water vapor sounding channels)
Clear sky ocean	Ch. 4–13	Ch. 3–5	Ch. 6–9, Ch. 18–22	Ch. 1–6	Ch. 9–11
Clear sky land/coast/sea-ice	Ch. 6–13	Ch. 3–5	Ch. 7–9, Ch. 18–22	n/a	n/a
Cloudy ocean	Ch. 7–13	Ch. 3–5	Ch. 8–9	n/a	n/a
Rainy ocean	Ch. 9–13	n/a	n/a	n/a	n/a

channel selection and gross error checking (see Subsection 2.3.1.2). Data passing this QC are thinned again to reduce the observation error correlation, and the thinned data are output for use in data assimilation systems. In global and local analysis, variational bias correction (VarBC, Derber and Wu 1998; Dee 2004) is used to reduce air-mass dependent biases. VarBC is an adaptive bias correction scheme in which a linear regression formula representing biases is embedded in the observation operator and regression coefficients are set as analysis variables. The formulations are described in Subsection 2.5.7.4. In meso-scale analysis, air-mass dependent biases are removed in pre-processing using the VarBC coefficients obtained in the latest global analysis. The satellite radiance data used are from MW sounders, MW imagers, CSR and hyperspectral IR sounders (GA only). The specific procedures for each data type are described in the following subsections.

2.3.4.1 MW sounders

The sets of channels used are defined in advance according to individual surface and atmospheric conditions. The sets are summarized in Table 2.3.2.

2.3.4.2 MW imagers

Vertically polarized-channel radiances that are less affected by cloud/rain are assimilated over ice-free ocean areas. In meso-scale analysis, precipitation retrieval is also assimilated for areas over the ocean surrounding Japan. Precipitation amount estimations are resampled onto inner model grids with spatial smoothing.

2.3.4.3 CSR

CSR data are horizontally thinned to divisions of 220 km for global analysis and 45 km for meso-scale analysis and local analysis. Hourly CSR data are used in these analysis types. Values with a low percentage of clear pixels and a large standard deviation of brightness temperature are excluded due to their low representation of the area. CSR data from high-altitude areas (above 4,000 m) are not used. In calculation with RTTOV-10, emissivity atlas data and retrieved surface temperatures from window channel radiance are used for areas over land.

2.3.4.4 Hyperspectral IR sounders

Clear sky radiance data from hyperspectral IR sounders are used in global analysis. Data are horizontally thinned to divisions of 200 km, and cloud top height estimation/cloud screening are applied in quality control. The related methods are based on the CO₂ slicing approach (Eyre and Menzel 1989).

2.3.5 GNSS-RO

Bending angle data for altitudes up to 60 km are used in global analysis with 500-m vertical intervals. Refractivity data are used in meso-scale analysis. No bias correction is applied for GNSS-RO data.

2.3.6 Ground-based GNSS

ZTD data are used in global analysis. Stations at an elevation of 5,000 m are not used, and those from which the absolute difference of elevation to the model surface exceeds 300 m are excluded. GNSS-ZTD values smaller than 1,000 mm or larger than 3,000 mm are rejected in climatological checking. Data with absolute differences of more than 50 mm are regarded as suspect. If the absolute difference of elevation to the model surface in values around suspect data exceeds 50 mm, the ZTD data are not used.

PWV data are used in meso-scale and local analysis. As Japan is characterized by steep mountainous terrain, large differences are found between actual ground surface elevations and model surface elevations especially in mountainous areas. In meso-scale analysis, stations at an elevation of 700 m or more above mean sea level are not used, and those from which the absolute difference of elevation to the model surface exceeds 200 m are excluded. GNSS-PWV values smaller than 1 mm or larger than 90 mm are rejected in climatological checking. The first guess PWV is then interpolated or extrapolated to the actual terrain surface and compared to the GNSS-PWV. Data with an absolute difference of more than 8 mm from the first guess are rejected in gross error checking. As there is a dense GNSS-PWV network for the analysis systems, data are thinned by 30 km for meso-scale analysis and 20 km for local analysis. GNSS-PWV data in rainy conditions are not used in meso-scale analysis.

2.3.7 Radar Reflectivity

To assimilate radar reflectivity data in meso-scale analysis and local analysis, an indirect assimilation technique called 1D+4DVAR (Ikuta and Honda 2011) is employed. This approach is based on Caumont *et al.* (2010). In 1D+4DVAR, radar reflectivity data are used to retrieve relative humidity (RH) values, which are assimilated as conventional observation data in 4D-Var. In this system, only retrieved RH values from below the melting layer are used because it is known that reflectivity inappropriately simulated in the ice phase with the operational MSM hydrometeor forecast, causing large biases in RH retrievals. In addition, data from around a height of 2000 m above sea level are also not used since these data are used for making R/A and are already assimilated in meso-scale analysis in another form (surface rainfall, see Subsection 2.3.9). For operation, reflectivity data from the JMA C-band radar network are used. Reflectivity data from space-based Dual frequency Precipitation Radar are used to retrieve RH values and assimilated in meso-scale analysis.

2.3.8 Radial Velocity

Hourly radial velocity data from WDRs and DRAWs are used in meso-scale analysis and local analysis. In pre-processing, the data are resampled into a 5 km range resolution and a 5.625° azimuthal resolution. The resampled data are checked with respect to the number of data sampled, radial velocity variance and the difference between maximum and minimum velocity. High elevation angle data ($\geq 5.9^\circ$) are not used to avoid the contamination of precipitation velocity values, and those from areas close to radar site ($< 10\text{km}$) are not used to avoid the influence of back scatter noise. Data showing wind speeds of less than 5 m/s are also not used to avoid ground clutter contamination.

2.3.9 R/A

Hourly R/A data are assimilated in meso-scale analysis. As R/A data are quality controlled, 1 km grid values are simply resampled into inner-model grid boxes (15 km) and input for this type of analysis.

2.3.10 Soil Moisture Content

Variable transformation using the cumulative distribution function (CDF) matching method is applied for soil moisture content (SMC). The CDF matching method involves fitting the probability density function (PDF) of observation to the PDF of model variables. This pre-conditioning via CDF matching helps to minimize the cost function because the innovation of SMC becomes Gaussian after the CDF matching. A variational bias correction method is used for SMC in local analysis.

2.3.11 CDA: Feedback Data Base

All information concerning the quality of observational data obtained during the quality control procedure is archived in the Comprehensive Database for Assimilation (CDA), which is extensively used for both real-time and non real-time data monitoring activities. All information contained in the CDA is managed in the form of integer two byte data. The database's format is quite simple and designed for flexible use so that information on observations can be archived easily, and is also user-friendly to facilitate data retrieval.

2.4 Typhoon Bogussing

For tropical cyclones (TCs) over the western North Pacific, typhoon bogus data are generated as a form of pseudo-observation information and assimilated for realistic TC structure analysis based on model resolutions. The data consist of pressures values at the mean sea level (P_{mst}) and vertical profiles of the wind (W_{prf}) around TCs. Wind profiles are placed at 850 and 300 hPa in global analysis, 1000, 925, 850, 800, 700, 600, 500, 400 and 300 hPa in meso-scale analysis. The bogus generated has an axially asymmetric structure in types of the analysis.

Firstly, symmetric bogus profiles are generated automatically from central pressure values and the 15m/s wind speed radius of a TC (R_{15}) as analyzed by forecasters. The surface pressure profile is defined using Fujita's formula (Fujita 1952). Gradient wind balance is assumed for calculation of surface pressure profile meeting the requirements at hand from the wind speed at the particular radius of R_{15} . Upper geopotential profiles are defined using an empirical formula based on the TC analysis described by Frank (1977). It is assumed that the temperature anomaly has its maximum at 250 hPa. The wind field on each level is derived from geopotential height profiles with gradient wind balance. The surface wind field is also derived from gradient wind balance, but is modified to include the effects of surface friction.

Secondly, asymmetric components are retrieved from the first guess fields and added to the symmetric bogus profile to generate the final asymmetric bogus structure. When the target area of bogussing is across the lateral boundary in the meso-scale analysis, asymmetric components are not added.

Finally, pseudo-observation data are generated from the resulting bogus structure at the TC center analyzed (P_{mst}), the TC center of the first guess (P_{mst}), and several points surrounding the TC center analyzed (P_{mst} and W_{prf}). The configuration for the surrounding point distribution is adaptable to the typhoon track error of the first guess.

2.5 Global Analysis

2.5.1 Introduction

A four-dimensional variational (4D-Var) data assimilation method is employed in analysis of the atmospheric state for the Global Spectral Model (GSM), and is performed at 00, 06, 12, 18 UTC. An early analysis with a short cut-off time is performed to prepare initial conditions for operational forecasting, and a cycle analysis with a long cut-off time is performed to maintain the quality of the global data assimilation system. In order to improve computational efficiency, an incremental method (Courtier *et al.* 1994) is adopted in which the analysis increment is evaluated first at a lower (inner) resolution (TL319L100: grid roughly equivalent to 0.5625° (55 km) and up to 0.01 hPa), and is then interpolated and added to the first-guess field at the original resolution (TL959L100: grid roughly equivalent to 0.1875° (20 km)). Specification of the JMA Global Analysis system is summarized in Section 2.1.

The three-dimensional variational (3D-Var) data assimilation system was operated as the first operational variational analysis system for GSM with the inner resolution of T106L40 (grid roughly equivalent to 110 km and up to 0.4 hPa) in September 2001 (Takeuchi and Tsuyuki 2002). In February 2005, the 4D-Var data assimilation system was operated with the inner resolution of T63L40 (grid roughly equivalent to 180 km) (Kadowaki 2005). The inner resolution was upgraded to T106L40 in March 2006 (Narui 2006), T159L60 (grid roughly equivalent to 75 km and up to 0.1 hPa) in November 2007, TL319L60 in October 2011 (Kadowaki and Yoshimoto 2012), and TL319L100 in March 2014.

2.5.2 Incremental 4D-Var Formulation

In 4D-Var, 3–9-hour forecasts from the GSM are used as a first guess (background), and all observations from within three hours of analysis time are organized in hourly time slots as described in the Subsection 2.5.3. The cost function is used to measure the distance between the model trajectory and observations over a six-hour assimilation window.

In an incremental method, the analysis increment is evaluated first at a low-resolution. To determine the low-resolution analysis increment Δx_i , minimization of the cost function J as defined by Eq. (2.5.1) is performed in the inner loop.

$$J(\Delta x_0) = \frac{1}{2} \Delta x_0^T \mathbf{B}^{-1} \Delta x_0 + \frac{1}{2} \sum_{i=1}^n (\mathbf{H}_i \Delta x_i - d_i)^T \mathbf{R}_i^{-1} (\mathbf{H}_i \Delta x_i - d_i) + J_C \quad (2.5.1)$$

$$\Delta x_{i+1} = \mathbf{M}_i \Delta x_i \quad (i = 0, \dots, n-1)$$

where the subscript $i(\geq 1)$ is the index of time slot and n is the number of time slots. Δx_0 (Δx_1) is the low-resolution analysis increment at the initial time before (after) initialization, $\Delta x_{i \geq 2}$ is the analysis increment evolved according to the tangent linear (TL) model to time of slot i . \mathbf{M}_0 is the TL operator of the nonlinear normal-mode initialization operator (Machenhauer 1977) and $\mathbf{M}_{i \geq 1}$ is the TL model of the low-resolution nonlinear (NL) forecast model M_i as detailed in Subsection 2.5.4 for integration from time of slot i to that of slot $i+1$. \mathbf{R}_i denotes the covariance matrix of observation errors at time of slot i , and \mathbf{B} is the covariance matrix of background errors as detailed in Subsection 2.5.6 and Subsection 2.5.7. \mathbf{H}_i is the TL operator of the observation operator H_i . The innovation vector is given for each assimilation slot by $d_i = y_i^0 - H_i x_i^b$, where x_i^b is the background state evolved using the high-resolution NL model, and y_i^0 is the observation data at time of slot i . J_C is a penalty term used to suppress gravity waves described in Subsection 2.5.5.

To minimize the cost function J , the limited memory Broyden-Fletcher-Goldfarb-Shanno (L-BFGS) algorithm (Liu and Nocedal 1989) with Veersé's preconditioner (Veersé *et al.* 2000) is applied. Here, the gradient of the cost function ∇J is required. This is determined via the adjoint procedures of Eq. (2.5.2)–Eq. (2.5.5) as computed in reverse time.

$$p_{n+1} = 0 \quad (2.5.2)$$

$$p_i = \mathbf{M}_i^T p_{i+1} + \mathbf{H}_i^T \mathbf{R}_i^{-1} (\mathbf{H}_i \Delta x_i - d_i) \quad (i = n, \dots, 1) \quad (2.5.3)$$

$$p_0 = \mathbf{M}_0^T p_1 + \mathbf{B}^{-1} \Delta x_0 \quad (2.5.4)$$

$$\nabla J(\Delta x_0) = p_0 \quad (2.5.5)$$

where p_i is a auxiliary variable, \mathbf{M}_i^T represents the adjoint (AD) model of the TL model \mathbf{M}_i , and \mathbf{H}_i^T is the AD operator of \mathbf{H}_i . Note that Eq. (2.5.2)–Eq. (2.5.5) should contain additional terms for the penalty term in Eq. (2.5.1), which are omitted here for simplicity.

The variables analyzed are relative vorticity ζ , divergence η , temperature T , surface pressure P_s , and the logarithm of specific humidity $\ln q$ in the spectral space on the model layers (eta coordinates). The observational data y_i^0 include, but are not limited to, wind vectors, temperature, relative humidity, and satellite radiances.

The low-resolution analysis increment Δx_i determined from minimization of the cost function in the inner loop is interpolated to the high-resolution analysis increment. Adding this analysis increment to the first guess field produces high-resolution analysis.

2.5.3 Procedural Description

The flow of 4D-Var is shown in Figure 2.5.1 for the 12 UTC analysis time, and is the same for the cycle and early analyses.

The procedure is as follows:

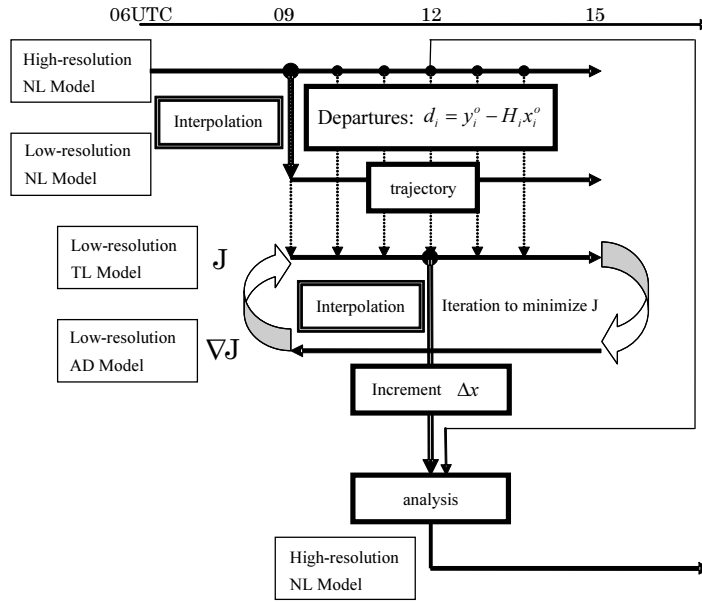


Figure 2.5.1: Flow of 4D-Var operation for the 12 UTC analysis time

1. The nine-hour forecast (09 – 15 UTC) of the high-resolution outer NL model (as per the GSM with a resolution of TL959L100) from previous analysis is used as a first guess (background), and departures between the model trajectory and observations $d_i = y_i^o - H_i x_i^b$ over a six-hour assimilation window (09 – 15 UTC) are measured. Observations are organized in six time slots with intervals of 0.5 hours for the first slot, 1.5 hours for the last slot, and 1 hour for the others (Figure 2.5.2). All observations in each time slot are regarded as taking place at each representative time.
2. The three-hour forecast field (valid at 09 UTC) of the first guess is interpolated into the field with the resolution of the inner model (TL319L100). The interpolation is performed not only horizontally but also vertically in consideration of the topography difference between TL959 and TL319.
3. The inner NL model is run from the interpolated field to calculate the background state in the low-resolution model space.
4. The TL and AD models are run to calculate the cost function J and its gradient ∇J with the innovation vector $d_i = y_i^o - H_i x_i^b$. These processes are iterated to minimize the cost function J . The iteration is performed up to about 70 times, and the background trajectory is not updated in the system.
5. After the minimization of J , the field of the three-hour forecast (valid at 12 UTC) of the TL model is chosen as the analysis increment. It is interpolated horizontally and vertically into the field with the resolution of the first-guess field (TL959L100). Finally, the analysis increment is added to the first-guess field (valid at 12 UTC) to obtain the final product.

2.5.4 Inner Model

The inner NL model is based on the GSM, but many processes are based on those of the older GSM for many reasons. Especially, moisture processes (the convection and cloud schemes) are based on those of GSM0103 (JMA 2002), mainly for the stability of inner TL model integration. Nonlinear normal-mode initialization (Machenhauer 1977) is also added.

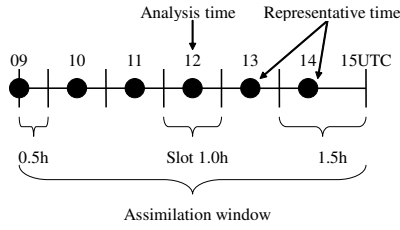


Figure 2.5.2: Schematic diagram of time slots for the analysis time 12 UTC. The black circles indicate the representative time of each time slot.

The inner TL model includes the following simple processes, most of which are based on the inner NL model:

1. **Initialization:** To control gravity waves, the TL version of the nonlinear normal-mode initialization is adopted.
2. **Horizontal Diffusion:** Horizontal diffusion is enhanced over that of the inner NL model based on Buizza (1998).
3. **Surface Turbulent Fluxes:** Surface turbulent fluxes are formulated as Monin-Obukhov bulk formulae based on the inner NL model. Sensible and latent heat flux are perturbed only over the sea.
4. **Vertical Turbulent Transports:** The vertical turbulent transports of momentum, heat, and moisture are formulated as the hybrid downgradient-type scheme based on the inner NL model (Subsection 3.2.7). Those diffusion coefficients are not perturbed in the inner NL model.
5. **Orographic Gravity Wave Drag:** The parameterization for orographic gravity wave drag consists of two components: one for long waves (wavelength > 100 km) and the other for short waves (wavelength ≈ 10 km) based on the inner NL model which is almost same as in the GSM (Subsection 3.2.8.1). The Richardson number is not perturbed in some parts for long waves for the stability of inner TL model integration.
6. **Long-wave Radiation:** Long-wave radiation in the TL model is based on Mahfouf (1999). The tendency of the perturbed temperature T' is given by

$$\frac{\partial T'}{\partial t} = -\alpha \frac{g}{C_p} \frac{\partial}{\partial p} \left(4F \frac{T'}{T} \right) \quad (2.5.6)$$

where $\alpha = 1/\{1 + (p_r/p)^{10}\}$, $p_r = 300$ hPa, F represents the net radiation fluxes calculated in the inner NL model, and g and C_p denote the gravitational constant and isobaric specific heat, respectively.

7. **Clouds and Large-scale Precipitation:** Clouds and large-scale precipitation are based on the inner NL model, in which the former are prognostically determined in a way similar to that proposed by Smith (1990). A simple statistical approach proposed by Sommeria and Deardorff (1977) is employed to compute cloud amounts and cloud water content. The parameterization of the conversion rate from cloud ice to precipitation follows the scheme proposed by Sundqvist (1978). These considerations are much simplified in the TL model. The cloud fraction, the amount of falling cloud ice and the dependence on water vapor of isobaric specific heat are not perturbed. Only certain variables are perturbed in computing the conversion from cloud water to precipitation and the evaporation of precipitation.

8. **Cumulus Convection:** Cumulus convection is formulated as the prognostic Arakawa-Schubert scheme (Arakawa and Schubert 1974; Moorthi and Suarez 1992; Randall and Pan 1993) based on the inner NL model, but is highly simplified. Vertical wind shear and the planetary mixing length are not perturbed. The magnitude of mass-flux perturbation is set bounds for the stability of inner TL model integration, and as a result this model is not exactly linear.

2.5.5 Penalty Term

The penalty term, which is the third term of Eq. (2.5.1), is given by

$$J_C = \alpha \left(|\mathbf{N}_G \Delta x_0|^2 + \sum_{i=2}^n |\mathbf{N}_G \Delta x_i|^2 \right) \quad (2.5.7)$$

where \mathbf{N}_G denotes an operator used to calculate the tendency of the gravity wave mode based on Machenhauer (1977). α is an empirically determined constant $3.0 \times 10^{-2} [\text{s}^4/\text{m}^2]$. Although this penalty term is primarily introduced to suppress gravity waves in the analysis increment Δx_i , it is also effective in stabilizing calculation.

2.5.6 Background Term

The background term, which is the first term on the right side of Eq. (2.5.1), dominates how the difference between observation data and the first guess is converted into correction for the first guess in the 4D-Var analysis procedure. The multivariate couplings in the analysis variables are based on the geostrophic linear balance between mass and wind. Control variables are introduced to reduce the correlations among the analysis variables, and additional statistical relations are considered in the algorithm. These include the lower geostrophic balance on smaller horizontal and vertical scales, the almost complete lack of geostrophic balance near the equator, the dependency of geostrophy on the vertical level, and the weak coupling between divergence and vorticity as well as between divergence and mass.

The control variables in 4D-Var are relative vorticity $\Delta \zeta$, unbalanced divergence $\Delta \eta_U$, unbalanced temperature and surface pressure $(\Delta T_U, \Delta P_{sU})$, and the logarithm of specific humidity $\Delta \ln q$ in the spectral space on the model layers. Δ denotes deviation from the first guess and subscript U means the term of "unbalanced". Autocovariances of the control variables are assumed to be homogeneous and isotropic. Correlation structures do not depend on geographical locations, but vertical correlations do depend on horizontal scale. The unbalanced variables $\Delta \eta_U$ and $(\Delta T_U, \Delta P_{sU})$ are defined as

$$\Delta \eta_U \equiv \Delta \eta - P \Delta \phi_B \quad (\Delta \phi_B = \Delta \phi_B(\Delta \zeta)) \quad (2.5.8)$$

$$\begin{pmatrix} \Delta T_U \\ \Delta P_{sU} \end{pmatrix} \equiv \begin{pmatrix} \Delta T \\ \Delta P_s \end{pmatrix} - Q \Delta \phi_B - R \Delta \eta_U \quad (2.5.9)$$

where P , Q , and R are regression coefficients, $\Delta \phi_B$ is a modified balance mass variable derived from relative vorticity as described in Subsection 2.5.6.1. This formulation is similar to that previously used in ECMWF (Derber and Bouttier 1999). The regression coefficients are computed statistically using the NMC method (Parrish and Derber 1992) with 24/48-hour forecast differences to enable estimation of the total covariances for each total spectral coefficient.

2.5.6.1 Modified Balance Mass Variable

The geostrophic balance is well kept at mid-levels in the troposphere in the extratropics. In other areas, the balance is weak. To incorporate consideration of these relationships, a modified balance mass variable is introduced. The statistical relationships linking relative vorticity, divergence, temperature, and surface pressure are calculated. First, singular value decomposition of the linear balance operator L is conducted.

$$\Delta \tilde{\phi}_B = L \Delta \zeta = U W V^T \Delta \zeta \quad (2.5.10)$$

where $\Delta\tilde{\phi}_B$ is the original balance mass variable, W is a positive semi-definite diagonal matrix, and U and V are orthogonal matrices. The decomposed modes depend on latitude (i.e., a singular mode with a small singular value has a large amplitude in the low latitudes). Each wave number component of L is denoted as

$$\begin{aligned}\Delta\tilde{\phi}_{B_n}^m &= c_n^m \Delta\zeta_{n-1}^m + c_{n+1}^m \Delta\zeta_{n+1}^m && ((n, m) \neq (0, 0), n = m, m + 1, \dots, N), \\ c_n^m &= -\frac{2\Omega a^2}{n^2} \sqrt{\frac{n^2 - m^2}{4n^2 - 1}}, && c_{N+1}^m = 0, \quad \Delta\tilde{\phi}_{B_0}^0 = 0\end{aligned}\quad (2.5.11)$$

here Ω is the angular velocity of the earth, a is the earth's radius, n is the total wavenumber, and m is the zonal wavenumber. Second, the coefficient of regression between mass variable $\Delta\Phi$ ³ (as derived from temperature and surface pressure) and balance mass variable is calculated as follows

$$D_j = \frac{\langle (U^T \Delta\Phi)_j^m (U^T \Delta\tilde{\phi}_B)_j^m \rangle}{\langle [(U^T \Delta\tilde{\phi}_B)_j^m]^2 \rangle}\quad (2.5.12)$$

where $\langle \rangle$ denotes statistical, zonal-wavenumber, and vertical-level mean, D_j denotes a positive definite diagonal matrix, and j denotes an index of singular vectors in latitudinal wave numbers, respectively. The regression coefficients D_j (0 – 1) indicate the extent to which geostrophic balance is satisfied. Modified balance mass variable $\Delta\phi_B$ is then constructed as follows:

$$\Delta\phi_B = UDU^T \Delta\tilde{\phi}_B = UDWV^T \Delta\zeta = \tilde{L}\Delta\zeta\quad (2.5.13)$$

Note that the modified balance operator \tilde{L} is based on 1) conversion from the spectral space to the singular vector space, 2) the product of the regression coefficients D , and 3) conversion from the singular vector space to the spectral space. The correlation between the modified mass variable and the unbalanced mass variable (i.e. $\Delta\Phi - \Delta\phi_B$) can be neglected in all regions including the tropics.

2.5.6.2 Regression Coefficients for $\Delta\eta_U$ and $(\Delta T_U, \Delta P_{sU})$

The regression coefficient matrices P , Q , and R are calculated for each total wavenumber n as follows:

$$P_n = \left\langle \Delta\eta_n^m (\Delta\phi_{B_n}^m)^T \right\rangle \left\langle \Delta\phi_{B_n}^m (\Delta\phi_{B_n}^m)^T \right\rangle^{-1}\quad (2.5.14)$$

$$Q_n = \left\langle \begin{pmatrix} \Delta T_n^m \\ \Delta P_{s_n}^m \end{pmatrix} (\Delta\phi_{B_n}^m)^T \right\rangle \left\langle \Delta\phi_{B_n}^m (\Delta\phi_{B_n}^m)^T \right\rangle^{-1}\quad (2.5.15)$$

$$R_n = \left\langle \left[\begin{pmatrix} \Delta T_n^m \\ \Delta P_{s_n}^m \end{pmatrix} - Q_n \Delta\phi_{B_n}^m \right] (\Delta\eta_{U_n}^m)^T \right\rangle \left\langle \Delta\eta_{U_n}^m (\Delta\eta_{U_n}^m)^T \right\rangle^{-1}\quad (2.5.16)$$

where $\langle \rangle$ denotes the statistical and zonal-wavenumber mean.

2.5.6.3 Background Error Covariance Matrix

The background error covariance matrices of the control variables are calculated for each total wavenumber n , and the matrix size is equivalent to the number of vertical levels for $\Delta\zeta$, $\Delta\eta_U$, and $\Delta \ln q$ or the number of vertical levels +1 for $(\Delta T_U, \Delta P_{sU})$.

$$B_{\zeta n} = \left\langle \Delta\zeta_n^m \overline{\Delta\zeta_n^m}^T \right\rangle, \quad B_{\eta_U n} = \left\langle \Delta\eta_{U_n}^m \overline{\Delta\eta_{U_n}^m}^T \right\rangle\quad (2.5.17)$$

³The mass variable $\Delta\Phi_k$ is defined as $\Delta\Phi_k = \Delta\phi_k + R_d \bar{T}_k \Delta p_k / \bar{p}$, where subscript k is the vertical level index, $\Delta\phi_k$ is the geopotential height, \bar{T}_k is the reference (global mean) temperature, \bar{p} is the reference (global mean) pressure at ground surface, Δp_k is the pressure, and R_d is the dry gas constant. In the calculation of $\Delta\phi_k$, \bar{T}_k and \bar{p} are also used and some approximation is done.

$$B_{\begin{pmatrix} T_U \\ p_{sU} \end{pmatrix}_n} = \left\langle \begin{pmatrix} \Delta T_{U_n^m} \\ \Delta p_{sU_n^m} \end{pmatrix} \overline{\begin{pmatrix} \Delta T_{U_n^m} \\ \Delta p_{sU_n^m} \end{pmatrix}^T} \right\rangle, \quad B_{\ln q_n} = \left\langle \Delta \ln q_n^m \overline{\Delta \ln q_n^m}^T \right\rangle \quad (2.5.18)$$

where $\langle \rangle$ denotes the statistical and zonal-wavenumber mean, and overline means complex conjugate. The total variances of the control variables are rescaled by a factor of 0.8836.

2.5.6.4 Cholesky Decomposition of Background Error Covariance Matrix

The background error covariance matrix mentioned above is decomposed using the Cholesky decomposition. This gives independent and normalized (i.e., preconditioned) control variables Δy_n^m as follows:

$$J_n^{(x)} = \sum_{m=-n}^n \frac{1}{2} (\overline{\Delta x_n^m})^T B_n^{-1} \Delta x_n^m = \sum_{m=-n}^n \frac{1}{2} (\overline{\Delta x_n^m})^T (L_n L_n^T)^{-1} \Delta x_n^m = \sum_{m=-n}^n \frac{1}{2} (\overline{\Delta y_n^m})^T \Delta y_n^m \quad (2.5.19)$$

$$\Delta y_n^m \equiv L_n^{-1} \Delta x_n^m \quad (2.5.20)$$

where $J_n^{(x)}$ is a background error term for the control variable x at the total wavenumber n , B_n is a background covariance matrix for x , and L_n is a lower triangular matrix.

In summary, normalized control variables $\Delta y_n^m(k)$ are completely independent and normalized based on background error variance. The background term of the cost function is simplified as a summation of the square of the normalized control variables.

2.5.7 Observation Terms

2.5.7.1 Observation Data

The assimilated observation types are shown in Table 2.1.1, and brief explanations for each data type as well as the quality control procedures are found in Section 2.2 and Section 2.3.

Observational data and related departures (observation minus first guess) are given with the location and time via the pre-analysis procedure. Reported surface pressure data from the station height and sea surface pressure data from surface observation are assimilated after conversion to the model surface height prior to assimilation. Scatterometer data are assimilated as wind data at the lowest model level, but are taken as wind data at 10 m above sea level. The zenith total delay from GNSS data are assimilated over land. Satellite radiance data from MW sounders, MW imagers, Hyperspectral IR sounders and CSRs are directly assimilated using the K matrix model of RTTOV-10 (Saunders *et al.* 2012). GNSS-RO data are assimilated in the form of bending angle at the tangent point using ROPP8 (Culverwell *et al.* 2015).

2.5.7.2 Observation Error

Observation errors (the diagonal part of the observation error covariance matrix) are estimated based on innovation statistics (Desroziers *et al.* 2005), and are summarized in Table 2.5.1 and Table 2.5.2. For ocean surface wind data from scatterometers, such errors are defined as those with values of 4 or 6 m/s. For bending angle data, errors are defined as a function of height only. The observation error threshold is 1% of the observed bending angle above 10 km, and varies linearly from 20% at 0 km to 1% at 10 km. For ground-based GNSS-ZTD data, the observation error is 20 mm. The error at an arbitrary reported pressure level is linearly interpolated in the logarithm of pressure ($\log(p)$). The cross correlations of observation errors (off the diagonal part of the observation error covariance matrix) are not considered explicitly in 4D-Var. To eliminate the cross-correlation term in the cost function, horizontally or vertically dense observations are thinned spatially in pre-analysis, and observation errors are inflated with predefined factors.

Table 2.5.1: Observation error tables used in operational global analysis for (a) conventional observation, (b) AMV, (c) AMSU-A, (d) ATMS, (e) SAPHIR, (f) MHS, (g) SSMIS, (h) GMI, (i) AMSR2 and (j) CSR from four geostationary satellites. P_s , u , v , T , RH and T_B denote surface pressure, zonal and meridional wind components, temperature, relative humidity and brightness temperature, respectively. “x” in (c) - (f) denotes that the channel is not used.

(a) conventional observation					(b) AMV	
Element	P_s (hPa)	u, v (m/s)	T (K)	RH (%)	Element	u, v (m/s)
Level(hPa)					Level(hPa)	
Surface	0.7				1,000	4.5
1,000		2.3	1.7	6.4	850	4.5
850		2.4	1.2	15.9	700	4.5
700		2.5	1.0	19.8	500	4.5
500		2.5	0.8	31.5	300	5.3
300		2.7	0.9	31.7	200	5.8
200		2.8	1.1	24.1	100	6.8
100		3.1	1.2	3.8	50	7.0
50		3.0	1.4	1.4	30	7.2
30		3.0	1.5	1.3	10	7.6
10		3.9	2.5	1.3	1	9.1
1		4.6	5.4	1.3	0.4	10.6
0.4		7.7	7.6	1.3	0.1	10.6
0.1		7.7	7.6	1.3		

(c) AMSU-A T_B (K)							(d) ATMS T_B (K)		(e) SAPHIR T_B (K)	
Satellite Channel	Aqua	Metop-A	Metop-B	NOAA-15	NOAA-18	NOAA-19	Satellite Channel	Suomi-NPP	Satellite Channel	Megha-Tropiques
4	x	0.45	x	0.45	0.45	0.45	6	0.3	1	13.5
5	x	0.3	x	0.3	x	0.3	7	0.3	2	11.25
6	x	0.3	0.3	x	0.3	0.3	8	0.3	3	11.25
7	x	x	0.3	0.3	0.3	0.3	9	0.3	4	11.25
8	0.3	x	0.3	0.3	x	x	18	9	5	9
9	0.3	0.3	0.3	0.3	x	0.3	19	11.25	6	9
10	0.3	0.3	0.3	0.45	0.3	0.3	20	13.5		
11	0.3	0.45	0.45	x	0.45	0.45	21	15.75		
12	0.64	0.64	0.64	0.64	0.64	0.64	22	18		
13	x	1.02	1.02	0.85	1.02	1.02				
14	x	2.83	2.83	x	2.63	2.63				

(f) MHS T_B (K)					(g) SSMIS T_B (K)			(h) GMI T_B (K)		(i) AMSR2 T_B (K)	
Satellite Channel	Metop-A	Metop-B	NOAA-18	NOAA-19	Satellite Channel	DMSP-F17	DMSP-F18	Satellite Channel	GPM	Satellite Channel	GCOM-W
3	18	18	18	x	9	18	18	3	7.2	7	6.4
4	13.5	13.5	13.5	13.5	10	18	18	5	10	9	10
5	9	9	9	9	11	27	27	6	7.2	11	6.4
					13	7.6	7.6	8	8.8	13	9.2
					14	10	10	12	13.5		
					16	8	8	13	9		
					17	8.8	8.8				

(j) CSR T_B (K)							
Satellite Central Wavelength (μm)	GOES-15	Satellite Central Wavelength (μm)	Meteosat-8	Satellite Central Wavelength (μm)	Meteosat-11	Satellite Central Wavelength (μm)	Himawari-8
6.55	1.5	6.25	1.5	6.25	1.5	6.25	1.5
		7.35	1.5	7.35	1.5	6.95	1.5
						7.35	1.5

Table 2.5.2: Observation error tables used in operational global analysis for hyperspectral IR sounders. (k) AIRS, (l) IASI and (m) CrIS.

(k) AIRS T_B (K)						(l) IASI T_B (K)								(m) CrIS T_B (K)		
Ch.	Aqua		Ch.	Ch.	Ch.	Ch.	Metop-A	Metop-B	Ch.	Metop-A	Metop-B	Ch.	Metop-A	Metop-B	Ch.	Suomi-NPP
6	1.59	144	1.29	226	1.02	32	1.35	1.23	217	0.90	0.84	332	1.02	0.99	27	0.90
7	1.62	145	1.38	227	0.96	38	1.32	1.17	219	0.93	0.87	335	0.84	0.75	31	0.75
15	1.59	150	1.38	232	1.02	44	1.29	1.14	224	0.90	0.84	345	1.44	1.44	37	0.60
20	1.56	151	1.38	239	0.93	49	1.23	1.11	226	0.93	0.84	347	0.90	0.84	51	0.54
21	1.59	156	1.26	250	1.02	51	1.26	1.14	230	0.93	0.84	350	0.93	0.90	69	0.45
22	1.56	157	1.32	251	0.96	55	1.20	1.08	232	0.90	0.84	354	0.84	0.75	73	0.42
27	1.56	162	1.29	252	0.99	57	1.23	1.11	237	0.90	0.84	356	0.90	0.87	75	0.45
28	1.56	168	1.26	253	0.96	61	1.17	1.05	239	0.90	0.81	360	0.87	0.75	79	0.45
39	1.53	169	1.17	256	1.08	63	1.20	1.08	243	0.96	0.93	363	0.96	0.93	80	0.45
40	1.53	173	1.14	257	1.05	85	1.11	0.99	246	0.87	0.78				81	0.42
51	1.50	174	1.20	261	1.11	109	1.11	0.99	249	0.90	0.81				83	0.45
68	1.44	175	1.20	262	1.08	116	1.11	0.99	252	0.87	0.81				85	0.48
69	1.47	179	1.14	267	1.11	122	1.08	0.96	259	0.84	0.78				87	0.51
71	1.44	180	1.14	272	1.11	128	1.08	0.96	262	0.87	0.78				88	0.51
92	1.44	185	1.17	295	1.23	135	1.05	0.93	265	0.84	0.78				93	0.54
93	1.44	186	1.17	299	1.02	141	1.05	0.93	267	0.87	0.78				96	0.51
98	1.44	190	1.11	1897	0.60	146	0.99	0.87	269	0.87	0.75				102	0.57
99	1.44	192	1.11	1901	0.54	148	1.02	0.90	275	1.02	0.99				106	0.57
104	1.47	193	1.08	1911	0.60	154	1.02	0.90	279	0.87	0.84				116	0.66
105	1.47	198	1.02	1917	0.54	159	0.99	0.87	282	0.84	0.75				123	0.75
110	1.47	201	1.05	1918	0.57	161	0.99	0.87	285	0.90	0.84				124	0.66
111	1.47	204	1.05	1921	0.72	167	0.99	0.90	294	0.87	0.78				125	0.72
116	1.47	207	1.11	1923	0.69	173	0.99	0.90	296	0.90	0.81				132	0.66
117	1.50	210	1.11	1924	0.66	180	0.99	0.90	306	0.87	0.78				136	0.63
123	1.41	213	1.02	1928	0.72	185	0.96	0.87	309	0.90	0.84				138	0.69
124	1.41	215	0.99			187	0.96	0.87	313	0.99	0.96				142	0.66
128	1.47	216	1.02			193	0.99	0.87	320	1.02	0.99				148	0.63
129	1.47	218	0.99			199	0.99	0.90	323	0.84	0.78					
138	1.32	221	0.99			205	0.96	0.87	326	1.02	0.96					
139	1.35	224	1.08			212	0.93	0.87	329	0.81	0.75					

2.5.7.3 Observation Operator

In 4D-Var, observation data at a given location and time are simulated using forecast variables for the surrounding grids in the nearest forecast hour with spatial inter/extrapolation and variable conversion. Observation operator involves these consecutive procedures. A fast radiative transfer model known as RTTOV-10 is used as the observation operator for satellite radiance data assimilation. ROPP is used as the observation operator for the assimilation of GNSS-RO bending angle data. These operators are provided as external libraries from EUMETSAT NWP-SAF and ROM SAF, respectively.

2.5.7.4 Variational Bias Correction

As mentioned in Subsection 2.3.4, variational bias correction (VarBC) is applied to satellite radiance data biases. In 4D-Var with VarBC, the observation operators are extended to include bias correction terms and the control (analysis) variables are extended to include bias correction (regression) coefficients. These coefficients are optimized as control variables.

The extended form of the cost function Eq. (2.5.1) is defined as follows.

$$J(\Delta z_0) = \frac{1}{2} \Delta x_0^T \mathbf{B}^{-1} \Delta x_0 + \frac{1}{2} \Delta \beta^T \mathbf{B}_\beta^{-1} \Delta \beta + \frac{1}{2} \sum_{i=1}^n \left(\mathbf{H}_i \Delta x_i + \sum_{j=0}^m \Delta \beta_j p_{i,j} - d_i \right)^T \mathbf{R}_i^{-1} \left(\mathbf{H}_i \Delta x_i + \sum_{j=0}^m \Delta \beta_j p_{i,j} - d_i \right) + J_C \quad (2.5.21)$$

where,

$$\Delta z_0 \equiv \left[\Delta x_0^T, \Delta \beta^T \right]^T, \quad \mathbf{B}_\beta \equiv \text{diag} \left(\frac{F_{inf}^2}{N_{var}}, \dots, \frac{F_{inf}^2}{N_{var}} \right), \quad N_{var} \equiv \begin{cases} \frac{N}{\log_{10} \frac{N}{N_0} + 1} & (N \geq N_0) \\ N_0 & (N < N_0) \end{cases}$$

Δz_0 represents extended increments, consisting of low resolution model variable increments Δx_0 and bias correction coefficient increment $\Delta \beta$, \mathbf{B}_β is the background error covariance matrix for the bias correction coefficient β , $p_{i,j}$ is predictors for bias correction, m is the number of predictors for all radiance observation types, F_{inf} is an arbitrarily defined inflation factor, N is the number of data and N_0 is the threshold for the valid number of data.

The second term on the right hand side of Eq. (2.5.21) is the background term for the bias correction coefficients, and $\sum_{j=0}^m \Delta \beta_j p_{i,j}$ is the bias correction term. This equation is used instead of Eq. (2.5.1) in 4D-Var.

2.6 Meso-scale Analysis

2.6.1 Introduction

Meso-scale Analysis (MA) produces initial conditions for the Meso-Scale Model (MSM, Subsection 3.5.1) every three hours with incorporation of information from observations into the model for better forecasting of weather phenomena with emphasis on high-impact events.

In March 2002, a four-dimensional variational (4D-Var) scheme was introduced as the data assimilation approach for MA (Ishikawa and Koizumi 2002) in place of the previous three-dimensional optimal interpolation (3D-OI) scheme to create the world's first operational limited-area 4D-Var system. Following the upgrade of the MSM forecast model to a non-hydrostatic type (JMA-NHM; Saito *et al.* 2006, 2007) in September 2004, the previous hydrostatic 4D-Var was replaced by a non-hydrostatic model-based 4D-Var system (known as JMA Nonhydrostatic model-based Variational Analysis Data Assimilation (JNoVA; Honda *et al.* 2005)) in April 2009. This development enabled MA to produce initial conditions more consistent with the upgraded MSM forecast model. A further upgrade of the model also took place in February 2017, replacing JMA-NHM with a newly developed non-hydrostatic model called ASUCA (Ishida *et al.* 2009, 2010; Hara *et al.* 2012) as described in Subsection 3.5.1. A 4D-Var system based on ASUCA is currently under development (Aranami *et al.* 2015), and the current MA is still based on JNoVA.

The domain of MA is the same as that of the MSM, covering Japan and its surrounding areas. Operation covering the current $4,080 \times 3,300$ km domain, extended from the previous $3,600 \times 2,880$ km, was introduced in March 2013 (Subsection 3.5.1).

Various observational data are used to help improve the accuracy of prediction for meso-scale weather events, including information from weather radars, satellite observations and ground-based GNSS (see Table 2.1.2). Thanks to the advanced data assimilation scheme of 4D-Var utilized with these data, MA produces initial conditions highly consistent with the balance of model equations.

2.6.2 Operational System

MA is performed using the JNoVA system and produces initial conditions for MSM forecasts every 3 hours (00, 03, 06, 09, 12, 15, 18 and 21 UTC). Figure 2.6.1 shows a schematic depiction of the MA process, which is as follows (the numbers correspond to those in Figure 2.6.1):

1. With the initial condition produced in the previous MA, the high-resolution (5 km) forecast model is run within the 3-hour data assimilation window to obtain the first guess.
2. Quality-control on observations (Section 2.3) and the calculation of their deviations from the first guess are conducted for preparing inputs to the next step.
3. The JNoVA 4D-Var analysis is performed to obtain the optimal model state by assimilating the observations in a low-resolution (15 km) space.
4. The low-resolution analysis increment is added to the high-resolution first guess via an interpolation process to prepare the initial condition for the next step.
5. Based on the initial condition set in the previous step, the high-resolution (5 km) forecast model is run within the data assimilation window to obtain the initial condition for the MSM.

In MA, the first and last steps in which the high-resolution forecast model is run are called the outer steps, and the step in which the JNoVA in the low-resolution space is executed is called the inner step. The forecast model used in the two outer steps is the JMA-NHM⁴. The analysis domain is shown in Figure 2.6.2 with a topographic map at a 5-km resolution as used in MA. The lateral boundary conditions are given by Global Spectral Model (GSM) forecasts, while the initial conditions of the first guess are taken from the previous MA (the 3-hour forecast in the last outer step). In other words, MA frames cycle analysis nested into the GSM.

The data assimilation window is set to 3 hours, and the end of the window corresponds to the analysis time. The cut-off time of input observation data for MA is 50 minutes after each analysis time. Observation data received by this time are distributed into four time slots by rounding off the observation time to the nearest hour (as represented by four star shapes under the curly braces in Figure 2.6.1). Accordingly, data observed within the period from 3.5 hours before to 0.5 hours after the analysis time are assimilated in the inner step.

As described previously, the JNoVA in the inner step is a data assimilation system based on the four-dimensional variational (4D-Var) method as detailed in Subsection 2.6.3. This approach is based on maximum likelihood estimation, and optimal values (i.e. analysis fields) are determined by minimizing the cost function (see Subsection 2.6.3.1 for details). This minimization procedure requires iterative calculation of the cost function and its gradient (about 32 times on average), which takes considerable computational resources. To reduce this burden, the operational JNoVA involves an incremental approach (Courtier *et al.* 1994) in which a model with a low-resolution relative to that used in the outer step is adopted in cost function minimization. Operational formulation of this minimization with the incremental approach is detailed in Subsection 2.6.3.1. The model used in the minimization process for the JNoVA is called the inner model, and its specifications are described in Subsection 2.6.3.3. The horizontal grid spacing is 5 km (817×661 grid points) with 48 vertical layers in the outer steps, whereas larger horizontal grid spacing of 15 km (273×221 grid points) with 38 vertical layers is used in the inner step. In daily operation, the calculation time of the inner step is about 21 minutes, and those of the outer steps are about 3 minutes each.

⁴The JMA-NHM has not been used as the MSM forecast model since February 2017 (Section 3.5), but MA is still based on the JMA-NHM.

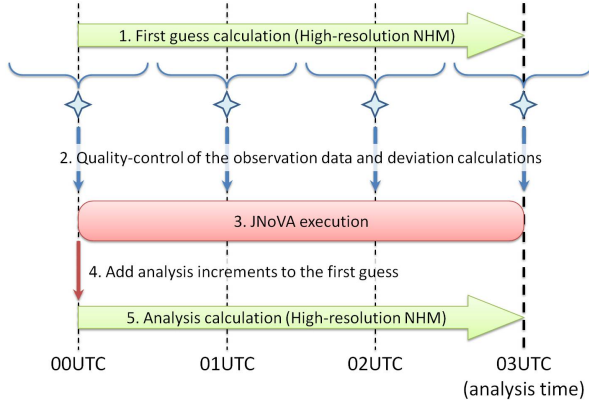


Figure 2.6.1: Schematic procedure of MA (example of 03 UTC analysis)

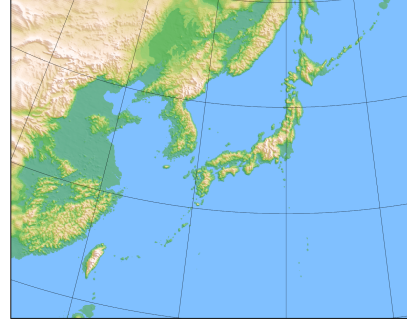


Figure 2.6.2: MA domain and topography

2.6.3 Basic Formulation

2.6.3.1 Cost Function

In the MA system, 4D-Var data assimilation is used to seek the optimal model trajectory in a phase space by minimizing its deviation from observations and the first guess. The deviation is measured using the cost function J , defined as

$$J(\mathbf{x}_0) = J_b + J_o + J_p = \frac{1}{2}(\mathbf{x}_0 - \mathbf{x}_0^b)^T \mathbf{B}^{-1}(\mathbf{x}_0 - \mathbf{x}_0^b) + \sum_{t=0}^N \frac{1}{2}(H_t(\mathbf{x}_t) - \mathbf{y}_t)^T \mathbf{R}_t^{-1}(H_t(\mathbf{x}_t) - \mathbf{y}_t) + J_p, \quad (2.6.1)$$

where the superscript T stands for transpose.

The first and the second terms of Eq. (2.6.1) are called the background and observation terms, and represent deviations from the first guess and the observations, respectively. \mathbf{x}_0 is the model state at the beginning of the data assimilation window (time level $t = 0$) to be optimized⁵, \mathbf{x}_0^b is the first guess of the model state at $t = 0$, \mathbf{y}_t is a column vector consisting of observational data available at t ($t = 0, \dots, N$), and \mathbf{x}_t is a model state at t as forecast from the initial condition \mathbf{x}_0 as

$$\mathbf{x}_t = M_t(\mathbf{x}_0), \quad (2.6.2)$$

where M_t denotes the forecast operator. H_t is an observation operator used to convert model state variables to observations, typically consisting of conversion from model variables to observed parameters and interpolation from model grid points to observation points. The error covariance matrixes \mathbf{B} and \mathbf{R}_t , specify the error profiles (uncertainty and error correlation) of \mathbf{x}_0^b and \mathbf{y}_t , respectively (see Subsection 2.6.3.2 and Subsection 2.6.4.2).

The third term of Eq. (2.6.1) J_p is a penalty term based on a digital filter to suppress high-frequency noise mainly caused by gravity waves (Gauthier and Thépaut 2001). It is given as

$$J_p = \frac{\lambda}{2} \left\| \delta \mathbf{x}_{N/2} - \delta \bar{\mathbf{x}}_{N/2} \right\|_E^2, \quad (2.6.3)$$

where λ is a weighting factor, $\delta \mathbf{x}_{N/2}$ is the model state analysis increment at the center of the data assimilation window ($t = N/2$), $\delta \bar{\mathbf{x}}_{N/2}$ is the digitally filtered analysis increment at $t = N/2$, and $\| \cdots \|_E$ is the moist total energy norm as proposed by Ehrendorfer *et al.* (1999).

⁵Lateral boundary conditions over the data assimilation window can be included in the vector to be optimized, \mathbf{x}_0 , but this is not adopted in the operational MA.

For a time series of model states over the data assimilation window $\{\mathbf{x}_0, \dots, \mathbf{x}_N\}$, the digitally filtered state $\bar{\mathbf{x}}$ at $t = N/2$ is given as

$$\bar{\mathbf{x}}_{N/2} = \sum_{k=0}^N h_{N/2-k} W_k \mathbf{x}_k, \quad (2.6.4)$$

where

$$h_k = \frac{\sin k\theta_c}{k\pi}, \quad (2.6.5)$$

is a low-pass filter that removes time oscillations exceeding the cutoff frequency θ_c . The Dolph-Chebyshev window function W_k (Lynch 1997) is also used to suppress the noise from the Fourier truncation (Gibbs oscillation).

In MA, the incremental approach is adopted to reduce the computational cost. Some implementations to reduce the computational cost of the 4D-Var scheme were proposed by Courtier *et al.* (1994). In MA, the stipulation of *Remark5 in Section3* of Courtier *et al.* (1994) is used⁶. In the incremental approach for MA, optimization is performed using an inner model (see Subsection 2.6.3.3) to obtain an analysis increment for the low-resolution model space (inner step). A low-resolution version of the analysis increment at $t = 0$, $\delta\mathbf{w}_0$ and the background error covariance \mathbf{B}_W are respectively given as

$$\delta\mathbf{w}_0 = \mathbf{S}(\mathbf{x}_0 - \mathbf{x}_0^b), \quad (2.6.6)$$

$$\mathbf{B}_W \approx \mathbf{S}\mathbf{B}\mathbf{S}^T, \quad (2.6.7)$$

where \mathbf{S} is an operator for conversion from the high to the low resolution. As a result, the low-resolution cost function can be expressed as

$$J(\delta\mathbf{w}_0) = \frac{1}{2} \delta\mathbf{w}_0^T \mathbf{B}_W^{-1} \delta\mathbf{w}_0 + \frac{1}{2} \sum_t [G_t(\mathbf{w}_t) - \hat{\mathbf{y}}_t]^T \mathbf{R}_t^{-1} [G_t(\mathbf{w}_t) - \hat{\mathbf{y}}_t] + J_p, \quad (2.6.8)$$

$$\mathbf{w}_t = L_t(\mathbf{w}_0), \quad (2.6.9)$$

where G_t is the low-resolution observation operator and L_t is the inner-model forecast operator. The observation vector \mathbf{y}_t is modified as

$$\hat{\mathbf{y}}_t = \mathbf{y}_t - H_t(\mathbf{x}_t^b) + G_t(\mathbf{w}_t^b), \quad (2.6.10)$$

for input of the high-resolution departures $\mathbf{y}_t - H_t(\mathbf{x}_t^b)$ to the inner step. The high-resolution analysis at $t = 0$ is given as

$$\mathbf{x}_0 = \mathbf{x}_0^b + \mathbf{S}^{-1} \delta\mathbf{w}_0, \quad (2.6.11)$$

where \mathbf{S}^{-1} is an operator for interpolation from low-resolution to the high-resolution model space. The final analysis \mathbf{x}_N is obtained by running forecast with the high resolution model over the data assimilation window (outer step).

$$\mathbf{x}_N = M_N(\mathbf{x}_0). \quad (2.6.12)$$

2.6.3.2 Background Error Covariance

As detailed previously, the background error covariance \mathbf{B} indicates the error profile of the first guess (Subsection 2.6.3.1). However, calculation using the complete form of \mathbf{B} is impractical due to the extremely large dimensions of the model state space. In practice, drastic simplification is applied to \mathbf{B} to make the problem tractable.

A group of parameters are defined as control variables, and their background errors are treated as being uncorrelated with each other. The control variables used in MA are as follows:

⁶This method involves the approximation of propagation in the time of perturbation ($\delta\mathbf{x}_0 = \mathbf{x}_0 - \mathbf{x}_0^b$) using the non-linear inner model in lower resolution by taking the finite difference of inner-model forecasts of observables from perturbed and unperturbed states ($\mathbf{w}_0 = \mathbf{w}_0^b + \delta\mathbf{w}_0$ and \mathbf{w}_0^b), or by taking $G_t(\mathbf{w}_t) - G_t(\mathbf{w}_t^b)$ as obtained from Eq. (2.6.9) and Eq. (2.6.10).

- u : x component of horizontal wind
- v : y component of horizontal wind
- (θ, p_s) : potential temperature and surface pressure
- $\tilde{q}_v = q_v/q_v^{bs}$: pseudo-relative humidity (q_v : specific humidity; q_v^{bs} : saturation specific humidity of first guess)

For each control variable (denoted by ϕ), the spatial structure of the background error covariance \mathbf{B}^ϕ is modeled as

$$\mathbf{B}^\phi = \mathbf{B}_v^{\phi 1/2} \mathbf{C}_h^{\phi 1/2} \mathbf{C}_h^{\phi 1/2 T} \mathbf{B}_v^{\phi 1/2 T}. \quad (2.6.13)$$

$\mathbf{B}_v^{\phi 1/2}$ is the square root of the vertical background error covariance $\mathbf{B}_v^\phi (= \mathbf{B}_v^{\phi 1/2} \mathbf{B}_v^{\phi 1/2 T})$, which is diagonal with respect to horizontal locations, and is assumed to be homogeneous over the domain, i.e., the matrix element $B_{v(i,j,k)(i',j',k')}^\phi$ corresponding to a pair of spatial points (i, j, k) and (i', j', k') is expressed as

$$B_{v(i,j,k)(i',j',k')}^\phi = \delta_{ii'} \delta_{jj'} B_{vkk'}^{\phi \text{col}}, \quad (2.6.14)$$

where (i, j) and (i', j') denote horizontal coordinates, and k and k' denote vertical levels. Accordingly, taking a single vertical column, the properties of $\mathbf{B}_v^{\phi \text{col}}$ are discussed here for simplicity. $\mathbf{B}_v^{\phi \text{col}}$ can be written as an eigenvalue decomposition

$$\mathbf{B}_v^{\phi \text{col}} = \mathbf{V} \mathbf{D} \mathbf{V}^T, \quad (2.6.15)$$

where the columns of \mathbf{V} are eigenvectors of $\mathbf{B}_v^{\phi \text{col}}$, and \mathbf{D} is the diagonal matrix of the eigenvalues. The form of $\mathbf{B}_v^{\phi \text{col} 1/2}$ as used in MA is given as

$$\mathbf{B}_v^{\phi \text{col} 1/2} = \mathbf{V} \mathbf{D}^{1/2} \mathbf{V}^T. \quad (2.6.16)$$

Thus, we obtain the matrix element of $\mathbf{B}_v^{\phi 1/2}$ as

$$B_{v(i,j,k)(i',j',k')}^{\phi 1/2} = \delta_{ii'} \delta_{jj'} B_{vkk'}^{\phi \text{col} 1/2}. \quad (2.6.17)$$

$\mathbf{C}_h^{\phi 1/2}$ is the square root of the horizontal background error correlation $\mathbf{C}_h^\phi (= \mathbf{C}_h^{\phi 1/2} \mathbf{C}_h^{\phi 1/2 T})$, which is diagonal with respect to the vertical levels. The matrix element of $\mathbf{C}_h^{\phi 1/2}$ is expressed as

$$C_{h(i,j,k)(i',j',k')}^\phi = \delta_{kk'} C_{h(i,j)(i',j')}^{\phi k}. \quad (2.6.18)$$

A Gaussian form is assumed for the horizontal correlation between points (i, j) and (i', j') for the vertical level k , i.e., the matrix element $C_{h(i,j)(i',j')}^{\phi k}$ is given as

$$C_{h(i,j)(i',j')}^{\phi k} = \exp\left[-\frac{1}{2}\left\{\left(\frac{i-i'}{\sigma_x^{\phi k}}\right)^2 + \left(\frac{j-j'}{\sigma_y^{\phi k}}\right)^2\right\}\right] = C_{hxii'}^{\phi k} C_{hyjj'}^{\phi k}. \quad (2.6.19)$$

The horizontal correlation length in the x and y directions $\sigma_x^{\phi k}$ and $\sigma_y^{\phi k}$ are also assumed to be homogeneous over the domain. In MA, $\mathbf{C}_h^{\phi 1/2}$ is given as

$$C_{h(i,j,k)(i',j',k')}^{\phi 1/2} = \delta_{kk'} C_{h(i,j)(i',j')}^{\phi k 1/2} = \delta_{kk'} C_{hxii'}^{\phi k 1/2} C_{hyjj'}^{\phi k 1/2}, \quad (2.6.20)$$

where $\mathbf{C}_{hx}^{\phi k 1/2}$ ($\mathbf{C}_{hy}^{\phi k 1/2}$) is taken as the Cholesky decomposition of $\mathbf{C}_{hx}^{\phi k}$ ($\mathbf{C}_{hy}^{\phi k}$), i.e., a lower triangular matrix.

The final background error covariance \mathbf{B} in the model state space is given as

$$\mathbf{B} = \mathbf{B}^{1/2} \mathbf{B}^{1/2 T}, \quad (2.6.21)$$

$$\mathbf{B}^{1/2} = \mathbf{K} \text{diag}(\mathbf{B}^{u 1/2}, \mathbf{B}^{v 1/2}, \mathbf{B}^{(\theta, p_s) 1/2}, \mathbf{B}^{\tilde{q}_v 1/2}), \quad (2.6.22)$$

where \mathbf{K} is a linearized transform from the control variable space to the model state space. In the transform \mathbf{K} , pressure is determined from the control variable (θ, p_s) assuming hydrostatic balance.

Based on this simplified form of \mathbf{B} , a variable transform is made from the analysis increment in the model state $\delta\mathbf{x}_0$ to a new variable λ , which is related to $\delta\mathbf{x}_0$ by $\mathbf{B}^{1/2}$

$$\delta\mathbf{x}_0 = \mathbf{x}_0 - \mathbf{x}_0^b = \mathbf{B}^{1/2}\lambda. \quad (2.6.23)$$

Variational optimization is performed with respect to λ thus obtained. This transform, called preconditioning, simplifies the background term of the cost function J (see Eq. (2.6.1) and Eq. (2.6.9); for simplicity, the present discussion does not deal with J_p). The cost function and its gradient after the transform are given as ⁷

$$J(\lambda) = \frac{1}{2}\lambda^T\lambda + \sum_{t=0}^N \frac{1}{2} \left(H_t(M_t(\mathbf{x}_0^b + \mathbf{B}^{1/2}\lambda)) - \mathbf{y}_t \right)^T \mathbf{R}_t^{-1} \left(H_t(M_t(\mathbf{x}_0^b + \mathbf{B}^{1/2}\lambda)) - \mathbf{y}_t \right), \quad (2.6.24)$$

$$\nabla_{\lambda} J = \lambda + \sum_{t=0}^N \mathbf{B}^{1/2T} \mathbf{M}_t^T \mathbf{H}_t^T \mathbf{R}_t^{-1} \left(H_t(M_t(\mathbf{x}_0^b + \mathbf{B}^{1/2}\lambda)) - \mathbf{y}_t \right), \quad (2.6.25)$$

where \mathbf{M}_t^T and \mathbf{H}_t^T are the adjoint model and the adjoint of the observation operator.

The parameters that characterize the error profile, \mathbf{B}_v^{ϕ} , $\sigma_x^{\phi k}$ and $\sigma_y^{\phi k}$, are estimated using the NMC method (Parrish and Derber 1992). Differences between 6h and 12h MSM forecasts valid at the same time are calculated for different cases and used as statistical samples of the background error. The samples are generated for the first ten days of each month from August 2014 to July 2015 with two pairs of MSM forecasts a day. The background error statistics \mathbf{B}_v^{ϕ} , $\sigma_x^{\phi k}$ and $\sigma_y^{\phi k}$ are taken as the averages over all these samples. MA involves the use of constant error statistics throughout the year without consideration of seasonal variations in error profiles.

2.6.3.3 Inner Model

In the JNoVA, a simplified nonlinear version of the JMA-NHM (NLM, M_t in Subsection 2.6.3.1 and Subsection 2.6.3.2) is used in the inner step to provide trajectories at every iteration instead of the tangent linear model (TLM) of the NLM due to the discontinuity and nonlinearity of the JMA-NHM. The adjoint model (ADM, \mathbf{M}_t^T in Subsection 2.6.3.2) of the NLM provides gradient information Eq. (2.6.25). The TLM has been developed only for use in the process of the ADM development. In addition to the use of a simplified model, the inner step of the JNoVA is executed using the NLM and ADM with a lower resolution (15 km horizontal grid spacing and 38 vertical layers) to reduce the computational cost. In this subsection, the specifications of the inner model in the JNoVA are surveyed focusing on schemes different from those of the JMA-NHM used in the outer step (outer-NLM). These are also briefly listed in Table 2.6.1.

The version of the JMA-NHM in the NLM and ADM differs from that in the outer-NLM. In order to develop the TLM and ADM, the version of the JMA-NHM was fixed when development started in 2002, and some improvements to the JMA-NHM were subsequently included in the JNoVA. However, only certain improvements were adopted due to the development costs of the TLM and ADM. The NLM in the JNoVA gives the basic fields of the ADM, and all the grid point values at each time step are saved within the memory upon operation. The forecast variables are momentum (three components), potential temperature, pressure and mixing ratio of water vapor. Additionally, temperature and evaporation efficiency on the land surface are predicted (those on the sea surface are assumed to be constant). The prognostic variables of the ADM are the same as those of the NLM except for evaporation efficiency.

In the NLM and ADM, only fully compressible elastic equations are supported as governing equations. The vertical coordinate is z^* -coordinate, which is different from hybrid coordinate of the outer-NLM. For this reason, the vertical interpolation from the outer step to the inner step is necessary. Regarding the advection scheme, the flux form fourth-order centered difference, same as the outer-NLM, is adopted, but the advection correction scheme is only used for the NLM. In the horizontally explicit, and vertically implicit (HE-VI) scheme of the NLM and ADM, gravity wave and sound wave are split and calculated in smaller time steps.

⁷Here, the resolution of the inner model is set to be the same as that of the outer model for simplicity. Actual operational implementation involves the use of two different resolutions based on the incremental approach discussed previously (see Eq. (2.6.9)). Formulation in line with operational implementation is obtained by making the replacements $(\delta\mathbf{x}_0, \mathbf{x}_0, \mathbf{x}_0^b, \mathbf{B}^{1/2}, \mathbf{y}_t, M_t, H_t) \rightarrow (\delta\mathbf{w}_0, \mathbf{w}_0, \mathbf{w}_0^b, \mathbf{B}_w^{1/2}, \hat{\mathbf{y}}_t, L_t, G_t)$ in Eq. (2.6.23), Eq. (2.6.24) and Eq. (2.6.25).

The smaller time step interval divides one larger time step (40 seconds) to seven small steps. In the JNoVA, all prognostic variables of the smaller time steps are reserved and used for the ADM integration. For the nonlinear computational diffusion,

$$D_{NL} \propto \frac{\partial}{\partial x} \left(\left| \frac{\partial \phi}{\partial x} \right| \frac{\partial \phi}{\partial x} \right)$$

is used in the outer-NLM and NLM(Nakamura (1978)), however, the perturbation of $\left| \frac{\partial \phi}{\partial x} \right|$ is not considered in the ADM. Targeted moisture diffusion is adopted in the NLM, but not available in the ADM. Except for the advection correction, the nonlinear computational diffusion and targeted moisture diffusion, the dynamical processes of the NLM are strictly linearized in the ADM.

For moist processes, the large-scale condensation (LSC) scheme for grid-scale precipitation is used in the NLM and ADM. As a sub-grid scale convective parameterization, while the NLM adopts the modified Kain-Fritsch scheme, the ADM does not consider.

In regard to turbulence scheme, the diagnostic-type Deardorff scheme (Deardorff 1980) is used in the NLM and ADM. As perturbations of turbulent kinetic energy are not considered in the ADM, there is no perturbation for the diffusive coefficients.

For a surface process in the NLM and ADM, the approaches proposed by Louis *et al.* (1982) for the land surface and Kondo (1975) for the sea surface (as previously used in the JMA-NHM) are adopted. In the both schemes of the ADM, the perturbations of the bulk coefficients are not considered.

The NLM and ADM also use the four-layer heat diffusive model for ground temperature as with the outer-NLM, but the perturbation is considered only at the highest layer which is close to the lowest layer of the atmosphere. Evaporation efficiency is given by the climate value in the NLM and ADM for simplicity.

The scheme for the radiation process in the NLM is slightly different from that in the outer-NLM. The cloud diagnosis is based on relative humidity rather than partial condensation scheme. The method to evaluate the effective radius of a cloud ice particle in the NLM is based on Ou and Liou (1995) without the modification proposed by McFarquhar *et al.* (2003). Additionally, the approach of Räisänen (1998) is adopted to calculate long wave radiation at each cloud layer in the outer-NLM, but not in the NLM. In the ADM, consideration of the radiation process is omitted for simplicity.

Table 2.6.1: Specifications of the outer-NLM and the models employed in the JNoVA

	outer-NLM	NLM	ADM
Resolution	5km, 48layers	15km, 38layers	15km, 38layers
Horizontal advection	Flux form fourth-order with advection correction	Flux form fourth-order with advection correction	Flux form fourth-order
Pressure equation solver	HE-VI	HE-VI	HE-VI
Targeted moisture diffusion	Considered	Considered	Not considered
Moist physics	3-ice bulk microphysics	LSC	LSC
Convection	Modified Kain-Fritsch	Modified Kain-Fritsch	None
Turbulence	Mellor-Yamada-Nakanishi-Niino level-three	Deardorff	Deardorff
Surface flux	Beljaars and Holtslag	Louis(land) and Kondo(sea)	Louis(land) and Kondo(sea)
Ground temperature	4-layer thermal diffusion	4-layer thermal diffusion	4-layer thermal diffusion
Radiation	Considered	Considered	Not considered

Table 2.6.2: Observation error tables used in the operational meso-scale analysis for (a) conventional observation and wind profiler data and (b) AMV. P_s , u , v , T and RH denote surface pressure, x and y wind components in an MSM Lambert projection space, temperature and relative humidity respectively.

(a) Conventional observation and wind profiler data						(b) AMV		
Element	P_s (hPa)	u (m/s)	v (m/s)	T (K)	RH (%)	Element	u (m/s)	v (m/s)
Level (hPa)						Level (hPa)		
Surface	0.7					1,000	4.1	3.3
	0.6 for SYNOP in Japan					850	2.9	2.3
1,000		2.1	1.9	1.3	9.8	700	3.2	2.6
925		2.0	1.9	0.9	10.3	500	3.7	3.0
850		2.0	2.0	0.9	12.7	300	4.6	3.7
700		2.0	1.9	0.9	12.8	200	3.8	4.9
500		1.9	1.9	0.7	12.9	100	4.4	6.0
400		2.2	2.2	0.7	13.3	50	3.5	5.1
300		2.6	2.6	0.9	13.5	30	5.1	6.2
250		2.7	2.6	1.0	14.4	10	6.2	7.2
200		2.7	2.6	1.1	13.7			
150		2.6	2.6	1.1	16.6			
100		3.2	3.0	1.5	15.1			
70		3.7	3.1	1.9	13.6			
50		3.2	2.8	1.9	12.1			
30		3.2	2.8	1.9	11.8			
10		3.2	2.8	1.9	12.2			

2.6.4 Observation Terms

2.6.4.1 Observation Data

Assimilated observation types are shown in Table 2.1.2, and brief outlines of each data type and related quality control procedures are given in Section 2.2 and Section 2.3.

2.6.4.2 Observation Error

The observation error covariance matrix \mathbf{R} in Eq. (2.6.1) is assumed to be diagonal, and cross-correlation between different observations is not considered. Estimation of observation errors (diagonal components of \mathbf{R}) is based on innovation statistics (Desroziers *et al.* 2005). Errors for conventional observations, wind profiler data and AMVs are summarized in Table 2.6.2. Errors for satellite radiance are the same as those in global analysis (See Table 2.5.1(c) - (j)). The errors for GNSS-PWV and radial velocity are 3 mm and 3.3 m/s, respectively. Errors for relative humidity data from DPR are 5%, and those from ground-based radar are approximately 30%. Errors for ocean surface wind data from scatterometers are 3 m/s. For GNSS-RO refractivity data, observation errors are defined as a function of height only. The observation error is calculated using a linear interpolation to be 10% of refractivity at 0 km, 0.5% at 10 km, 0.18% at 20 km and 0.04% at 30 km. Errors for R/A are based on the precipitation amount (Koizumi *et al.* 2005). The error at an arbitrary reported pressure level is linearly interpolated in the logarithm of pressure ($\log(p)$). The cross-correlations of errors between different observations are not considered explicitly in 4D-Var. To eliminate consideration of cross-correlation terms in the cost function, dense observations are thinned spatially and observation errors are inflated in the pre-analysis procedure.

2.6.4.3 Observation Operator

The RTTOV-10 fast radiative transfer model is used as the observation operator for satellite radiance data assimilation. ROPP8 is used as the observation operator for the assimilation of refractivity data from GNSS-

RO. These observation operators are provided as external libraries from EUMETSAT NWP-SAF and ROM SAF, respectively.

2.6.4.4 Special Treatment for Precipitation Data

For the observation terms of the cost function Eq. (2.6.1), it is assumed that the probability density function (PDF) for observation errors is Gaussian. However, PDF for precipitation amount data does not follow a Gaussian function. Accordingly, the following observation term is used for one-hour precipitation amount data (Koizumi *et al.* 2005).

$$J_o^{PREC}(x) = \sum_{j \text{ (where } r_j^o \geq 0.5)}^n \frac{(H_j(x) - r_j^o)^2}{2\sigma_o(r_j^o)^2} \quad (2.6.26)$$

where, $H_j(x)$ is an observation operator used to convert the state variables x to one-hour cumulative precipitation values at the j -th grid point, r_j^o is the precipitation observed at the grid point and n is the number of grid points in the inner model. $\sigma_o(r_j^o)$ is the observation error which is defined as follows:

$$\sigma_o(r_j^o) \equiv \begin{cases} C_{sat} \max(r_{min}, r_j^o) & (H_j(x) \leq r_j^o) \\ C_{sat} C_a \max(r_{min}, r_j^o) & (H_j(x) > r_j^o) \end{cases}, \begin{cases} C_a = 3, C_{sat} = 1 & \text{for R/A} \\ C_a = 5, C_{sat} = 2 & \text{for satellite retrievals} \end{cases}, r_{min} \equiv 1 \text{mm/h} \quad (2.6.27)$$

where, C_{sat} is an observation error inflation factor for satellite retrievals and C_a is a tuning factor for the asymmetric structure of the departure frequency distribution around 0.

One-hour precipitation observation showing values of less than 0.5 mm are not assimilated, since the quality of such data is rather poor for snowfall. The observation error of satellite retrievals is considered to be larger than that of R/A because retrieval is from instantaneous observations rather than from one-hour cumulative observations.

2.6.4.5 Variational Quality Control

Variational quality control (VarQC, Andersson and Järvinen 1999) is applied in 4D-Var for conventional observations. With VarQC, the PDF of the observation error is assumed to be a summation of a Gaussian function and a positive constant value in the certain range. The constant value means the probability of rough error within the range.

The following observation term and its gradient are used for conventional observations in the cost function Eq. (2.6.1) in 4D-Var with VarQC.

$$j_o^{VarQC} = -\log\left(\frac{\gamma + \exp(-j_o)}{\gamma + 1}\right), \quad \gamma \equiv \frac{A \sqrt{2\pi}}{(1-A)2d} \quad (2.6.28)$$

$$\nabla j_o^{VarQC} = W^{VarQC} \nabla j_o, \quad W^{VarQC} \equiv 1 - \frac{\gamma}{\gamma + \exp(-j_o)} \quad (2.6.29)$$

where, A is the prior probability of rough error (e.g. $A = 0.05$ for SYNOP), d is the maximum standard deviations below which rough error is possible (e.g. $d = 9$ for SYNOP), j_o^{VarQC} is the observation term for a single observation component with VarQC and j_o is the term without VarQC.

Eq. (2.6.29) shows that ∇j_o^{VarQC} is almost the same (effective) as ∇j_o when j_o is small ($W^{VarQC} \approx 1$) and ∇j_o^{VarQC} is almost 0 (not effective) when j_o is large ($W^{VarQC} \ll 1$). Observations satisfying $W^{VarQC} < 0.25$ are regarded to be rejected by the VarQC.

2.7 Local Analysis

2.7.1 Introduction

Local Analysis (LA) produces initial conditions for the Local Forecast Model (LFM) (Subsection 3.6.1) at a horizontal resolution of 2 km. Its operation started in August 2012, with eight runs per day on an area of Japan measuring $2,200 \times 2,500$ km to initialize LFM forecasts over a domain covering the eastern part of the country. An enhancement in the operation of LA, along with the LFM, was implemented in May 2013, extending its domain to cover Japan and the surrounding areas ($3,160 \times 2,600$ km) and increasing its daily operations to 24 runs per day.

To provide initial conditions for this high-resolution forecast model targeting small-scale severe weather events, LA is designed to allow rapid production and frequent updating of analysis at a resolution of 5 km (Subsection 2.7.2). In each LA run, an analysis cycle with hourly three-dimensional variational (3D-Var) data assimilation is executed for the previous three hours to incorporate information from newly received observational data in each case. The analysis cycle was originally based on JMA-NHM (Saito *et al.* 2006, 2007) and JNoVA 3D-Var (the 3D-Var version of JNoVA (Honda *et al.* 2005)), which was replaced by the new-generation version based on ASUCA (Ishida *et al.* 2009, 2010; Hara *et al.* 2012) and ASUCA-3DVar in January 2015 (Aranami *et al.* 2015).

As with MA, high-density remote sensing data (including information from weather radars and ground-based GNSS) are assimilated on an hourly basis in LA as important sources of detailed information that can contribute to better forecasting of high-impact phenomena (see Table 2.1.3). LA was also adopted to make extensive use of satellite observations in January 2017, with application of the variational bias correction technique. The capacity of high-resolution NWP to capture small-scale variations in topography is expected to help to reduce representativeness errors in the assimilation of surface observations. In association, LA also assimilates automated surface station (AMeDAS) data ahead of other operational data assimilation systems with lower resolutions in order to appropriately reflect the effects of local-scale environments near the surface.

Based on these features, LA is characterized as a data assimilation system for high-resolution and high-frequency NWP.

2.7.2 Operational System

To satisfy the requirements outlined in Subsection 2.7.1, operational LA incorporates an analysis cycle with 3D-Var, which can meet the relevant demand for prompt and frequent product updates within a limited time frame (all processes, including data quality control, are completed within around 15 minutes) with far fewer computational resource requirements than 4D-Var.

LA involves the running of an analysis cycle on a domain identical to that of the LFM (see Figure 2.7.1), following the flow chart shown in Figure 2.7.2. The cycle consists of four successive instances of 3D-Var analysis (3, 2, 1 and 0 hours prior to the initial time of the LFM; (a), (b), (c) and (d) in Figure 2.7.2), where quality controlled observation data rounded to the nearest hour are assimilated. After each instance of analysis except the last one ((d) in Figure 2.7.2), a one-hour forecast (LF1) with a horizontal resolution of 5 km is executed using the analysis for initial conditions, thereby providing the first guess for the next 3D-Var analysis. In the first 3D-Var ((a) in Figure 2.7.2), the first guess is supplied from the MSM (Subsection 3.6.2), which also provides lateral and upper boundary conditions for LF1 throughout the three-hour data assimilation period. The analyzed field in the last 3D-Var ((d) in Figure 2.7.2) is used as the initial condition of the LFM. Further specifications of LA are provided in Table 2.1.3.

2.7.3 Basic Formulation

2.7.3.1 Cost Function

Local Analysis involves four time slots for the assimilation of observations using 3D-Var. The related cost function in the i -th time slot is defined as

$$J(\delta\mathbf{x}_i) = \frac{1}{2} \delta\mathbf{x}_i^T \mathbf{B}^{-1} \delta\mathbf{x}_i \quad (2.7.1)$$



Figure 2.7.1: domains of LA and LFM.

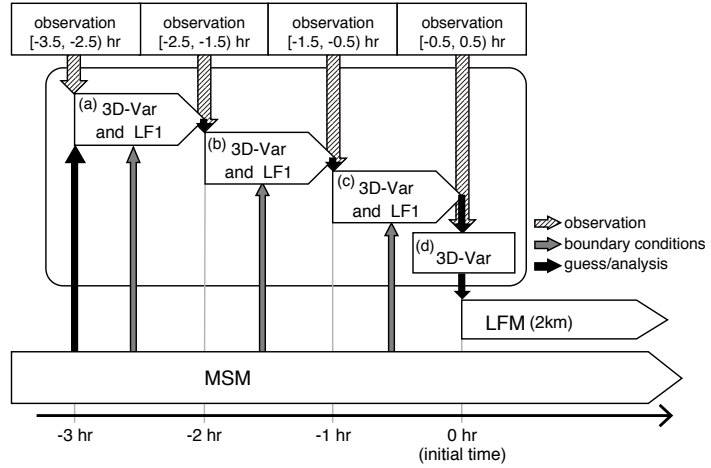


Figure 2.7.2: Schematic representation of LA analysis.

$$+\frac{1}{2} (\mathbf{H}\delta\mathbf{x}_i - \mathbf{d}_i + \mathbf{P}\delta\mathbf{b}_i)^T \mathbf{R}^{-1} (\mathbf{H}\delta\mathbf{x}_i - \mathbf{d}_i + \mathbf{P}\delta\mathbf{b}_i) \quad (2.7.2)$$

$$+\frac{1}{2} \delta\mathbf{b}_i^T \mathbf{S}_i^{-1} \delta\mathbf{b}_i \quad (2.7.3)$$

Here $\delta\mathbf{x}_i$ is the increment, $\delta\mathbf{b}_i$ is the increment of the observation bias vector, and the subscript i is the hourly time-index. The innovation vector \mathbf{d}_i is given by

$$\mathbf{d}_i = \mathbf{y}_i - \mathcal{H}(\mathbf{x}_i^b) - \mathcal{P}(\mathbf{b}_i^b), \quad (2.7.4)$$

where \mathbf{y}_i is the vector of observations, \mathbf{x}_i^b is the background state and \mathbf{b}_i^b is the background state of observation bias in the i -th time slot. The other symbols are as follows: \mathcal{H} is the nonlinear observation operator, \mathbf{H} is the tangent linearized observation operator, \mathcal{P} is the nonlinear operator of predictors for VarBC, \mathbf{P} is the matrix of predictors for VarBC, \mathbf{B} is the covariance matrix of background error, \mathbf{R} is the diagonal matrix of observation error and \mathbf{S}_i is the diagonal matrix of parameters used to control the adaptivity of observation bias vector estimation in VarBC.

2.7.3.2 Solution Procedure

The analysis \mathbf{x}^a at the initial time t of LFM, corresponding to $i = 4$, is calculated by repeating 3D-Var and one-hour forecasting. The one-hour forecast operator $\mathcal{M}_{i,i+1}$ is configured specifically for Local Analysis. In this configuration, the horizontal resolution is set to 5 km as in the MSM, but the physics schemes differ from those used in the MSM and the LFM.

The Local Analysis cycle is conducted as outlined below.

1. $\mathbf{x}_{i=1}^b$ valid at $t - 3\text{h}$ is provided by the MSM, and $\mathbf{b}_{i=1}^b$ is equal to the analysis variables of observation bias \mathbf{b}^a in the previous Local Analysis.
2. $\delta\mathbf{x}_i$ is optimized by minimization of the cost function, $J(\delta\mathbf{x}_i)$.
3. The i -th background state is updated with the one-hour forecast:

$$\mathbf{x}_{i+1}^b = \mathcal{M}_{i,i+1}(\mathbf{x}_i^b + \delta\mathbf{x}_i), \quad (2.7.5)$$

and the observation bias is given by

$$\mathbf{b}_{i+1}^b = \mathbf{b}_i^b + \delta \mathbf{b}_i. \quad (2.7.6)$$

4. Steps 2 and 3 are repeated three times.

5. For the initial time, the cost function $J(\delta \mathbf{x}_{i=4})$ is minimized and the analysis increments are added to the background state. The low-resolution (5 km) analysis variables are given by

$$\mathbf{x}^a = \mathbf{x}_4^b + \delta \mathbf{x}_4, \quad (2.7.7)$$

and the analysis variables of observation bias \mathbf{b}^a are given by

$$\mathbf{b}^a = \mathbf{b}_4^b + \delta \mathbf{b}_4. \quad (2.7.8)$$

Finally, the low-resolution analysis \mathbf{x}^a is interpolated to the high resolution of 2 km with consideration of ancillaries (i.e. topography, soil type and land usage) to be used as the initial condition with the LFM.

2.7.3.3 Analysis variables

The analysis increment is defined as $\delta \mathbf{x} = (\delta u, \delta v, (\delta T_g, \delta p_s, \delta \theta), (\delta W_g, \delta \mu_p))^T$.

- u : x -component of horizontal wind.
- v : y -component of horizontal wind.
- (T_g, p_s, θ) : underground temperature and skin temperature, surface pressure and potential temperature.
- (W_g, μ_p) : soil moisture and pseudo-relative humidity.

T_g is perturbed only on land grids, as that on ocean grids is not a prognostic variable in the LFM.

2.7.3.4 Background Error Covariance

$\delta \mathbf{x}$ is transformed from uncorrelated control variables χ , as follows

$$\delta \mathbf{x} = \mathbf{B}^{1/2} \chi, \quad (2.7.9)$$

$\mathbf{B}^{1/2}$ is the square root of \mathbf{B} and has the form

$$\mathbf{B}^{1/2} = \mathbf{K}_p \mathbf{B}_s^{1/2} \quad (2.7.10)$$

$$= \mathbf{K}_p \text{diag} \left(\mathbf{B}_{s,u}^{1/2}, \mathbf{B}_{s,v}^{1/2}, \mathbf{B}_{s,(T_g,p_s,\theta)}^{1/2}, \mathbf{B}_{s,(W_g,\mu_p)}^{1/2} \right) \quad (2.7.11)$$

Here \mathbf{K}_p is a linearized parameter transform from the control variables to the model variables, $\mathbf{B}_s^{1/2}$ is a spatial transform and $\mathbf{B}_{s,\chi_i}^{1/2}$ is the spatial transform for the i th sub-group χ_i of parameters in χ . The spatial structure of background error covariance \mathbf{B}_{s,χ_i} is modeled as

$$\mathbf{B}_{s,\chi_i}^{1/2} = \mathbf{C}_v \mathbf{B}_{v,\chi_i}^{1/2} \mathbf{B}_{h,\chi_i}^{1/2} \quad (2.7.12)$$

Here, \mathbf{C}_v denotes a vertical coordinate transformation introduced to limit the terrain effect of the vertical coordinate within the lower troposphere.

The vertical background error covariance matrix \mathbf{B}_{v,χ_i} is assumed to be horizontally homogeneous over the domain. The square root of \mathbf{B}_{v,χ_i} is given by

$$\mathbf{B}_{v,\chi_i}^{1/2} = \mathbf{V} \mathbf{\Lambda}^{1/2} \mathbf{V}^T, \quad (2.7.13)$$

where $\mathbf{\Lambda}$ is a diagonal matrix whose elements are the eigenvalues of \mathbf{B}_{v,χ_i} , and \mathbf{V} is the orthogonal matrix ($\mathbf{V}^T \mathbf{V} = \mathbf{I}$) whose columns are the related eigenvectors.

The horizontal background error correlation \mathbf{B}_{h,χ_i} is defined on the vertical level of the model. In the calculation of $\mathbf{B}_{h,\chi_i}^{1/2}$, the recursive filter technique (Purser *et al.* 2003) is used in each of the x - and y -directions.

The background error statistics \mathbf{B}_{v,χ_i} and \mathbf{B}_{h,χ_i} are estimated using the NMC method (Parrish and Derber 1992). However, the error profiles in the lower levels are modified artificially to localize spatial correlations so that surface observations are assimilated more appropriately. The seasonal variation of background error statistics is not taken into account.

2.7.4 Observation Terms

2.7.4.1 Observation Data

Assimilated observation types and brief outlines of each data type are provided in Table 2.1.3.

2.7.4.2 Observation Error

The observation error covariance matrix is assumed to be diagonal, and cross-correlation between different observations is not considered as it is in MA.

2.7.4.3 Observation Operators

The observation operator for surface observations (i.e. wind at a height of 10 m and temperature, relative humidity at a height of 1.5 m) is based on the surface diagnostic scheme (Beljaars and Holtslag 1991) in the LFM. In the tangent linear operator for surface observation, perturbations of transfer coefficients for surface fluxes are not considered in diagnostic equations.

The observation operator for brightness temperature is RTTOV10.2 (Saunders *et al.* 2012), but perturbations of surface elements and ozone are ignored as inputs to the RTTOV tangent-linear model.

The observation operator for soil moisture is provided using a linear regression equation with coefficients estimated from cumulative distribution function matching.

2.7.5 Parallelization

The data assimilation domain is two-dimensionally decomposed into blocks, each of which is assigned to an MPI process. The loops for the y-direction and certain fused horizontal loops are forked via OMP parallelization similar to that of the LFM (see Section 3.5.10). Observations are distributed to blocks according to their location, and innovations are calculated in each process. In addition, the recursive filter method is parallelized using decomposed blocks.

In calculation of minimization, the inner product of general vectors in the L-BFGS (Nocedal and Wright 2006) is determined for each block, and loops for observations and model variables are forked via OMP parallelization.

2.8 Snow Depth Analysis

2.8.1 Global Snow Depth Analysis

Global snow depth analysis is executed every day separately from global atmosphere analysis. Global snow depth data with a 1.0° latitude/longitude resolution are analyzed using SYNOP snow depth data on the same day. Analysis involves two-dimensional optimal interpolation (2D-OI). Analysis snow depth data are interpolated into GSM model grids (TL959) that are converted to snow water equivalents as an initial condition for the land-surface process (see Subsection 3.2.9) in the GSM.

The first guess is produced for 2D-OI as follows:

$$G = C + \frac{1}{2}A_C \quad (2.8.1)$$

where G is the first guess, C is the climatological value, and A_C is the analysis anomaly from the climatological value for the previous day. C is interpolated to the analysis day from monthly climatological data. The monthly climatological data used from September to June also come from the climatology compiled by USAF/ETAC (Foster and Davy 1988), and those for July and August are interpolated from the same climatology for June and September.

Snow depth analysis excludes grids with evergreen broadleaves as the tree vegetation type and those with no climatological potential for snow cover as estimated from past SSM/I observation statistics. The snow

depths of these grids are set to 0 cm. Grids covering land ice are also excluded from such analysis. The snow depths of these grids are set to C , the climatological values.

2.8.2 Mesoscale Snow Depth Analysis

In the Meso Scale Model (MSM, Section 3.5), snow cover data are used to gauge the status of ground snow, with depths exceeding 5 cm in individual grid squares being classified as snow-covered. Snow depth analysis data are produced via a two-dimensional OI (2D-OI) in the high-resolution snow depth analysis system. The first guesses for 2D-OI are set using an offline version of the land surface model (LSM) with the same domain and grid spacing as the MSM. The offline LSM and the 2D-OI are outlined below.

The offline LSM, which includes a multi-layer snowpack model, simulates typical snow processes such as accumulation, compaction and ablation. The atmospheric forcing data necessary to drive the LSM are air temperature and wind velocity at the lowest atmospheric model level and radiative fluxes toward the surface as predicted by the MSM. Radar/Raingauge Analyzed Precipitation data (see Subsection 4.4.1) are used as rain and snowfall inputs to the LSM.

The model first guesses and observations reported from SYNOP and AMeDAS stations are handed over to the 2D-OI system in the snow depth analysis system. The methodology of OI is based on Brasnett (1999), where the correlation coefficient, μ_{kl} , is given by

$$\mu_{kl} = \alpha(r_{kl})\beta(\Delta z_{kl}) \quad (2.8.2)$$

with the horizontal and vertical separation, r_{kl} and Δz_{kl} between points k and l . $\alpha(r_{kl})$ and $\beta(\Delta z_{kl})$ are the horizontal and vertical structure functions:

$$\alpha(r_{kl}) = \left(1 + \frac{r_{kl}}{L}\right) \exp\left(-\frac{r_{kl}}{L}\right), \quad (2.8.3)$$

$$\beta(\Delta z_{kl}) = \exp\left\{-\left(\frac{\Delta z_{kl}}{h}\right)^2\right\}, \quad (2.8.4)$$

where L and h are set to 25 km and 500 m, respectively. The standard deviations of observation and background errors are set at 4 and 3 cm, respectively.

2.9 Non-real-time Quality Control

2.9.1 GDPFS-RSMC Operational Activities

JMA is designated as a Regional Specialized Meteorological Center (RSMC) of the World Meteorological Organization (WMO) Global Data-processing and Forecast System (GDPFS), and is known in this role as RSMC Tokyo. In March 1991, WMO Commission for Basic Systems (CBS) designated RSMC Tokyo as a lead center for monitoring the quality of land surface observations in Region II (Asia). As a part of its operational activities, JMA produces a six-monthly report containing a consolidated list of stations suspected of reporting low-quality observation data on station level pressure, mean sea level pressure and geopotential height during the six-month periods ending June and December. This report is available on JMA's website⁸.

RSMC Tokyo also produces monthly statistics on the quality of all observations received in time for use in its final global analyses. Copies of these reports are provided to major GDPFS centers and to the WMO Secretariat. The reports are also available on JMA's website⁹.

Data quality evaluation is based on differences between observations and first guess fields (three to nine-hour forecasts) from the global model. Standard procedures and formats for the exchange of monitoring results are given in the Manual on GDPFS (WMO-No.485).

⁸<http://qc.kishou.go.jp/clsf.html>

⁹<http://qc.kishou.go.jp/mmr.html>

2.9.2 WDQMS Operational Activities

The WMO Integrated Global Observing System (WIGOS) is a framework for all WMO observing systems and WMO contributions to co-sponsored observing systems in support of all WMO Programmes and activities. At WIGOS workshops on Quality Monitoring and Incident Management held in December 2014 and December 2015, plans were developed for a WIGOS Data Quality Monitoring System (WDQMS). The Task Team on WDQMS (TT-WDQMS) is working to develop the WDQMS under the Inter-Commission Coordination Group on WIGOS. Four NWP centers (European Centre for Medium-Range Weather Forecasts (ECMWF), National Centers for Environmental Prediction (NCEP), JMA and Deutscher Wetterdienst (DWD)) contribute to the NWP Quality Monitoring Pilot Project on WDQMS, providing monitoring output in near-real time to the WMO Secretariat. Contributions began with surface pressure data, and now also include information on surface humidity, wind, temperature and upper-air soundings.

2.9.3 Blacklist Management

As mentioned in Section 2.3, low quality observational data can result in significant forecast degradation. The cause of low quality may be instrumental failure, which can continue for a long time. Such observation data should be excluded in the first step of QC, and a blacklist is kept to meet this need. Blacklist management is one of the most important activities in QC. The quality of all observations is evaluated based on differences between observations and first guess fields from the global model (three to nine-hour forecasts), the meso-scale model (zero to three-hour forecasts) and the local forecast model (one-hour forecasts). Providers of problematic observation data are added to the blacklist.

2.10 Climate Data Assimilation System

For climate system monitoring and seasonal prediction, it is essential to comprehend both current and past climate conditions accurately and consistently over a long period of time. To this end, a high-quality, spatio-temporally homogeneous long-term climate dataset based on reanalysis of past observations using a state-of-the-art NWP technique is maintained.

JMA conducted its second global atmospheric reanalysis (the Japanese 55-year Reanalysis, or JRA-55; Kobayashi *et al.* 2015) covering the period from 1958, when regular radiosonde observations began on a global basis. Production is continued to the present on a near-real-time basis as an operational climate data assimilation system, thereby supporting real-time climate monitoring and seasonal prediction. JRA-55 data are also used in various research and development activities at JMA and in wider communities of fundamental academic fields such as meteorology, climatology and oceanography, practical application fields such as agricultural meteorology and renewable energy, and other areas.

The data assimilation system used to produce JRA-55 data is described in Kobayashi *et al.* (2015). The system is based on the TL319 version of JMA's operational data assimilation system as of December 2009 (JMA 2007, 2013), and features numerous improvements made since Japan's first global atmospheric reanalysis (the Japanese 25-year Reanalysis, or JRA-25; Onogi *et al.* 2007). These include a revised longwave radiation scheme, a 4D-Var data assimilation system and a variational bias correction scheme for satellite radiances. The system also incorporates several newly available observational datasets produced as a result of ongoing efforts to improve the quality of past observation data, including homogenization of radiosonde temperature observations (Haimberger *et al.* 2008, 2012) and reprocessing of satellite data at major meteorological satellite centers (e.g. van de Berg *et al.* 2002; Oyama 2010).

These improvements make JRA-55 products considerably superior to JRA-25 products. Two major issues with JRA-25 (a cold bias in the lower stratosphere and a dry bias in the Amazon basin) are mitigated in JRA-55, and the temporal consistency of temperature analysis is considerably better than in previous reanalysis products.

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)

System	
Model (version)	Global Spectral Model (GSM1705)
Date of implementation	December 1987 (Lastest version: 25 May 2017)
Configuration	
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
Initial conditions	
Data assimilation	Four-dimensional variational (4D-Var) method
Surface boundary conditions	
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
Other details	
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 strato- sphere 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)

System	
Model (version)	Meso-scale Model (forecast model: ASUCA)
Date of implementation	1 March 2001 (ASUCA: 28 February 2017)
Configuration	
Domain	Japan, Lambert projection, 817×661 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)
Initial conditions	
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
Surface boundary conditions	
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
Lateral boundary conditions	
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
Other details	
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 mono- tonicity, 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)

System	
Model (version)	Local Forecast Model (forecast model: ASUCA)
Date of implementation	30 August 2012 (ASUCA: 29 January 2015)
Configuration	
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)
Initial conditions	
Data assimilation	LA produces initial conditions via a three-hour analysis cycle based on hourly assimilation with 3D-Var and one-hour forecasts
Surface boundary conditions	
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
Lateral boundary conditions	
Model providing lateral boundary conditions	MSM
Lateral boundary condition update frequency	8 times/day, 00 – 13-hour forecasts using the latest MSM information
Other details	
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° 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 η 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 η vertical coordinate system, which is a hybrid between pressure p and σ ($\sigma = p/p_S$, where p_S is surface pressure), implicitly defined as $p = A(\eta) + B(\eta)p_S$. η 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 η -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 ∇ 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 η using the boundary conditions of $\dot{\eta} = 0$ at $\eta = 0$ and $\eta = 1$, η -velocity and ω 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 Φ 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 η (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, $kmax$ 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

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

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

Here, Φ_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

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

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

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

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+} :

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

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 δt described earlier give the following equations with the omission of terms related to the semi-implicit method for reasons of simplicity:

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

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

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

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

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

Eq. (3.2.20) rewritten for δQ_k is found to be similar to Eq. (3.2.19) for δp_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 δp_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 $p''_{k-1/2}$ ($k = 2, 3, \dots, kmax$) 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 δt 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 (λ, φ) 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 ζ , D and T_V :

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

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

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

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 η -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 η -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:

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

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

Here, K_0 is the base diffusion coefficient for second-order horizontal diffusion, p_{top} is the pressure at the highest model level (0.01hPa), and p_{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											
H2O(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)
CO2							C-k(16)	C-k(16)			
O3											
H2O(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											
CO2						S-k(6,pf)					
N2O			S-k(6,pf)						S-k(2,no)		
CH4								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 $k - 1/2$, τ_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 H₂O 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 CO₂ and H₂O in the 540–800 cm⁻¹ 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 τ is the optical thickness, ω_0 is the single scattering albedo and S_0 is the incident solar irradiance in the direction μ_0 (the cosine of the solar zenith angle). The coefficients γ_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 τ , ω_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 τ_{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 ₂	O ₂	CH ₄	N ₂ 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 $r_{e,liq}$ [μm] 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 [gm^{-3}], N_{TOT} is the number concentration of water cloud droplets [cm^{-3}], ρ_w is the density of water [gm^{-3}] and k is a constant. N_{TOT} is set as 100 and 300 cm^{-3} over ocean and land areas, respectively. The effective radius of cloud ice particles $r_{e,ice}$ [μm] depends on temperature T [K] and cloud ice content IWC [gm^{-3}] as follows (Wyser 1998):

$$B = -2 + 10^{-3}(273 - T)^{1.5} \log_{10} \frac{\text{IWC}}{\text{IWC}_0}$$

$$r_{e,ice} = 377.4 + 203.3B + 37.91B^2 + 2.3696B^3 \quad (3.2.34)$$

where 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 ρ 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 δE_s and δE_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

$$M_n = M_{Bn}(t)\eta_n(z) \quad (3.2.37)$$

where M_B is the mass flux at the cloud base and η 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 η 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 λ is the entrainment rate and z_b is the cloud base height. λ 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. δE_s and δE_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), Δp is the depth of model cloud top layer, Δp_{eff} is the effective depth of the cloud top, and τ_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 ω 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 ω_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 λ_{\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 λ_{\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 Δt (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_{LFC} and z_{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:

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

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

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

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

$$\Delta q_w = \frac{a_L}{2} \left(\overline{q_w'^2} - 2b \overline{q_w' s_l'} + b^2 \overline{s_l'^2} \right)^{\frac{1}{2}} \quad (3.2.50)$$

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. $\overline{q_w'^2}$, $\overline{q_w' s_l'}$, and $\overline{s_l'^2}$ are calculated using the level 2 turbulence closure model of Mellor and Yamada (see Subsection 3.2.7) in the boundary layer scheme. Δq_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 θ 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 ≥ 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:

$$O = \frac{v_{\text{cice}} q_c}{\Delta z} + D_{12S} q_c \quad (3.2.55)$$

where Δz is the layer thickness, v_{cice} is the terminal velocity of a small ice particle in cloud, and D_{I2S} 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):

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

$$= O$$

where τ_p represents a characteristic time scale for the conversion of cloud droplets into raindrops and snowflakes, and q_c^{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 τ_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

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

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

where b is the clear-sky precipitation fraction (set to 0.5), τ_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:

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

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

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

Here $\mathbf{v} = (u, v)$ represents horizontal wind, θ 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 κ is von Kármán's constant ($= 0.4$), g is the standard acceleration due to gravity ($= 9.80665 \text{ [m/s}^2\text{]}$), z_1 is the height of the lowest level of the model grid above the ground, and θ_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 Ψ_m and Ψ_h are parameterized by Beljaars and Holtslag (1991) as follows:

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

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

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

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_* ($\equiv \sqrt{|\overline{w'\mathbf{v}'}|_s}$) is the friction velocity, ν is the kinematic viscosity of air ($= 1.5 \times 10^{-5} \text{ m}^2/\text{s}$) and α 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 κ 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:

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

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

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

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

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

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

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

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

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

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):

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

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 τ_r excited by subgrid-scale variances of topography σ^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 (1.6×10^{-5} for type A and 6.4×10^{-4} for type B), ρ 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 $v = |v|$. 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 σ^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 (σ_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 σ in the type A scheme where h denotes the model topography, while the average of σ_m in the Gaussian grid box is regarded as σ 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

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

ϵ is a function of the Richardson number (R_i)

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

$$R_i = N^2 \left/ \left[\frac{\partial}{\partial z} \left(\mathbf{v} \cdot \frac{\boldsymbol{\tau}_r}{|\boldsymbol{\tau}_r|} \right) \right]^2 \right. \quad (3.2.86)$$

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 ϕ 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,

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

ρ 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^ϕ as the velocity in the direction of the azimuth ϕ , 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.

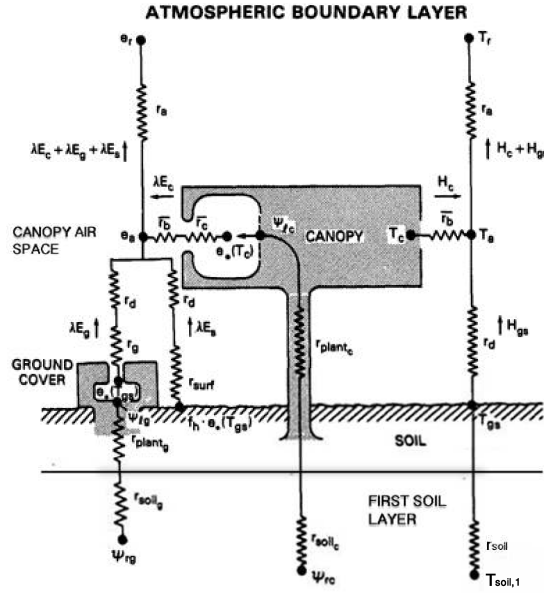


Figure 3.2.1: Schematic illustration of the land surface model (modified from Sellers *et al.* (1986)). Details are provided in the original paper.

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, τ_x and τ_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 evaporations 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 α 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 α_s can be determined as an average of the canopy albedo α_c and the ground albedo α_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 α_g is an average of the grass albedo α_{grs} and the bare soil albedo α_{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 α_{bs} is replaced by the value for snow α_{sn}

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

α_c and α_{grs} are calculated with radiative transfer equations (Sellers *et al.* 1986) for leaf and stem area. α_{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. α_{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 λ 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

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

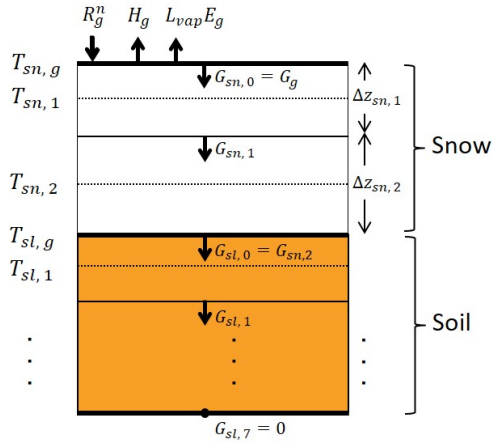


Figure 3.2.2: Schematic diagram of numerical discretization used to solve for snow temperatures $T_{sn,k}$ with two snow layers on top of soil. The subscripts sn , sl and k denote snow, soil and the k -th layer, respectively.

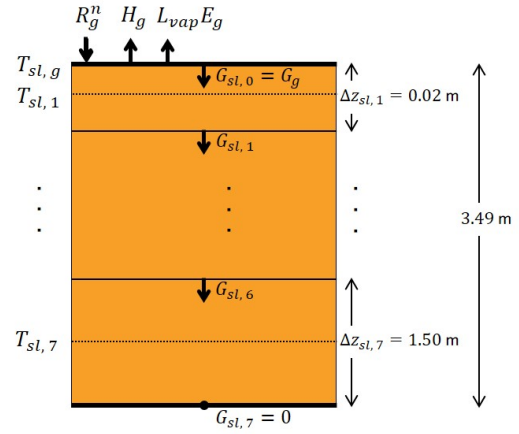


Figure 3.2.3: Schematic diagram of numerical discretization used to solve for soil temperatures $T_{sl,k}$ in snow-free areas.

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

where the subscript sl denotes soil, and Λ 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

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

where $Q_{sn,infl}$ is infiltration into snow including rainfall and water drip falling from leaves, $Q_{sn,drng}$ 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 ρ_{wtr} is water density, θ_{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 product¹ of NASA (Schaaf *et al.* 2002).

Each grid point on land is classified by a specific vegetation type provided from GLC2000² (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 product³ of NASA (Myneni *et al.* 2002). Fractions of canopy and grass covers are calculated using the 1 km MODIS-based Maximum Green Vegetation Fraction⁴ of USGS (Broxton *et al.* 2014), the 1 km Tree Cover Continuous Fields product⁵ of GLCF (DeFries *et al.* 2000), and the Cropland and Pasture Area fraction⁶ 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)⁷. The soil property is from HWSO (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 Hemholtz equations in the semi-implicit scheme. The Horizontal Advection stage is used in the implementation of the semi-Lagrangian advection scheme.

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

²<http://forobs.jrc.ec.europa.eu/products/glc2000/glc2000.php>

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

⁴https://archive.usgs.gov/archive/sites/landcover.usgs.gov/green_veg.html

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

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

⁷<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.

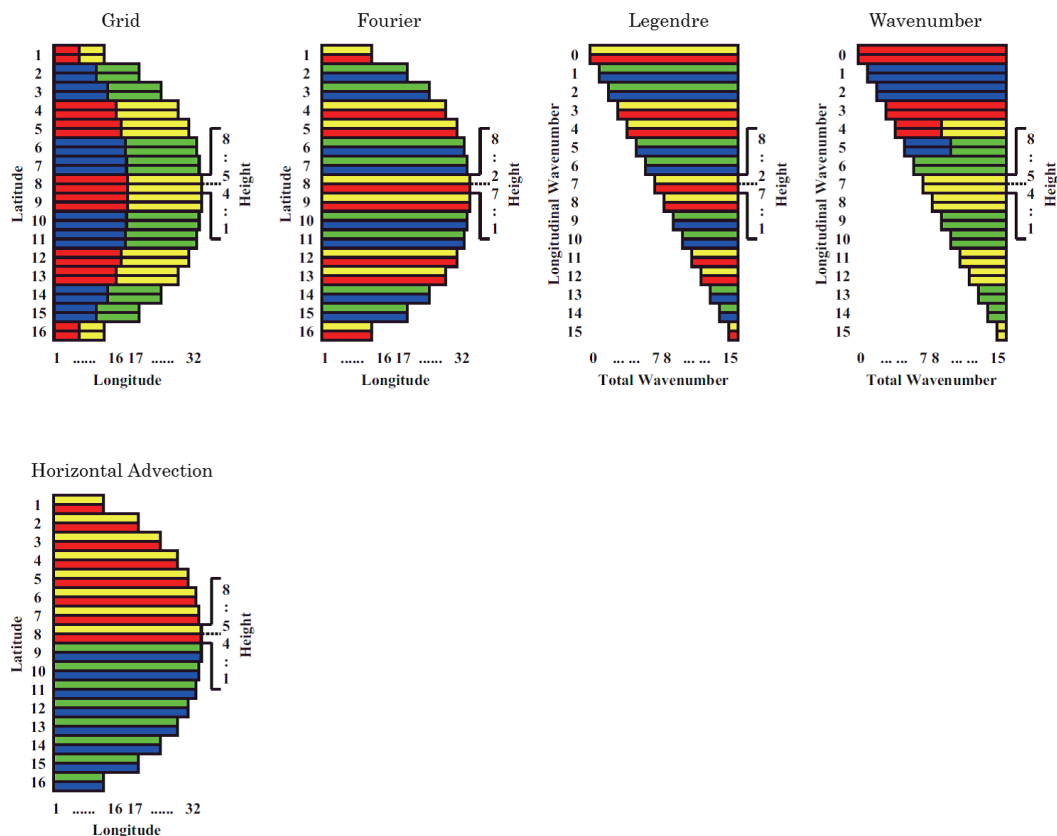


Figure 3.2.4: Schematic design of the parallelization. The number of processes used is assumed to be 4 in this example. Colors in the figure represent the rank for the computation in that area; red is rank 0, yellow is rank 1, blue is rank 2 and green is rank 3.

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 or 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 α_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 μ is the cosine of the solar zenith angle. The diffused albedo α_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⁻³K⁻¹] is the volumetric ice heat capacity, T_{ice} is the ice temperature, and $\lambda = 2.03$ [Wm⁻¹K⁻¹] 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 (α^N) and visible (α^V) are parameterized as follows:

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

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

$$\alpha^N = 0.86F + 0.01, \quad (3.2.117)$$

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

where T_{skin} is the ice surface temperature, and μ 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	Forecast guess
	Density	
	Liquid water content in voids	
	Albedo	
Sea ice	Ice concentration	Sea ice analysis (see Section 5.6)
	Temperature	Forecast guess
Open sea	Sea surface temperature	SST analysis (see Section 5.2)
	Friction velocity	Forecast guess

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.

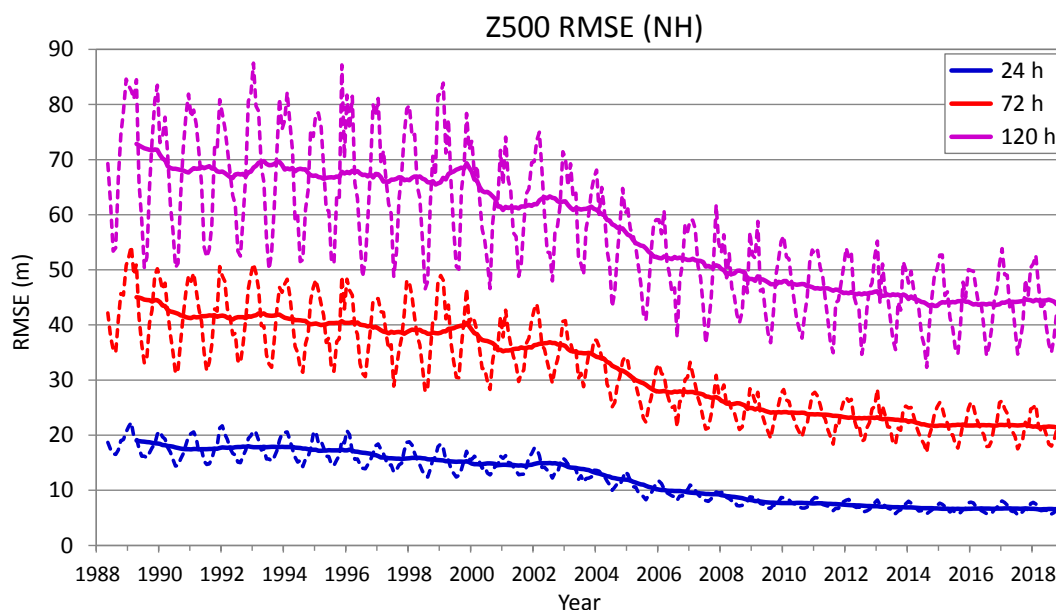


Figure 3.2.5: Root mean square error of GSM 500 hPa geopotential height (Z500) predictions against analysis in the Northern Hemisphere extra-tropics (20°N – 90°N). Dashed lines indicate monthly means, and solid lines represent running means calculated for the previous 12 months.

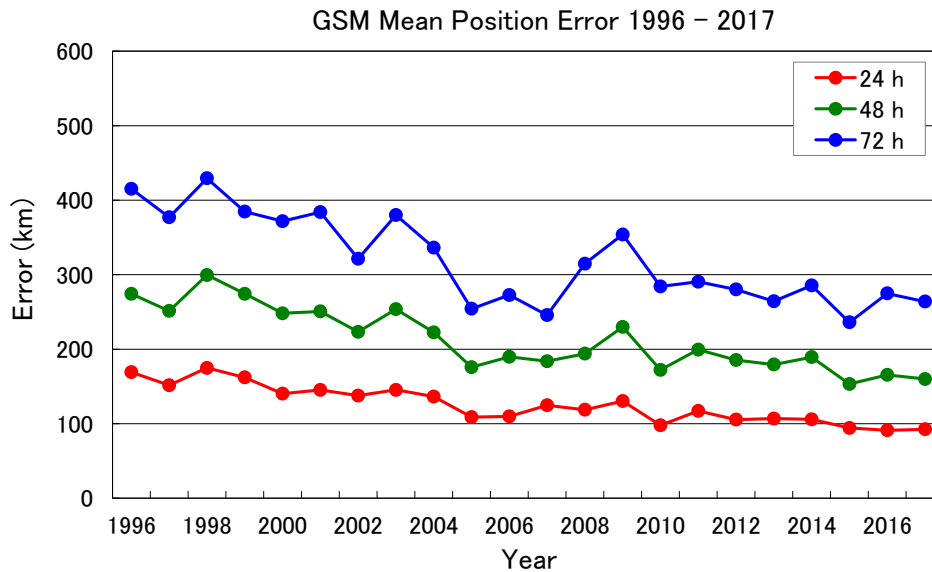


Figure 3.2.6: Mean position error of GSM TC track predictions in the western North Pacific from 1996 to 2017. The lines represent 24- (red), 48- (green) and 72-hour forecasts (blue).

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°N – 90°N) and the tropics (20°S – 20°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 ensemble⁸ with a forecast range of a month is used for the One-month Forecast issued on Thursdays. A 50-member lagged ensemble⁹ 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 (TS¹⁰) intensity or higher is present in the RSMC Tokyo - Typhoon Center's area of responsibility (0°–60°N, 100°E–180°).
- 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

⁸Composed 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.

⁹Composed 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.

¹⁰A 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 $2\sqrt{10}/3l$. 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 α 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 σ_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 η 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

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

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°N–90°N) and the tropics (20°S–20°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. λ and ϕ 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 α 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 (ΔSST_i^f) in the forecast from the initial time T_0 is constructed as

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

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 ± 27 days from that of the current initial time. ΔT and α 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 Center¹¹.

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 days¹². 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.

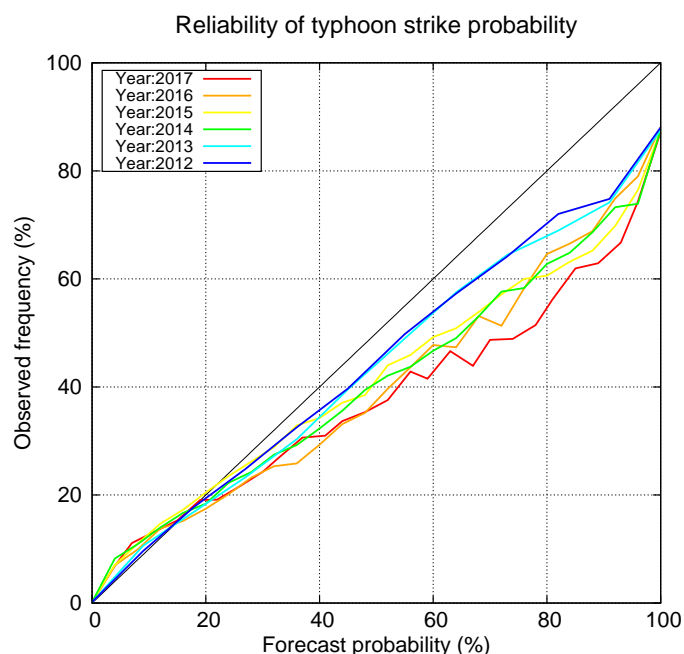


Figure 3.3.1: Reliability diagram for probabilistic verification of typhoon position forecasts as derived from the EPS over a six-year period. The target years for verification are 2012 (blue), 2013 (sky blue), 2014 (light green), 2015 (yellow), 2016 (orange) and 2017 (red). RSMC Tropical Cyclone Best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

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

¹²The 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 Verification¹³.

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.

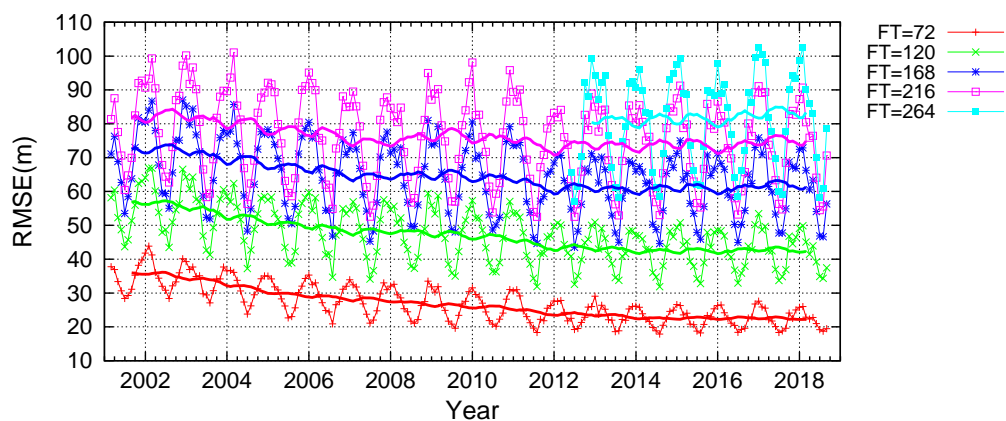


Figure 3.3.2: Time-series representation of ensemble mean scores for the EPS (where the score is the monthly-averaged RMSE of the ensemble mean) for Northern Hemisphere (20°N–90°N) 500-hPa geopotential height forecasts with lead times of 72 (red), 120 (green), 168 (blue), 216 (violet) and 264 (sky blue) 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 website¹⁴. 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).

¹³<http://epsv.kishou.go.jp/EPsv/>

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

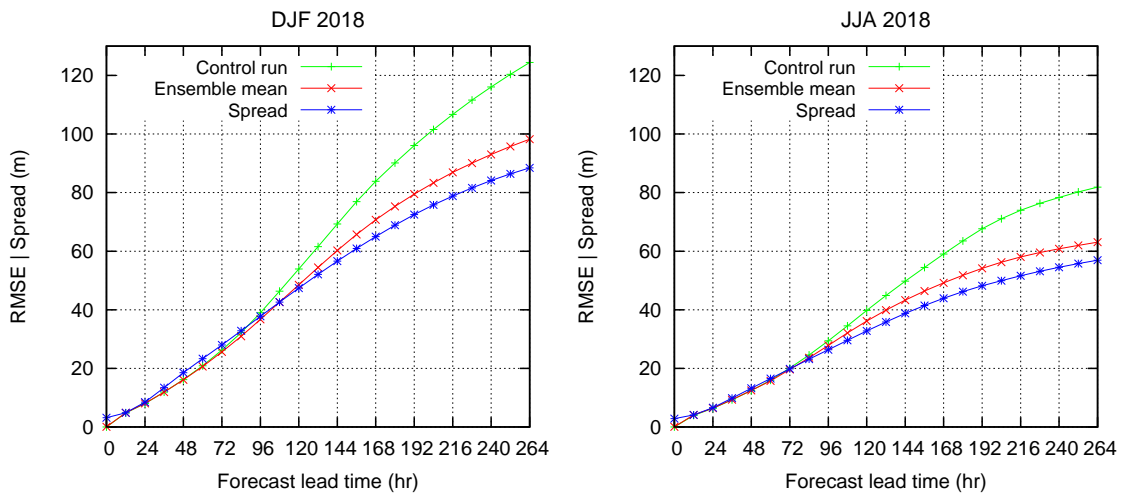


Figure 3.3.3: RMSEs for Northern Hemisphere (20°N–90°N) 500-hPa geopotential height forecasts of the ensemble mean (red) and unperturbed members (green) for DJF and JJA 2018 from the EPS. The spread of the ensemble (blue) is also shown.

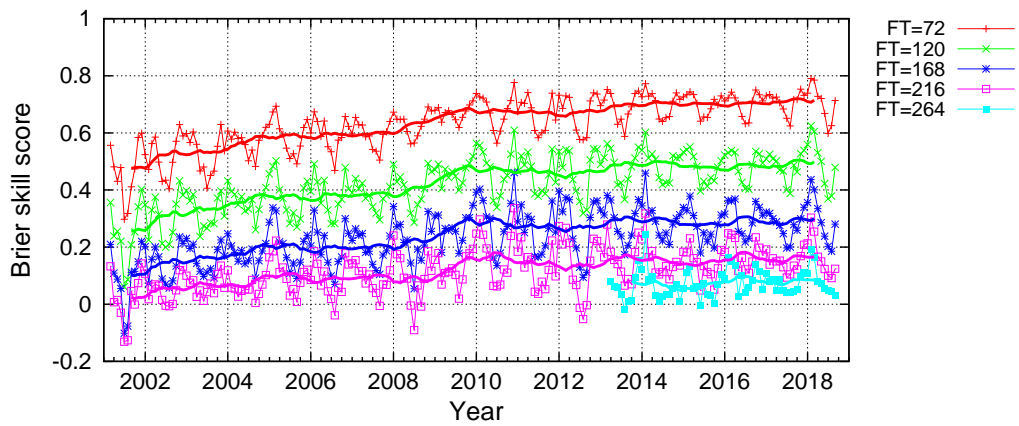


Figure 3.3.4: Same as Figure 3.3.2, but for Brier skill score for probabilistic forecasts of 500-hPa geopotential height negative anomalies with magnitudes less than one climatological standard deviation from JMA's EPS.

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.

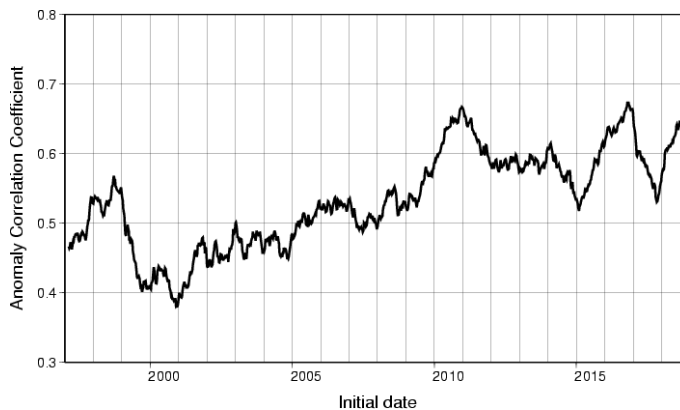


Figure 3.3.5: Time-series representation of the Northern Hemisphere (NH; 20°N–90°N) 500-hPa geopotential height anomaly correlation coefficient in ensemble mean forecasts averaged over 28 days from day 2 to day 29 (the running mean of 52 forecasts) based on operational forecasting conducted from 1997 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 x100 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

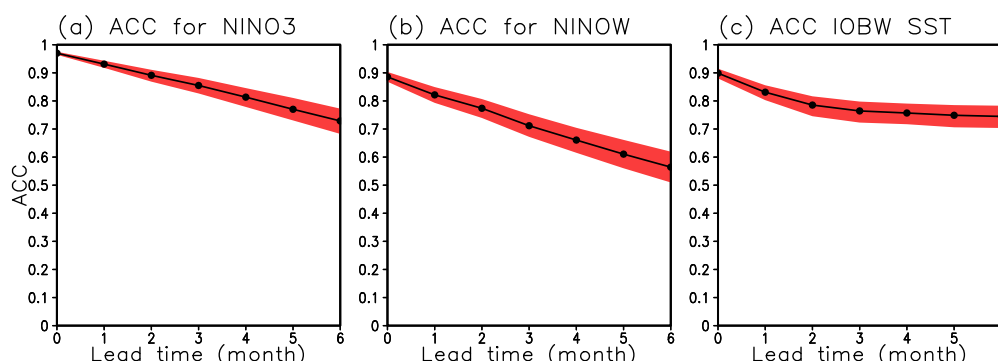


Figure 3.3.6: Anomaly correlations for SSTs, averaged over the 30 years, in (a) NINO.3 (5°S–5°N, 150°W–90°W), (b) NINO.WEST (0°–15°N, 130°E–150°E) and (c) IOBW (20°S–20°N, 40°E–100°E). Shading indicates a 90% confidence interval as estimated using the bootstrap method (1,000 samples).

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
Oceanic component	Basic equation	Primitive, free surface
	Domain	Global
	Resolution	1° (lon) \times $0.5^\circ - 0.3^\circ$ (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 Ujiie 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 (CO₂, N₂O, CH₄, 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° longitude and 0.5° latitude in the extratropics with meridional refinement from near the equator to 0.3° (see Figure 5.3.1). The model has 52 vertical layers, with an ocean bottom boundary layer (Nakano and Sugimoto 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 super-computer 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 Library¹⁵, 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

¹⁵The 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

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

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

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

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 (ξ, η, ζ) , 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 ξ_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 ρ 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.

π 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 ρ' , $(\rho \theta_m)'$ and π' represent perturbation from the hydrostatic state. g is gravity acceleration, and f is the Coriolis parameter. q_{α} is the ratio of the density of water substances α to the total mass density ($\alpha = v, c, r, i, s, g$). $W_{t_{\alpha}}$ is the terminal fall velocity of water substance α . θ_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 ϵ 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 φ 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 η coordinate (Simmons and Burridge 1981) is adopted to reduce the influences of topography as height increases (Ishida 2007). The vertical coordinate ζ 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_{ρ} 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_{θ} 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:

$$\left(\frac{1}{J}\rho w\right)^{\tau+\Delta\tau} = \left(\frac{1}{J}\rho w\right)^{\tau} - \left\{ \gamma R_d \pi^t \frac{1}{J} \zeta_z \frac{\partial}{\partial \zeta} (\rho \theta_m)^{\tau+\Delta\tau} + \frac{\rho^{\tau+\Delta\tau}}{J} g - \frac{\pi^t \bar{\rho}}{\pi} \frac{\bar{\rho}}{J} g \right\} \Delta\tau + R_w^t \Delta\tau, \quad (3.5.22)$$

$$\begin{aligned} \left(\frac{1}{J}(\rho\theta_m)'\right)^{\tau+\Delta\tau} &= \left(\frac{1}{J}(\rho\theta_m)'\right)^\tau - \left\{\frac{\partial}{\partial\zeta}\left(\frac{1}{J}\zeta_z\theta_m^\tau(\rho w)^{\tau+\Delta\tau}\right)\right\}\Delta\tau \\ &\quad - \left\{\frac{\partial}{\partial\xi}\left(\frac{1}{J}\theta_m^\tau(\rho\overline{U})\right) + \frac{\partial}{\partial\eta}\left(\frac{1}{J}\theta_m^\tau(\rho\overline{V})\right) + \frac{\partial}{\partial\zeta}\left(\frac{1}{J}\theta_m^\tau(\rho\overline{W})\right)\right\}\Delta\tau + \frac{1}{J}F'_{\rho\theta_m}\Delta\tau, \end{aligned} \quad (3.5.23)$$

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

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

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

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

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

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

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

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

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

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_ξ, C_η and C_ζ are the Courant number in the ξ, η and ζ 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 Δt , 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_ξ and F_η first, and the vertical flux F_ζ 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.

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

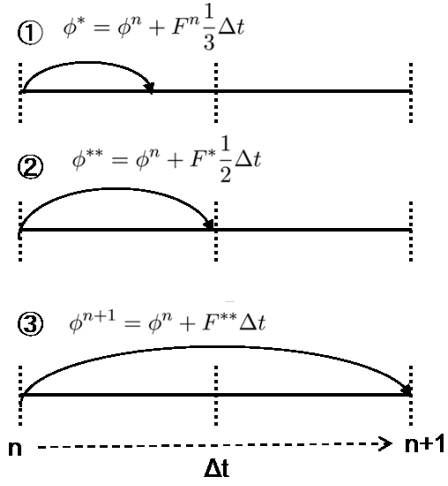


Figure 3.5.1: Schematic of RK3 time integration scheme.

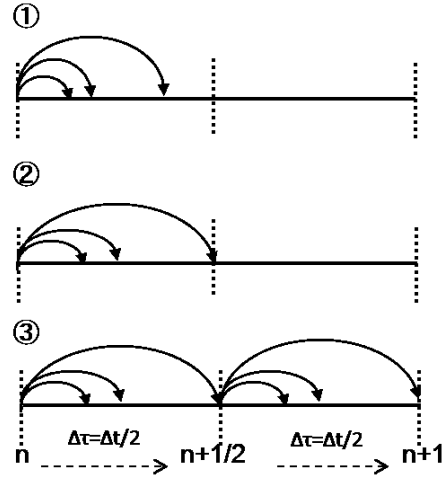


Figure 3.5.2: Schematic of time-splitting of vertical advection. The case needs to split ③ stage into 2 substeps.

where

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

Here, W^n is W at the time level n , β 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 $C'_{t\zeta} = (W^n + W_{t\alpha}^{n+\tau_1}) \Delta t' / \Delta \zeta$ 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 ϕ is the prognostic variable and ϕ_{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 ϕ relating to one cloud microphysical process in each particle gives the grid-mean temporal tendency of ϕ 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 ϕ 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 ρ_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 (ρ_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 λ_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 (= 273.15K).

The number concentration, the 0-th 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 λ_r are calculated as

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

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

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

4. Cloud water

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

$$n_c(D_c) = N_c \delta(D_c - \overline{D_c}), \quad (3.5.62)$$

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 ρ_0 is the density of the reference air, and α_{ux} , β_{ux} and γ_{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 ϕ associated with subgrid convection under the isobaric coordinate is generally described as

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

where $\overline{\phi' w'}$ and S_{ϕ} 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 q_r [kg kg ⁻¹]	Snow q_s [kg kg ⁻¹]	Graupel q_g [kg kg ⁻¹]	Cloud ice q_i [kg kg ⁻¹]	Cloud water q_c [kg kg ⁻¹]
Size distribution [m ⁻⁴]	$n_r(D_r) = N_{00r} \lambda_r^{\beta_r} \exp(-\lambda_r D_r)$ $N_{00r} = 0.22$ $\beta_r = 2.2$	not directly used	$n_g(D_g) = N_{0g} \exp(-\lambda_g D_g)$ $N_{0g} = 1.1 \times 10^6$	$n_i(D_i) = N_{0i} \exp(-\lambda_i D_i)$ $N_{0i} = 4.0 \times 10^7$	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 ⁻³]	$\rho_r = 1.0 \times 10^3$	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]	$U_r(D_r) = \left(\frac{\rho_0}{\rho_a} \right)^{\frac{1}{2}} \sum_{k=1}^2 c_k D_r^{d_k} e^{-f_k D_r}$ $c_1 = 4854.1$ $d_1 = 1.0$ $f_1 = 195.0$ $c_2 = -446.009$ $d_2 = 0.782127$ $f_2 = 4085.35$	$\alpha_{us} = 17$ $\beta_{us} = 0.5$ $\gamma_{us} = 0.5$	$\alpha_{ug} = 124$ $\beta_{ug} = 0.64$ $\gamma_{ug} = 0.5$	$\alpha_{ui} = 71.34$ $\beta_{ui} = 0.6635$ $\gamma_{ui} = 0.5$	not considered
			$U_x(D_x) = \alpha_{ux} D_x^{\beta_{ux}} \left(\frac{\rho_0}{\rho_a} \right)^{\gamma_{ux}}$		

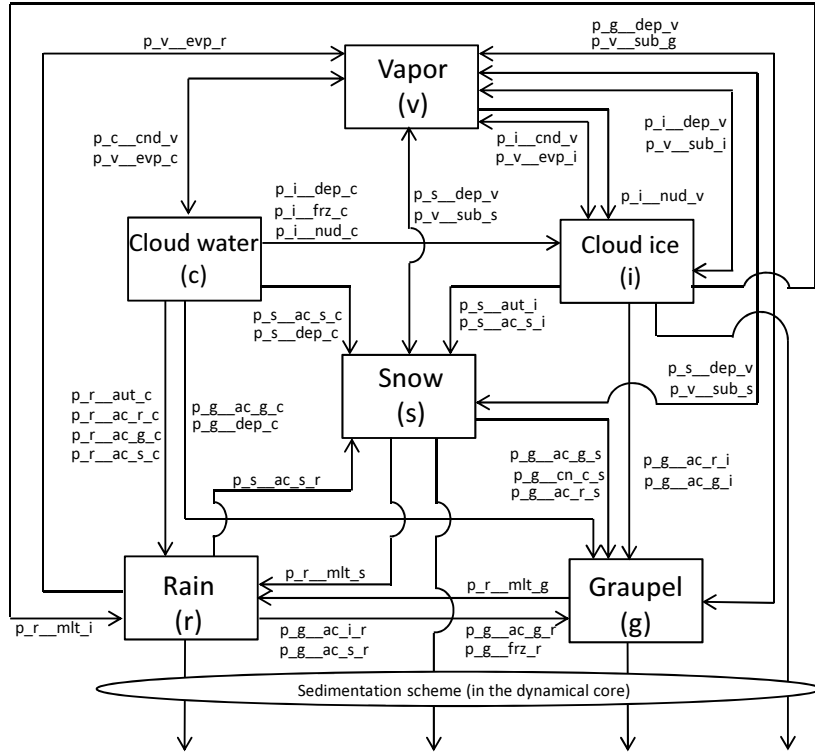


Figure 3.5.3: Cloud microphysical processes in the MSM. For a list of symbols, see Table 3.5.1.

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 ϕ , 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

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

Substituting Eq. (3.5.70), Eq. (3.5.71) and Eq. (3.5.72) into Eq. (3.5.69), the temporal tendency of ϕ 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 ϕ^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 δM_e , δM_u and δM_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

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

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

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

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.

δM_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 δP is the vertical grid thickness in the pressure coordinate, and the factor a is set to 0.03 m Pa^{-1} 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 detrain 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:

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

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

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:

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

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} [$\text{kgm}^{-2}\text{s}^{-1}$], is initially given as follows with the assumption that vertical velocity w_{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 τ_{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 TKE_{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 $20 \text{ m}^2 \text{ s}^{-2}$. Δp_{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 ϕ 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 τ_{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_{diag} is defined as the lifted air mass at the LCL T_{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 Π , 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 $G_{\text{max}} = 10^{-3}$ 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_{LCL} . If $T_{\text{diag}} > \overline{T}$ is not satisfied, the base of the potential USL is moved up to the next model level and comparison of T_{diag} with \overline{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 $T_{\text{diag}} > \overline{T}$ 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

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

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

where T [$^{\circ}\text{C}$] 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 (θ'_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 ($\overline{q_l}$) are given by

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

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

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_1 and Q_1 are given by

$$a_1 = \left[1 + \frac{L}{C_p} \left(\frac{\partial q_{sat}}{\partial T} \right)_{T=T_1} \right]^{-1}, \quad (3.5.93)$$

$$Q_1 = \frac{a_1 \Delta\bar{q}}{2\sigma_s}, \quad (3.5.94)$$

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, σ_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:

$$\sigma_s^2 = \frac{1}{4} \left(a_1^2 \overline{q_w'^2} - 2a_1 b_1 \overline{\theta_1' q_w'} + b_1^2 \overline{\theta_1'^2} \right), \quad (3.5.95)$$

$$b_1 = a_1 \Pi \left(\frac{\partial q_{sat}}{\partial T} \right)_{T=T_1}, \quad (3.5.96)$$

where Π 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 $\phi (= u, v, \theta, q_w)$ 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, θ_1 the liquid water potential temperature, q_w the total water content, and ℓ the mixing length (see Subsection 3.5.7.4). $\overline{\phi}$ represents an ensemble-averaged ϕ , and ϕ' is the turbulent fluctuation of ϕ . The buoyancy flux, $(g/\overline{\theta_v})\overline{w'\theta_v'}$, and the dissipation terms ε_X are described in Subsections 3.5.7.2 and 3.5.7.3, respectively. Here, q_1 is the mixing ratio of liquid water (including the ice phase), and

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

$$\theta_1 = \theta - \frac{L}{C_p} \frac{\theta}{T} q_1, \quad (3.5.106)$$

$$q_w = q_v + q_1, \quad (3.5.107)$$

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 θ_1 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 ($\overline{q_1}$) determined as moments of the PDF (see Subsection 3.5.6). Following Sommeria and Deardorff (1977) and Mellor (1977), the value is given by

$$\frac{g}{\theta_v} \overline{w'\theta_v'} = \frac{g}{\theta_v} (\beta_\theta \overline{w'\theta_1'} + \beta_q \overline{w'q_w'}), \quad (3.5.112)$$

$$\beta_\theta = 1 + 0.61\overline{q_w} - 1.61\overline{q_1} - \widetilde{R}a_1b_1c_1, \quad (3.5.113)$$

$$\beta_q = 0.61\overline{\theta} + \widetilde{R}a_1c_1, \quad (3.5.114)$$

$$\widetilde{R} = R - \frac{\overline{q_1}}{2\sigma_s} \frac{1}{\sqrt{2\pi}} \exp\left(-\frac{Q_1^2}{2}\right), \quad (3.5.115)$$

$$c_1 = (1 + 0.61\overline{q_w} - 1.61\overline{q_1})\frac{\overline{\theta}}{T}\frac{L}{C_p} - 1.61\overline{\theta}. \quad (3.5.116)$$

Here, a_1 , b_1 , σ_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 ε_X appearing in the equations are parameterized on the basis of Kolmogorov's local isotropy assumption as

$$\varepsilon = \frac{q}{B_1\ell}q^2, \quad \varepsilon_\theta = \frac{q}{B_2\ell}\overline{\theta_1^2}, \quad \varepsilon_q = \frac{q}{B_2\ell}\overline{q_w^2}, \quad \varepsilon_{\theta q} = \frac{q}{B_2\ell}\overline{\theta_1'q_w'}, \quad (3.5.117)$$

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

3.5.7.4 Mixing Lengths

The mixing length ℓ 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 , $q_c = [(g/\overline{\theta_v})\overline{w'\theta_v'}L_T]^{1/3}$, 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 , $\overline{\theta_1^2}$, $\overline{q_w^2}$ and $\overline{\theta_1'q_w'}$). Following Nakanishi and Niino (2004), S_{θ_1} , $S_{\theta q}$, and S_{q_w} are assumed to be the same as S_q , and $S_q = S_{\theta_1} = S_{\theta q} = S_{q_w} = 2(S_{M2.5} + S'_M)$. 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, θ_{v1} , and q_{v1}) and the surface (θ_{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, θ_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, β , 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, β 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, β is set to 1.

The transfer coefficients are formulated as

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

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

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, ψ_m , and heat, ψ_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 $x = (1 - 16z)^{1/4}$. 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 s m^{-1} from April to October and 60 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

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

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 ν the kinematic viscosity ($= 1.5 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$). u_* is the friction velocity as defined by

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

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, σ is the Stefan-Boltzmann constant, α 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 ρ 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 λ 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, τ 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, ρ_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).

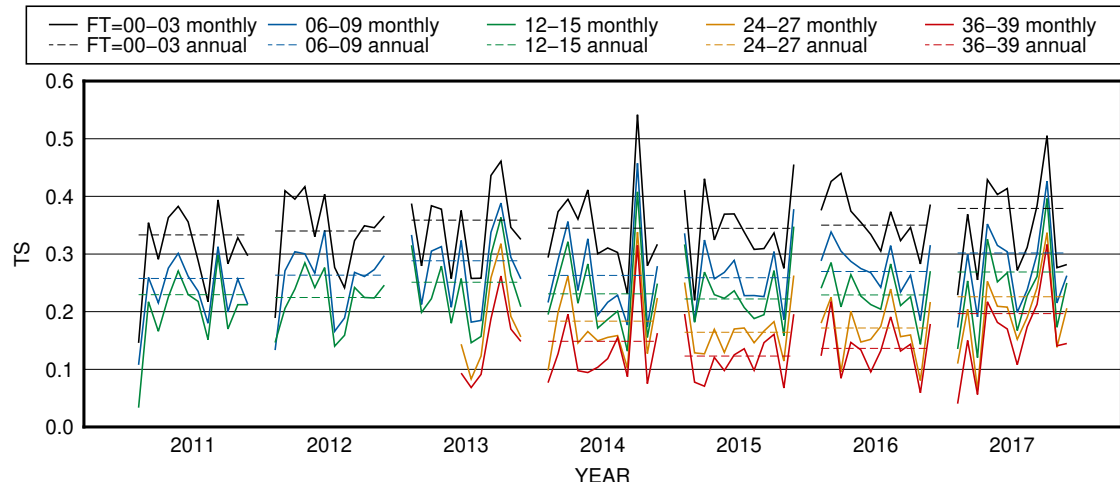


Figure 3.5.4: Monthly and annual threat scores of 3-hour cumulative precipitation at the 10mm threshold, against the R/A within 20km verification grids. The solid and dashed lines represent monthly and annual scores for each, FT represents the forecast range (hours). The verification period is from January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

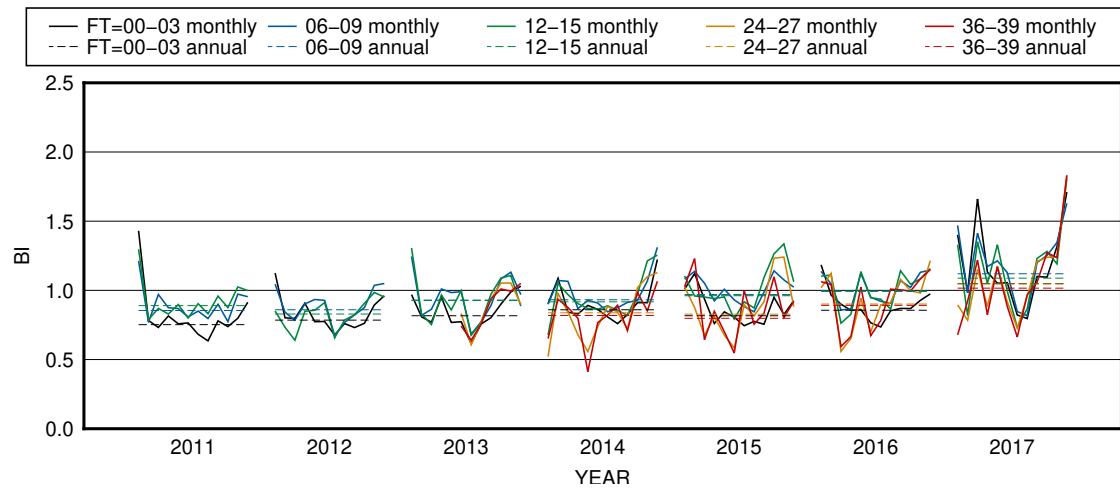


Figure 3.5.5: As per Figure 3.5.4, but for bias scores

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.

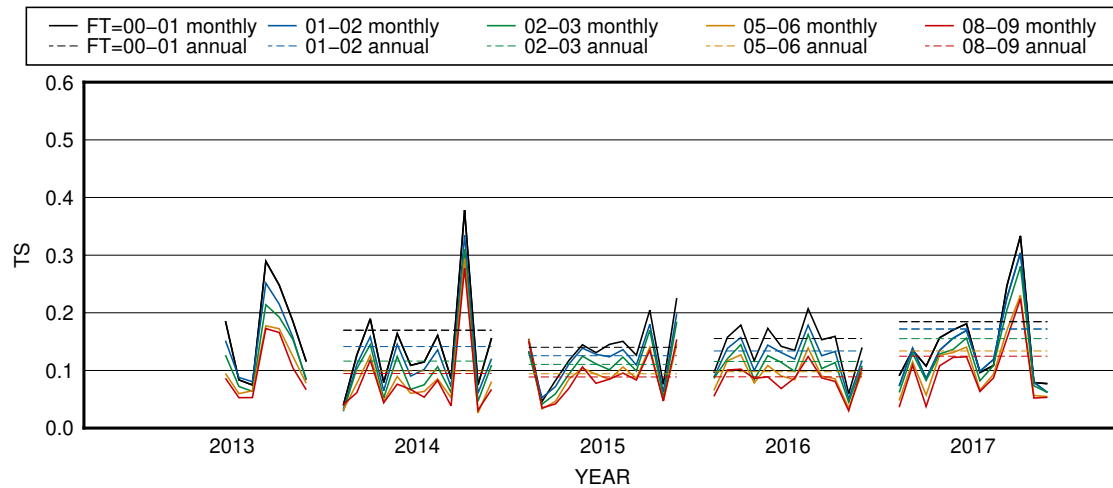


Figure 3.6.1: Monthly and annual threat scores for one-hour cumulative precipitation at the 10mm threshold against the R/A within 20km verification grids. The solid and dashed lines represent monthly and annual scores for each, and FT represents the forecast range (hours). The verification period is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

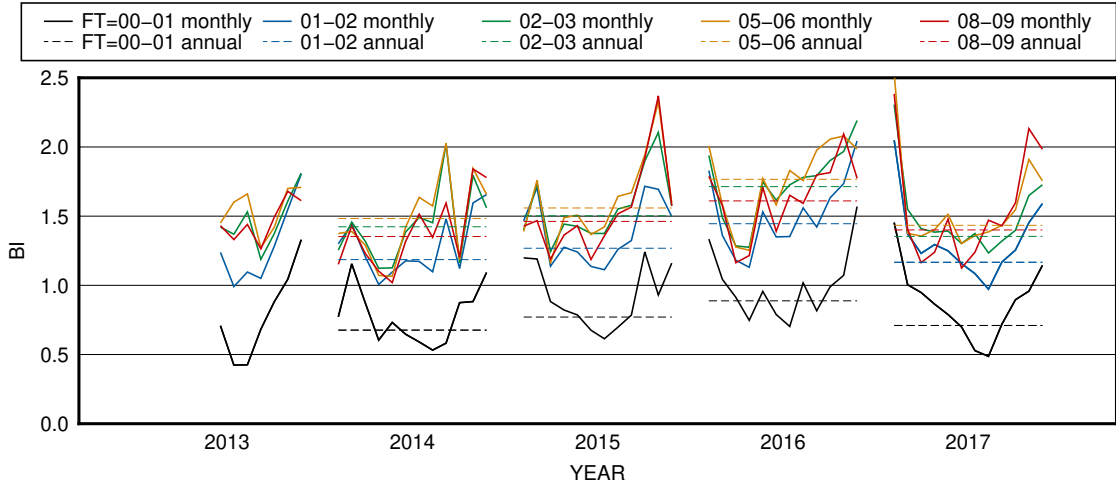


Figure 3.6.2: As Per Figure 3.6.1 but for bias scores

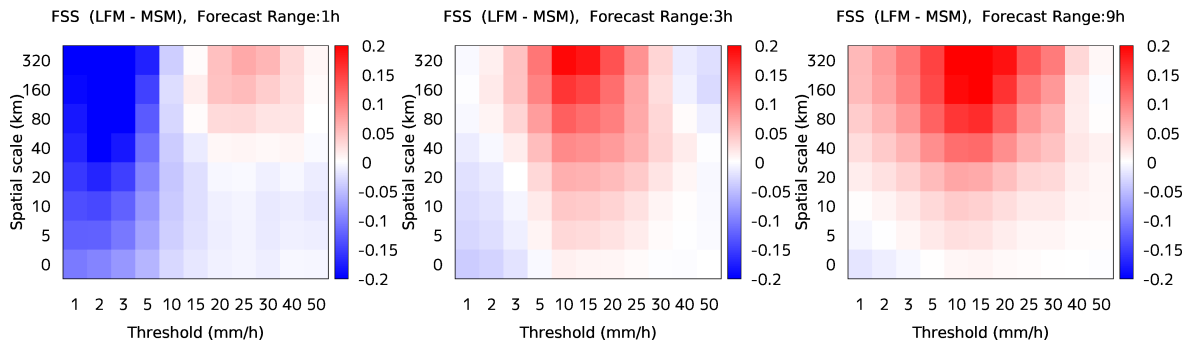


Figure 3.6.3: Monthly averaged subtraction of Fraction Skill Scores for MSM one-hour cumulative precipitation from those of the LFM. The forecast ranges are 1, 3 and 9 hours. The verification period is August 2017.

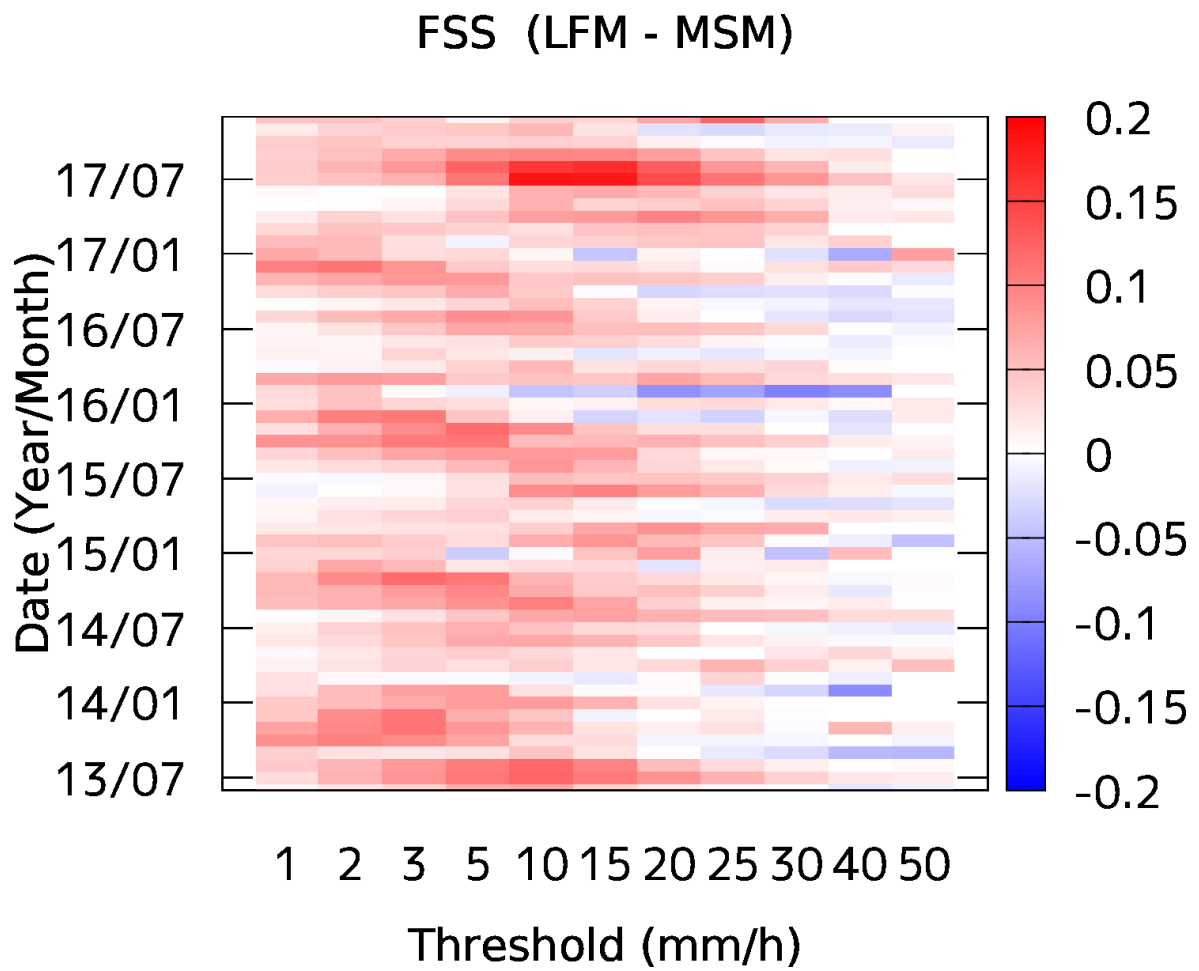


Figure 3.6.4: Monthly averaged subtraction of Fraction Skill Scores for MSM one-hour cumulative precipitation from those of the LFM. The forecast range is nine hours and the spatial scale is 80 km. The verification period is from June 2013 to December 2017.

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 δt is the single time step length and T_L is the Lagrangian time scale. σ 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 δx and δy 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 δz for a single time step δt 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 κ 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 Λ [1/h] is approximated as a function of precipitation intensity P [mm/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}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

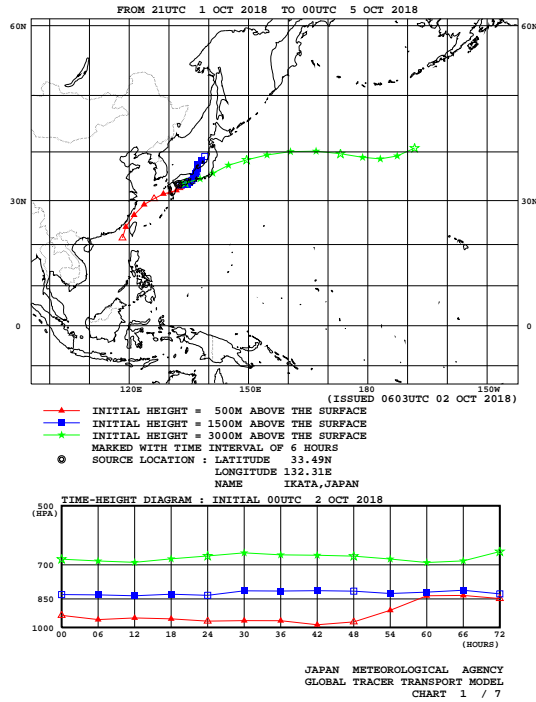
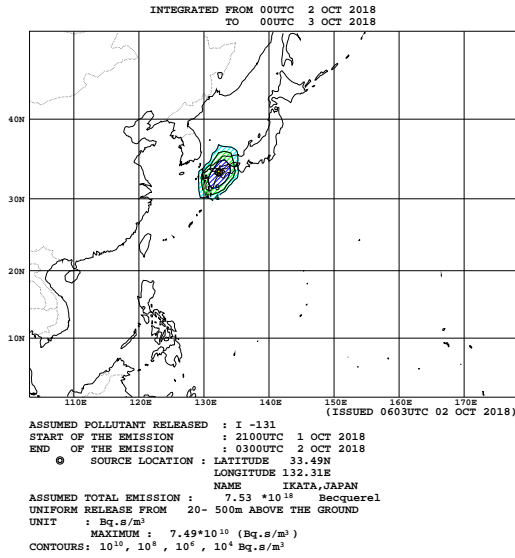


Figure 3.7.1: Example of the 3D trajectories

EXERCISE-EXERCISE-EXERCISE

EXERCISE-EXERCISE-EXERCISE

TIME INTEGRATED SURFACE - 500m LAYER CONCENTRATION



TIME INTEGRATED SURFACE - 500m LAYER CONCENTRATION

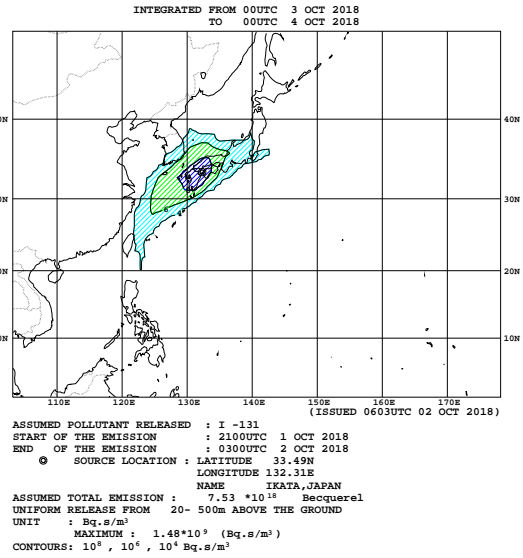
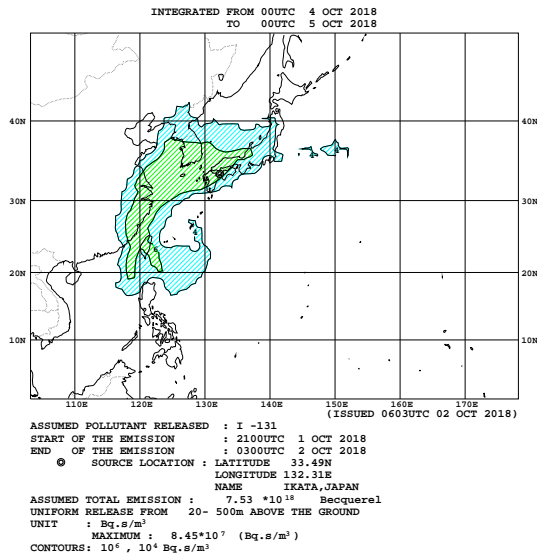


Figure 3.7.2: Example of time-integrated concentration in forecasts of up to 24 hours

Figure 3.7.3: Example of time-integrated concentration in forecasts of up to 48 hours

EXERCISE-EXERCISE-EXERCISE

TIME INTEGRATED SURFACE - 500m LAYER CONCENTRATION

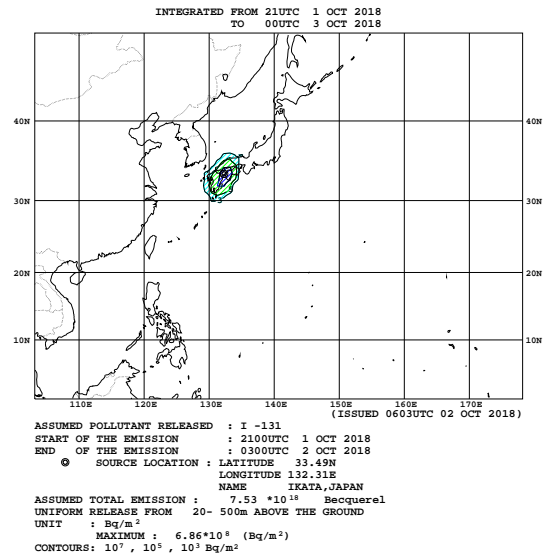


JAPAN METEOROLOGICAL AGENCY
GLOBAL TRACER TRANSPORT MODEL
CHART 4 / 7

Figure 3.7.4: Example of time-integrated concentration in forecasts of up to 72 hours
EXERCISE-EXERCISE-EXERCISE

EXERCISE-EXERCISE-EXERCISE

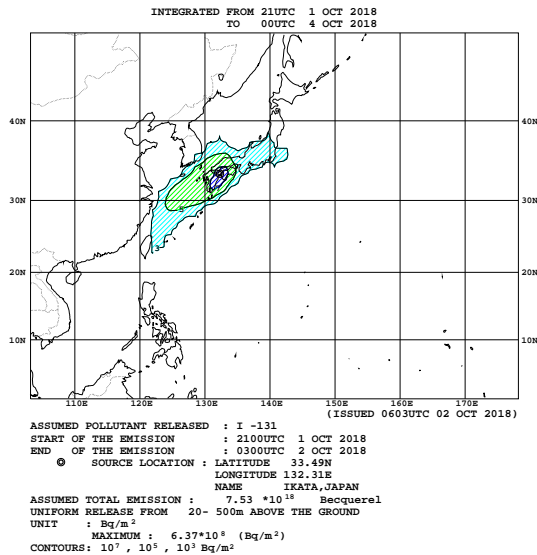
TOTAL (WET AND DRY) DEPOSITION



JAPAN METEOROLOGICAL AGENCY
GLOBAL TRACER TRANSPORT MODEL
CHART 5 / 7

Figure 3.7.5: Example of total deposition in forecasts of up to 24 hours
EXERCISE-EXERCISE-EXERCISE

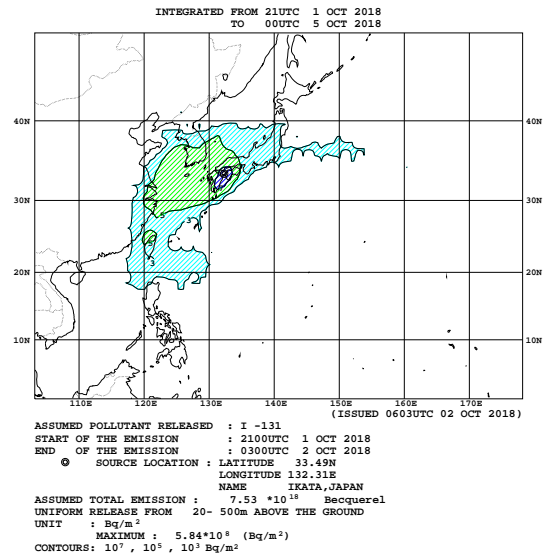
TOTAL (WET AND DRY) DEPOSITION



JAPAN METEOROLOGICAL AGENCY
GLOBAL TRACER TRANSPORT MODEL
CHART 6 / 7

Figure 3.7.6: Example of total deposition in forecasts of up to 48 hours

TOTAL (WET AND DRY) DEPOSITION



JAPAN METEOROLOGICAL AGENCY
GLOBAL TRACER TRANSPORT MODEL
CHART 7 / 7

Figure 3.7.7: Example of total deposition in forecasts of up to 72 hours

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.

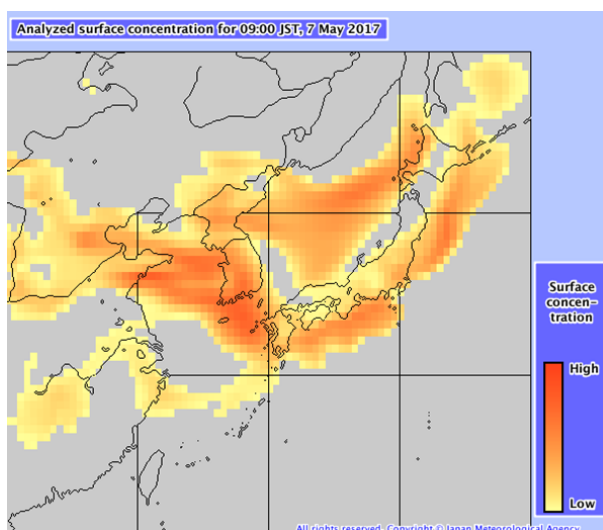


Figure 3.8.1: Kosa (Aeolian dust) prediction web page (<https://www.jma.go.jp/en/kosafcst/>).

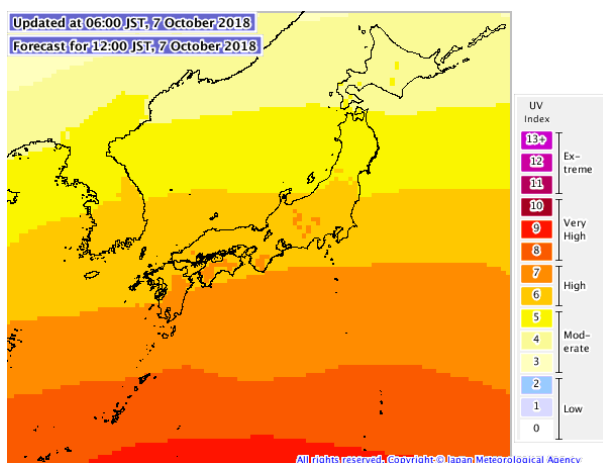


Figure 3.8.2: Clear-sky UV index forecast web page (<https://www.jma.go.jp/en/uv/>).

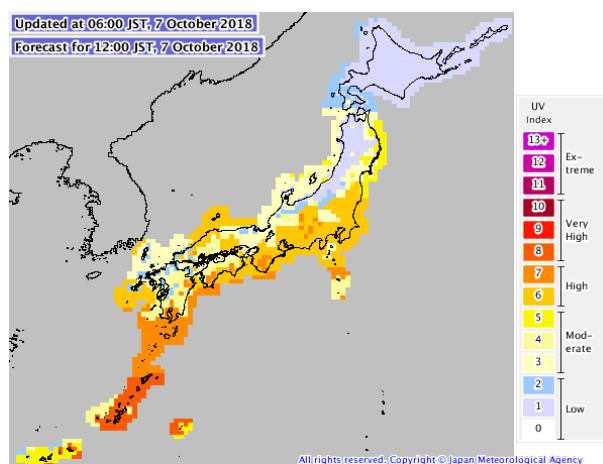


Figure 3.8.3: UV index forecast web page (<https://www.jma.go.jp/en/uv/>).

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 τ is a relaxation time of 6-24 hours. This technique enables the CTM to simulate meteorological fields realistically during the prediction period.

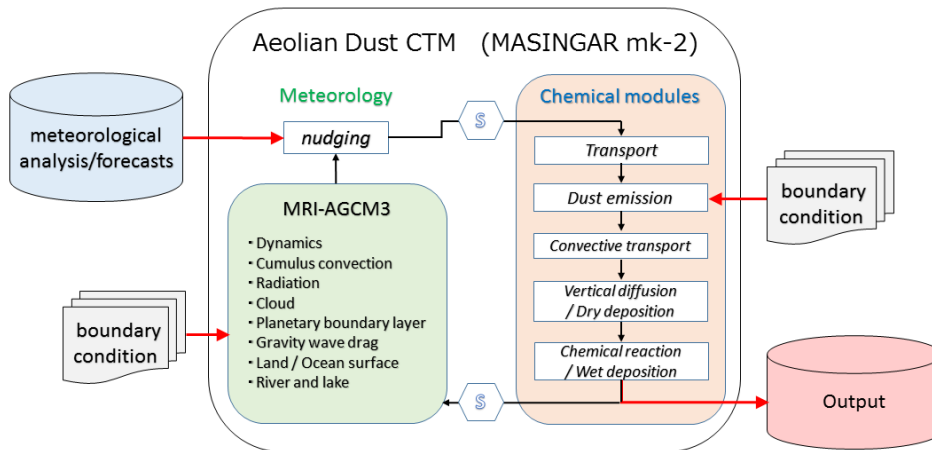


Figure 3.8.4: Schematic illustration of the CTM for Kosa (Aeolian dust) prediction. (The symbol mark S indicates the Scup coupler.)

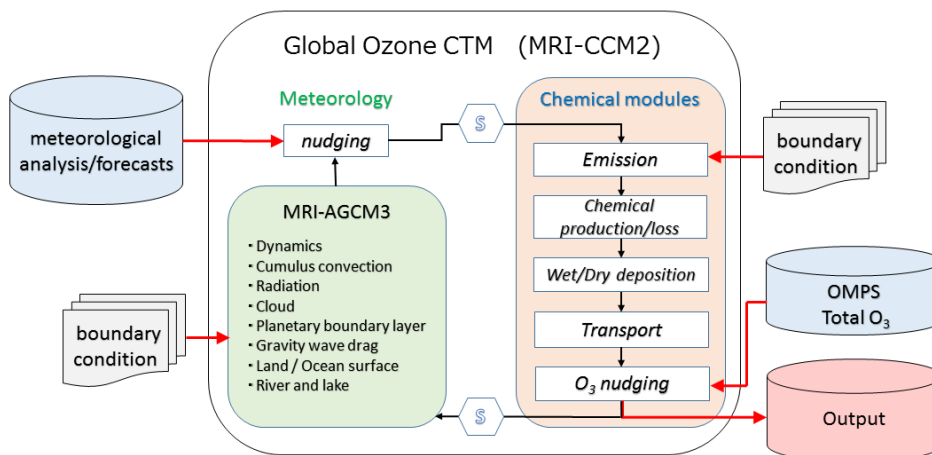


Figure 3.8.5: Same as Figure 3.8.4 except for the CTM for UV index forecast.

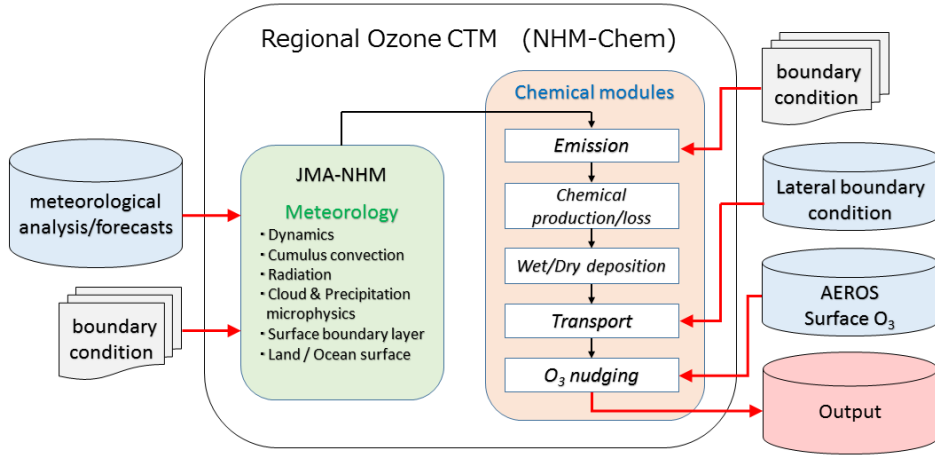


Figure 3.8.6: Same as Figure 3.8.4 except for the CTM for photochemical oxidant information.

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 α 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 Γ are certain constants, ρ is soil particle density, ρ_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, ρ_s^{dust} : predicted concentration of surface dust)

	Observed (<i>ww</i> = 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

Long-lived			
01: N ₂ O	02: CH ₄	03: H ₂ O	04: NO _y
05: HNO ₃	06: N ₂ O ₅	07: Cl _y	08: O _x
09: CO	10: OClO	11: CO ₂	12: Passive tracer
13: HCl	14: ClONO ₂	15: HOCl	16: Cl ₂
17: H ₂ O ₂	18: ClNO ₂	19: HBr	20: BrONO ₂
21: NO _x	22: HO ₂ NO ₂	23: ClO _x	24: BrO _x
25: Cl ₂ O ₂	26: HOBr	27: CCl ₄ (CFC-10)	28: CFCl ₃ (CFC-11)
29: CF ₂ Cl ₂ (CFC-12)	30: Br _y	31: CH ₃ Cl	32: CH ₃ Br
33: CF ₂ ClBr (Halon1211)	34: CF ₃ Br (Halon1301)	35: COF ₂	36: HF
37: CH ₂ O	38: CH ₃ OOH	39: C ₂ H ₆	40: CH ₃ CHO
41: C ₂ H ₅ OOH	42: PAN (CH ₃ C(O)OONO ₂)	43: CH ₃ C(O)OOH	44: C ₃ H ₈
45: ACET (CH ₃ C(O)CH ₃)	46: C ₃ H ₇ OOH	47: HACET (CH ₃ C(O)CH ₂ OH)	48: MGLY (CH ₃ C(O)CHO)
49: C ₂ H ₄	50: GLY ALD (HOCH ₂ CHO)	51: GPAN (HOCH ₂ C(O)OONO ₂)	52: GC(O)OOH (HOCH ₂ C(O)OOH)
53: C ₃ H ₆	54: ONIT (CH ₃ C(O)CH ₂ ONO ₂)	55: POOH (HOC ₃ H ₆ OOH)	56: C ₄ H ₁₀
57: C ₅ H ₈ (isoprene)	58: MACR	59: ISON	60: ISOPOOH
61: NALD	62: MACROOH	63: MPAN	64: C ₁₀ H ₁₆ (terpenes)
Short-lived			
01: O(1D)	02: OH	03: Cl	04: O(3P)
05: O ₃	06: HO ₂	07: NO ₂	08: NO
09: Br	10: N	11: ClO	12: BrO
13: NO ₃	14: BrCl	15: H	16: CH ₃ O ₂
17: C ₂ H ₅ O ₂	18: CH ₃ C(O)O ₂	19: C ₃ H ₇ O ₂	20: ACETO ₂ (CH ₃ C(O)CH ₂ O ₂)
21: EO ₂ (HOC ₂ H ₄ O ₂)	22: EO (HOC ₂ H ₄ O)	23: GC(O)O ₂ (HOCH ₂ C(O)O ₂)	24: PO ₂ (HOC ₃ H ₆ O ₂)
25: ISOPO ₂	26: MACRO ₂		
Chemical families			
O _x = O ₃ + O(3P) + O(1D)			
ClO _x = Cl + ClO			
Cl _y = ClO _x + OClO + 2Cl ₂ O ₂ + HCl + ClONO ₂ + HOCl + 2Cl ₂ + ClNO ₂ + BrCl			
NO _x = NO + NO ₂ + NO ₃			
NO _y = NO _x + N + HNO ₃ + 2N ₂ O ₅ + HO ₂ NO ₂ + ClONO ₂ + ClNO ₂ + BrONO ₂ + PAN + GPAN + ONIT + ISON + NALD + MPAN			
BrO _x = Br + BrO + BrCl			
Br _y = BrO _x + HBr + HOBr + BrONO ₂			

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.

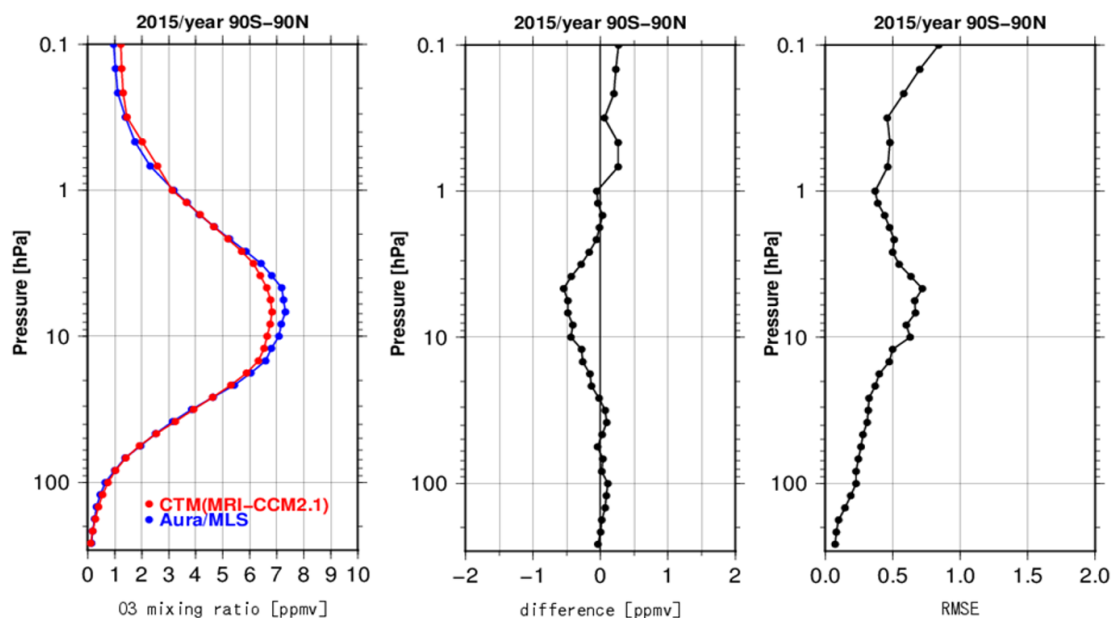


Figure 3.8.7: Annual mean vertical ozone profile in 2015 averaged over the globe. The left panel shows the ozone mixing ratio of the CTM (red) and satellite observation (blue, Aura/MLS). The middle panel shows the difference of the mixing ratios, and the right panel indicates the root mean square difference.

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.

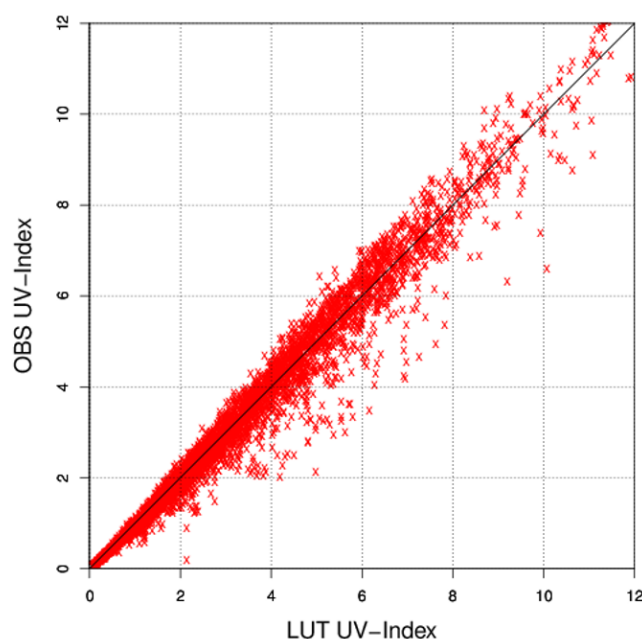


Figure 3.8.8: Relationship between calculated clear-sky UV indices and observed UV indices in clear-sky conditions at three JMA stations from 2015 to 2017. In UV index calculation, climatological aerosol optical depth is applied.

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	East Asia
GFED3	(Global Fire Emission Database, version 3)	Giglio <i>et al.</i> (2010)	natural	global
MEGAN2	(Model of Emission of Gases and Aerosols from Nature, version 2)	Guenther <i>et al.</i> (2006)	natural	global

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 γ 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° × 1.5° (latitude/longitude)	
Elements	Extra-tropics:	Geopotential height (Z), temperature (T) and wind at 850, 500 and 250 hPa; Mean sea-level pressure (MSLP)
	Tropics:	Z, T and wind at 850 and 250 hPa
Scores	Wind:	Root mean square vector wind error and mean error of wind speed
	Other elements:	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°N – 20°N), SH extra-tropics (20°S – 90°S), Tropics (20°N – 20°S), North America (25°N – 60°N, 50°W – 145°W), Europe/North Africa (25°N – 70°N, 10°W – 28°E), Asia (25°N – 65°N, 60°E – 145°E), Australia/New Zealand (10°S – 55°S, 90°E – 180°E), Northern polar region (90°N – 60°N) and Southern polar region (90°S – 60°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:	Z, T and wind at 850, 500 and 250 hPa
	Tropics:	Z, T and wind at 850 and 250 hPa
Scores	Wind:	Root mean square vector wind error and mean error of wind speed
	Other elements:	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° × 1.5° 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 ±1, ±1.5 and ±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

$$f(D) = \frac{1}{\sqrt{2\pi\sigma_D^2}} \exp\left[-\frac{\log_{10}^2(D/D_m)}{2\sigma_D^2}\right] \quad (3.10.1)$$

where D_m is the mean diameter (set to 0.0316 mm) and σ_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 δt (set to 10 minutes) is given by

$$x(t + \delta t) = x(t) + \delta x \quad (3.10.2)$$

$$y(t + \delta t) = y(t) + \delta y \quad (3.10.3)$$

$$z(t + \delta t) = z(t) + \delta z - V_f \delta t \quad (3.10.4)$$

where δx , δy and δz 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 κ 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 ρ_p is particle density, ρ_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 η_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 λ_a . η_a and λ_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). η_0 (18.18 $\mu\text{Pa s}$) and λ_0 (0.0662 μm) are the standard values for the reference atmosphere ($T_0 = 293.15$ K and $p_0 = 1013.25$ hPa).

The density of volcanic ash particles ρ_p [kg/m^3] is defined as a function of diameter D [m]:

$$\rho_p(D) = \frac{0.48 + \rho_{pm} D}{2.0 \times 10^{-4} + D} \quad (3.10.10)$$

where ρ_{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 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 Λ [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

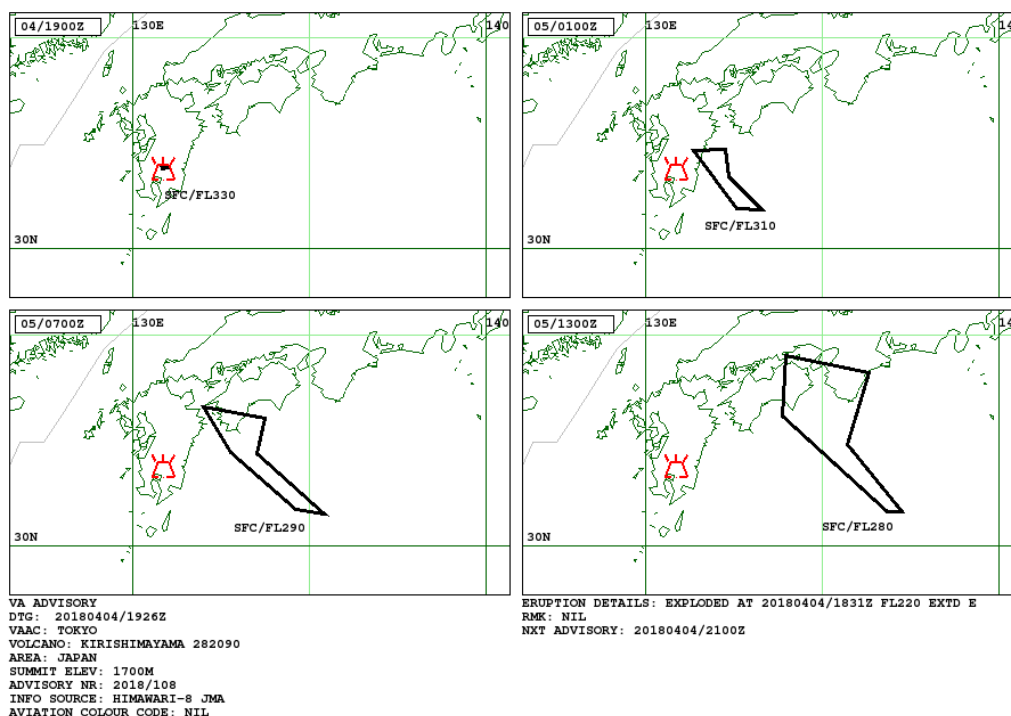


Figure 3.10.2: Sample VAA in graphical form

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 β (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 [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 δt 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 Λ_s [h^{-1}] is considered in addition to that associated with rain, and is given as follows:

$$\Lambda_s \approx 0.1P^{0.3} \quad (3.11.6)$$

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).

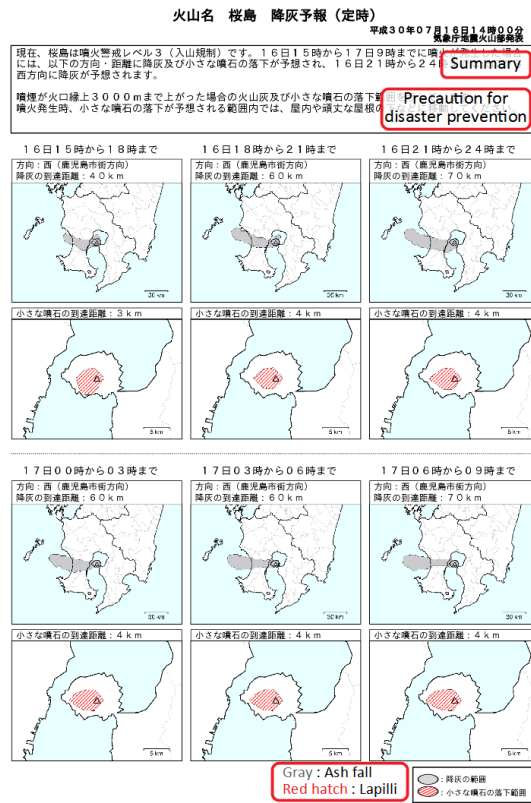


Figure 3.11.1: Sample VAFF (Scheduled)

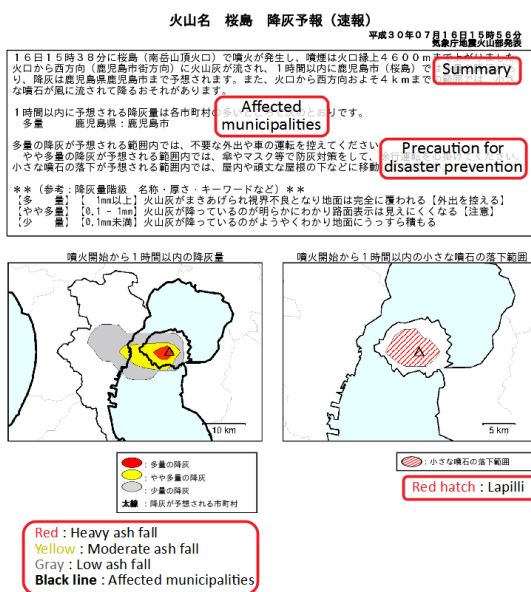


Figure 3.11.2: Sample VAFF (Preliminary)

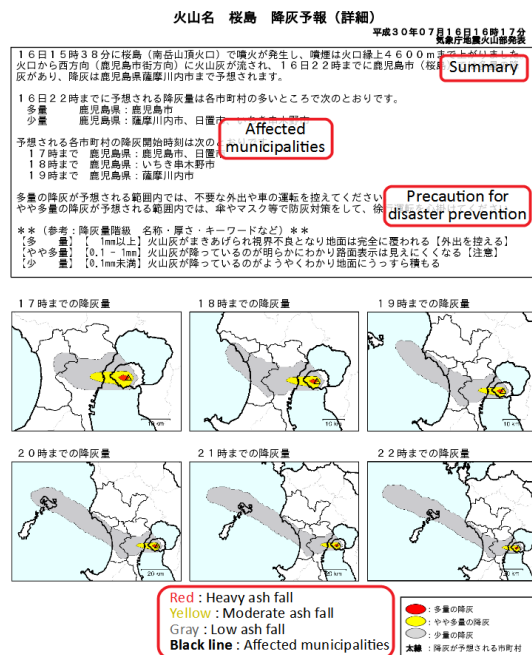


Figure 3.11.3: Sample VAFF (Detailed)

Chapter 4

NWP Application Products

4.1 Summary

NWP results provide useful information both for the general public and for special applications. Against this background, JMA provides these data in real time to its own local observatories, to private companies and to related organizations both in Japan and abroad. Facsimile charts served for a long time as the primary means of NWP output provision, but the development of telecommunication infrastructure and sophisticated visualization systems has now made dissemination based on gridded values the essential method.

In addition to raw NWP data, application products derived from NWP output are also disseminated. An example of such products is information on parameters not explicitly calculated in NWP models, such as probabilistic forecasts, data on turbulence potential for the aviation sector, and error-reduced estimation of NWP output. These are calculated based on the statistical relationship between NWP output and corresponding observations. JMA disseminates Very-short-range (15-hour) Forecasts of Precipitation, Hourly Analysis of horizontal wind and temperature fields, and a number of forms of guidance for short-range forecasting. To support mid-to-long-range forecasting, various forecast charts and gridded data for weekly, monthly and seasonal forecasts are also disseminated.

In the following sections, the specifications of NWP application products and their utilization by JMA are described.

4.2 Weather Chart Services

Facsimile chart provision is a conventional service operated to disseminate the results of NWP in graphical form. Under this service, JMA facsimile charts are sent to national meteorological services via the Global Telecommunication System (GTS) and to ships via the shortwave radio transmission (call sign:JMH).

Table 4.2.1 and Figure 4.2.1 give summaries of weather charts readily accessible by international users (i.e., those provided through GTS and JMH).

The development of the Web complements and supports innovation in these services, and a number of related projects are under way worldwide. JMA takes part in international initiatives such as the Project on the Provision of City-Specific Numerical Weather Prediction (NWP) Products to Developing Countries via the Internet in WMO Regional Association II (RA II) and the Severe Weather Forecast Demonstration Project (SWFDP) involving WMO RAs II and V. The Agency's own projects in this regard include the JMA Pilot Project on EPS Products and SATAID Services on the WMO Information System.

4.3 Gridded Data Products

As part of JMA's general responsibility in meteorological information service provision, gridded data products are distributed to domestic and international users. In line with the requirements of the WMO Information

Table 4.2.1: List of facsimile charts provided through GTS and radio facsimile JMH. Symbols for vertical level: Surf: surface, Trop: tropopause, numbers (850, 700, ... 100): level of pressure in hPa; Symbols for contours: D: dewpoint depression ($T - T_d$), E: precipitation (over the past 12 h for 24 h forecast, and over the past 24 h for others), H: geopotential height, J: wave height, O: vertical velocity (ω), P: MSL pressure, T: temperature, W: wind speed (isotachs), Z: vorticity; Symbols for other drawings: a: wind arrow from gridded data, b: observation plots, d: hatch for area $T - T_d < 3$ K, g: arrow for prevailing wave direction, j: jet axis, m: wave period, s: daily mean sea surface temperature, t: temperature numbers, x: streamlines; Symbols for dissemination and temporal speciality: ' : sent to GTS, * : sent to JMH, ¶: only for 00 UTC, §: only for 12 UTC.

Model	Area (see Figure 4.2.1)	Forecast Time							
		Analysis	12h	24h	36h	48h	72h	96h 120h	144h 168h 192h
GSM	A' (Far East)	500 (H, Z)' 850 (T; a)+700 (O)'		500 (T)+700 (D)'* 500 (H, Z)'* 850 (T; a)+700 (O)'* Surf(P, E; a)'*					
	C (East Asia)	300 (H, W; a, t, b)'¶ 500 (H, T; a, b)'* 700 (H, T; b, d)' 850 (H, T; b, d)'*				500 (H, Z)' 850 (T, a)+700 (O)' § Surf(P, E)'*		Surf(P, E)'* §	
	O (Asia)							500 (H, Z)' § Surf(P)+850 (T)' §	
	Q (Asia-Pacific)	200 (H, W; t, a, j)+Trop(H)' 250 (H, W; t, a)'		250 (H, W; t, a)' 500 (H, W; t, a)'					
	W (West Pacific)	200 (x)' 850 (x)'		200 (x)' 850 (x)'		200 (x)' 850 (x)'			
	D (N Hem.)	500 (H, T)' §							
Ocean Wave	X (Japan)	Surf(J; b, g, m)'*		Surf(J; b, g, m)'*					
	C'' (NW Pacific)	Surf(J; g, m)'*	Surf(J; g, m)'*			Surf(J; g, m)'*			
Sea Surface Temperature	C''2 (NW Pacific)	Surf(s)'*¶							

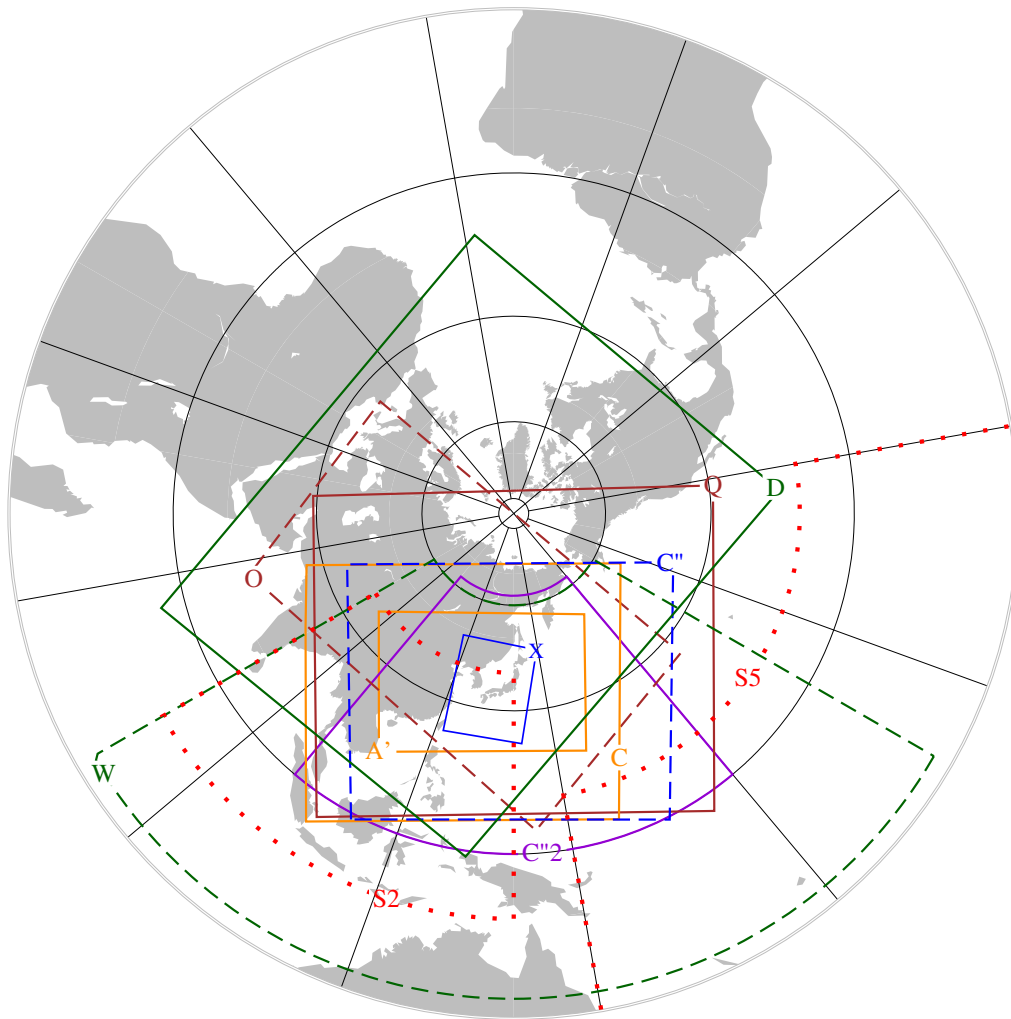


Figure 4.2.1: Areas for charts disseminated through GTS and radio facsimile JMH (symbols A', C, C'', C''2, D, O, Q, W and X). The dotted boxes labeled S2 and S5 are areas of SWFDP products for WMO Regional Associations II and V, respectively (for information).

System (WIS), this data service utilizes both dedicated and public (i.e., Internet) network infrastructure.

The dedicated infrastructure consists of an international part called GTS, together with domestic parts inside JMA (including the Meteorological Satellite Center and the Meteorological Research Institute) and provision to government agencies and the Meteorological Business Support Center, which is in charge of managed services for general users including those in the private sector.

The portal to JMA's international services over the Internet is the website of the Global Information System Centre (GISC) Tokyo¹. Currently, the international service for gridded data products includes the GSM, the Global Ensemble Forecast (One-week) and the Ocean Wave Model as listed in Table 4.3.1.

¹<https://www.wis-jma.go.jp>

Table 4.3.1: List of gridded data products transmitted through GTS and the GISC Tokyo website. Symbols for content: C_L : low cloud amount, C_M : middle cloud amount, C_H : high cloud amount, D: dewpoint depression ($T - T_d$), E: precipitation (from initial time), E_6 : precipitation over the past 6h, G: prevailing wave direction, H: geopotential height, J: wave height, M: wave period, N: total cloudiness, O: vertical velocity (ω), P: MSL pressure, P_S : surface pressure, R: relative humidity, T: temperature, U: eastward wind speed, V: northward wind speed, W: vertical wind shear, X: stream function, Y: velocity potential, Z: vorticity, μ : average over ensemble, σ : standard deviation over ensemble. The symbols $^\circ$, $*$, § , $^\#$, † , ‡ are notes on availability, as detailed in the table.

Model	GSM	GSM	GSM	GSM
Service Channel	GTS and GISC	GTS and GISC	GTS and GISC	GISC
Code form	GRIB Edition 1	GRIB Edition 1	GRIB Edition 1	GRIB Edition 2
Area	Whole Globe	20°S–60°N 60°E–160°W	Whole Globe	Whole Globe and also 5°S–90°N, 30°E–165°W
Resolution	1.25° × 1.25°	1.25° × 1.25°	2.5° × 2.5°	0.5° × 0.5° (0.25° × 0.25° for surface)
Contents	10, 20 hPa 30, 50, 70hPa 100 hPa 150 hPa 200 hPa 250 hPa 300 hPa 400 hPa 500 hPa 600 hPa 700 hPa 800 hPa 850 hPa 900 hPa 925 hPa 950, 975 hPa 1000 hPa Surface	H, U, V, T H, U, V, T H, U, V, T, W [#] H, U, V, T, W [#] H, U, V, T, X, Y, W [#] H, U, V, T, W [#] H, U, V, T, R, O, W [#] H, U, V, T, R, O, W [#] H, U, V, T, R, O, Z, W [#] H, U, V, T, R, O H, U, V, T, R, O, W [#] H, U, V, T, R, O, X, Y H, U, V, T, R, O H, U, V, T, D, O H, U, V, T, R, O	H, U, V, T H, U, V, T H, U, V, T H, U, V, T H [§] , U [§] , V [§] , T [§] , X, Y H, U, V, T H, U, V, T, D H, U, V, T, D H [§] , U [§] , V [§] , T [§] , D [§] , Z H [§] , U [§] , V [§] , T [§] , D [§] , O H [§] , U [§] , V [§] , T [§] , D [§] , O, X, Y H, U, V, T, D P [¶] , U [¶] , V [¶] , T [¶] , D [¶] , E [¶]	H*, U*, V*, T* H [°] , U [°] , V [°] , T [°] H [°] , U [°] , V [°] , T [°] H*, U*, V*, T* H, U, V, T H [°] , U [°] , V [°] , T [°] H, U, V, T, D* [‡] H*, U*, V*, T*, D* [‡] H, U, V, T, D* [‡] H, U, V, T, D H, U, V, T, D H, U, V, T, R, O, X, Y H, U, V, T, R, O H, U, V, T, R, O H, U, V, T, R, O H, U, V, T, R, O H, U, V, T, R, O H, U, V, T, R, O H, U, V, T, R, O H, U, V, T, R, O H, U, V, T, R, O, X, Y P, U, V, T, D* [‡] , E [†] P, U, V, T, R, E [†] , P _S , N, C _L , C _M , C _H
Forecast time range (<i>from–until/interval</i>)	0–84h/6h †: except for analysis #: 0-36h/6h, †: 6-36h/6h	0–84h/6h	0–72h/24h *: Analysis only	0–84h/3h
Extension on 12UTC	96–192h/12h	§: 96–192h/24h ¶: 90–192h/6h	96–192h/24h °: 96–120h/24h	90–264h/6h
Initial times	00, 06, 12, 18UTC	00, 06, 12, 18UTC	00UTC and 12UTC ‡: 00UTC only	00, 06, 12, 18UTC

Model	Global Ensemble Forecast (One-Week)	Ocean Wave Model
Service Channel	GTS and GISC	GTS and GISC
Code form	GRIB Edition 1	GRIB Edition 2
Area	Whole Globe	75°S–75°N, 0°E–359.5°E
Resolution	2.5° × 2.5°	0.5° × 0.5°
Contents	250 hPa 500 hPa 850 hPa 1000 hPa Surface	$\mu U, \sigma U, \mu V, \sigma V$ $\mu H, \sigma H$ $\mu U, \sigma U, \mu V, \sigma V, \mu T, \sigma T$ $\mu H, \sigma H$ $\mu P, \sigma P$ J, M, G
Forecast time range	0–192h/12h	0–84h/6h
Extension on 12UTC	(none)	96–192h/12h
Initial times	00UTC and 12UTC	00,06,12,18UTC

4.4 Very-short-range Forecasting of Precipitation

JMA has operated a fully automated system for analysis and very-short-range forecasting of precipitation since 1988 to provide the following products for monitoring and forecasting of local severe weather conditions:

1. Radar/Raingauge-Analyzed Precipitation (R/A; a type of precipitation analysis)
2. Very-Short-Range Forecasting of Precipitation (VSRF; a type of precipitation forecast)
3. Extended VSRF (ExtVSRF; a type of extended precipitation forecast)

The data in these products show a close correlation with rainfall amounts observed using raingauges. From R/A and VSRF, indices with close ties to landslides, flooding and inundation are produced and used to issue advisories and warnings for these phenomena. The products are provided to local meteorological offices, local governments and broadcasting stations which have responsibility for disaster mitigation. This section outlines how the products are created.

Table 4.4.1: Specifications of Radar/Raingauge-Analyzed Precipitation (R/A), Very-Short-Range Forecasting of Precipitation (VSRF) and Extended VSRF (ExtVSRF)

	R/A	VSRF	ExtVSRF
Spatial resolution	1 km	1 km	5 km
Update interval	10 min.	10 min.	1 hour
Analysis/Forecast element	1 hour accumulated rainfall amount	1 hour accumulated rainfall amount	1 hour accumulated rainfall amount
Forecast time	-	Up to 6 hours ahead	From 7 to 15 hours ahead
Forecast interval	-	1 hour	1 hour
Time required to execute	About 5 min. after observation time	About 8 min. after observation time	About 18 min. after observation time

4.4.1 Analysis of Precipitation (R/A)

R/A enables estimation of accurate one-hour precipitation amounts based on precipitation intensity as observed using radars and rainfall amounts observed using raingauges. It involves the use of data from 46 radars (20 JMA units and 26 operated by other organizations) to cover large areas at a higher spatial resolution than the raingauge network as well as data from up to 10,000 raingauges (1,300 AMeDAS units and 8,700 operated by other organizations) to determine actual amounts of precipitation. JMA uses the Z-R relationship to convert the radar reflectivity factor to precipitation intensity.

The one-hour cumulative precipitation amounts estimated using radar observation usually differ from those observed with raingauges. Radar amount calculation is based on cumulative precipitation intensity over an hour, and values are calibrated using raingauge data to enable more accurate estimation (Makihara 2000). A schematic diagram of this procedure is shown in Figure 4.4.1. First, the radar data are quality controlled to remove ground and sea clutter, bright band and weak echoes unrelated to rainfall on the ground. Satellite imagery and NWP gridded data are used for this purpose. Then, the primary calibration factor over the entire detection range of each radar is calculated by comparing radar precipitation to that of neighboring radars and raingauge data with differences in radar beam height taken into account. Next, the secondary calibration factor is calculated by comparing radar precipitation calibrated with the primary factor to raingauge data at grids where raingauges are located to evaluate local heavy precipitation more accurately. For grids with no raingauges, factor calculation is based on weighted interpolation of values for surrounding grids where raingauges are present. A nationwide composite chart of all radar calibrated precipitation data is created using the maximum value method, in which the largest value is selected if a grid has several data from observation by multiple radars.

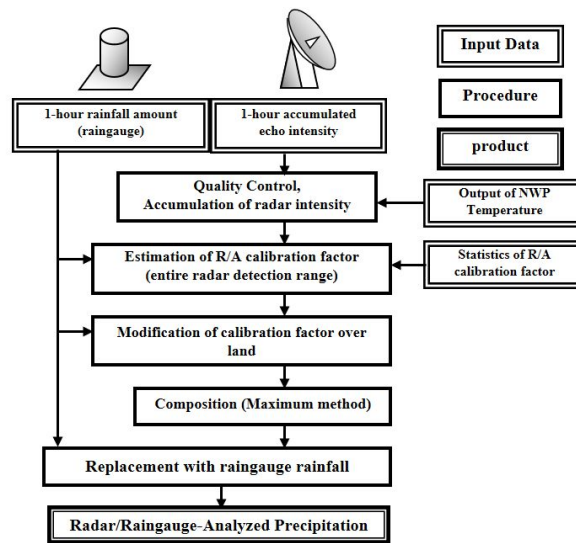


Figure 4.4.1: Flow of Radar/Raingauge-Analyzed Precipitation

4.4.2 Forecasting of Precipitation up to 6 hours ahead (VSRF)

VSRF, which employs the calibrated precipitation intensity determined in the course of R/A as the initial value and is formulated from extrapolation and model forecasts, is a superior estimate of precipitation.

A schematic diagram of the related procedure is shown in Figure 4.4.2. Two methods are used for VSRF. One is the extrapolation of movements of analyzed precipitation systems (i.e., extrapolation forecasts; referred to here as EX6). In the course of extrapolation, the growth and decay of precipitation systems caused by orographic effects and echo intensity trends are taken into account. The other method involves precipitation forecasts of the MSM and LFM, which are available after around two hours from the initial time for the MSM and one hour for the LFM. EX6 is more skillful than MSM and LFM forecasts, although the forecast time is short and skill rapidly diminishes. Meanwhile, the skill of MSM and LFM forecasts degrades gradually and is comparable to that of EX6 when the forecast time reaches a few hours. To produce better model forecasts (referred to here as BLD), JMA introduced a blending technique involving weighted averaging of MSM forecasts and LFM forecasts. This is similar to the merging technique outlined below (see Subsection 4.4.2.2)

The merging technique essentially involves weighted averaging of the EX6 and the BLD. As merging weights are set close to zero for the BLD in the first hour, the products are similar to EX6 output. Thereafter, merging weights for the BLD increase with forecast time. These are determined by comparing the skills of the BLD and the EX6 with R/A. The forecast time steps are two or five minutes, and forecast precipitation is accumulated to produce hourly forecasts up to six hours ahead.

4.4.2.1 Processes assumed in EX6

Extrapolation vectors (i.e., the movement vectors of precipitation systems) are evaluated using a generalized cross-correlation method involving comparison of precipitation system locations at the initial time with those at 0.5, 1, 2 and 3 hours before.

As the seeder-feeder mechanism is assumed to work in regions of orographic updraft, precipitation systems are allowed to grow in the course of extrapolation over such regions. Precipitation systems that have passed over mountains higher than the echo top height are decayed when the following conditions are met:

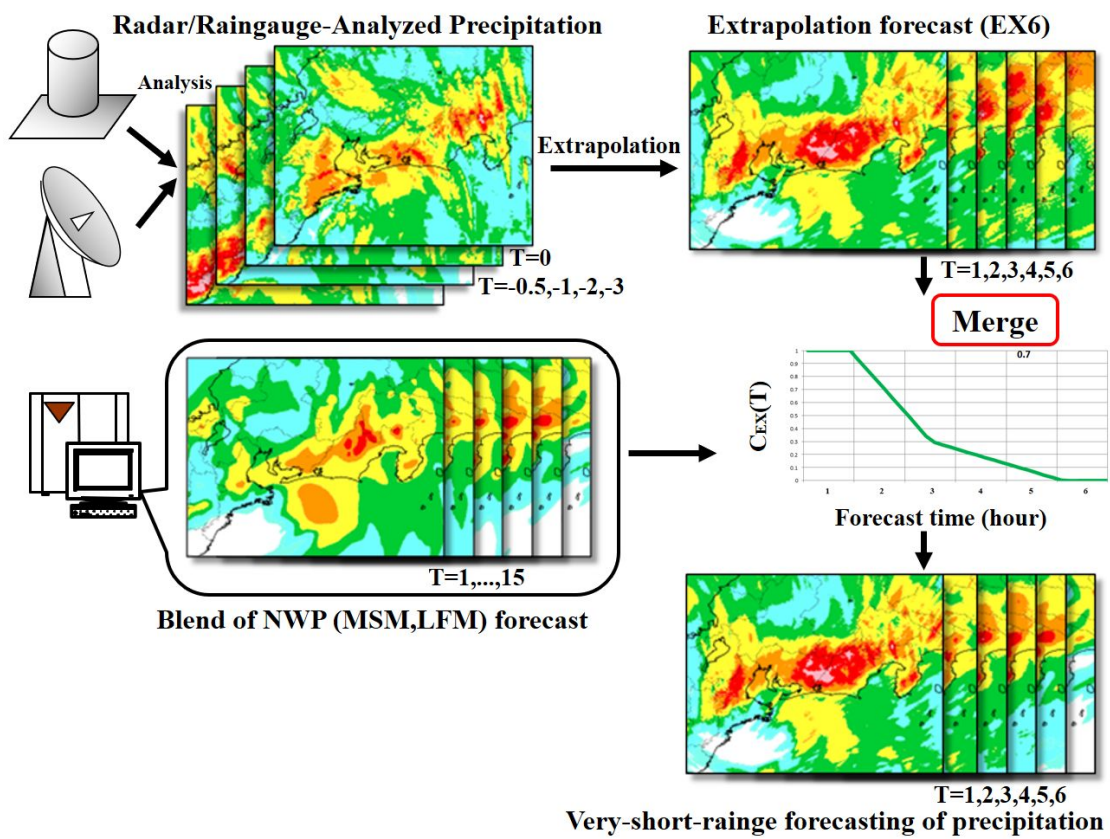


Figure 4.4.2: Schematic representation of very-short-range precipitation forecasting

1. Orographic downslope motion of the rain system is expected from the low-level wind of the MSM,
2. The direction of rain system movement or that of the 700 hPa wind of the MSM is largely parallel to that of the 900 hPa wind of the MSM.

Echo intensity trends can also be determined by comparing the current area average of echo intensity to a past one. Movement vectors for intensity trends are calculated in addition to extrapolation vectors. The vectors move echo intensity trends, which in turn change forecast precipitation.

4.4.2.2 Merging Technique

First, the relative skill of the EX6 and the BLD are estimated. The EX6 from three hours before is verified against the current analysis. For the BLD forecast, the latest available data are verified against the current analysis. The relative reliability coefficient C_{RR} is defined as follows:

$$C_{RR} = \min\left(1, \frac{D_{EX}}{D_{BLD}}\right) \quad (4.4.1)$$

where D_{EX} is the two-dimensional pattern distance, or the two-dimensionally extended Levenshtein distance, between the EX6 and the analysis, and D_{BLD} is the two-dimensional pattern distance between the BLD forecast and the analysis.

The relative weight of the extrapolation forecast $C_{EX}(T)$ is then determined using C_{RR} and the function $C(T, BR)$, where BR ² is blend reliability and T denotes the forecast time as indicated in the merge process of Figure 4.4.3

$$C_{EX}(T) = 1 - C_{RR} \cdot (1 - C(T, BR)) \quad (4.4.2)$$

Finally, the merged forecast $R_{MRG}(T)$ is calculated with the following equation:

$$R_{MRG}(T) = C_{EX}(T) \cdot R_{EX}(T) + (1 - C_{EX}(T)) \cdot R_{BLD}(T) \quad (4.4.3)$$

where $R_{EX}(T)$ denotes extrapolation forecasting of precipitation at the forecast time T and $R_{BLD}(T)$ denotes the BLD forecast of precipitation from the latest initial time at the same valid time T .

4.4.2.3 Example and Verification Score of R/A and VSRF

R/A and VSRF examples are shown in Figure 4.4.4. R/A for the Kinki region in the central western area of Japan for 14:50 UTC on 23 August 2018 is shown in the left panel (a), and the three-hour VSRF forecast for the same valid time (i.e., initial time 11:50 UTC 23 on August 2018) is shown in the right panel (b). The intense rain band is well forecasted.

VSRF accuracy has been statistically verified with the critical success index (CSI)³. Forecasts are compared with precipitation analysis after both fields are averaged in 5×5 km grids. Indices from 1- to 6-hour forecasts for July 2018 are shown in Figure 4.4.5.

It can be seen that scores deteriorate with longer forecast times. EX6 maintains its superiority to BLD up to three hours, but the relationship reverses after this time. VSRF exhibits the best performance for all forecast times.

4.4.3 VSRF Forecast Range Extension to 15 hours (ExtVSRF)

4.4.3.1 Basic Concept of ExtVSRF

In June 2018, JMA launched an extended VSRF forecast called ExtVSRF to support early judgement on the need for evacuation and other measures by clarifying the tendency of rainfall toward dawn when heavy rain

² $BR \equiv \text{Score}(BLD) / (\text{Score}(BLD) + \text{Score}(EX6))$

³The CSI is the number of correct "Yes" forecasts divided by the total number of occasions on which the event was forecast and/or observed. It is also cited as the threat score (see Subsection A.2.9).

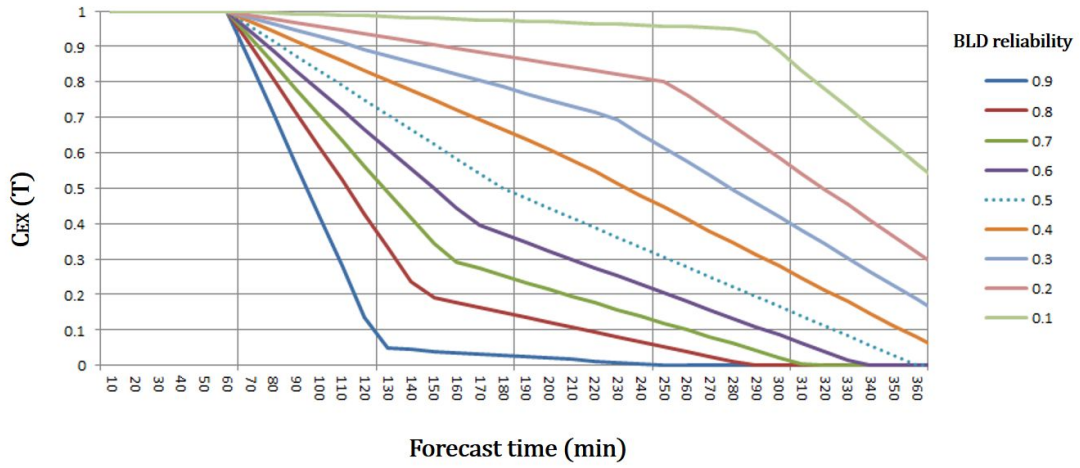


Figure 4.4.3: Dependence of $C_{EX}(T)$ on forecast time and BLD reliability

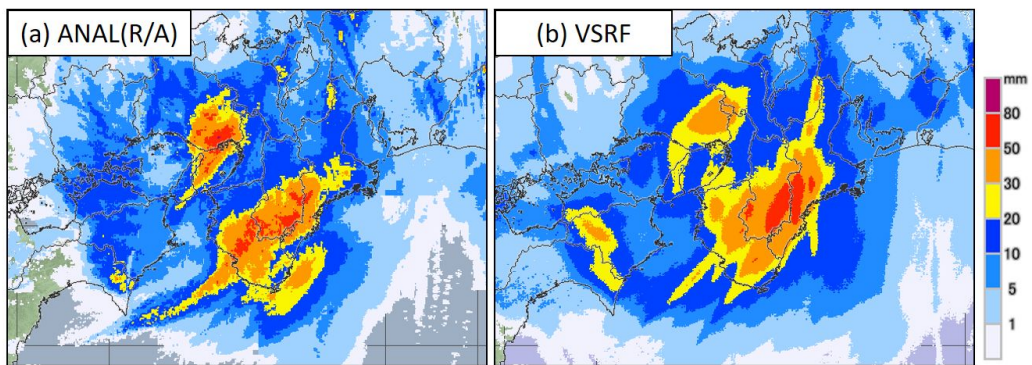


Figure 4.4.4: (a) the Radar/Raingauge-Analyzed Precipitation for 11:50 UTC on 23 August 2018, and (b) 3-hour VSRF forecast of precipitation for the same valid time.

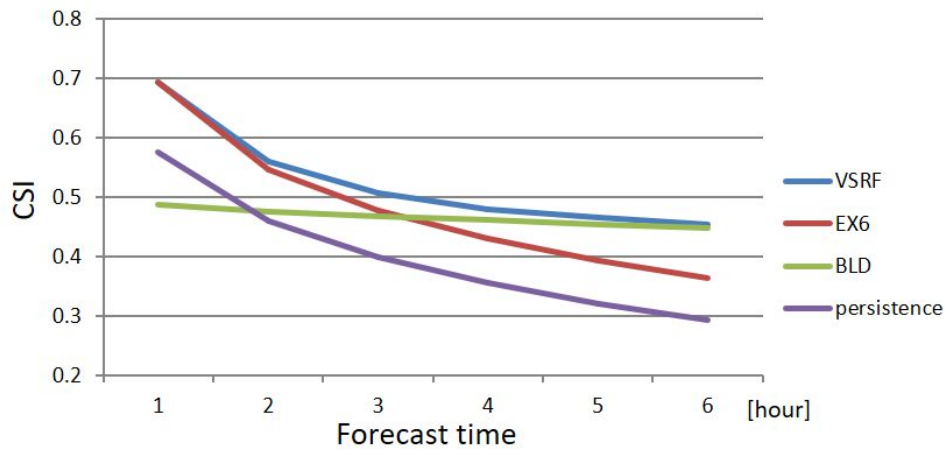


Figure 4.4.5: CSI of very-short-range forecasting (VSRF) of precipitation averaged within 5×5 km grids over land for July 2018, together with that of EX6, BLD and persistence forecasting. The threshold value is 1.0mm.

falls in the evening. The extended forecast facilitates understanding of overall precipitation distribution as a trend, and was developed as a separate product from VSRF.

The flow of ExtVSRF calculation is shown in Figure 4.4.6. The forecast is derived from a combination of MSM precipitation amount forecasts, MSM Guidance for mean and maximum precipitation amounts and LFM Guidance for maximum precipitation amounts, and is not merged with the EX6 because the latter’s precision from 7 to 15 hours ahead is significantly poorer than that produced by the combination of these guidance forecasts.

The latest available guidance forecasts for mean precipitation amounts and maximum precipitation amounts are divided into two groups, and are verified with current analysis using the fraction skill score (FSS)⁴. The forecast with the best score from each group is chosen and mixed with the weighted average.

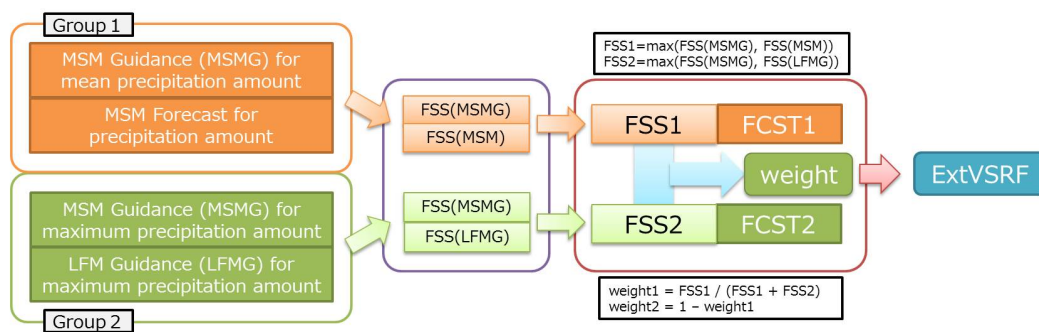


Figure 4.4.6: Extended VSRF flow chart

⁴In FSS, verification incorporates consideration of position gaps and indicates the accuracy of precipitation distribution (see Subsection A.2.12).

4.4.3.2 ExtVSRF Verification Score and Example

An example of R/A and ExtVSRF is shown in Figure 4.4.7. The R/A for the western area of Japan at 23 UTC 19 June 2018 (08 JST 20 June 2018) is shown in the left panel (a), and the 14-hour forecast of ExtVSRF at the same valid time (i.e., initial time 09 UTC 19 June 2018 (18 JST 19 June 2018)) is shown in the right panel (b). The intense rain band is well forecast, and it is understandable that this morning rainfall will be heavy at the moment of previous early evening.

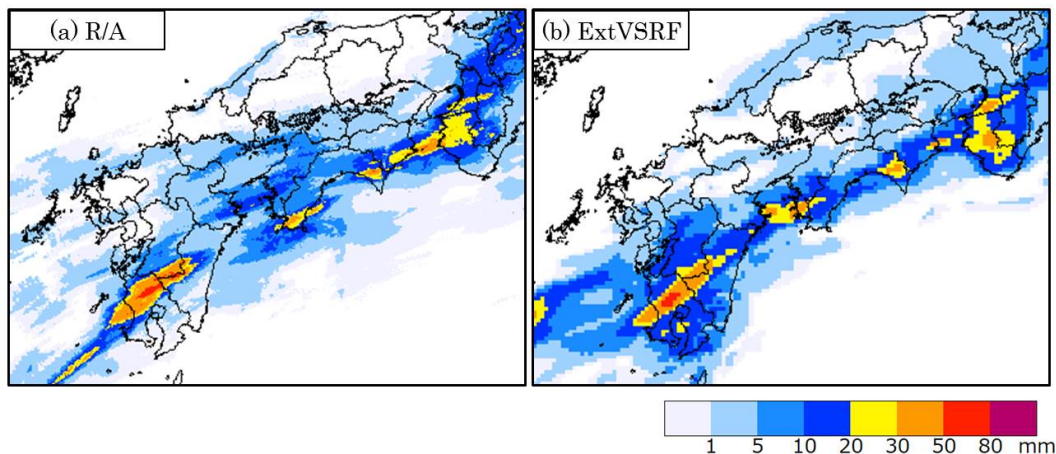


Figure 4.4.7: (a) Radar/Rain gauge-Analyzed Precipitation at 23 UTC on 19 June 2018, and (b) 14-hour ExtVSRF forecast of precipitation for the same valid time.

The accuracy of ExtVSRF has been statistically verified with FSS data. Forecasts are compared with precipitation analysis in 5×5 km grids. Indices from 1- to 15-hour forecasts for July/October 2017 and January 2018 are shown in Figure 4.4.8 together with those of VSRF and MSM Guidance forecasts for mean precipitation amounts. It can be seen that ExtVSRF is superior to MSM Guidance for all forecast times.

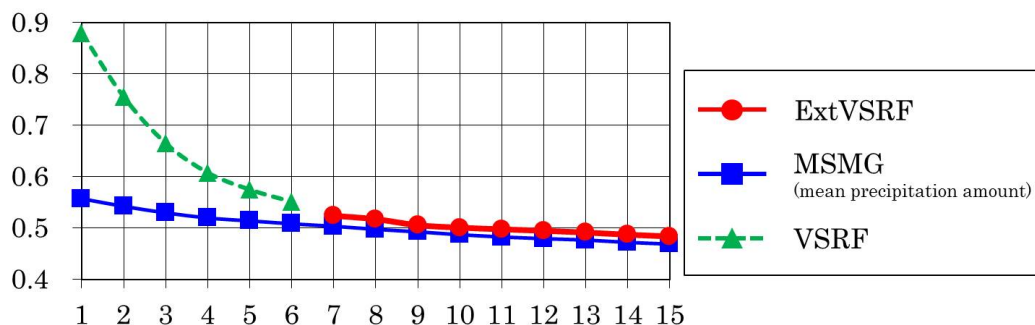


Figure 4.4.8: FSS of the Extended VSRF (ExtVSRF) for June 2017, October 2017 and January 2018 together with that of VSRF and MSM Guidance forecast for mean precipitation amount. The threshold value is 1 mm.

4.5 Hourly Analysis

Hourly analysis provides hourly gridded data on three-dimensional temperature and wind, to help forecasters in continuous atmospheric monitoring. Meteorological imagery products are also provided to the aviation sector online.

The configuration of the hourly analysis system, which is based on 3D-Var, is given in Table 4.5.1. Analysis involves the use of the latest Meso-scale Model (MSM; Section 3.5) forecast as the first guess (a 2-4 hour forecast depending on the analysis time). The domain of the hourly analysis covers Japan and its surrounding area at a horizontal resolution of 5 km with 721×577 grid points. There are 48 vertical layers defined for hybrid terrain-following coordinates (Subsection 3.5.3) with the top of the domain at 21,801 m.

The observations assimilated in the analysis are from AMeDAS (the Automated Meteorological Data Acquisition System; Japan surface station data on wind and temperature), Wind Profiler (wind), Weather Doppler radar (radial velocity), AIREP, AMDAR (wind and temperature), and AMVs from Himawari-8 (Atmospheric Motion Vectors, wind). The data cut-off time is set to 18 minutes past the hour so that products can be distributed by 30 minutes past the hour.

In order to obtain a good fit to surface observations on land, the 3D-Var in the hourly analysis adopts a short background error correlation distance and a small observation error on the surface. Thus, the surface field on land typically has large local increments. Outlined below are modifications of the 3D-Var scheme and additional post-processing performed to address this situation.

Table 4.5.1: Hourly analysis specifications

Analysis time	Every hour (on the hour)
Analysis scheme	3D-Var
Data cut-off time	18 minutes past the hour
First guess	2, 3 or 4-hour forecast by the MSM
Domain configuration	Japan and its surrounding area Lambert projection: 5 km at 60°N and 30°N, 721×577 Grid point (1, 1) is at the northwest corner of the domain. Grid point (489, 409) is at 140°E, 30°N.
Vertical coordinate	z - z^* hybrid
Vertical levels	48 levels up to 22 km
Analysis variables	Wind, temperature, surface wind and surface temperature
Observations (as of 31 December 2017)	AMeDAS, Wind Profiler, Weather Doppler radar (radial velocity), AIREP, AMDAR, and AMVs from Himawari-8
Post-processing	Surface filtering (followed by adjustment of the increment within the boundary layer)
Product distribution	Before 30 minutes past the hour

- The control variables at the bottom level are treated as uncorrelated to those at other levels in 3D-Var analysis because the large analysis increments at the surface need to be adjusted independently with a surface filter as described below.
- After 3D-Var analysis is complete, a surface filter is applied to the surface temperature and wind fields to attenuate surface increments over the sea with distance from the coastline and reduce excessive increments in sea regions near the coastline within the range of correlation from land observations. Analysis increments on the surface and in the upper air are not consistent at this point.
- After application of the surface filter, the increments on the surface and in the upper air are merged in each vertical column within the boundary layer of the first guess to make the surface and upper-air increments consistent. The weight of the surface increment attenuates with height above the ground, and approaches zero around the top of the boundary layer.

4.6 Guidance for Short-range Forecasting

4.6.1 Overview

JMA provides various kinds of forecast guidance to support forecasters in their work. The objective is to reduce systematic errors in NWP output (via guidance on variables such as temperature and wind) and to enable the derivation of quantitative values not directly calculated in NWP models (via guidance on variables such as probability and weather categories). Table 4.6.1 lists the guidance provided for short-range forecasting (up to 84 hours) at JMA.

To cope with frequent model upgrades, JMA uses methods that allow ongoing adjustment of statistical equations. These adaptive approaches, which are based on Kalman filtering and a neural network, replaced the previous non-successive multivariate regression method in 1996 and have since been applied to various guidance values.

The Kalman filtering and neural network used in the guidance system are outlined in Subsection 4.6.2 and Subsection 4.6.3, respectively, and utilization of guidance at forecasting offices is summarized in Subsection 4.6.4.

4.6.2 Guidance Based on Kalman Filtering

4.6.2.1 Kalman Filtering

As a statistical post-processing method for NWP output, JMA developed guidance using Kalman filtering (KF) on the basis of earlier work conducted by Persson (1991) and Simonsen (1991). KF evolves coefficients based on the following equations:

$$y_{\tau} = \mathbf{c}_{\tau} \mathbf{X}_{\tau} + v_{\tau} \quad (4.6.1)$$

$$\mathbf{X}_{\tau+1} = \mathbf{A}_{\tau} \mathbf{X}_{\tau} + \mathbf{u}_{\tau} \quad (4.6.2)$$

where y represents predictand (i.e., the target of forecasting), \mathbf{c} represents predictors ($1 \times n$ matrix), \mathbf{X} represents coefficients ($n \times 1$ matrix) with covariance matrix \mathbf{Q} ($n \times n$ matrix), v represents observation noise with variance \mathbf{D} , \mathbf{u} represents system noise ($n \times 1$ matrix) with the covariance matrix \mathbf{U} ($n \times n$ matrix), and \mathbf{A} ($n \times n$ matrix) describes the evolution of the coefficients in time.

Eq. (4.6.1) relates to observation, and is a linear expression relating the predictand and predictors. Eq. (4.6.2) is a system expression denoting the time evolution of the coefficients. In guidance, the time evolution matrix \mathbf{A} can be treated as a unit matrix:

$$\mathbf{A}_{\tau} \equiv \mathbf{I} \quad (4.6.3)$$

The objective of KF is to determine the most likely estimation of the coefficients $\mathbf{X}_{\tau+1/\tau}$, whose subscript denotes an estimate at $\tau + 1$ based on observation at τ . In contrast, the single subscripts in these two equations denote true values at τ . $\mathbf{X}_{\tau+1/\tau}$ is determined from the previous estimate $\mathbf{X}_{\tau/\tau-1}$ and the forecast error:

Table 4.6.1: Parameters of the guidance products for short-range forecasting

Parameters	Target	Model	Forecast hour	Method*	
Categorized weather over 3 hours (fair, cloudy, rainy, sleety, snowy)	Grids	20km	GSM	KT=6, 9, 12, ..., 81, 84	NN
		5km	MSM	KT=3, 6, 9, ..., 36, 39	
Mean precipitation amount over 3 hours	Grids	20km	GSM	KT=6, 9, 12, ..., 81, 84	KF & FBC
		5km	MSM	KT=3, 6, 9, ..., 36, 39	
Maximum precipitation amount over 1, 3 hours	Grids	20km	GSM	KT=6, 9, 12, ..., 81, 84	NN
		5km	MSM	KT=3, 6, 9, ..., 36, 39	
Maximum precipitation amount over 24 hours	Grids	20km	GSM	KT=27, 30, 33, ..., 81, 84	MLR
		5km	MSM	KT=24, 27, 30, 33, 36, 39	
Mean precipitation amount over 1 hours	Grids	5km	LFM	KT=1, 2, 3, ..., 8, 9	LAF & FBC
Maximum precipitation amount over 1 hours	Grids	5km	LFM	KT=1, 2, 3, ..., 8, 9	—
Probability of precipitation over 6 hours > 1mm/6h	Grids	20km	GSM	KT=9, 15, 21, ..., 75, 81	KF
		5km	MSM	KT=6, 12, 18, ..., 23, 39	
Maximum temperature in the daytime (09-18 local time)	Points	AMeDAS	GSM	Today to 3 days after	KF
			MSM	Today and tomorrow	
Minimum temperature in the morning (00-09 local time)	Points	AMeDAS	GSM	Today to 3 days after	KF
			MSM	Today and tomorrow	
Time-series temperature	Points	AMeDAS	GSM	KT=3, 4, 5, ..., 81, 84	KF
			MSM	KT=1, 2, 3, ..., 38, 39	
Time-series humidity	Points	SYNOG	GSM	KT=3, 4, 5, ..., 81, 84	KF
			MSM	KT=1, 2, 3, ..., 38, 39	
Wind speed and direction	Points	AMeDAS	GSM	KT=3, 6, 9, ..., 81, 84	KF & FBC
			MSM	KT=1, 2, 3, ..., 38, 39	
Maximum wind speed and direction over 3hours	Points	AMeDAS	GSM	KT=3, 6, 9, ..., 81, 84	KF & FBC
			MSM	KT=3, 6, 9, ..., 36, 39	
Daily minimum humidity	Points	SYNOG	GSM	Today to 3 days after	NN
			MSM	Today and tomorrow	
Snowfall amount over 6,12 hours	Points	AMeDAS	GSM	KT=9, 12, 15, ..., 81, 84	NN
			MSM	KT=6, 9, 12, ..., 36, 39	
Snowfall amount over 24 hours	Points	AMeDAS	GSM	KT=27, 30, 33, ..., 81, 84	NN
			MSM	KT=24, 27, 30, 33, 39	
Snowfall amount over 3,6,12 hours	Grids	5km	GSM	KT=9, 12, 15, ..., 81, 84	DIAG
			MSM	KT=6, 9, 12, ..., 36, 39	
Snowfall amount over 24 hours	Grids	5km	GSM	KT=27, 30, 33, ..., 81, 84	DIAG
			MSM	KT=24, 27, 30, 33, 36, 39	
Probability of thunderstorm over 3 hours	Grids	20km	GSM	KT=6, 9, 12, ..., 81, 84	LR & LAF
			MSM	KT=6, 9, 12, ..., 36, 39	

* NN: neural network, KF: Kalman filter, FBC: frequency bias correction, MLR: multiple linear regression, DIAG: diagnostic method, LR: logistic regression, LAF: lagged averaged forecast method

$$\mathbf{X}_{\tau+1/\tau} = \mathbf{X}_{\tau/\tau} \quad (4.6.4)$$

$$= \mathbf{X}_{\tau/\tau-1} + \mathbf{K}_{\tau}(\mathbf{y}_{\tau} - \mathbf{c}_{\tau}\mathbf{X}_{\tau/\tau-1}) \quad (4.6.5)$$

where

$$\mathbf{K}_{\tau} = \mathbf{Q}_{\tau/\tau-1}\mathbf{c}_{\tau}^T(\mathbf{c}_{\tau}\mathbf{Q}_{\tau/\tau-1}\mathbf{c}_{\tau}^T + D_{\tau})^{-1} \quad (4.6.6)$$

\mathbf{K}_{τ} is the kalman gain. \mathbf{Q}_{τ} the covariance of \mathbf{X} , is updated as follows:

$$\mathbf{Q}_{\tau+1/\tau} = \mathbf{Q}_{\tau/\tau} + \mathbf{U}_{\tau} \quad (4.6.7)$$

$$= \mathbf{Q}_{\tau/\tau-1} - \mathbf{K}_{\tau}\mathbf{c}_{\tau}\mathbf{Q}_{\tau/\tau-1} + \mathbf{U}_{\tau} \quad (4.6.8)$$

Eq. (4.6.4) and Eq. (4.6.7) are derived from Eq. (4.6.2) and Eq. (4.6.3), respectively.

Finally, the forecast value is calculated with the updated coefficients and predictors at $\tau + 1$;

$$y_{\tau+1/\tau} = \mathbf{c}_{\tau+1} \mathbf{X}_{\tau+1/\tau} \quad (4.6.9)$$

For some forecast parameters (such as temperature), the predictand y is defined as the difference between NWP output and observations, while for others (such as precipitation amount), y is the observation itself.

In the forecast guidance system with KF, D_τ in Eq. (4.6.6) and \mathbf{U}_τ in Eq. (4.6.8) are treated as empirical parameters for control of the adaptation rate.

4.6.2.2 Frequency Bias Correction

With KF, the most likely estimation of the predictand that minimizes the expected root mean square error is obtained. However, the output has a tendency toward lower-than-actual frequency for forecasting of relatively rare events such as strong wind and heavy rain. To compensate for this unfavorable characteristic, a frequency bias correction scheme is applied to the KF output.

The basic concept involves multiplying the estimation of KF, y , by a correction factor $F(y)$ to obtain the final output y^b :

$$y^b = y \cdot F(y)$$

To determine $F(y)$, certain thresholds t^i are set to span the given observation data. The corresponding thresholds f^i for the forecast data set are then calculated so that the number of observation data smaller than t^i approximates to that of forecast data smaller than f^i . Finally, the correction factors are computed as follows:

$$F(f^i) = t^i / f^i$$

$$F(y) \text{ for } f^i < y < f^{i+1} \text{ is linearly interpolated between } F(f^i) \text{ and } F(f^{i+1}).$$

Since KF is an adaptive method, f^i is also updated each time the observation y_τ corresponding to the estimates of KF $y_{\tau/\tau-1}$ is available. The update procedure is as follows:

$$f_{\tau+1}^i = \begin{cases} f_\tau^i(1 + \alpha) & \text{if } y_\tau < t^i \text{ and } y_{\tau/\tau-1} > f^i \\ f_\tau^i(1 - \alpha) & \text{if } y_\tau > t^i \text{ and } y_{\tau/\tau-1} < f^i \\ f_\tau^i & \text{otherwise} \end{cases}$$

where α is an empirical parameter used to determine the adaptation rate.

4.6.2.3 Example of Guidance (GSM) Based on Kalman Filtering (3-hour Precipitation Amount)

In this guidance, the predictand is the observed three-hour cumulative precipitation amount averaged within a 20×20 km square. The following parameters derived from NWP output are used as predictors.

1. NW85: NW – SE component of wind speed at 850hPa
2. NE85: NE – SW component of wind speed at 850hPa
3. SSI: Showalter's stability index (between 850 and 500hPa)
4. OGES: Orographic precipitation index
5. PCWV: Precipitable water content \times Wind speed at 850hPa \times Vertical p-velocity at 850hPa
6. QWX: Σ (Specific humidity \times Vertical p-velocity \times Relative humidity) between 1000 and 300hPa
7. EHQ: Σ (Excess from reference humidity \times Specific humidity \times Depth of wet layer) between 1000 and 300hPa
8. DXQV: Precipitation index in winter synoptic pattern

9. FRR3: Mean precipitation amount over three hours based on the GSM

In the sample precipitation forecasts shown in Figure 4.6.1, the NWP model (GSM) predicted very little or no precipitation in area M, where observation shows light precipitation. The guidance predicted light precipitation in this area, representing a better results. Examination of the coefficient values shows an orographic effect relating to the OGES enhanced precipitation amount in this area.

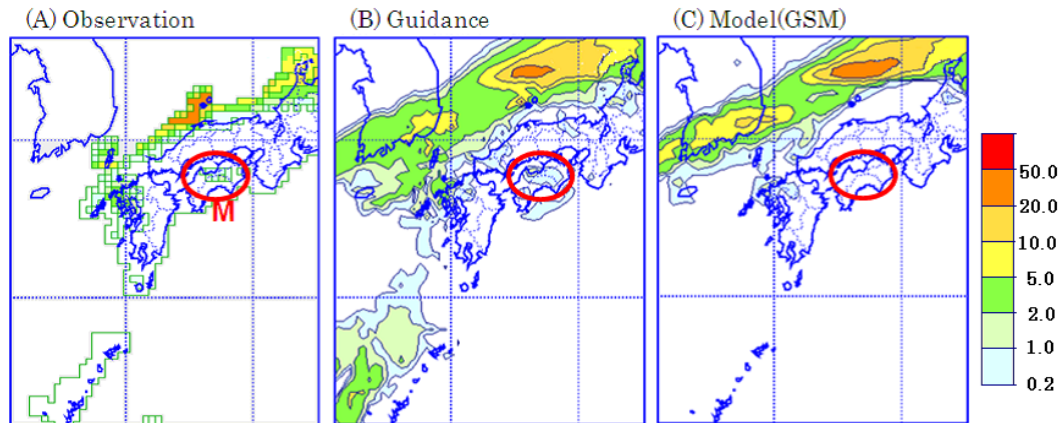


Figure 4.6.1: Mean precipitation amount over three hours: (A) observation, (B) guidance forecast, (C) NWP model (GSM) forecast

4.6.3 Guidance Based on a Neural Network

4.6.3.1 Neural Network

The neural network (NN) approach involves machine learning to support analysis of non-linear relationships between predictors and predictands (Yanagino and Takada 1995). In this method, multiple layers of neurons are linked to construct a hierarchical neural network as shown in Figure 4.6.2. The first layer is called the input layer and the last layer is called the output layer. Those between them are called the hidden layers.

A signal put into the input layer is propagated to the next layer via inter-neuron connections. The signal is subjected to simple processing by the neurons of the receiving layer prior to its propagation to the next layer. This process is repeated until the signal reaches the output layer.

A schematic diagram of a neuron is shown in Figure 4.6.3. The input of each neuron is a weighted sum of the outputs of other neurons, and the output is a function of its input. This is called an activation function, and a sigmoid function as shown in Figure 4.6.4 is usually used for this purpose.

NN weights are iteratively adjusted using numerous sets of input/output data. The most popular adjustment method involves the back-propagation of error algorithm described as follows:

1. First, weights are initialized with randomized values.
2. The network calculates output values using a given set of input values.
3. Weights are adjusted to make the NN output close to the supervisor data (correct values of the output variable).
4. The processes of 2 and 3 are iterated until the error measure falls below a specified value or the number of iterations reaches a specified maximum.

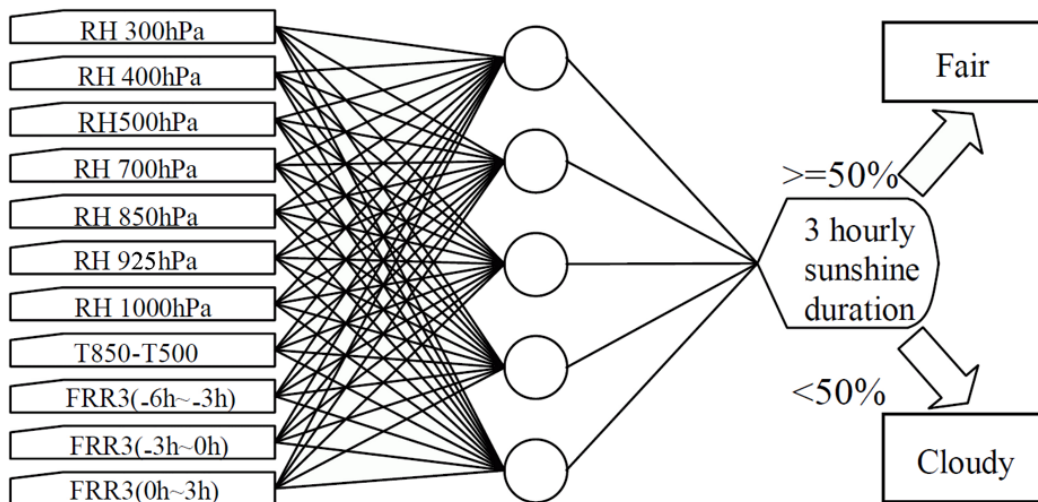


Figure 4.6.2: Neural network for fair/cloudy determination. RH: relative humidity, FRR3: precipitation over 3 hours

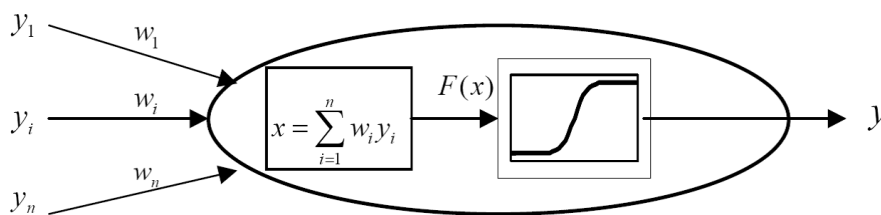


Figure 4.6.3: Schematic representation of a neuron

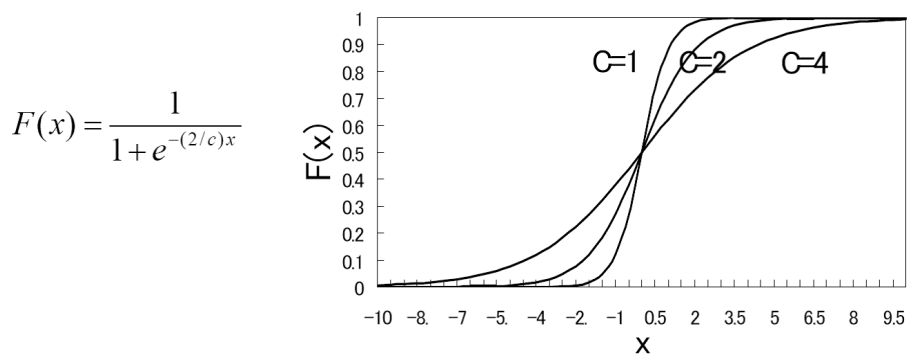


Figure 4.6.4: Examples of the sigmoid function

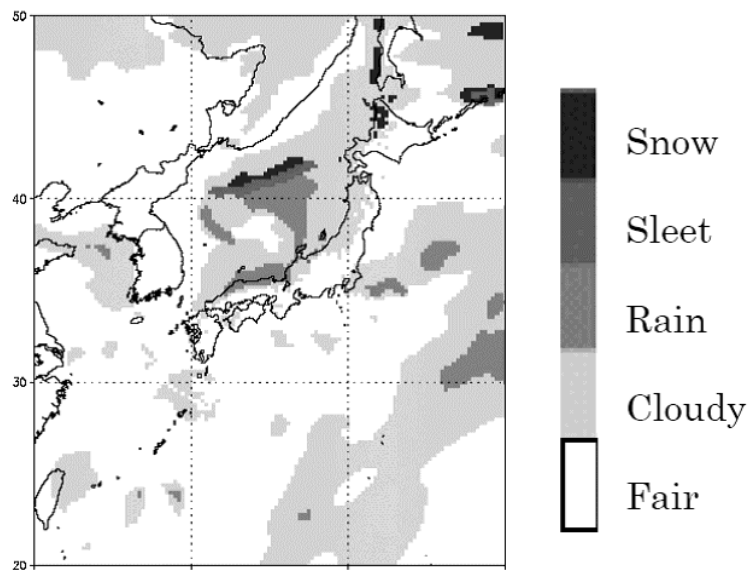


Figure 4.6.5: Sample output of categorized weather guidance

4.6.3.2 Example of Guidance Based on a Neural Network (Categorized Weather)

In the forecast guidance system, a neural network model is constructed at each grid or observation point from sets of NWP output and observed weather elements. Categorized weather is one of the forecast guidance parameters to which the NN is applied. Figure 4.6.5 shows an example of output categorized weather guidance, in which an NN model is used to derive sunshine duration, which is in turn used to determine non-precipitation weather categories (fair or cloudy). The NN is constructed for each AMeDAS station, and output values (three-hourly sunshine durations) are interpolated to grid points. Precipitation weather categories (rain, sleet, snow) are determined from precipitation amount guidance and precipitation type guidance. The constitution of the sunshine duration NN model is shown in Figure 4.6.2, and its characteristics are summarized as follows:

1. The model incorporates a three-layer feed-forward neural network.
2. A linear activation function is used in the output layer and sigmoid activation functions are used in the hidden layer.
3. In the learning processes, NWP output is used as input data and sunshine durations observed at each AMeDAS point are used as supervisor data.
4. The weights of the network are modified when the observation data are obtained.

4.6.4 Utilization of Guidance at Forecasting Offices

Forecast guidance products are disseminated to observatories and used for drafting weather forecasts in editing software. Figure 4.6.6 shows an example of the data entry screen. Forecasters revise elements (time-series data of weather categorization, PoP, temperature etc.) on the display consideration of current weather conditions and empirical knowledge. The processed data are then composed to create a forecast bulletin and provided to users.

An algorithm incorporating the steps shown below is used to draft weather forecast bulletins automatically.

1. Three-hourly dominant weather categories are derived from the majority of weather categorization on grids in the forecast area.

- The weather forecast bulletin for the day is derived from the sequence of three-hourly dominant weather categories over the forecast area. Examples of the algorithm are shown in Table 4.6.2.

The screenshot shows a web-based interface for editing weather forecast data. The main content is a grid of data points for a specific date and time. Annotations with arrows point to various parts of the interface:

- List of input data:** Points to the top right section containing a table of parameters and their initial values or save times.
- Time series data:** Points to the top row of the main data grid, which includes time slots (e.g., ~18, ~21, ~24).
- Dominant weather categories:** Points to the '天気' (Weather) row in the main data grid.
- PoP:** Points to the '降水確率' (Precipitation Probability) row.
- Wind:** Points to the '風向' (Wind Direction) and '風速' (Wind Speed) rows.
- Local weather:** Points to the '特定域' (Specific Area) section, which includes '陸上' (Land) and '海上' (Sea) weather conditions.
- Local winds:** Points to the '風向' and '風速' rows within the '特定域' section.
- Ocean wave height:** Points to the '波の高さ' (Wave Height) row.
- Swell condition:** Points to the '波浪種別' (Wave Type) row.
- Temperature:** Points to the '気温' (Temperature) row, which includes '最高' (Maximum) and '最低' (Minimum) values.
- highs/lows:** Points to the '独自子報文' (Independent Report) row, which contains numerical data.
- Draft of forecast bulletin:** A large arrow points from the right side of the data grid towards the right edge of the window.

Figure 4.6.6: Sample data entry screen in forecast editing software

Table 4.6.2: Sample of the algorithm for weather forecast bulletin drafting

Sequence of 3-hourly categorized weather*								Draft of a weather forecast bulletin
0 - 3	3 - 6	6 - 9	9 - 12	12-15	15-18	18-21	21-24	
F	F	F	F	C	F	F	F	Fair
R	R	R	R	R	S	S	S	Rain, snow from the evening
C	R	F	R	C	F	R	C	Cloudy, occasional rain
C	R	C	C	C	C	R	C	Cloudy, rain in the morning and the evening

* F:Fair C:Cloudy R:Rain S:Snow

4.7 Application Products for Aviation Services

4.7.1 Aerodrome Forecast Guidance

Aerodrome forecast guidance (TAF guidance) is derived from the output of the MSM eight times a day, providing hourly predictions up to 39 hours ahead. The predicted values in this guidance are listed in Table 4.7.1.

Table 4.7.1: Parameters of TAF guidance. (for 95 airports in Japan)

Parameters	TAF guidance
Visibility	Minimum and mean visibility during an hour Probability of minimum visibility < 5km and 1.6km during 3 hours
Cloud	Cloud amount and height of 3 layers at minimum ceiling during an hour Probability of minimum ceiling < 1000ft and 600ft during 3 hours
Weather	Categorized weather every hour
Temperature	Maximum temperature in the daytime, minimum temperature in the morning and temperatures every hour
Wind	Wind speed and direction every hour Wind speed and direction of hourly maximum peak wind
Gust	Probability of gust during 3 hours Gust speed and direction of hourly maximum peak gust
Thunderstorm	Probability of thunder during 3 hours
Snow	Snowfall amount during 3 hours

4.7.1.1 Visibility

Minimum visibility and probability in minimum visibility guidance (VIS) are based on statistical interpretation of NWP output. VIS is calculated using linear equations whose coefficients are adapted via Kalman filtering (see Subsection 4.6.2.1) with the predictors and METAR reports. VIS consists of three linear equations classified by weather type (rain, snow, no precipitation). The following predictors from the output of the MSM are used for each equation:

- No precipitation: $(1 - RH)^{1/2}, Qc^{1/2}$, where RH is surface relative humidity ($0 \sim 1$), Qc is cloud water content near surface(kg/kg).
- Rain: $RR^{1/2}, (1 - RH)^{1/2}, Qc^{1/2}$, where RR is precipitation amount over an hour (mm).
- Snow: $RR^{1/2}, (1 - RH)^{1/2}, VV \times T$, where VV is surface wind speed (m/s, only < 15), T is surface temperature ($^{\circ}C, -10 < T < 0$)

Frequency bias correction (see Subsection 4.6.2.2) is applied to parameters calculated using the three equations. One parameter is subsequently chosen depending on the weather category predicted in weather guidance (described later).

4.7.1.2 Cloud

TAF cloud guidance involves statistical interpretation of NWP output. First, cloud amounts in each of 38 layers (0, 100, . . . , 1000, 1500, . . . , 5000, 6000, . . . , 10000, 12000, . . . , 30000 ft) are calculated using an NN (see Subsection 4.6.3), and the lowest three cloud layers are then extracted as with METAR reports. The input data (predictors) are relative humidity, the lapse rate between the surface and 925hPa, and the precipitation amount.

Guidance on minimum ceiling probability is based on statistical interpretation of NWP output, predicting the probability of the minimum ceiling during a three-hour period being below 1000ft and below 600ft. The predictors are precipitation amount over three hours (total precipitation and snow precipitation), the lapse rate between the surface and 925hPa, relative humidity, the E-W component of wind speed, the S-N component of wind speed, cloud amount, cloud ice content and cloud water content. Logistic regression (Agresti 2002) was introduced to predict minimum ceiling probability in December 2010.

4.7.1.3 Weather

Weather guidance predicts weather conditions based on a diagnostic method for the interpretation of MSM output into categories (fine, cloudy, rainy, snowy and precipitation intensity) (JMA 1997). To determine the precipitation type (rain or snow), hourly temperature guidance is used instead of MSM temperature. To determine precipitation type (rain or snow), the hourly temperature guidance is used instead of MSM temperature. This improves the accuracy of precipitation type prediction.

4.7.1.4 Wind and Temperature

Wind and temperature guidance are calculated using the same methods as guidance for short-range forecasting (see Section 4.6).

4.7.1.5 Gust Winds

Gusting wind guidance predicts the probability of gusting during three-hour periods as well as the speed and direction of the hourly maximum peak gust. Its production involves Kalman filtering, frequency bias correction and logistic regression. Predictors are gust speed as per the MSM, surface wind speed, maximum wind speed in the boundary layer, vertical wind shear between the surface and the boundary layer, SSI, and vertical p-velocity at 925hPa. TAF gust guidance was introduced in December 2012.

4.7.1.6 Thunderstorms

Probability of thunderstorm (PoT) guidance predicts the probability of thunder during three-hour periods around airport. In its production, which involves logistic regression. 6 of 12 potential predictors are selected with focus on SSI, CAPE and precipitation amounts over the three-hour period. PoT guidance was introduced in May 2007

4.7.1.7 Snow

Snowfall amounts at airports are calculated from the grid-type snowfall amount guidance shown in Table 4.6.1. Snowfall amount guidance for TAF was introduced in October 2015.

4.7.2 Products for Domestic Area Forecast

4.7.2.1 Gridded Values of Significant Weather

Guidance variables for domestic area forecasts for flights, including turbulence, icing, CB clouds, tropopause height and vertical wind shear, are derived from MSM output. These values, together with common meteorological variables such as temperature, wind and humidity are used to produce domestic area forecasts and SIGMET information at JMA. This purely aviation-oriented dataset is called SIGGV (Gridded Values of Significant weather), with the specifications listed in Table 4.7.2. The parameter VWS, which is an indicator of clear air turbulence (CAT), is calculated as vertical wind shear between the model levels in kt/1000ft. The parameter TIndex (Kudo 2011) is a combined index of multiple turbulence indices, and is an indicator of various kinds of turbulence including CAT, mountain waves and cloud-related turbulence. CB cloud amount and top height calculation are based on the Percel method. The icing parameter is an indicator of aircraft icing derived from an empirical equation incorporating temperature and dew-point temperature. As illustrated in Figure 4.7.1, SIGGV, which is distributed as binary data, can be visualized on terminals at aviation forecast offices. it is also used for the production of the fax charts detailed below.

Table 4.7.2: SIGGV specifications

Base model	MSM
Forecast time	T=0-39, 1 hourly
Grid coordinate	Polar Stereographic, 111 × 93
Parameters	U, V, T, RH, Psea, Rain, CB cloud amount, CB top height, Tropopause height, Icing, VWS, TIndex

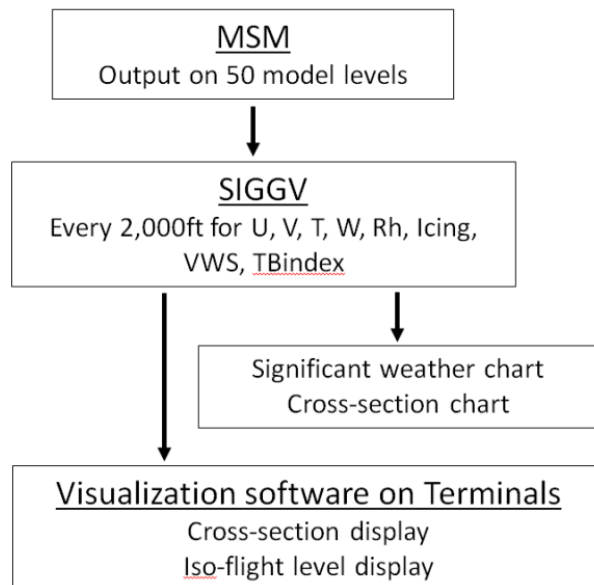


Figure 4.7.1: Data flow of products for domestic area forecast

4.7.2.2 Domestic Significant Weather Chart

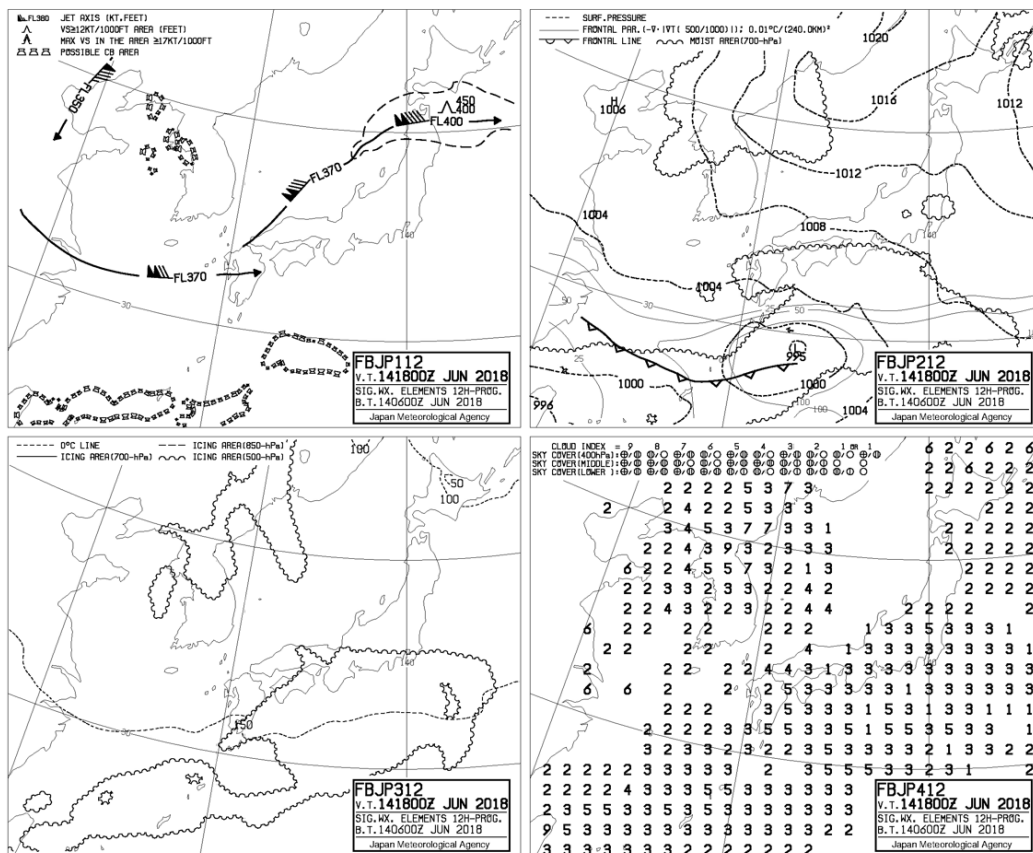


Figure 4.7.2: Sample domestic significant weather chart

This chart shows 12-hour forecast fields of the parameters listed below in four panels: (Figure 4.7.2)

- Upper-left:
 - Jet stream axes.
 - Possible CAT areas.
 - Possible CB areas.
- Lower-left:
 - Contours of 0°C height.
 - Possible icing areas at 500, 700 and 850hPa based on the -8 D method (Godske 1957)
- Upper-right:
 - Contours of sea level pressure.
 - Moist areas at 700 hPa.
 - Front parameters $DDT = -\nabla_n |\nabla_n T|$, where T is mean temperature below 500hPa and ∇_n denotes the horizontal gradient perpendicular to the isotherms.

- “NP fronts” drawn along the maxima of *DDT*.

- Lower-right:

- Cloud indices indicating low, middle and upper cloud amounts.

4.7.2.3 Domestic Cross-section Chart

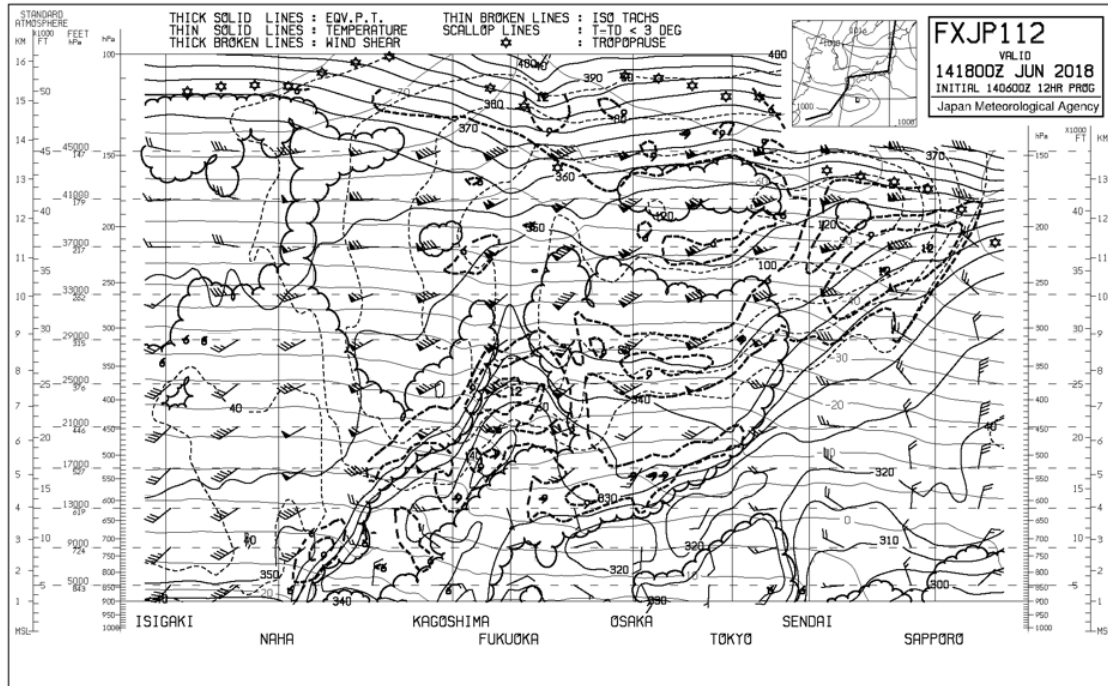


Figure 4.7.3: Sample domestic cross-section chart. Only the lower part of the fax, corresponding to the 12-hour forecast, is shown.

This chart shows 6- and 12-hour forecast fields along the major domestic route, and illustrates temperature, equivalent potential temperature, wind barbs and isotachs, moist areas, vertical wind shear and tropopause height (Figure 4.7.3).

4.7.3 Products for International Area Forecast

Global Grid Point Values are derived from the GSM four times a day and distributed in thinned GRIB code, a format compatible with products from the World Area Forecast Centers (WAFC). In addition to the parameters included in WAFC products, TBindex, Icing, VWS and Cb cloud top height are derived using the method used for domestic SIGGV (see Subsection 4.7.2).

JMA produces 13 significant weather (SIGWX) charts and 18 wind and temperature (WITEM) charts based on WAFS significant weather data provided by World Area Forecast Centers (WAFCs).

4.8 Ensemble Prediction System Products

4.8.1 EPS Products for One-week Forecasting

To assist forecasters in issuing one-week weather forecasts, ensemble mean products are made from EPS output.

An example of an ensemble chart showing average mean sea level pressure and precipitation is shown in Figure 4.8.1.

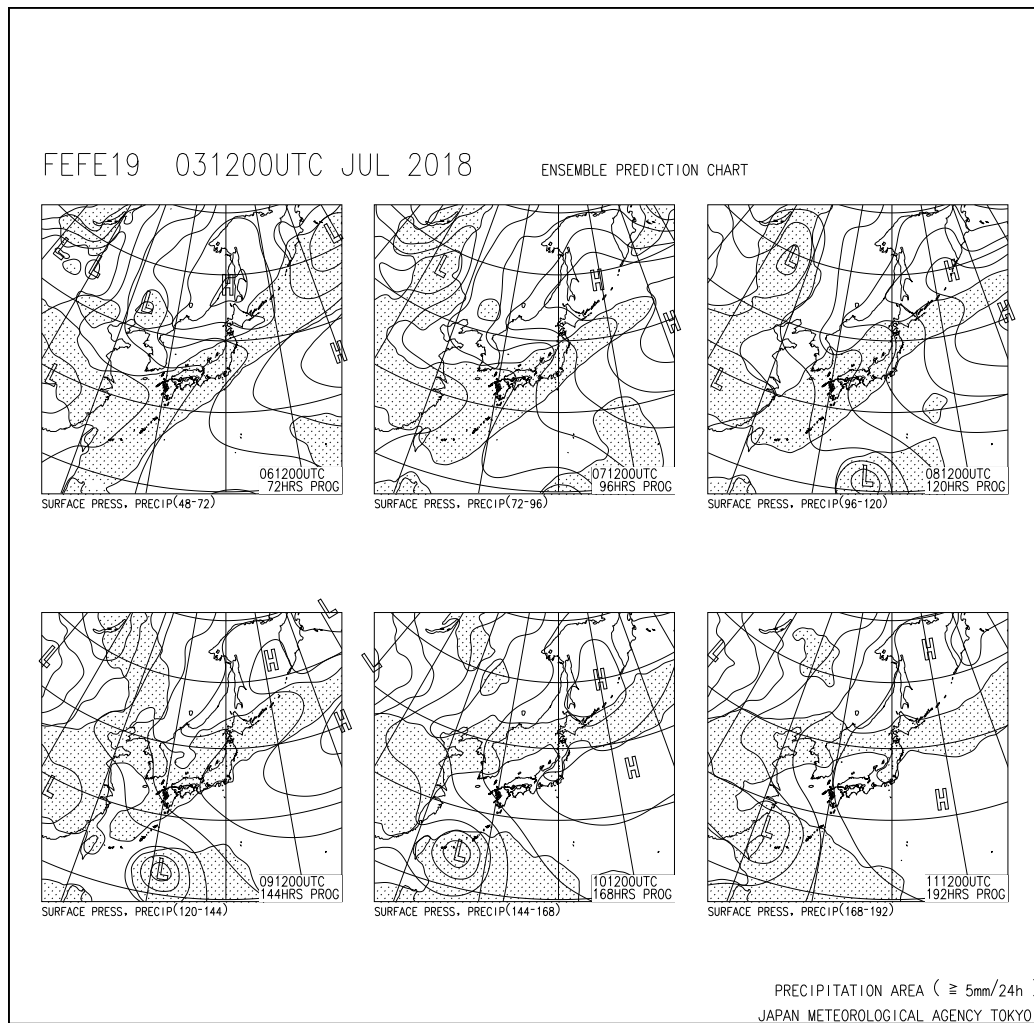


Figure 4.8.1: Ensemble prediction chart showing average mean sea level pressure and precipitation from day 2 to day 7. This schematic representation is produced by averaging with all members.

4.8.2 EPSs Products for One-month and Seasonal Forecasting

4.8.2.1 Standard Products

The following operational forecast and hindcast products from the Global and Seasonal EPSs are provided via the WMC (World Meteorological Center) Tokyo web-page on the TCC (Tokyo Climate Center) website.

- Ensemble mean maps
- Ensemble spread maps (for operational forecasts only)
- Verification maps and scores

4.8.2.2 Gridded Datasets

The following operational forecast and hindcast gridded datasets of the Global and Seasonal EPSs are provided to registered users via the WMC Tokyo web-page.

- Global EPS
 - Daily mean ensemble statistics (for operational forecasts only)
 - Daily mean forecast of individual ensemble member
- Seasonal EPS
 - Monthly mean ensemble statistics (for operational forecasts only)
 - Monthly mean forecast of individual ensemble member

4.8.2.3 El Niño Outlook

Outlooks of sea surface temperature deviations in the Pacific and the Indian Ocean produced from the Seasonal EPS are provided via the El Niño Monitoring and Outlook web-page on the TCC website (Figure 4.8.2).

4.8.2.4 Probabilistic Forecast Products for Seasonal Forecasts

To support NMHSs (National Meteorological and Hydrological Services) in their production of seasonal forecasts, probabilistic forecast products with three-categories (e.g., above-, near-, and below-normal) are produced from the Seasonal EPS and provided via the WMC Tokyo web-page (Figure 4.8.3).

4.8.2.5 Forecast Products in Support of Early Warnings for Extreme Weather Events

To support NMHSs in their production of early warnings for extreme weather events, Extreme Forecast Index (EFI) warning maps and meteograms for major cities are produced from the Global EPS and provided to registered users via the WMC Tokyo web-page (Figure 4.8.4 and Figure 4.8.5). The EFI is a measure of EPS forecast deviation from climatological probability distribution (Lalurette 2003). JMA uses a revised version of the EFI to add weight for the tails of probability distribution (Zsótér 2006).

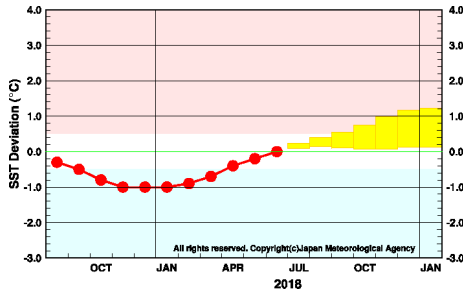


Figure 4.8.2: Five-month running mean of SST deviation for NINO.3. Red dots indicate analysis values, and yellow boxes indicate values predicted by the Seasonal EPS. Each box denotes the range in which the value is expected to be included with a probability of 70% or more.

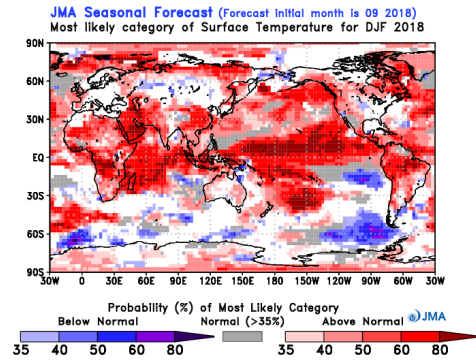


Figure 4.8.3: Probabilistic forecast map of surface air temperature for seasonal forecasting. Probabilities are estimated using numerical guidance with application of the Model Output Statistics (MOS) technique based on hindcast results.

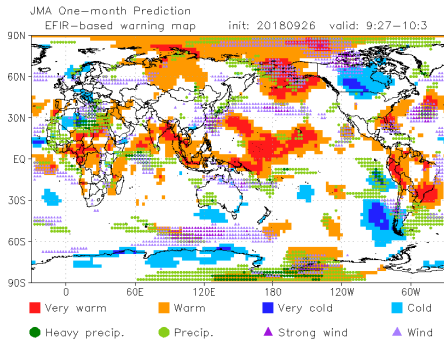


Figure 4.8.4: EFI warning map for temperature, precipitation and wind. Pale symbols indicate grids where the EFI is above 0.5, and dark symbols indicate grids where the EFI is above 0.8.

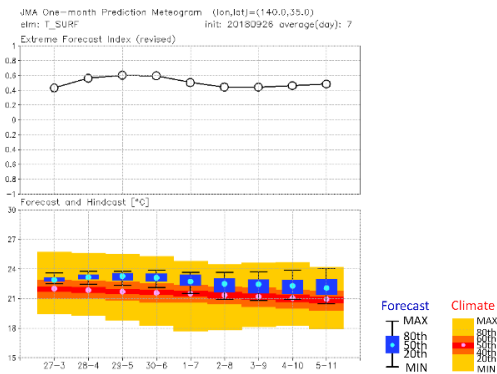


Figure 4.8.5: Meteogram of the Global EPS for temperature. The upper graph is an EFI timeseries representation, and the lower one is a timeseries representation of forecast ranges with cold color box-and-whisker plots and climatological values with warm shading.

4.9 Atmospheric Angular Momentum Functions

Atmospheric Angular Momentum (AAM) functions were proposed to support evaluation of the earth's rotational variations based on precise estimation of variations in atmospheric angular momentum. For monitoring of atmospheric effects associated with the earth's rotation, JMA sends AAM products to NCEP (a sub-bureau of the International Earth Rotation Service (IERS)) via GTS. The AAM functions are expressed as follows (Barnes *et al.* 1983):

$$\begin{aligned} \chi_1 = & -1.00 \left[\frac{r^2}{(C-A)g} \right] \int P_S \sin \phi \cos \phi \cos \lambda dS \\ & - 1.43 \left[\frac{r}{\Omega(C-A)g} \right] \iint (u \sin \phi \cos \lambda - v \sin \lambda) dPdS, \end{aligned} \quad (4.9.1)$$

$$\begin{aligned} \chi_2 = & -1.00 \left[\frac{r^2}{(C-A)g} \right] \int P_S \sin \phi \cos \phi \sin \lambda dS \\ & - 1.43 \left[\frac{r}{\Omega(C-A)g} \right] \iint (u \sin \phi \sin \lambda + v \cos \lambda) dPdS, \end{aligned} \quad (4.9.2)$$

$$\chi_3 = -0.70 \left[\frac{r^2}{Cg} \right] \int P_S \cos^2 \phi dS - 1.00 \left[\frac{r}{\Omega Cg} \right] \iint u \cos \phi dPdS. \quad (4.9.3)$$

In Eq. (4.9.1) to Eq. (4.9.3), P is pressure, $\int dS$ is the surface integral over the globe, (ϕ, λ) are latitude and longitude, u and v are the eastward and northward components of wind velocity, P_S is surface pressure, g is the mean acceleration of gravity, r is the mean radius of the earth, C is the polar moment of inertia of the solid earth, A is the equatorial moment of inertia, and Ω is the mean angular velocity of the earth.

The functions χ_1 and χ_2 represent equatorial components, and the function χ_3 is the axial component. All components are non-dimensional. The first term of each one is a pressure term related to the redistribution of air masses. The second is a wind term related to the relative angular momentum of the atmosphere.

Variations in AAM functions calculated from JMA global analysis data have been reported to correspond closely to variations in the earth's rotation. Figure 4.9.1 shows seasonal variations in observed earth rotation and atmospheric relative angular momentum (the wind term of χ_3) calculated by the National Astronomical Observatory of Japan (Naito and Kikuchi 1992).

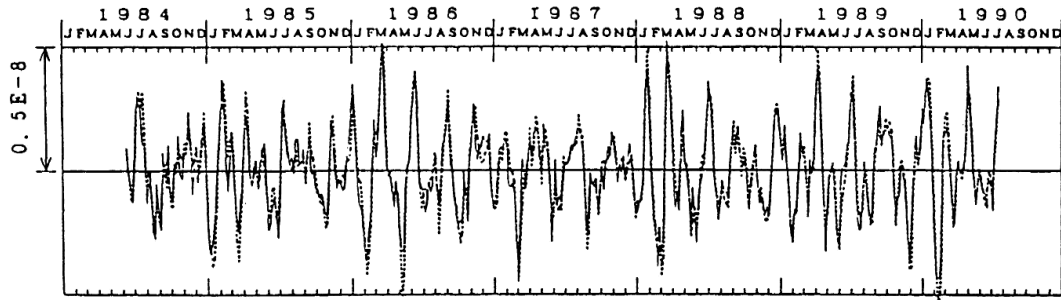


Figure 4.9.1: Seasonal variations in observed earth rotation (solid line) and calculated atmospheric angular momentum (broken line). Both sets of data are 150 days' high-pass filtered.

AAM functions calculated from JMA global analysis data at 00, 06, 12 and 18UTC have been provided operationally since early 1993. AAM functions calculated from JMA global 8-day forecast data at 12UTC are now also provided.

AAM functions calculated for a test period between 21 June and 30 September 1992 are shown in Figure 4.9.2, where days 1 - 102 correspond to this period. Each term of the AAM functions is multiplied by 10^7 .

The broken lines show 6-hourly values of the functions (i.e., the difference from the period mean values), and the solid lines show band-pass filtered values for periods of 5-10 days. Oscillation with a 5-10 day period is notable in each term of each component, implying a corresponding oscillation with a similar period in the global-scale atmosphere.

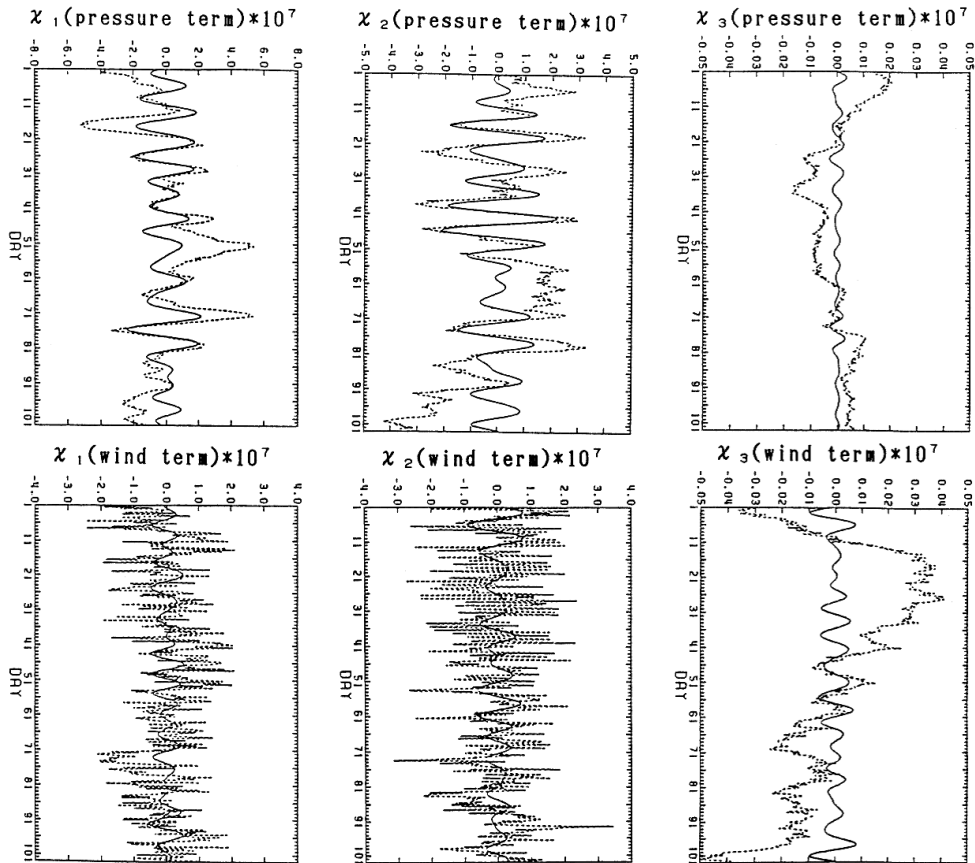


Figure 4.9.2: Pressure terms (top) and wind terms (bottom) of AAM functions. The left panels are the χ_1 component, the center ones are χ_2 and the right ones are χ_3 . Days 1 - 102 correspond to 21 June - 30 September 1992. The broken lines show 6-hourly values of the functions, and the solid lines show band-pass filtered values for periods of 5-10 days. Each value is multiplied by 10^7 .

Chapter 5

Ocean Models

5.1 Summary

- (i) In regard to sea surface temperatures, JMA engages in high-resolution analysis based on satellite remote sensing data and *in-situ* observation data to provide real-time ocean information, and also conducts analysis based on *in-situ* observation data to monitor long-term ocean variations such as El Niño events and global warming.
- (ii) JMA launched a global ocean data assimilation system for the monitoring of El Niño and the Southern Oscillation (ENSO) in 1995. The MOVE/MRI.COM-G system introduced in 2008 was updated in 2015 to MOVE/MRI.COM-G2, which consists of an ocean general circulation model (MRI.COM) and an objective analysis scheme. Its output, along with atmospheric analysis, is fed to a coupled ocean-atmosphere model for ENSO prediction and seasonal forecasting of climate in Japan.
- (iii) Another ocean data assimilation system for the Western North Pacific (MOVE/MRI.COM-WNP) has been operated since 2008 to analyze and predict variations in sea water temperature, salinity and currents associated with eddy-scale oceanic phenomena such as the Kuroshio, the Oyashio and mid-scale eddies in seas adjacent to Japan.
- (iv) JMA operates the Global Wave Model (GWM), the Coastal Wave Model (CWM) and the Wave Ensemble System (WENS) as ocean wave models. All three are based on the MRI-III developed by the Agency's Meteorological Research Institute (MRI). The WENS has been developed to provide probabilistic information on ocean waves in medium-range forecasts. The Shallow-water Wave Model (SWM) is also operated in trial mode to predict ocean waves in bays and near-shore areas around Japan with a higher horizontal resolution of one arc minute.
- (v) JMA operates two storm surge models. One covers the Japan region and predicts storm surges generated by tropical and extra-tropical cyclones. The model runs eight times a day and provides 39-hour forecasts for more than 400 points along the Japanese coast. The other is for the Asian region, and is developed within the framework of the WMO Storm Surge Watch Scheme. The model runs four times a day and predicts storm surges up to 72 hours ahead. Horizontal storm surge maps and time-series charts for selected points are issued to Typhoon Committee Members.
- (vi) JMA operates a numerical sea ice model to support sea ice forecasting for the southern part of the Sea of Okhotsk. The model forecasts distributions and concentrations of sea ice with a one-week lead time based on dynamic and thermodynamic equations during the winter season.
- (vii) JMA's oil spill prediction model was introduced in 1999. Operation is triggered when a large-scale oil spill occurs offshore. The effects of transport by sea surface winds, ocean waves and sea surface currents, turbulent diffusion, evaporation and emulsification are considered.

5.2 Sea Surface Temperature Analysis

5.2.1 Merged Satellite and *In-situ* Data Global Daily Sea Surface Temperature (MGDSST)

Daily sea surface temperatures (SSTs) in the global ocean on a grid of $0.25 \times 0.25^\circ$ are objectively analyzed to support ocean information services (Kurihara *et al.* (2006)). The data are also used as boundary conditions for atmospheric short- and medium-range prediction models (see Section 3.2, Section 3.3 and Section 3.5) and as observational data in the ocean data assimilation system for the Western North Pacific (see Section 5.3). SST data obtained from satellite infrared sensors (NOAA/AVHRR, MetOp/AVHRR) and microwave sensors (Coriolis/WINDSAT, GCOM-W1/AMSR2) are used together with *in-situ* SST observations. Many *in-situ* data are obtained through the Global Telecommunication System, but domestic organizations also provide large amounts of data by e-mail.

Satellite-derived SST anomalies (SSTA) from daily SST climatologies are decomposed into long- and short-time scales with a cutoff period of 53 days and large/middle/small scales with cutoff wavelengths of 580 and 143km. The middle scale is intended to represent SST signals caused by mesoscale (eddy-scale) phenomena, and the small scale is intended to represent sub-mesoscale signals. Signals varying with a period shorter than 27 days are cut off due to the significant data noise they contain. Long-time scale signals represent intra-seasonal variations, and short-time scale signals represent variations influenced by atmospheric conditions such as tropical cyclones.

The large scale and long-time scale components of SSTA from satellites are calibrated with *in-situ* SSTA using Poisson's equation (Reynolds (1987)). Space-time optimum interpolation (OI) is applied to each component, and a zero value is adopted as the first guess. Space-time correlation coefficients and RMS values of the first guess error and satellite observation errors are statistically estimated a priori from satellite data using the method of Kuragano and Kamachi (2000). The daily SST is the sum of components of interpolated SSTA and daily climatologies (Figure 5.2.1).

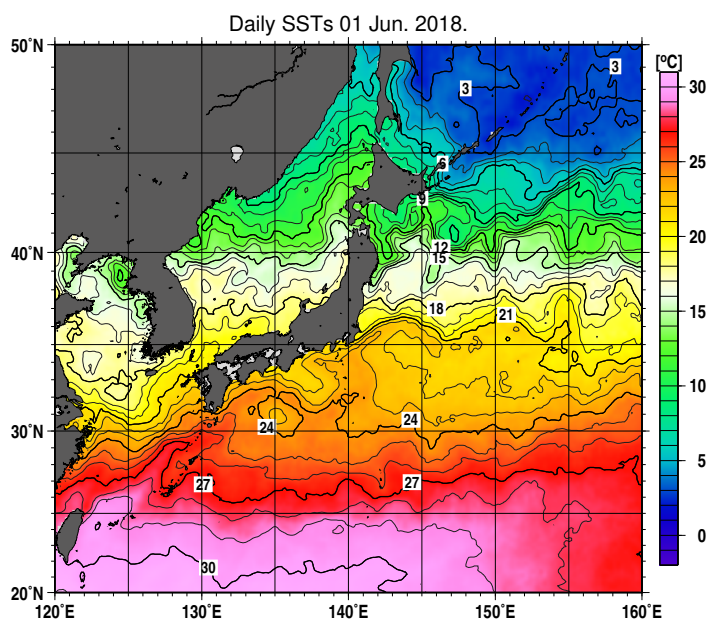


Figure 5.2.1: Sample SST analysis

5.2.2 Daily Sea Surface Analysis for Climate Monitoring

Sea Surface Temperature (SST) analysis for climate monitoring at JMA (COBE-SST) is based on the method described in Ishii *et al.* (2005), which is summarized below.

SST analysis has a resolution of 1° latitude and 1° longitude. The east-west grid points start at 0.5°E and end at 0.5°W. The north-south grid points start at 89.5°S and end at 89.5°N. Analysis is based on the optimum interpolation method, and the deviation of the previous day's analysis from the 1950–2000 normal is multiplied by 0.95 and used as a first guess. Analysis is performed on a daily basis with marine meteorological data for seven days centered on the day of interest. Daily observation data are averaged in 1.5° × 1.5° boxes for analysis using the optimum interpolation method to minimize processing time.

Bias correction for past SST observation reports is based on the method of Folland and Parker (1995). Observation data quality control is performed by checking ship tracks, dates and positions of reports, and erroneous data are automatically corrected in the compilation of marine meteorological data at JMA. Based on observation data deviations from 1950–2000 normals for three months including the day of interest, the biases of data with identical ship call signs are estimated, and call signs associated with large data biases are automatically blacklisted through daily analysis. Daily (final) analysis is performed with a delay of 31 days from real time to allow appropriate use of delayed observations. Daily analysis for the 30-day period following the final analysis is also performed for real-time utilization.

Information on sea ice concentration is used in estimation of SSTs for polar oceans.

The daily updated operational SST data are utilized as described below with historical values.

1. Monitoring of equatorial Pacific SSTs, El Niño/ La Niña evolutions and global warming over 100 years.
2. Input of the operational Ocean Data Assimilation System (MOVE/MRI.COM-G2) and historical oceanic analysis (see Section 5.3).
3. Input of the Japanese 55-year Reanalysis (JRA-55) (see Section 2.10).

Monthly averaged SST data are provided on the Tokyo Climate Center website¹. The characteristics of the data are described in Japan Meteorological Agency (2006), which is available on the Tokyo Climate Center website².

5.3 Ocean Data Assimilation System

Following the successful development of an ocean data assimilation system (MOVE/MRI.COM) by JMA's Meteorological Research Institute (MRI), JMA has operated two instances of the system since 2008. One is the global system (MOVE-G) for the monitoring of El Niño and the Southern Oscillation (ENSO), which was updated to MOVE-G2 in 2015. The other is the Western North Pacific system (MOVE-WNP) for analysis of eddy-scale oceanic phenomena in seas adjacent to Japan. This is the successor to the COMPASS-K resource described in the previous report. The following subsections give a brief outline of MOVE/MRI.COM and examples of MOVE-G2 and MOVE-WNP.

5.3.1 Ocean General Circulation Model and Objective Analysis Scheme

MOVE/MRI.COM consists of an ocean general circulation model (MRI.COM) and an objective analysis scheme (MOVE). The details of MOVE/MRI.COM are described in Usui *et al.* (2006).

MRI.COM (Tsujino *et al.* 2010) is a multilevel model that solves primitive equations under hydrostatic and the Boussinesq approximation. The σ - z vertical coordinate with which layer thickness near the surface follows surface topography is adopted to allow free surface elevation (Hasumi 2006). For nonlinear momentum advection, a generalized enstrophy-preserving scheme (Arakawa 1972) and a scheme involving the concept of diagonally upward/downward mass momentum fluxes along a sloping bottom are applied. Vertical viscosity

¹http://ds.data.jma.go.jp/tcc/tcc/products/el_nino/cobesst/cobe-sst.html

²http://ds.data.jma.go.jp/tcc/tcc/library/MRCS_SV12/index_e.htm

and diffusivity are determined using the turbulent closure scheme of Noh and Kim (1999). In MOVE-G2, isopycnal mixing (Gent and McWilliams 1990) is used for horizontal turbulent mixing, and harmonic viscosity with the parameterization of Smagorinsky (1963) is used for momentum. A tracer advection scheme based on conservation of second-order moments (Prather 1986) is newly adopted in MOVE-G2. In MOVE-WNP, a biharmonic operator is used for horizontal turbulent mixing, and a biharmonic friction with Smagorinsky-like viscosity (Griffies and Hallberg 2000) is used for momentum. In MOVE-G2 and MOVE-WNP, a sea ice model with the thermodynamics of Mellor and Kantha (1989) and the elastic-viscous-plastic rheology of Hunke and Dukowicz (1997) is also applied, with enhancements in areas such as thickness categories, ridging and rheology also implemented following the Los Alamos sea ice model (CICE; Hunke and Lipscomb 2006). Surface forcing to drive the model is based on the Japanese 55-year ReAnalysis (JRA-55, Kobayashi *et al.* (2015); see Section 2.10).

The analysis scheme adopted in MOVE is a multivariate three-dimensional variational (3DVAR) type with vertical coupled Temperature-Salinity (T-S) Empirical Orthogonal Function (EOF) modal decomposition (Fujii and Kamachi 2003). Amplitudes of T-S EOF modes above 1750 m for MOVE-G2 and 1500 m for MOVE-WNP are employed as control variables and the optimal temperature and salinity fields are represented via linear combination of the EOF modes. In this system, the model domain is divided into several subregions and vertical T-S EOF modes are calculated from the observed T-S profiles for each subregion. The 3DVAR results are inserted into the model temperature and salinity fields via incremental analysis updates (Bloom *et al.* 1996) with assimilation windows of 10 days for MOVE-G2 and 5 days for MOVE-WNP.

These *in-situ* observations are reported from ships, profiling floats and moored or drifting buoys via the GTS and other communication systems. In MOVE/MRI.COM, *in-situ* observations of subsurface temperature and salinity, as well as satellite altimeter data are assimilated into the model. The results of sea surface temperature analysis, performed independently from MOVE/MRI.COM, are also assimilated as observational data. COBE-SST (Ishii *et al.* (2005); see Subsection 5.2.2) grid-point values are used for MOVE-G2, and MGDSST (Kurihara *et al.* (2006); see Subsection 5.2.1) grid-point values are used for MOVE-WNP.

5.3.2 Ocean Data Assimilation System for Global oceans (MOVE-G2)

The horizontal resolution is 0.5° latitude and 1.0° longitude except for the 15°S-15°N band, where the latitudinal grid spacing decreases to the minimum of 0.3° between 6°S and 6°N (see Figure 5.3.1). The model has 52 vertical levels and bottom boundary layer (Nakano and Suginozawa 2002), 23 of which are placed above 200 m (see Figure 5.3.2). The model has realistic bottom topography, and the maximum depth of the bottom is set to 6300 m. The computational domain is the global ocean, including the Arctic Ocean with use of tri-polar horizontal coordinates (see Figure 5.3.1). New schemes are introduced in MOVE-G2, such as a bias correction scheme (Fujii *et al.* 2012), a first guess at appropriate time (FGAT) scheme (Lorenz and Rawlins 2005), and a global water mass correction scheme (Kuragano *et al.* 2014).

The latest assimilation results are obtained once every five days, and the targeted term is 3-7 days before assimilation. Assimilation data for the same term are updated every five days using additional delayed-mode observation data until the term reaches 39-43 days before the latest assimilation.

MOVE-G2 output is used in various forms for the monitoring of ENSO at JMA, and products for the equatorial Pacific region are distributed in publications titled Monthly Highlights on the Climate System and El Niño Outlook. Figure 5.3.3 shows one such chart from MOVE-G2 indicating depth-longitude sections of temperature and related anomalies. Report charts are also provided on the Tokyo Climate Center Web page³.

5.3.3 Ocean Data Assimilation System for the Western North Pacific (MOVE-WNP)

The model domain spans from 117°E to 160°W zonally and from 15 to 65°N meridionally. Horizontal resolution is variable, with values of 1/10° from 117 to 160°E and 1/6° from 160°E to 160°W, and 1/10° from 15 to 50°N and 1/6° from 50 to 65°N. There are 54 levels in the vertical direction, with thickness increasing from 1 m at the surface to 600 m near the bottom (see Figure 5.3.4). Oceanic states at the side boundaries are replaced by those from a North Pacific model with a horizontal resolution of 1/2° (one-way nesting).

³<http://ds.data.jma.go.jp/tcc/tcc/index.html>

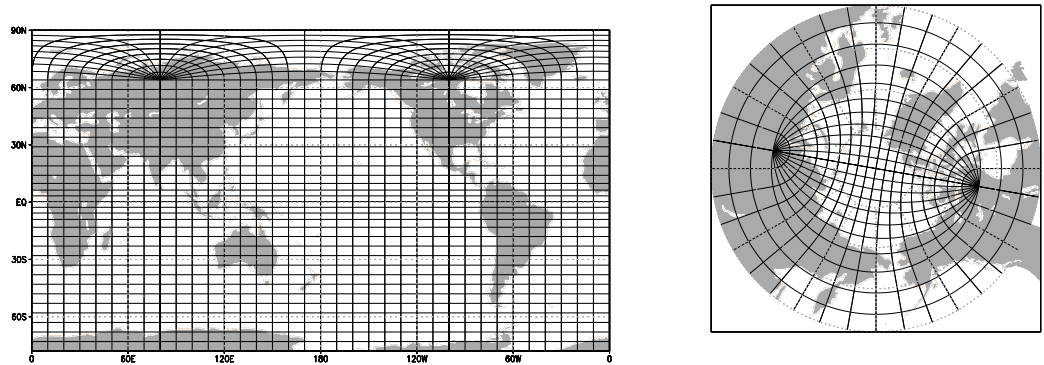


Figure 5.3.1: MOVE-G2 horizontal grids

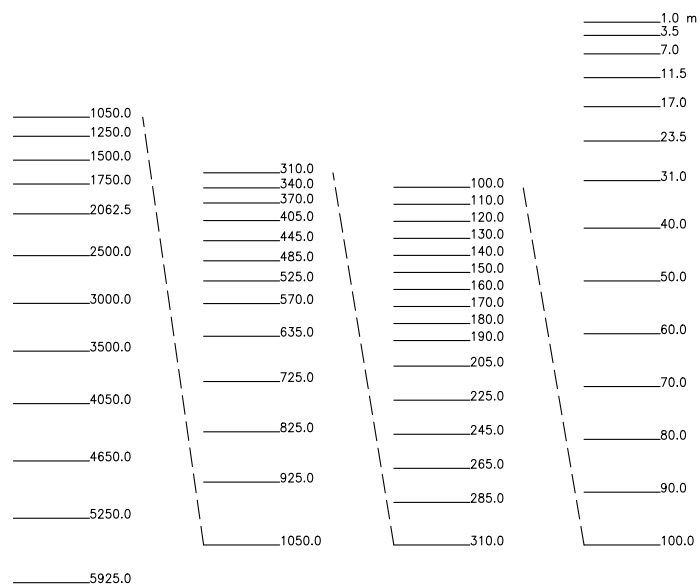


Figure 5.3.2: MOVE-G2 vertical levels with depths in meters

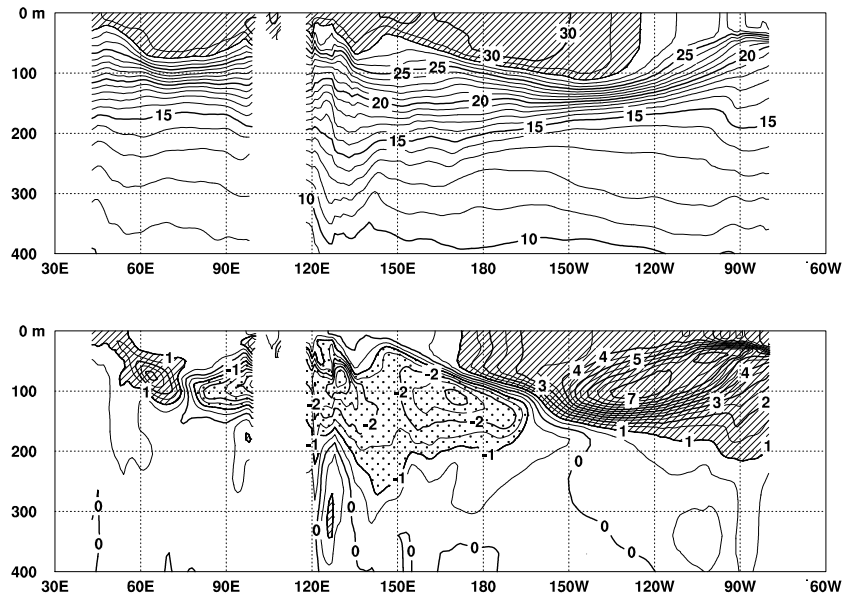


Figure 5.3.3: Depth-longitude cross sections of monthly mean temperature and temperature anomalies along the equator in the Indian and Pacific Ocean area for November 2015 based on MOVE-G2. The base period for the normal is 1981–2010.

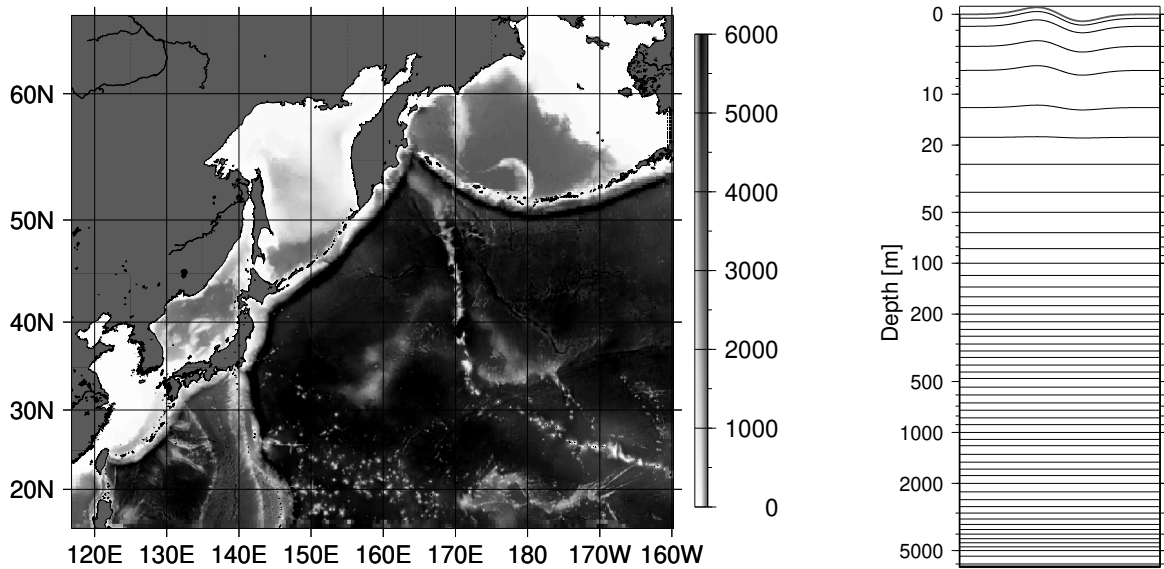


Figure 5.3.4: Bottom topography (left) and vertical levels (right) of the OGCM for the Western North Pacific.

Figure 5.3.5 shows predicted ocean current fields at a depth of 50m, which was calculated by the forecast run from the initial condition of 9 August 2017. The assimilated fields are also shown in the figure. It can be seen that the large meander of the Kuroshio path was successfully predicted in the forecast run. The delayed assimilation run is implemented every five days and the prompt assimilation run is implemented every day. The output from MOVE-WNP is used as the initial condition of the ocean forecasting model with a one month prediction period. The results of the operational assimilation, nowcast and forecast runs are provided on JMA's web site and those of assimilation and nowcast runs are available on the NEAR-GOOS RRTDB⁴.

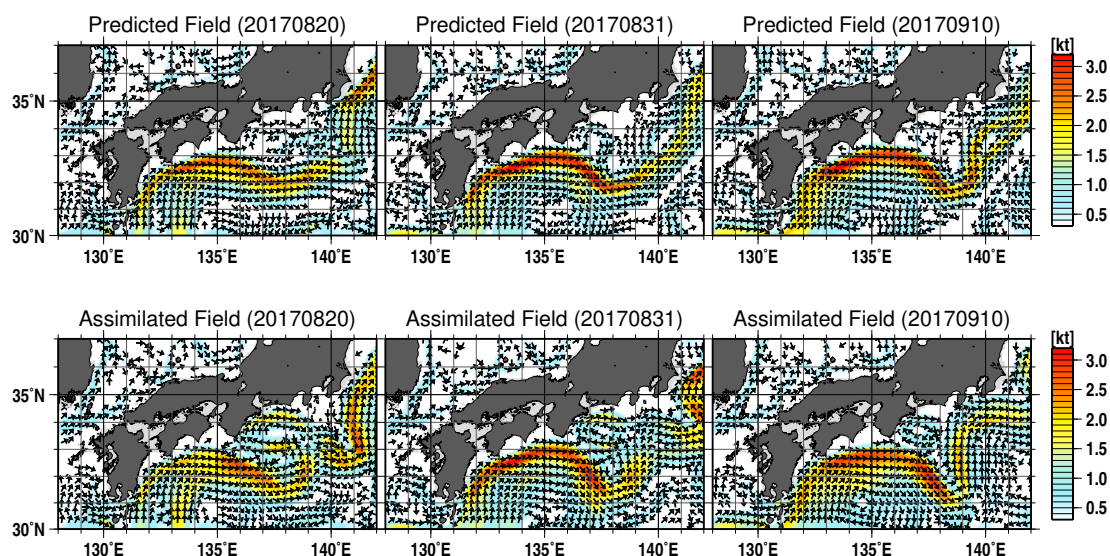


Figure 5.3.5: Horizontal current fields at a depth of around 50m. The top panels show the results of prediction using the initial condition of 9 August 2017, and the bottom ones show assimilation results.

5.3.4 Future plan

JMA's Meteorological Research Institute (MRI) is currently developing a new ocean data assimilation system involving a new version of MRI.COM (Tsuji *et al.* 2017) and an improved MOVE system. The new system will be based on a four-dimensional variational (4DVAR) analysis method and assimilate sea ice concentration with the 3DVAR technique.

MOVE-WNP will also be adapted with a 4DVAR analysis method (Usui *et al.* 2015) and give initial conditions for a new ocean model with higher horizontal resolution (≈ 2 km) for the area around Japan. The new model (MRI.COM-JPN) explicitly represents tidal motion and is expected to provide information on sub-meso scale phenomena (e.g., Kuroshio frontal waves and coastal events).

5.4 Ocean Wave Models

5.4.1 Introduction

The Japan Meteorological Agency (JMA) operates the Global Wave Model (GWM), the Coastal Wave Model (CWM), and the Wave Ensemble System (WENS). The Shallow-water Wave Model (SWM) is also operated in trial mode. All the models are classified as third-generation wave models.

⁴<https://www.data.jma.go.jp/gmd/goos/data/database.html>

Table 5.4.1: Specifications of the Global Wave Model, the Coastal Wave Model and the Wave Ensemble System.

Model name	Global Wave Model	Coastal Wave Model	Wave Ensemble System
Type of wave model	spectral model (third-generation wave model, MRI-III)		
Area	global 75°N – 75°S 180°W – 0° – 180°E	coastal sea of Japan 50°N – 20°N 120°E – 150°E	global 75°N – 75°S 180°W – 0° – 180°E
Grid size	0.5° × 0.5° (720 × 301)	0.05° × 0.05° (601 × 601)	1.25° × 1.25° (288 × 121)
Time step (advection term) (source term)	10 minutes 30 minutes	1 minutes 3 minutes	30 minutes 60 minutes
Forecast range (from 06UTC, 18UTC) (from 00UTC) (from 12UTC)	132 hours 132 hours 264 hours	132 hours 132 hours 132 hours	264 hours 264 hours
Spectral component	900 components 25 frequencies from 0.0375 to 0.3 Hz (logarithmically partitioned) 36 directions		
Initial condition	Analysis by Optimal Interpolation		
Boundary condition	Global Wave Model		
Wind field	Global Spectral Model (GSM)		Global Ensemble Prediction System (GEPS)
	Fujita's empirical formula and a corresponding gradient wind for a typhoon		
Shallow-water effects	Refraction and bottom friction		

The GWM, the CWM and the WENS are based on the MRI-III, which was originally developed by the Meteorological Research Institute of JMA (Ueno 2004). The current versions of the GWM and the CWM, which include shallow water effects, have been used for short-range forecasts since May 2017. The WENS, which is a prediction system with probability information and is used for middle-range forecasts, has been in operation since June 2016. The specifications of the three models are given in Table 5.4.1, and their domains are shown in Figure 5.4.1.

The SWM is based on the WAM (The WAMDI Group 1988), but has been modified by the National Institute for Land and Infrastructure Management of the Ministry of Land, Infrastructure, Transport, and Tourism (MLIT) and was put into quasi-operation under a cooperative framework with Water and Disaster Management Bureau of MLIT. It has high resolution of 1 minute (see Table 5.4.2 for specifications), and is applied only to limited areas. The addition of a final 11 areas since 2013, has resulted in today's operation over 22 areas. SWM products are used exclusively within JMA and Regional Development Bureaus of MLIT.

5.4.2 Ocean Wave Model Structure

The ocean wave models forecast the wave energy density (spectrum) of each frequency and direction (i.e., the two-dimensional (directional) wave spectrum). The basic equation is the energy balance expression:

$$\frac{\partial F}{\partial t} + \nabla \cdot (C_g F) + \frac{\partial}{\partial \theta}(\Omega F) = S_{net} = S_{in} + S_{nl} + S_{ds} + S_{btm} \quad (5.4.1)$$

where

$$\Omega = \frac{C_g}{C_p} \left(-\frac{\partial C_p}{\partial x} \cos \theta + \frac{\partial C_p}{\partial y} \sin \theta \right),$$

which represents refraction in shallow water. $F(f, \theta, \mathbf{x}, t)$ is a two-dimensional spectrum dependent on the frequency f and the wave direction θ , $C_g(f, \theta, \mathbf{x})$ is the group velocity, and $C_p(f, \mathbf{x})$ is the phase speed. The group velocity is simply $C_g(f)$ for deep-water waves but depends on water depth for shallow-water waves. S_{net} is a net source function consisting of S_{in} , S_{nl} , S_{ds} , and S_{btm} , which are briefly outlined below. Only the model

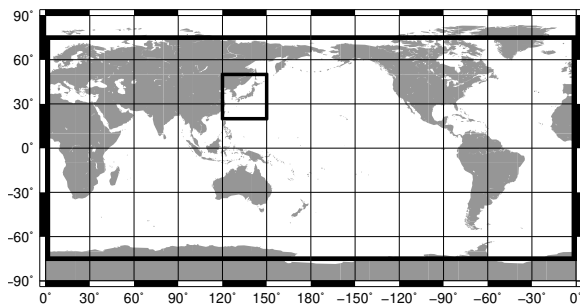


Figure 5.4.1: Calculation areas for the Global Wave Model (outer thick lines) and the Coastal Wave Model (inner thick lines).

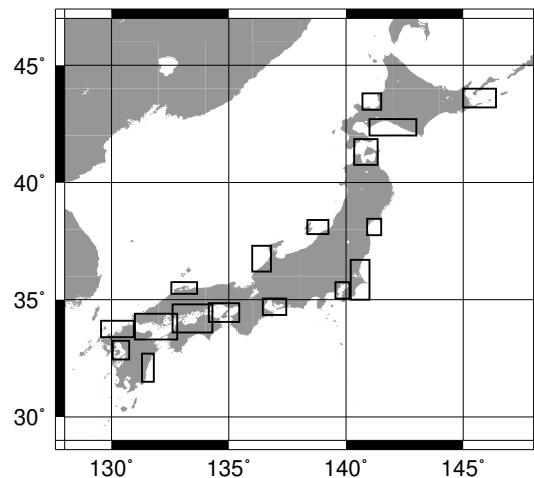


Figure 5.4.2: Calculation areas for the Shallow-water Wave Model.

Table 5.4.2: Specifications of the Shallow-water Wave Model.

Type of wave model	spectral model (third-generation wave model, WAM)					
Areas	Domain name	Grid size	Integration domain	Domain name	Grid size	Integration domain
	Tokyo Bay	37 × 43	35.75°N – 35.05°N 139.55°E – 140.15°E	Off Shimane	67 × 31	35.75°N – 35.25°N 132.55°E – 133.65°E
	Ise Bay	61 × 43	35.05°N – 34.35°N 136.45°E – 137.45°E	Ishikari Bay	49 × 43	43.80°N – 43.10°N 140.70°E – 141.50°E
	Harima-Nada	79 × 49	34.85°N – 34.05°N 134.15°E – 135.45°E	Off Ishikawa	49 × 67	37.30°N – 36.20°N 136.00°E – 136.80°E
	Osaka Bay			Off Nemuro	85 × 49	44.00°N – 43.20°N 145.00°E – 146.40°E
	Ariake Sea	43 × 49	33.25°N – 32.45°N 130.05°E – 130.75°E	Off Miyazaki	31 × 73	32.70°N – 31.50°N 131.30°E – 131.80°E
	Shiranui Sea			Tsugaru Strait	61 × 67	41.85°N – 40.75°N 140.35°E – 141.35°E
	Off Niigata	55 × 37	38.40°N – 37.80°N 138.35°E – 139.25°E	Off Ibaraki	49 × 103	36.70°N – 35.00°N 140.20°E – 141.00°E
	Sendai Bay	37 × 43	38.45°N – 37.75°N 140.90°E – 141.50°E	Off Boso		
	Off Tomakomai	121 × 43	42.70°N – 42.00°N 141.00°E – 143.00°E	Genkai-Nada	83 × 43	34.10°N – 33.40°N 129.55°E – 130.95°E
	Suo-Nada	109 × 67	34.40°N – 33.30°N 131.00°E – 132.80°E			
	Iyo-Nada					
	Aki-Nada					
	Hiuchi-Nada	103 × 73	34.80°N – 33.60°N 132.60°E – 134.30°E			
Grid resolution	1' × 1'					
Time step (advection term) (source term)	1 minutes 1 minutes					
Forecast range	39 hours					
Spectral component	1260 components 35 frequencies from 0.0418 to 1.1 Hz (logarithmically partitioned) 36 directions					
Initial condition	Coastal Wave Model					
Boundary condition	Coastal Wave Model					
Wind field	Meso-Scale Model (MSM) Fujita's empirical formula and a corresponding gradient wind for a typhoon					

numerics of the MRI-III are described here, as those of the WAM are already extensively referenced elsewhere (e.g. Janssen 2004).

1. S_{in} : energy input from wind. This value generally takes the form $S_{in} = A + BF$, where A is linear wave growth and BF is exponential growth. In the MRI-III, the formula of Cavaleri and Rizzoli (1981) is used for linear growth

$$A = 1.5 \times 10^{-3} \left(u_*^4 / 2\pi g^2 \right) \exp[-(f_{PM}/f)^4] (\max(0, \cos(\theta - \theta_W)))^4 \quad (5.4.2)$$

where u_* is the friction velocity of wind, θ_W is the wind direction, and g is the gravitational acceleration. In general, the linear term has little influence on wave growth except in the very early stages. Meanwhile, the exponential term BF has a key role in wave growth. In the MRI-III, The B is expressed as

$$B(f, u_*, \theta_W - \theta) = c_{in} \left(\frac{u_*}{C_{pd}} \right)^2 \cos^3(\theta_W - \theta) / |\cos(\theta_W - \theta)|. \quad (5.4.3)$$

where C_{pd} is the phase speed of deep water waves, i.e., $C_{pd} = \frac{g}{\omega} = \frac{g}{2\pi f}$.

This expression is based on Mitsuyasu and Honda (1982) and Plant (1982).

2. S_{nl} : nonlinear energy transfer associated with resonant interaction. Since rigorous calculation is highly time-consuming, a practical scheme known as discrete interaction approximation (DIA) (Hasselmann *et al.* 1985) is commonly used in operational wave models. This approach involves the use of only one parameter for the set of four resonant waves:

$$\left. \begin{aligned} f_1 &= f_2 = f, \\ f_3 &= f(1 + \lambda) = f_+, \\ f_4 &= f(1 - \lambda) = f_-, \\ \theta_1 &= \theta_2 = \theta, \\ \theta_3 - \theta &= \pm \cos^{-1} \left\{ \frac{1 + 2\lambda + 2\lambda^3}{(1 + \lambda)^2} \right\}, \\ \theta_4 - \theta &= \mp \cos^{-1} \left\{ \frac{1 - 2\lambda - 2\lambda^3}{(1 - \lambda)^2} \right\}. \end{aligned} \right\} \quad (5.4.4)$$

$$\begin{aligned} \left(\begin{array}{c} \delta S_{nl} \\ \delta S_{nl}^+ \\ \delta S_{nl}^- \end{array} \right) &= \left(\begin{array}{c} -2(\Delta f \Delta \theta) / (\Delta f \Delta \theta) \\ (1 + \lambda)(\Delta f \Delta \theta) / (\Delta f_+ \Delta \theta) \\ (1 - \lambda)(\Delta f \Delta \theta) / (\Delta f_- \Delta \theta) \end{array} \right) \\ &\times C f^{11} g^{-4} \left[F^2 \left\{ \frac{F_+}{(1 + \lambda)^4} + \frac{F_-}{(1 - \lambda)^4} \right\} - 2 \frac{F F_+ F_-}{(1 - \lambda^2)^4} \right] \end{aligned} \quad (5.4.5)$$

where $F \equiv F(f, \theta)$, $F_+ \equiv F(f_+, \theta_3)$, $F_- \equiv F(f_-, \theta_4)$. The coefficient C is determined to fit exact calculation for the JONSWAP spectrum. Hasselmann *et al.* (1985) defined the related parameters as $\lambda = 0.25$, corresponding to $\theta_3 - \theta = \pm 11.5^\circ$, $\theta_4 - \theta = \mp 33.6^\circ$ and $C = 3 \times 10^7$. DIA calculation is found to support highly accurate estimation based on parameter multiplication. In the MRI-III, S_{nl} is calculated using the DIA scheme with three configurations. The parameters used are $\lambda_1 = 0.19$ ($C_1 = 1.191 \times 10^7$), $\lambda_2 = 0.23$ ($C_2 = 6.835 \times 10^6$), and $\lambda_3 = 0.33$ ($C_3 = 1.632 \times 10^6$).

3. S_{ds} : energy dissipation associated with wave breaking and other influences. In the MRI-III, dissipation terms are expressed as local energy dissipation as proposed by Ueno (1998).

$$S_{ds} = -c_b \frac{u_*}{g^3} f^7 (F(f, \theta))^2 \quad (5.4.6)$$

where c_b is a coefficient determined to fit wave generation. In the MRI-III, a slightly artificial swell decay process is included.

$$S_{sds} = -2.96 \times 10^{-6} \tanh \left[4(f_s - f) / f_p \right] F(f, \theta), \quad (5.4.7)$$

where $f_p = 0.156g/U_{10N}$ represents the peak frequency of the Pierson-Moskowitz (PM) spectrum from the 10m height wind speed U_{10N} . This decay function is applied to the spectrum of frequencies lower than $f_s = 1.8f_p$ when the significant wave height exceeds 1.5m.

4. S_{btm} : the energy loss associated with bottom friction. This effect needs to be considered for shallow water. In the MRI-III, this term is taken from Hasselmann *et al.* (1973).

$$S_{btm} = -\frac{0.038}{g^2} \cdot \frac{(2\pi f)^2}{\sinh^2(kd)} \cdot F(f, \theta) \quad (5.4.8)$$

where k is the wave number and d is depth. Energy loss increases for shallow water and long-period waves.

5.4.3 Wind Field

Wind fields for the GWM and the CWM are given by the Global Spectral Model (GSM), while the SWM uses Meso-Scale Model (MSM) winds. For the WENS, 27-member wind fields of the Global Ensemble Prediction System (GEPS) are employed.

In the GWM, the CWM and the SWM, wind fields around typhoons are modified using an empirical method. As typhoons contribute significantly to extremely high waves in the western North Pacific, accurate wave forecasts are crucial to the prevention of shipwrecks and coastal disasters. Since NWP models occasionally fail to predict typhoon conditions such as intensity and location accurately, wind fields based on operational typhoon analysis and forecasting are imposed onto NWP winds (known as bogus wind) when a typhoon is present over the western North Pacific. Changes in the course of a typhoon may also result in drastic wave field alterations, especially in the small region covered by the SWM. Accordingly, wave fields are also predicted with the typhoon assumed to move along a five-point path (center, faster, slower, right end and left end) in the typhoon forecast error circle.

To create bogus wind data, sea level pressure distribution near a typhoon is assumed to have a profile expressed by Fujita's empirical formula (Fujita 1952)

$$P(r) = P_\infty - \frac{P_\infty - P_0}{\sqrt{1 + (r/r_0)^2}} \quad (5.4.9)$$

where P_∞ , P_0 and r_0 denote the ambient pressure, the central pressure of the typhoon, and the scaling factor of the radial distribution of the pressure, respectively. Surface winds near the typhoon are estimated from the pressure field by assuming the gradient wind balance with modifications based on the typhoon movement and surface friction effects.

5.4.4 Wave Analysis

An assimilation scheme (Kohno *et al.* 2012) for the GWM and the CWM was introduced in October 2012. In this system, initial conditions (wave spectra) are modified based on significant wave heights under the Objective Wave Analysis System (Kohno *et al.* 2009), which objectively analyzes wave heights using optimal interpolation (OI) with observations from radar altimeters of satellites, buoys, coastal wave recorders and ships. The specifications are shown in Table 5.4.3. Introduction of data assimilation improves the prediction of ocean wave fields, especially in terms of shorter forecast times and swell propagation.

5.4.5 Products

JMA issues many ocean wave products based on GWM, CWM, and WENS data.

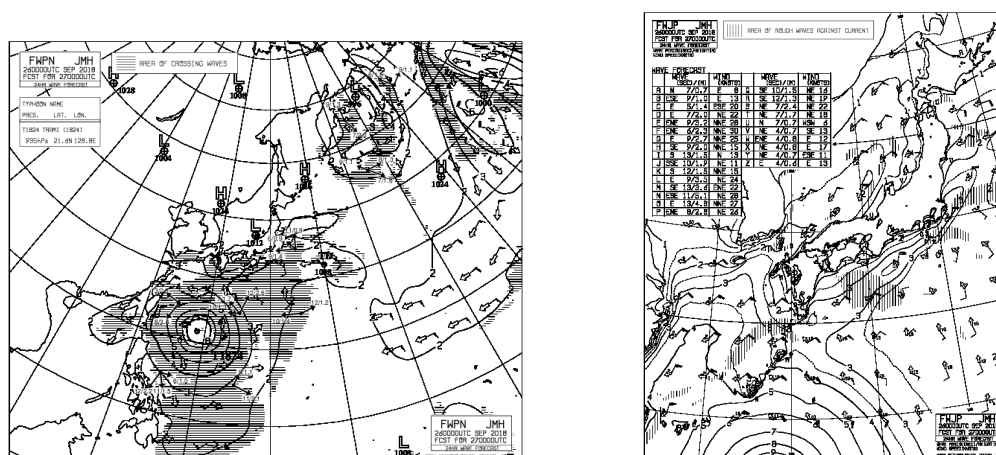
Charts of analysis and 24-hour ocean wave forecasts are served twice a day via the JMH radio facsimile broadcast service and the JMA website for the Western North Pacific and seas around Japan (Figure 5.4.3). The

Table 5.4.3: JMA Objective Wave Analyses System specifications.

Analysis scheme	Optimal interpolation
Data cut-off time	6 hours and 25 minutes for early run analysis 12hours for delay analysis
First guess	6-hour forecast by the GWM
Analysis variables	Significant wave height
Grid size	0.5°×0.5° grid
Integration domain	Global oceans
Observational data	BUOY, SHIP, Nowphas, GPS wave meter, JASON3, SARAL
Assimilation window	6 hours

charts indicate significant wave heights, peak wave directions, and peak wave periods. In addition, information on rough sea areas, that may hinder maritime navigation was incorporated in 2017. The areas of horizontal hatching in Figure 5.4.3(a) indicate areas of crossing waves that may give rise to unexpectedly high-sea conditions. The areas of vertical hatching in Figure 5.4.3(b) show areas in which wave heights and steepness increase due to the effects of opposing ocean currents.

Statistical products detailing significant wave heights and peak wave periods with probability for medium-range forecasts are produced from WENS output and provided on the JMA website for the WMO Severe Weather Forecasting Demonstration Project (SWFDP; <https://www.wis-jma.go.jp/swfdp/>). Figure 5.4.4-Figure 5.4.7 show examples for TY Trami (1824).



(a) Wave forecast chart for the Western North Pacific (FWPN) created from the GWM.

(b) Wave forecast chart for coastal region of the Japan (FWJP) created from the CWM.

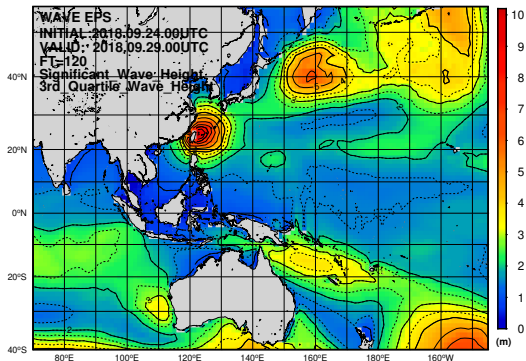
Figure 5.4.3: Wave forecast charts based on 24-hour model predictions at the initial time of 00UTC 26 September 2018.

5.4.6 Improvement and Development

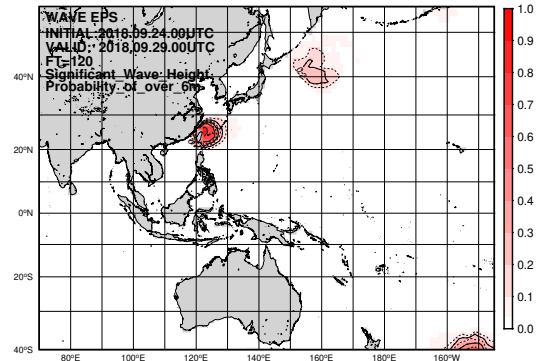
The Main improvements made to JMA ocean wave models since 2013 are as follows:

1. The shallow water effect was introduced into the GWM and the CWM in 2017.
2. The WENS was put into operation in 2016.
3. SWM target regions were added by 2016.

Figure 5.4.8 shows root mean square errors (RMSEs) for the significant wave height of the GWM against monthly average observations from 2013 to 2017. Although not apparent score changes, the error amplitude has been smaller since 2013.

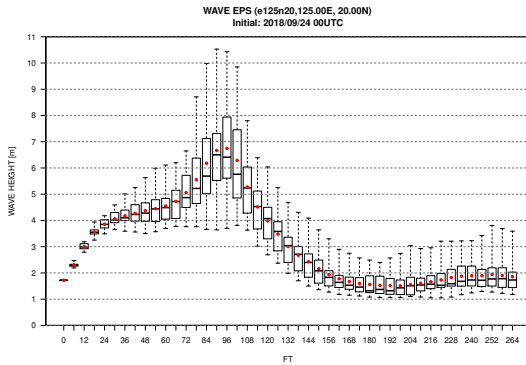


(a) 3rd quartile of significant wave height.

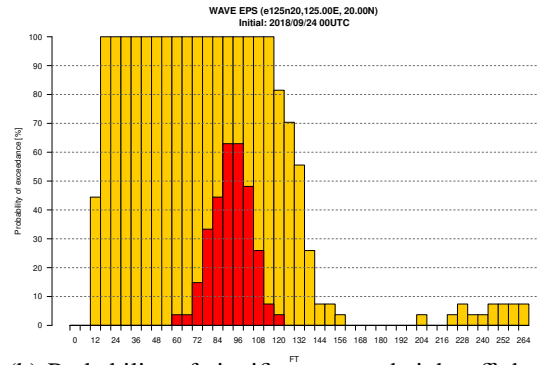


(b) Probability of significant wave height over 6m.

Figure 5.4.4: Statistics on significant wave height from 120-hour WENS predictions at the initial time of 00UTC 24 on September 2018.

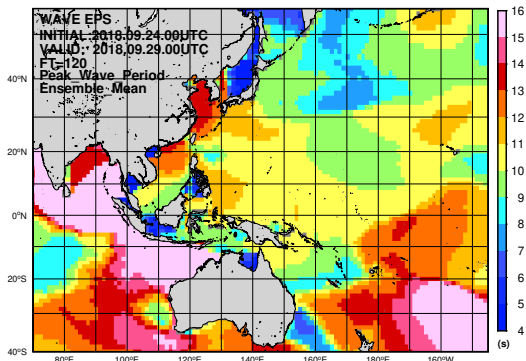


(a) Boxplots of significant wave height off the northeast coast of the Philippines.

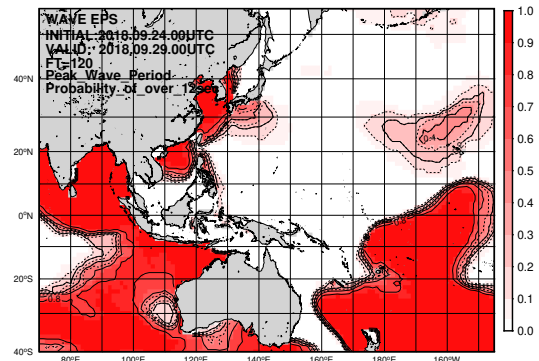


(b) Probability of significant wave height off the northeast coast of the Philippines. Yellow > 3m, Red > 6m.

Figure 5.4.5: Time-series representation of significant wave height from WENS prediction at the initial time of 00UTC 24 on September 2018.



(a) Ensemble mean of wave period.



(b) Probability of wave period over 12sec.

Figure 5.4.6: Statistics on peak wave period from 120-hour WENS predictions at the initial time of 00UTC 24 on September 2018.

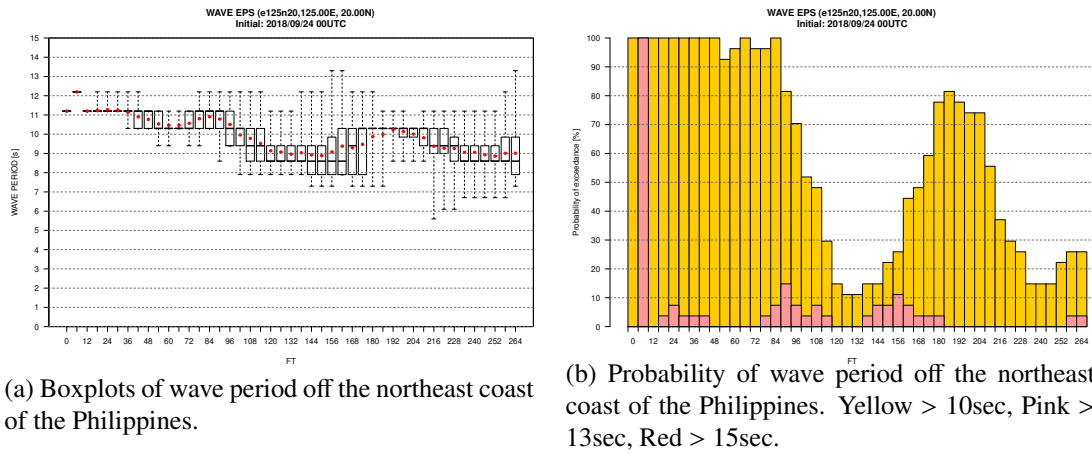


Figure 5.4.7: Time-series representation of peak wave period from WENS predictions at the initial time of 00UTC 24 on September 2018.

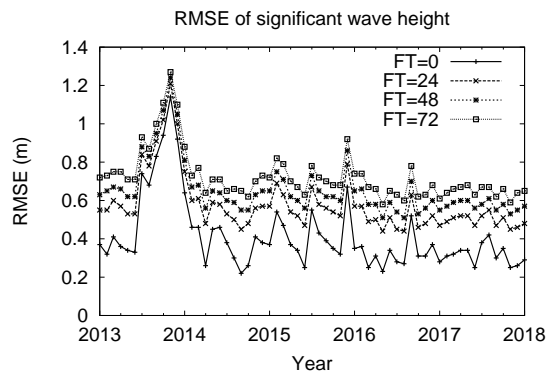


Figure 5.4.8: Time-series representation of analysis and forecast scores for the GWM.

Wave forecast errors are occasionally observed in relation to typhoons. For example:

1. Wave distribution resembling ball pairs sometimes appears due to the bogus wind embedding method.
2. As the model cannot predict asymmetric patterns for typhoons during extratropical transition and similar, initial winds with "bogus wind" need to be improved.

The grid resolution of the GWM and the WENS will be enhanced. In future work, a multi-grid model incorporating the GWM, the CWM and a higher resolution wave model for coastal regions of Japan will be operated.

JMA also plans to introduce a shallow water effect into the WENS based on the module currently used in the GWM and the CWM.

5.5 Storm Surge Model

5.5.1 Japan Area Storm Surge Model

5.5.1.1 Introduction

The Japanese Archipelago is vulnerable to storm surges because of its topography (characterized by a gulf open to the south and a shallow coast), making accurate and timely forecasts/warnings crucial in mitigating related threats to life and property.

The Japan Meteorological Agency (JMA), which is responsible for issuing storm surge warnings, has operated a numerical storm surge model since July 1998 to provide basic related information. At first, the model was run four times a day when a typhoon was present in the vicinity of Japan. It is continuously improved in areas such as enlarging the model domain, predicting extratropical cyclone conditions, extending forecast times, adding advection terms, and etc. Since May 2010, a new storm surge model with higher resolution (an approx. 1-km mesh) and a gridded astronomical tide analysis method have been operated in storm tide calculation for more detailed information and warnings.

5.5.1.2 Dynamics

Storm surges are mainly caused by the effects of wind setup due to strong onshore winds on the sea surface and inverse barometer effects associated with pressure drops in low-pressure systems. The effects of wind setup are proportional to the square of wind speed and inversely proportional to water depth, and are related to coastal topography, meaning that they are amplified in open bays against the wind.

The JMA storm surge model is similar to that described by Higaki *et al.* (2009), and is based on two-dimensional shallow water equations driven by meteorological fields. These equations incorporate vertically integrated momentum equations in two horizontal directions:

$$\frac{\partial U}{\partial t} + u \frac{\partial U}{\partial x} + v \frac{\partial U}{\partial y} - fV = -g(D + \eta) \frac{\partial(\eta - \eta_0)}{\partial x} + \frac{\tau_{sx}}{\rho_w} - \frac{\tau_{bx}}{\rho_w} \quad (5.5.1a)$$

$$\frac{\partial V}{\partial t} + u \frac{\partial V}{\partial x} + v \frac{\partial V}{\partial y} + fU = -g(D + \eta) \frac{\partial(\eta - \eta_0)}{\partial y} + \frac{\tau_{sy}}{\rho_w} - \frac{\tau_{by}}{\rho_w} \quad (5.5.1b)$$

and the continuity equation:

$$\frac{\partial \eta}{\partial t} + \frac{\partial U}{\partial x} + \frac{\partial V}{\partial y} = 0 \quad (5.5.2)$$

where U and V are volume fluxes in the x - and y -directions, defined as:

$$U \equiv \int_{-D}^{\eta} u dz \quad (5.5.3a)$$

$$V \equiv \int_{-D}^{\eta} v dz \quad (5.5.3b)$$

f is the Coriolis parameter; g is gravity acceleration; D is the water depth below mean sea level; η is surface elevation; η_0 is the inverse barometer effect converted into the equivalent water column height; ρ_w is the density of water; τ_{sx} and τ_{sy} are the x - and y -components of wind stress on the sea surface, respectively; and τ_{bx} and τ_{by} are the x - and y -components of the stresses of bottom friction, respectively. The drag coefficient is based on adjustment of the model in reference to the results of Smith and Banke (1975) and Frank (1984):

$$c_d = \begin{cases} (0.63 + 0.1185W) \times 10^{-3} & (W < 20m/s) \\ \{3.00 + 0.0120(W - 25)\} \times 10^{-3} & (W \geq 20m/s) \end{cases} \quad (5.5.4)$$

The equations are solved via numerical integration using the explicit finite difference method. A staggered (or Arakawa-C) grid (Arakawa and Lamb 1977) is adopted for the grid system.

5.5.1.3 Meteorological Forcing

The fields of surface wind and atmospheric pressure dynamically predicted by the Meso-Scale Model (MSM) are used for external forcing with the storm surge model. When a tropical cyclone (TC) is present in the area around Japan, a simple parametric TC model is also used for forcing.

The simple parametric TC model (referred to as bogus) is introduced to account for TC track forecast error and its influence on storm surge forecasting, but a single result is insufficient for risk management because storm surge behavior strongly depends on TC tracks. To consider the influence of TC track uncertainty on the occurrence of storm surge, five runs of the storm surge model are conducted with possible TC tracks prescribed at the center of and at four points on the forecast circle within which the TC is forecast to be with a probability of 70% (Figure 5.5.1): Center track (1), Fastest track (2), Rightward-biased track (3), Slowest track (4) and Leftward-biased track (5). The five tracks are used to create meteorological fields with the parametric TC model.

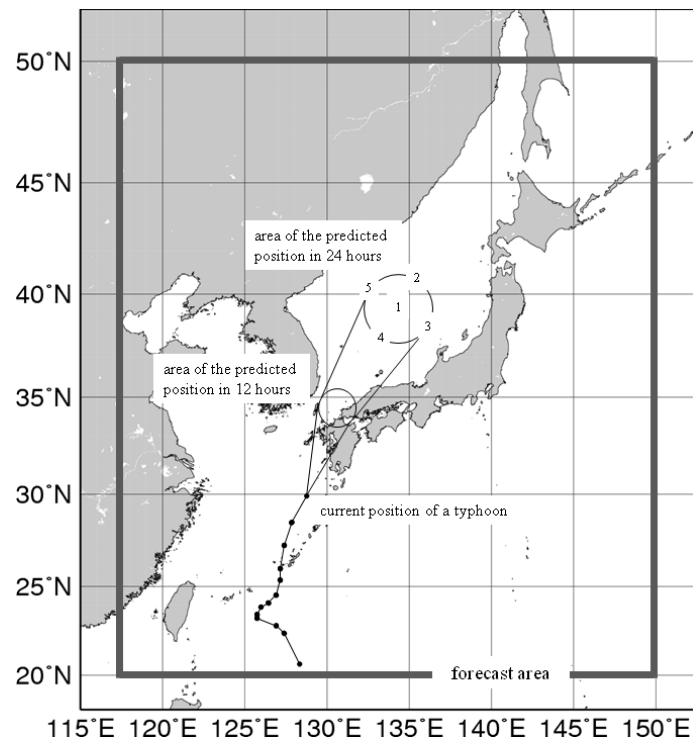


Figure 5.5.1: Bogus TC tracks and the domain of the Japan area storm surge model

The simple parametric TC model utilizes Fujita's formula (Fujita 1952), which represents radial pressure distribution in a TC:

$$P = P_{\infty} - \frac{P_{\infty} - P_c}{\sqrt{1 + (r/r_0)^2}} \quad (5.5.5)$$

and the gradient wind relation:

$$-\frac{v_g^2}{r} - f v_g = -\frac{1}{\rho} \frac{\partial P}{\partial r} \quad (5.5.6)$$

In Eq. (5.5.5), P is atmospheric pressure at distance r from the center of the TC, P_{∞} is environmental pressure, P_c is the central pressure of the TC and r_0 is a scaling factor for radial distribution of pressure. In Eq. (5.5.6), ρ is the density of air and v_g is the gradient wind.

To represent the asymmetry of the wind field \mathbf{w} in a TC, the moving velocity vector of the TC multiplied by a weight that decays exponentially with distance from the TC center is added to the gradient wind:

$$\mathbf{w} = C_1 \left\{ \mathbf{v}_g + \mathbf{C} \cdot \exp\left(-\pi \frac{r}{r_e}\right) \right\} \quad (5.5.7)$$

\mathbf{C} is the TC velocity vector, and r_e is the coefficient of decay.

TC analysis and forecast information, such as the center position, central pressure and maximum wind, is applied to these formulas to synthesize the wind and pressure fields (Konishi 1995).

5.5.1.4 Model Specifications

Table 5.5.1 gives the specifications of the storm surge model, whose domain covers the whole of Japan (Figure 5.5.1).

Table 5.5.1: Japan area storm surge model specification

Model	2-dimensional model
Grid	Lat-Lon Arakawa-C grid
Region	20°N - 50°N, 117.5°E - 150°E
Resolution	approximately 1, 2, 4, 8, 16 km (Adaptive mesh)
Time step	4 seconds
Initial time	00, 03, 06, 09, 12, 15, 18, 21 (UTC)
Forecast time	39 hours
Member	TC case: 6 members (MSM+5 bogus) no TC case: 1 member (MSM)

Since storm surge is essentially a long wave, its phase speed is proportional to the square root of water depth. It is inefficient to set the same resolution for all grids in consideration of computer resources. Accordingly, the Adaptive Mesh Refinement (Berger and Olinger 1984), in which the mesh is fine over shallow water and coarse over deep water, is adopted. The resolution is varied over five levels (1, 2, 4, 8 and 16 km) with water depth (Figure 5.5.2). This method makes storm surge calculation more efficient than with the normal lat-lon grid system.

The storm surge model runs eight times a day (every 3 hours) and calculates storm surge predictions up to 39 hours ahead. Initial values of surface elevation (η) and volume fluxes (U and V) are generated from previous calculation using the newest MSM prediction for forcing (hindcast). Since the initial values are not as important as those in atmospheric models, assimilation of observation data is not conducted.

The model computes only storm surges, i.e. anomalies from the level of astronomical tides. However, storm tides (storm surge plus astronomical tide) are required to issue a storm surge warnings. Astronomical tides are predicted using harmonic analysis of sea level observations based on JMA's gridded astronomical tide method, in which astronomical tide calculation is performed even for no-observation grid areas (Subsection 5.5.3). After storm surge model computation, the astronomical tide level for the coastal area is added to the predicted storm surge.

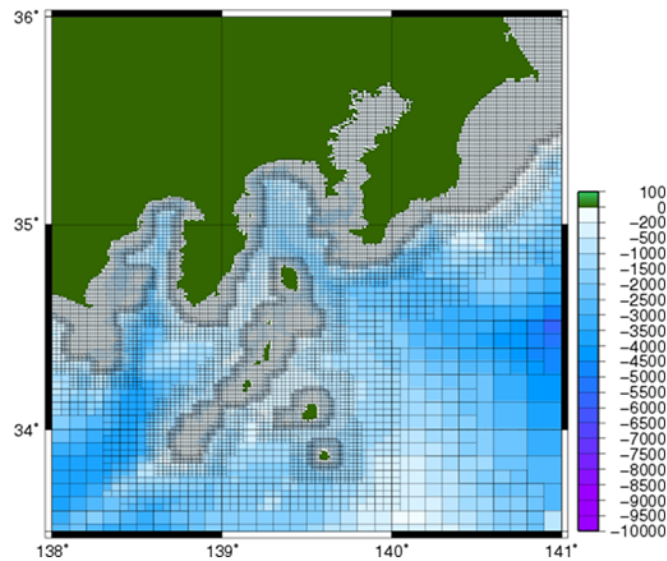


Figure 5.5.2: Storm surge model horizontal grid system and water depth (around the Kanto region)

5.5.1.5 Verification

The accuracy of storm surge prediction depends on the accuracy of the storm surge model itself and atmospheric model conditions. To help eliminate the influence of atmospheric model uncertainty, the accuracy of the model was evaluated using storm surge predictions driven by atmospheric analysis data.

Figure 5.5.3 shows a scatter diagram of storm surge hindcasts against observation values from 214 tide stations for the statistical period is from June 2015 to December 2017. The stations are managed by organizations including JMA, the Ports and Harbours Bureau, the Japan Coast Guard, the Geographical Survey Institute and etc. The figure shows that storm surge prediction errors lie in the range of ± 50 cm, although large errors (hindcast values exceeding observation by more than 50 cm) are also observed. These errors are assumed to generally result from factors excluded from the storm surge model, such as the effect of wave setup, ocean currents and sea water stratification.

Storm surge associated with Typhoon Jebi (T1821) is presented here as an example of related prediction. Figure 5.5.4 shows the track of the typhoon, which passed over central Japan and caused storm surge conditions in Osaka Bay and elsewhere on 4 September 2018. Figure 5.5.5 (a) shows storm surge distribution around the bay as of 06 UTC on 4 September as predicted by the storm surge model, which reasonably forecasted the extreme surge conditions associated with wind setup in the bay's inner part. Figure 5.5.5 (b) shows a time-series chart of storm surge at the port of Osaka. The peak was slightly underestimated and the forecast of its timing was delayed by about an hour, but in general the extreme storm surge conditions were fairly represented.

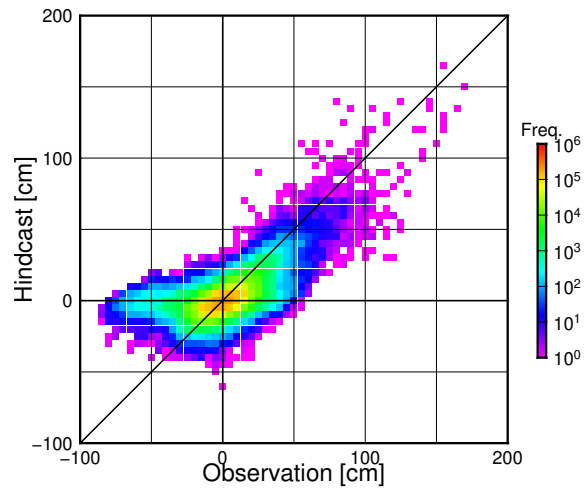


Figure 5.5.3: Scatter diagram of storm surge hindcast values against observed values

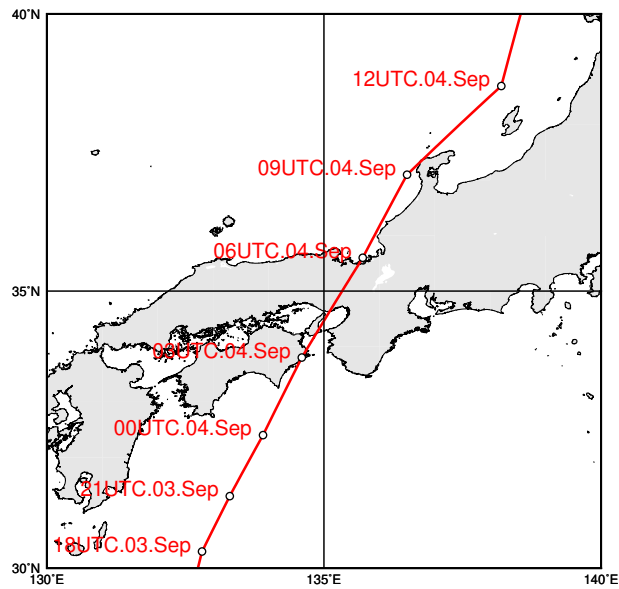
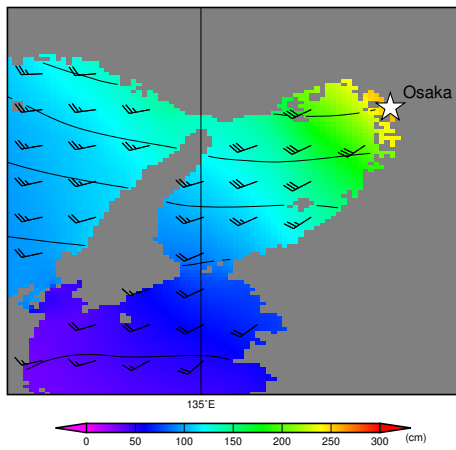
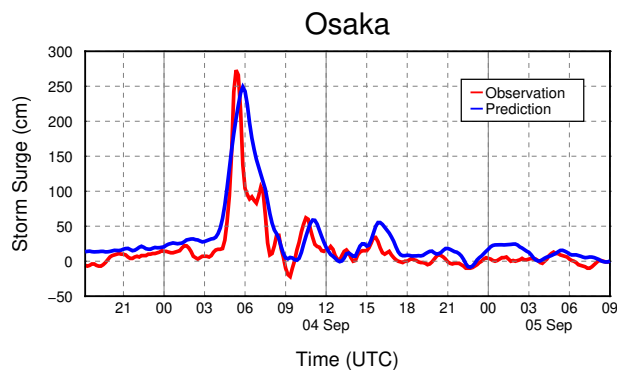


Figure 5.5.4: The track of Typhoon Jebi (T1821)



(a) Distribution of storm surge at 06 UTC on 4 September around Osaka Bay



(b) Time-series chart of storm surge at the port of Osaka

Figure 5.5.5: Storm surge distribution and time-series chart for the port of Osaka as predicted by the storm surge model with an initial time of 18 UTC on 3 September 2018

5.5.2 Asia Area Storm Surge Model

5.5.2.1 Introduction

The late 2000s saw severe storm surge disasters worldwide, including on the coast of the Gulf of Mexico (caused by Hurricane Katrina) in 2005, on the coast of Bangladesh (caused by Cyclone Sidr) in 2007, and on the coast of Myanmar (caused by Cyclone Nargis) in 2008.

In response to a request by the WMO Executive Council (60th session, June 2008), WMO initiated the development of the regional Storm Surge Watch Scheme (SSWS) for areas affected by tropical cyclones. In relation to the western North Pacific and the South China Sea, the ESCAP/WMO Typhoon Committee (41st session, January 2009) endorsed a commitment by the RSMC Tokyo - Typhoon Center to produce storm surge forecasts with the aim of strengthening the storm surge warning capabilities of National Meteorological and Hydrological Services (NMHSs) in the region. JMA began development of a storm surge model for the Asia region in 2010 in collaboration with Typhoon Committee Members providing sea level observation and sea bathymetry data. Horizontal distribution maps of predicted storm surges and time-series charts are published on JMA's Numerical Typhoon Prediction website (Hasegawa *et al.* 2017).

More recently, Hurricane Sandy made landfall on the eastern coast of USA in 2012, causing serious damage including the traffic paralysis, massive blackouts and cessation of economic activity in New York. Typhoon Haiyan also caused more than 6,000 fatalities in the Philippines in 2013. Against such a background, storm surge and inundation countermeasures play pivotal roles in efforts to prevent typhoon-related disaster condition.

5.5.2.2 Dynamics

The basic equations of the Asia area storm surge model are similar to those of the Japan area storm surge model (Subsection 5.5.1), but with advection terms omitted. The expressions incorporate vertically integrated momentum fluxes associated with the influence of the earth's rotation with gravity acceleration:

$$\frac{\partial U}{\partial t} - fV = -g(D + \eta) \frac{\partial(\eta - \eta_0)}{\partial x} + \frac{\tau_{sx}}{\rho_w} - \frac{\tau_{bx}}{\rho_w} \quad (5.5.8a)$$

$$\frac{\partial V}{\partial t} + fU = -g(D + \eta) \frac{\partial(\eta - \eta_0)}{\partial y} + \frac{\tau_{sy}}{\rho_w} - \frac{\tau_{by}}{\rho_w} \quad (5.5.8b)$$

and the continuity equation:

$$\frac{\partial \eta}{\partial t} + \frac{\partial U}{\partial x} + \frac{\partial V}{\partial y} = 0 \quad (5.5.9)$$

Definitions of the various variables and constants are as per those of the Japan area storm surge model. Wind stresses are expressed as:

$$\tau_{sx} = c_d \rho_a W u_w \quad (5.5.10a)$$

$$\tau_{sy} = c_d \rho_a W v_w \quad (5.5.10b)$$

where c_d is the drag coefficient, ρ_a is the density of air, $W \equiv \sqrt{u_w^2 + v_w^2}$ is wind speed, and (u_w, v_w) is wind velocity. The drag coefficient is set from the results of Smith and Banke (1975) and Frank (1984):

$$c_d = \begin{cases} (0.63 + 0.066W) \times 10^{-3} & (W < 25m/s) \\ \{2.28 + 0.033(W - 25)\} \times 10^{-3} & (W \geq 25m/s) \end{cases} \quad (5.5.11)$$

5.5.2.3 Data

Bathymetry data for the storm surge model mostly come from 30-second-interval grid datasets of the General Bathymetric Chart of the Oceans (GEBCO) (Figure 5.5.6). These data are partially modified using local bathymetry data provided by Typhoon Committee Members to enable more accurate forecasts.

Astronomical tides are determined via harmonic analysis using past tide observation data provided by Typhoon Committee Members.

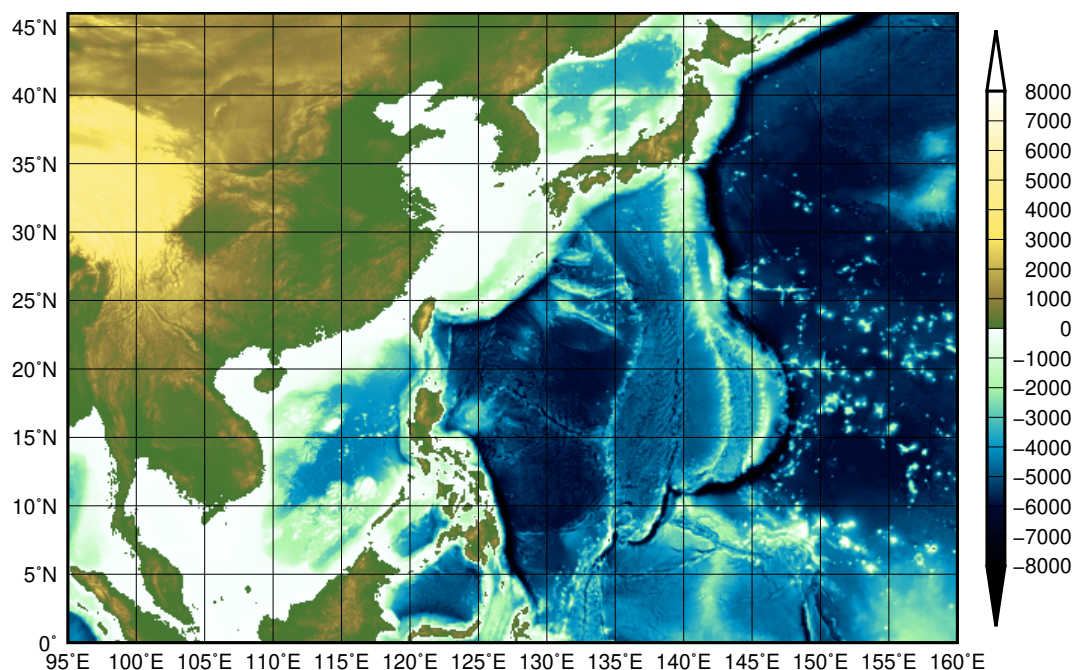


Figure 5.5.6: Model domain and topography of the Asia area storm surge model

5.5.2.4 Meteorological Forcing

Operation of the Asia area storm surge model involves the blending of a simple parametric TC model (typhoon bogus) and output of JMA's operational Global Spectral Model (GSM) as meteorological forcing fields. The simple parametric TC model in this resource is as per that of the Japan area storm surge model (Subsection 5.5.1). Related calculation requires an atmospheric model covering the Asian region, but the resolution of the atmospheric model (20 km) is insufficient for adequate expression of TC intensity. Accordingly, meteorological forcing is generated by planting bogus information into atmospheric model gridded data.

5.5.2.5 Multi-scenario Prediction

Storm surge model calculation was previously based on one scenario involving the GSM and typhoon bogus. However, deterministic forecasting is insufficient for risk management because the occurrence and the intensity of storm surge strongly depend on TC tracks. Against this background, JMA introduced multi-scenario predictions (Hasegawa *et al.* (2017)) determined from the Global Ensemble Prediction System (GEPS) (Kyouda and Higaki (2015)), which has 27 members. To cover the most representative storm surge conditions with minimal calculation, five typical scenarios are selected from all members with cluster analysis (the K-means method):

$$C_k = \frac{1}{N_k} \sum x_i, (k = 1, \dots, K) \quad (5.5.12a)$$

$$x_i = (lat_i, lon_i), (i = 1, \dots, N) \quad (5.5.12b)$$

where C_k is the cluster center and the TC location, $N = 27$, $K = 5$. Five scenarios are assumed in order to provide appropriate variance for coverage of representative scenarios. As the horizontal resolution of the

GEPS (40 km) is considered too coarse for adequate TC prediction, typhoon bogus is introduced into TC tracks of selected scenarios.

5.5.2.6 Model Specifications

Table 5.5.2 outlines the specifications of the Asia area storm surge model. The horizontal grid resolution is 2 minutes, corresponding to a distance of about 3.7 km. The model covers most of the RSMC Tokyo - Typhoon Center's area of responsibility (Figure 5.5.6), running every six hours and calculating storm surge predictions up to 72 hours ahead. If no TC is present, a single calculation is conducted with GSM prediction. If one or more TCs are present or expected, five predictions are carried out based on possible scenarios from the GEPS with bogus-modified wind and pressure fields.

Three-hourly distribution maps of the whole domain and enlarged versions showing only areas around the TC are available up to 72 hours ahead. The time-series charts provided include data on predicted/astronomical tides, storm surge, sea level pressure and surface wind. Time-series charts for 78 locations are currently provided to Typhoon Committee Members.

Table 5.5.2: Asia area storm surge model specifications

Model	2-dimensional linear model
Grid	Lat-Lon Arakawa-C grid
Region	0° - 46°N, 95°E - 160°E
Resolution	2-minutes mesh (approximately 3.7 km mesh)
Time step	8 seconds
Initial time	00, 06, 12, 18 (UTC)
Forecast time	72 hours
Member	TC case: 6 members (GSM + 5 bogus) no TC case: 1 member (GSM)

5.5.2.7 Verification

To evaluate the performance of the Asia area storm surge model, accuracy was verified by comparing predicted and observed values of hourly storm surge for eight tide stations. Predictions were calculated using GSM analysis data for 2017 and the parametric TC model with typhoon best-track data produced by the RSMC Tokyo - Typhoon Center. Figure 5.5.7 shows a scatter diagram of storm surges hindcasts against observation values. Most surge prediction errors lie in the range of ± 50 cm with a maximum of 150 cm. Yearly verification details are provided in the Annual Report on the Activities of the RSMC Tokyo - Typhoon Center ⁵.

Figure 5.5.8 and Figure 5.5.9 show storm surge distribution mapping for Typhoon Hato (T1713) and a time-series chart for Quarry Bay (Hong Kong). The typhoon generated extremely high storm surges in Hong Kong and made landfall on the coast of southern China. The results for Scenario 1 correspond closely to the observation values.

⁵<http://www.jma.go.jp/jma/jma-eng/jma-center/rsmc-hp-pub-eg/annualreport.html>

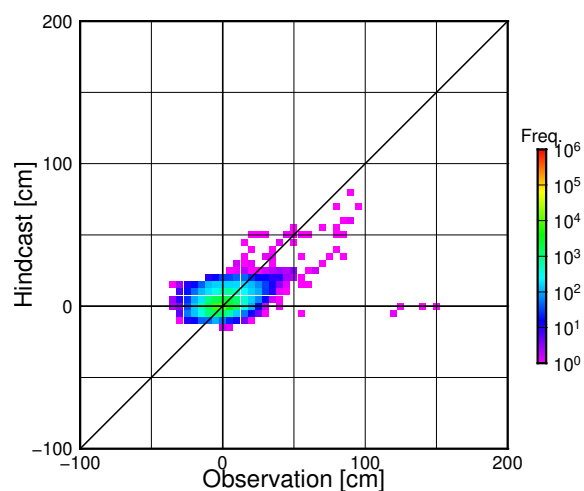


Figure 5.5.7: Scatter diagram of storm surge hindcast values against observation values

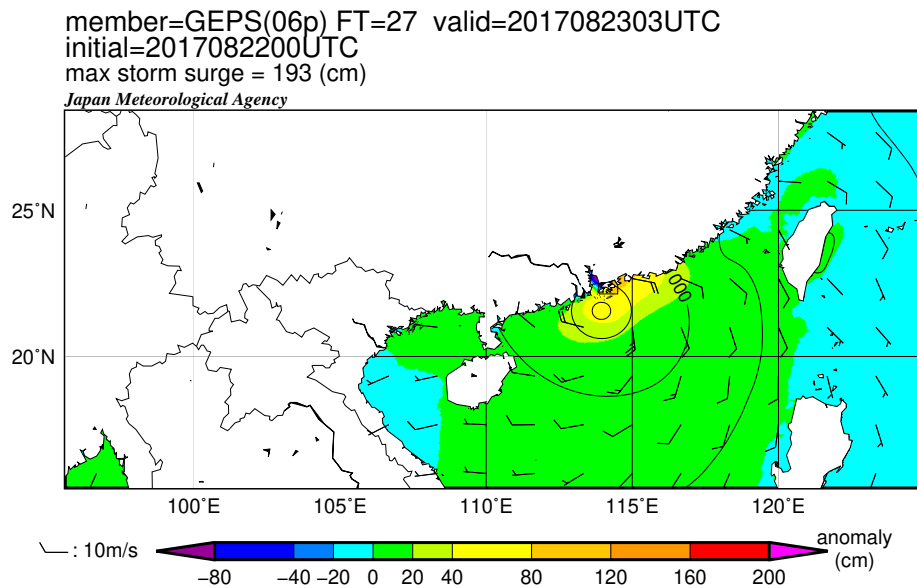


Figure 5.5.8: Distribution of storm surge at 03 UTC on 23 August as predicted by the Asia area storm surge model with an initial time of 00 UTC on 22 August

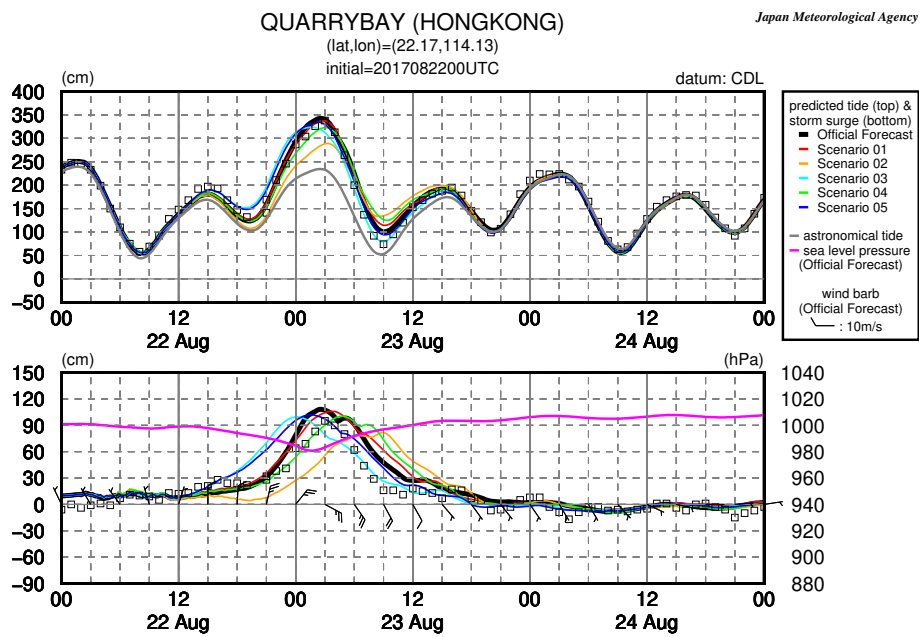


Figure 5.5.9: Time-series charts of storm tide and astronomical tide (top), storm surge, sea level pressure and surface wind (bottom) for Quarry Bay (Hong Kong) as predicted by the Asia area storm surge model with an initial time of 00 UTC on 22 August. Squares show hourly observations.

5.5.3 Astronomical Tide Analysis

5.5.3.1 Introduction

The model described in Subsection 5.5.1 calculates only storm surges, defined as anomalies from the astronomical tide level. However, prediction of storm tides (i.e., storm surge plus astronomical tides) is needed for storm surge warning issuance. In 2010, JMA changed its storm surge warning criteria to cover issuance for all coastal areas of Japan in consideration of inundation risk at all points. Appropriate issuance of warnings requires calculation to determine astronomical tides in all coastal areas.

5.5.3.2 Analysis Method

Tidal variations are expressed as a composite of periodic oscillations with various frequencies, as observed with semi-diurnal, diurnal and annual tides. Semi-diurnal and diurnal tides are caused by lunar and solar gravitational forces, while annual tides are brought by seasonal variations in seawater temperature and sea surface pressure. Harmonic constants are sets of amplitudes and phases for individual tidal constituents. Harmonic constants at tide station points can be derived by analyzing hourly tidal observation data, but cannot be determined for arbitrary coastal points where such data are unavailable using this method.

To enable analysis of astronomical tides for the whole of Japan's coast, both short-period (semi-diurnal and diurnal) and long-period (annual) tides are considered as shown in Figure 5.5.10. The constituents used in this method (Takasa *et al.* (2011)) are shown in Table 5.5.3.

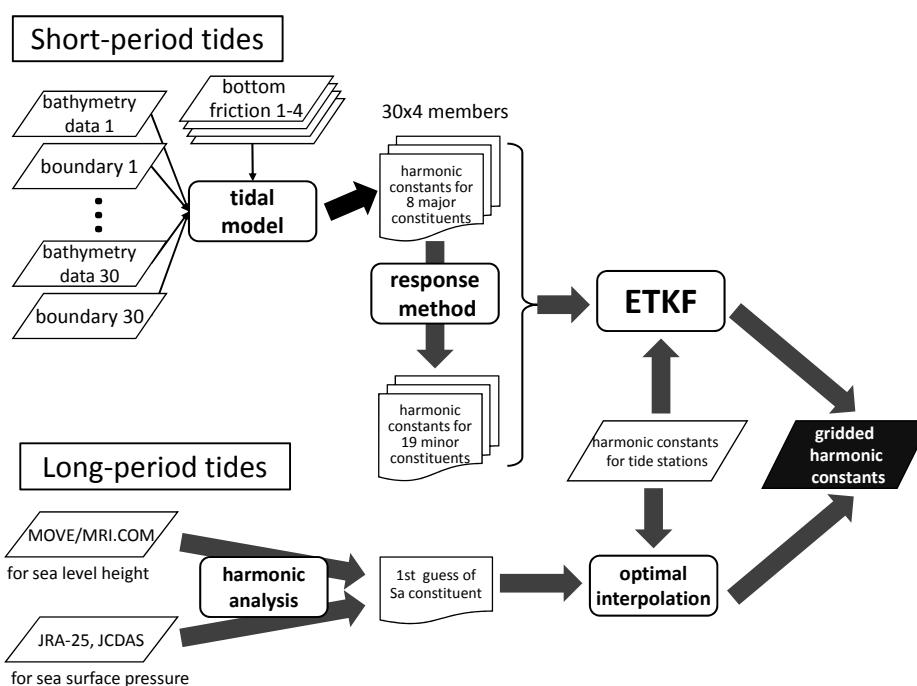


Figure 5.5.10: Flow of astronomical tide analysis

5.5.3.3 Short-period Tides

Eight major constituents with relatively large amplitudes (K1, O1, P1, Q1, K2, M2, N2 and S2; Table 5.5.3) are estimated using the Oregon State University Tidal Inversion Software (OTIS) tidal model package (Egbert and Erofeeva 2002), which involves data preparation, ocean dynamics and data assimilation. Only the linearized

Table 5.5.3: Constituents used in astronomical tide analysis.

Name	Type	Major/Minor	Name	Type	Major/Minor
S_a	annual	-	θ_1	diurnal	minor
$2Q_1$	diurnal	minor	J_1	diurnal	minor
σ_1	diurnal	minor	OO_1	diurnal	minor
Q_1	diurnal	major	$2N_2$	semi-diurnal	minor
ρ_1	diurnal	minor	μ_2	semi-diurnal	minor
O_1	diurnal	major	N_2	semi-diurnal	major
MP_1	diurnal	minor	ν_2	semi-diurnal	minor
M_1	diurnal	minor	M_2	semi-diurnal	major
χ_1	diurnal	minor	λ_2	semi-diurnal	minor
π_1	diurnal	minor	L_2	semi-diurnal	minor
P_1	diurnal	major	T_2	semi-diurnal	minor
K_1	diurnal	major	S_2	semi-diurnal	major
ψ_1	diurnal	minor	R_2	semi-diurnal	minor
ϕ_1	diurnal	minor	K_2	semi-diurnal	major

version of ocean dynamics (a tidal model), in which Fourier transform is applied to eliminate time variation, is used:

$$i\omega U - fV + gH \frac{\partial \zeta}{\partial x} + \kappa U = F_U \quad (5.5.13)$$

$$i\omega V + fU + gH \frac{\partial \zeta}{\partial y} + \kappa V = F_V \quad (5.5.14)$$

$$\left(\frac{\partial U}{\partial x} + \frac{\partial V}{\partial y} \right) + i\omega \zeta = 0 \quad (5.5.15)$$

where ω is the tidal constituent frequency, U and V are the x and y components of current integrated from the sea surface to the bottom, respectively, f is the Coriolis parameter, g is gravity acceleration, H is depth, ζ is the anomaly from mean sea level, κ is the dissipation coefficient of bottom friction, and F is the tide-generating force.

The Ensemble Transform Kalman Filter (ETKF) (Bishop *et al.* 2001) is used to assimilate harmonic constants at tide stations. As the model results contain uncertainty due to a lack of resolution and accuracy in bathymetry data and lateral boundary conditions, perturbations are added to these conditions to create an ensemble. There are 30 sets of bathymetry data (incorporating random errors) and boundary condition data generated by blending results from four tidal models (NAO.99Jb (Matsumoto *et al.* 2000), FES2004 (Lyard *et al.* 2006), GOT00.2 (an update to Ray (1999)) and TPXO (Egbert and Erofeeva 2002)). Four sets of bottom friction data are also used in consideration of the influence of such friction on tidal amplitude (Yano *et al.* (2010) and An (1977)). The 120 ensemble members are associated with a combination of the 30 sets of bathymetry and boundary condition data and 4 sets of bottom friction data.

A total of 19 minor constituents (see Table 5.5.3) are estimated from major constituents of similar frequency using the response method (Munk and Cartwright 1966).

5.5.3.4 Long-period Tides

The first guess of annual constituents (S_a) is derived from the results of harmonic analysis of reanalyzed sea level height from MOVE-WNP (see Section 5.3 and Usui *et al.* (2006)) corrected with sea surface pressure from the Japanese 25-year Reanalysis (JRA-25) and the JMA Climate Data Assimilation System (JCDAS) (see Section 2.10 and Onogi *et al.* (2007)) assuming hydrostatic balance. This is modified using harmonic constants for tide stations with the optimal interpolation (OI) method.

5.5.3.5 Verification

To verify astronomical tide analysis based on the method described, the outcomes are compared with those from harmonic constants at tide stations. Figure 5.5.11 shows a comparison of root mean square errors (RMSEs) for the respective distributions. For most stations, the error is less than 3 cm, although larger values are observed for some stations, especially in bays and inland sea areas.

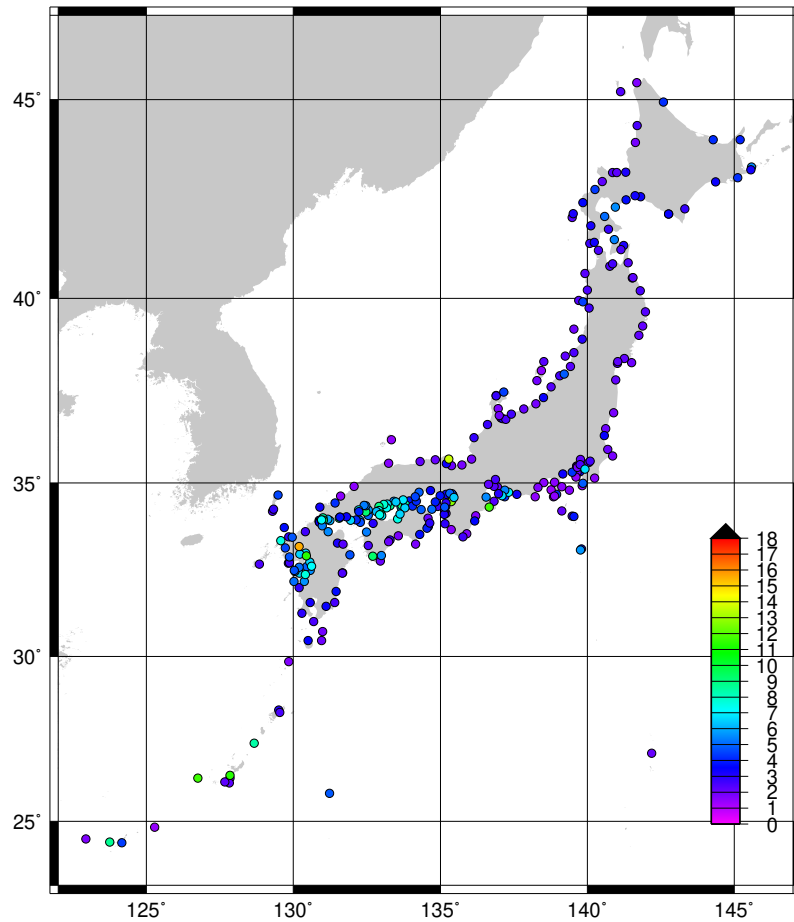


Figure 5.5.11: Distribution of RMSEs from astronomical tide analysis. The unit is cm.

5.6 Sea Ice Model

5.6.1 Introduction

The Japan Meteorological Agency (JMA) has operated a numerical sea ice model since December 1990 to support sea ice forecasting for the southern part of the Sea of Okhotsk in the winter season. Based on dynamics and thermodynamics, the model forecasts distribution and concentration of sea ice for the coming seven-day period. Its output is operationally disseminated twice a week online⁶ and via the JMH broadcast system while sea ice is present around Japan's northern island of Hokkaido.

5.6.2 Model Structure

5.6.2.1 Forecast Area

Figure 5.6.1 shows the forecast area, which is a grid of 71×71 squares each with side dimensions representing 12.5km. The model calculates the volume, concentration, velocity and thickness (=volume/concentration) of sea ice for each square based on initial data on sea ice concentration and sea surface temperature (SST), meteorological forecast data and ocean current statistics.

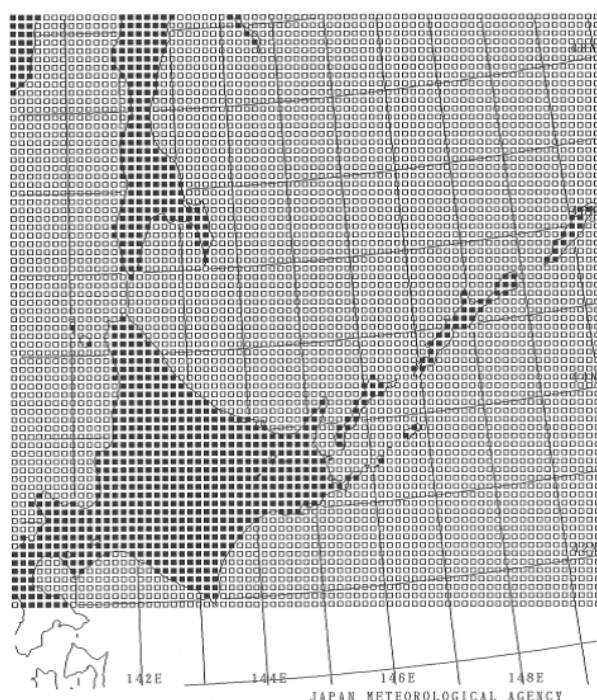


Figure 5.6.1: Forecast area of the sea ice model. ■: land grid square □: sea grid square

5.6.2.2 Calculation of Sea Ice Conditions

The volume (M_i) and concentration (A_i) of sea ice for each grid square are governed by the following equations:

$$\frac{\partial M_i}{\partial t} = -\text{div}(M_i V_i) + P_M$$

⁶<https://www.data.jma.go.jp/gmd/kaiyou/db/seaiice/forecast/nsif.html> (in Japanese) and <https://www.jma.go.jp/jmh/jmhmenu.html> (in Japanese)

$$\frac{\partial A_i}{\partial t} = -\text{div}(A_i V_i) + P_A + D_A \quad (5.6.1)$$

where V_i is sea ice velocity, as determined in the dynamical process described in Subsection 5.6.2.3. P_M and P_A denote changes in volume and concentration, respectively, caused by formation or melting of sea ice and snowfall. These values are determined in the thermodynamic process described in Subsection 5.6.2.4. D_A is a term related to the development of hummocks caused by sea-ice convergence. D_A is given by calculating the convergence of V_i : (Udin and Ullerstig 1976)

$$D_A = \begin{cases} \text{div}(A_i V_i) & A_i = 1 \quad \text{and} \quad \text{div}(V_i) < 0 \\ 0 & 0 < A_i < 1 \quad \text{or} \quad \text{div}(V_i) \geq 0. \end{cases} \quad (5.6.2)$$

5.6.2.3 Dynamical Processes

The momentum equation for sea ice is as follows: (Hibler 1979)

$$\rho_i H_i \frac{\partial V_i}{\partial t} = \tau_a(V_a) + \tau_w(V_w, V_i) + C(V_i) + G(V_w) + F_i \quad (5.6.3)$$

τ_a : wind stress,

τ_w : water stress,

C : Coriolis force,

G : pressure gradient force caused by sea surface tilt,

F_i : internal ice stress.

Here, ρ_i and $H_i (=M_i / A_i)$ are sea ice density and thickness, respectively. V_a , V_w , and V_i denote the velocity of wind, ocean currents and sea ice, respectively. As the left-side term of Eq. (5.6.3) is smaller than the other terms by more than one order of magnitude, V_i can be approximated on the assumption that the terms on the right of the equation are in balance. V_a is given by the Global Spectral Model (GSM, see Section 3.2) and V_w is given by the climatology described in Subsection 5.6.3.3. We simplified Hibler's viscous-plastic method to calculate F_i because F_i is such a quite complex term that we used a lot of computational resources. The alternate method is that a provisional sea ice velocity calculated by the assumption that the first four terms of Eq. (5.6.3) are balanced is modified with the non-slip condition for coastal grid squares.

5.6.2.4 Thermodynamic Processes

The thermodynamic processes in the model affect the formation or melting of sea ice caused by heat exchange among the atmosphere, the ocean and sea ice. In the sea ice area, heat exchange between the atmosphere and sea ice causes changes in sea ice thickness. The heat balance equation for the sea ice surface is as follows (Semtner 1976):

$$(1 - Al)R_s \downarrow + R_a \downarrow + SH(T_i) \downarrow + LH(T_i) \downarrow - FL(T_i, H_i) \downarrow - R_i(T_i) \uparrow = 0 \quad (5.6.4)$$

R_s : solar radiation,

Al : albedo of sea ice or snowfall,

R_a : atmospheric radiation,

SH : sensitive heat flux,

LH : latent heat flux,

FL : vertical heat flux in sea ice,

R_i : radiation emitted from sea ice,

T_i : surface temperature of sea ice.

R_s and R_a are given by the GSM, and T_i can be calculated from Eq. (5.6.4). If $T_i < -1.8^\circ\text{C}$, sea ice gains thickness in an amount estimated from FL . If $T_i > 0^\circ\text{C}$, sea ice loses thickness in an amount estimated from the sum of all terms on the left of Eq. (5.6.4) after T_i is set to 0°C . If $-1.8^\circ\text{C} \leq T_i \leq 0^\circ\text{C}$, sea ice remains unchanged.

In open water areas, heat exchange between the atmosphere and the ocean causes changes in sea water temperature that affect sea ice melting conditions. The ocean in the model consists of a thin surface layer and a mixed layer. The amount of heat exchange between the ocean and the atmosphere for each grid square is described as follows:

$$Q_w \downarrow = (1 - Al_w)R_s \downarrow + R_a \downarrow + SH(T_s) \downarrow + LH(T_s) \downarrow - R_w(T_s) \uparrow \quad (5.6.5)$$

Here R_w denotes radiation emitted from the sea surface, and Al_w is the albedo of sea water. T_s is the temperature of the surface layer. Heat exchange between the sea surface layer and the mixed layer is calculated as follows:

$$T_s = \frac{(T_s - T_f)D_s + (T_m - T_f)D_m}{D_s + D_m} + T_f \quad (5.6.6)$$

Here D_s and D_m denote the depth of the surface layer and the mixed layer, respectively, and are fixed for each sea grid square. T_m is the temperature of the mixed layer, and T_f is the freezing point (-1.8°C) of sea water. Here, calculation of sea water temperature change is based on the assumptions that direct heat exchange between sea ice and sea water occurs only through the surface layer and that heat exchange between sea ice and the surface layer occurs to drive T_s to the melting point (0°C).

5.6.3 Data Used in the Model

5.6.3.1 Initial Data on Sea Ice and Sea Surface Temperature

Initial fields of sea ice concentration are subjectively estimated on the basis of data from satellites (mainly HIMAWARI and NOAA/Metop), aircraft, ships and coastal observations, while initial fields of sea ice thickness are derived from the previous forecast and daily SST analysis data for the seas around Japan (given by MGDSSST; Subsection 5.2.1) are used as initial fields for SST.

5.6.3.2 Meteorological Data

Air pressure, air temperature, wind, dew point, solar radiation, atmospheric radiation and precipitation on the sea surface for each grid square are given from interpolation of predictions made by the atmospheric numerical model (GSM).

5.6.3.3 Ocean Current Data

The distribution of ocean currents used in the model is obtained from the Japan Maritime Safety Agency (1983) as shown in Figure 5.6.2. The value is fixed throughout the winter season.

5.6.4 Numerical Sea Ice Model Results

In the example of seven-day forecast results shown in Figure 5.6.3, the model projects that sea ice will move southward to the Sea of Okhotsk coast of Hokkaido.

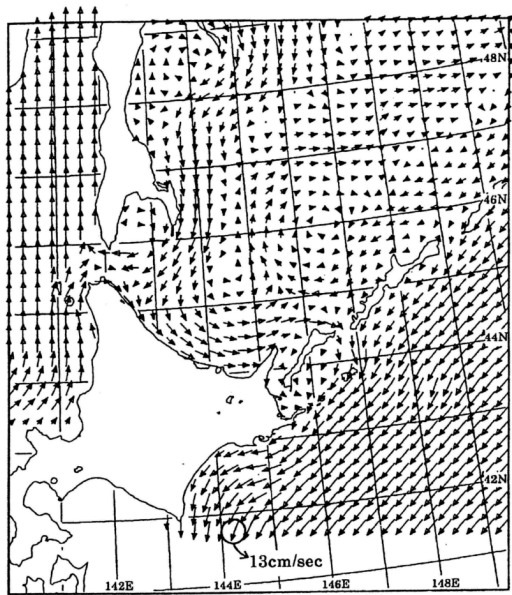


Figure 5.6.2: The ocean currents used in the model.

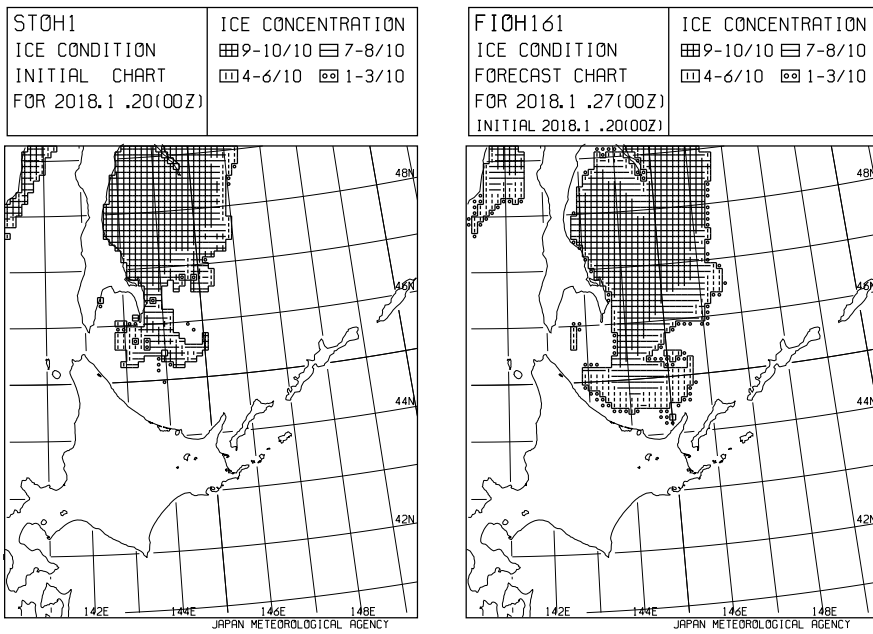


Figure 5.6.3: An example of the results of the model.

5.7 Oil Spill Prediction Model

5.7.1 Introduction

In the 1990s, large-scale oil spills (such as those involving the vessels ABT Summer, MB Braer and Sea Empress) frequently occurred around the world. In Japan, the wrecking of the Russian tanker Nakhodka resulted in a serious oil spill in the Sea of Japan in January 1997, causing major environmental damage along Japan's western coast. Following the Japanese Government's subsequent consideration of countermeasures for large-scale oil spills, JMA has operated its Oil Spill Prediction Model since October 1999.

The model predicts the large-scale behavior of oil spilled in offshore seas, where tidal currents are negligible. Based on accident information from the Japan Coast Guard, JMA operates the model to produce forecasts with lead times of up to 192 hours. The results are provided to personnel responsible for emergency response operations.

The model is applicable to the entire western North Pacific. The domain of calculation is selected from seven settings from $0.8^\circ \times 0.8^\circ$ to $12^\circ \times 12^\circ$ in latitude and longitude based on consideration of incident conditions.

5.7.2 Basic Equation

The oil spill prediction model is generally described by the following equation including terms of advection and diffusion,

$$\frac{dC}{dt} = \frac{\partial C}{\partial t} + \mathbf{V} \cdot \nabla C = \nabla \cdot (K \nabla C) + S \quad (5.7.1)$$

where C is pollutant concentration, t is time, \mathbf{V} is advection velocity, K is the turbulent diffusion coefficient, and S (referred to as the source term) represents the process involved in changing the total amount of oil spilled through changes in oil properties.

Equation (5.7.1) in general can be numerically solved either by calculating C directly using the finite difference method or by simulating the behaviors of a large number of particles representing oil components. The latter approach is used in JMA's Oil Spill Prediction Model. Spilled oil is expressed as numerous particles $C_n(n = 1, 2, \dots)$ using:

$$C_n \{ \mathbf{x}(t + \delta t), s(t + \delta t); t + \delta t \} = \Phi [C_n \{ \mathbf{x}(t), s(t); t \}, \delta t] \quad (5.7.2)$$

where $\mathbf{x} = (x, y, z)$ indicates the position of each particle and s is the chemical status of the oil. Φ is a general function describing oil property changes over time.

In the advection term, the effects of surface winds, ocean waves, and ocean currents are taken into account as potentially major factors. Ekman drift current generated by sea surface winds is an example of such an influence. In the JMA model, surface flows are determined as 2.5% of the wind speed with an angle of 15° clockwise with respect to the wind direction. As another example, Stokes drift involves forward movement of particles at the sea surface in the wave direction as a result of wave motion back and forth in each wave cycle. This effect is more significant when high waves are present, and is independent of wind when swell is predominant. Accordingly, Stokes drift is included explicitly and calculated from conditions predicted by JMA ocean wave models. Ocean currents are provided by the JMA Ocean Data Assimilation System for the Western North Pacific (MOVE-WNP; Subsection 5.3.3).

The three-dimensional diffusion of oil is basically calculated via the shear diffusion treatment proposed by Elliott (1986). Surface flow is assumed to have a logarithmic profile in the vertical direction, and spilled oil is assumed to be carried at a particular horizontal speed in each water level. The shear mechanism is also associated with vertical diffusion. Spilled oil is divided into a large number of droplets with varying levels of buoyancy in line with their size. Consideration of this buoyancy and the present depth of oil drops allows determination of oil motion in the vertical direction and clarification of whether the oil floats on the surface.

In addition to the above shear diffusion process, isotropic diffusion may also be generated by small scale eddies and similar influences as estimated using the constant diffusion coefficient $K_h = 95.0 \text{ m}^2/\text{s}$. As such

Table 5.7.1: Oil Spill Prediction Model specifications

Applicable area	10°S – 65°N, 120°E – 180°E	
Domain of calculation	7 options (0.8° × 0.8° – 12° × 12°)	
Grid spacing	7 options (2–30km), according to the domain of calculation	
Number of grids	41 × 41	
Prediction period	192 hours	
Physical and chemical process	Advection	Ekman drift (estimated from wind field of Global Spectrum Model) Stokes drift (estimated from wave field of Global/Coastal Wave Models) Ocean current (MOVE/MRI.COM-NWP)
	Diffusion	Elliott (1986) etc.
	Evaporation	Fingas (2010)
	Emulsification	Reed (1989)

diffusion may be greater in conditions of strong wind or high waves, the influence is parameterized with additional diffusion coefficients:

$$\begin{cases} \text{waves: } K_{wv} = 500.0H_w^2/T_w \\ \text{winds: } K_{wnd} = 5.0W^3/g \end{cases} \quad (5.7.3)$$

where H_w and T_w are the wave height and period, W is wind speed, and g is gravitational acceleration. The coefficients are empirically determined on the basis of actual cases.

Additional diffusion of oil parcels is estimated from the total value of the diffusion coefficients (K_h , K_{wv} , and K_{wnd}). Specific values are calculated using the random walk method with such diffusion assumed to be horizontal.

Due to the complex behavior of spilled oil, consideration of all related chemical processes is largely impractical. Accordingly, only evaporation and emulsification are considered as major processes. Evaporation is estimated using empirical formulae (Fingas 2010), in which the evaporation rate E_v (%) of most oils can be expressed by the form of either the logarithmic or the root profile over time.

$$E_v = \begin{cases} (a + b \cdot T) \ln t \\ (a + b \cdot T) \sqrt{t} \end{cases} \quad (5.7.4)$$

The constant coefficients a and b are based on experimental results and are listed in the Environment Canada oil data catalogue. T represents oil temperature, and is assumed to be equivalent to sea surface temperature (SST). t is the number of minutes elapsed since the spill.

Emulsification is calculated using the formula of Reed (1989), by which the water content F_{wc} is estimated as:

$$\frac{dF_{wc}}{dt} = 2.0 \times 10^{-6} (W + 1)^2 \cdot \left(1 - \frac{F_{wc}}{C_3}\right) \quad (5.7.5)$$

where W (m/s) is wind speed. C_3 is a constant parameter for the upper limit of water content, and differs among oil types. Oil density is also calculated in consideration of water content, which can change the behavior of oil.

The specifications of the Oil Spill Prediction Model and related processes are summarized in Table 5.7.1.

5.7.3 Products

The model is operated in the event of a large-scale oil spill in offshore deep-water seas, where short-term tidal currents can be negligible. The results of oil spill prediction are provided to the Japanese Government and/or the Japan Coast Guard along with various marine meteorological charts. An example of prediction is shown in Figure 5.7.1.

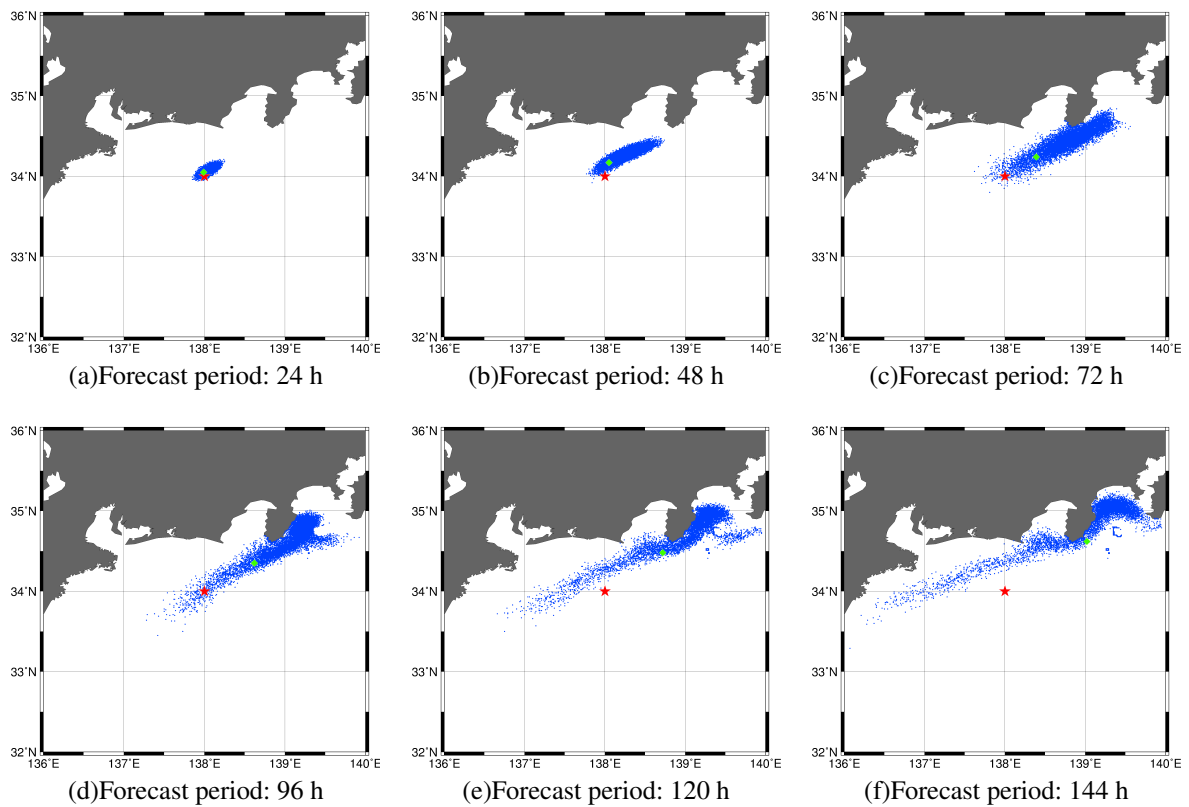


Figure 5.7.1: A sample test simulation for the sea south of Japan supposing an accident at 00UTC on 17 May 2018 at 34.0 °N 138.0 °E. The stars show the accident location, and the rhombuses show the source of the spillage (assuming the source is carried by ocean currents). The area and amount of spilled oil are indicated by dot distribution.

Appendix A

Verification Indices

This appendix highlights a number of verification indices referenced in this document. The indices are also used in international verification via the Global Data-processing and Forecasting System (GDPFS) of the World Meteorological Organization (WMO 2017).

A.1 Basic Verification Indices

A.1.1 Mean Error

Mean Error (ME), also called Bias, represents the mean value of deviations between forecasts and verification values, and is defined by

$$\text{ME} \equiv \left(\sum_{i=1}^n w_i D_i \right) / \sum_{i=1}^n w_i, \quad (\text{A.1.1a})$$

$$D_i = F_i - A_i, \quad (\text{A.1.1b})$$

$$w_i = \frac{1}{n} \text{ (or } \cos \phi_i, \text{ and so on),} \quad (\text{A.1.1c})$$

where F_i , A_i , and D_i represent the forecast, the verification value, and the deviation between the forecast and the verification value, respectively. w_i represents the weighting coefficient, n is the number of samples, and ϕ_i is latitude. In general, observational values, initial values or objective analysis values are often used as verification values. When the forecast is fully correct, called a *perfect forecast*, ME is equal to zero.

Calculation of an average over an extensive region such as the Northern Hemisphere requires evaluation with weighting coefficients in consideration of latitude-related differences among areas. By way of example, to evaluate objective analysis in an equirectangular projection, the weighting coefficient “ $w_i = 1/n$ ” is often replaced with the cosine of latitude “ $\cos \phi_i$ ”. The other indices in Section A.1 are handled in the same manner.

A.1.2 Root Mean Square Error

Root Mean Square Error (RMSE) is often used to represent forecast accuracy, and is defined by

$$\text{RMSE} \equiv \sqrt{\sum_{i=1}^n w_i D_i^2} / \sqrt{\sum_{i=1}^n w_i}, \quad (\text{A.1.2})$$

where D_i represents deviation between forecast and verification values in Eq. (A.1.1b), w_i represents the weighting coefficient in Eq. (A.1.1c), and n is the number of samples. Proximity of the RMSE to zero indicates that forecast values are closer to verification values. For a perfect forecast, RMSE is equal to zero. With the components of ME and random error separated, RMSE is expressed as follows:

$$\text{RMSE}^2 = \text{ME}^2 + \sigma_e^2, \quad (\text{A.1.3})$$

where σ_e represents standard deviation (SD) for the deviation D_i , and is given by

$$\sigma_e^2 = \left(\sum_{i=1}^n w_i (D_i - \text{ME})^2 \right) / \sum_{i=1}^n w_i. \quad (\text{A.1.4})$$

A.1.3 Anomaly Correlation Coefficient

The anomaly correlation coefficient (ACC) is one of the most widely used measures in the verification of spatial fields (Jolliffe and Stephenson 2003), and represents the correlation between anomalies of forecasts and those of verification values with reference values such as climatological data. ACC is defined as follows:

$$\text{ACC} \equiv \frac{\sum_{i=1}^n w_i (f_i - \bar{f})(a_i - \bar{a})}{\sqrt{\sum_{i=1}^n w_i (f_i - \bar{f})^2 \sum_{i=1}^n w_i (a_i - \bar{a})^2}}, \quad (-1 \leq \text{ACC} \leq 1), \quad (\text{A.1.5})$$

where n is the number of samples, and f_i, \bar{f}, a_i and \bar{a} are given by the following equations:

$$f_i = F_i - C_i, \quad \bar{f} = \left(\sum_{i=1}^n w_i f_i \right) / \sum_{i=1}^n w_i, \quad (\text{A.1.6a})$$

$$a_i = A_i - C_i, \quad \bar{a} = \left(\sum_{i=1}^n w_i a_i \right) / \sum_{i=1}^n w_i, \quad (\text{A.1.6b})$$

where F_i, A_i , and C_i represent the forecast value, the verification value and a reference such as a climatological value, respectively. \bar{f} is the mean of f_i , \bar{a} is the mean of a_i , and w_i represents the weighting coefficient in Eq. (A.1.1c). If the variation pattern of forecast anomalies is perfectly coincident with that of verification anomalies, the ACC will take the maximum value of 1. Conversely, if the variation pattern is completely reversed, it will take the minimum value of -1.

A.1.4 Ensemble Spread

Ensemble Spread is a familiar measure representing the degree of forecast uncertainty in the ensemble forecast. It is the standard deviation of the ensembles as defined by

$$\text{Ensemble Spread} \equiv \sqrt{\frac{1}{N} \sum_{i=1}^N \left(\frac{1}{M} \sum_{m=1}^M (F_{m,i} - \bar{F}_i)^2 \right)}, \quad (\text{A.1.7})$$

where M is the number of ensemble members, N is the number of samples, $F_{m,i}$ represents the forecast of the m th member, and \bar{F}_i is the ensemble mean, defined by

$$\bar{F}_i \equiv \frac{1}{M} \sum_{m=1}^M F_{m,i}. \quad (\text{A.1.8})$$

Table A.2.1: Schematic contingency table for categorical forecasts of a binary event. The numbers of outcomes in each category are indicated by FO , FX , XO and XX , and N is the total number of events.

	Observed	Not Observed	Total
Forecasted	FO (hits)	FX (false alarms)	$FO + FX$
Not Forecasted	XO (misses)	XX (correct rejections)	$XO + XX$
Total	M	X	N

A.1.5 S1 Score

The S1 Score is often used to measure the degree of error in the depiction of forecast pressure fields, and is defined by

$$S1 \equiv 100 \times \frac{\sum_{i=1}^n w_i \{|\partial_x D_i| + |\partial_y D_i|\}}{\sum_{i=1}^n w_i [\max(|\partial_x F_i|, |\partial_x A_i|) + \max(|\partial_y F_i|, |\partial_y A_i|)]}, \quad (\text{A.1.9})$$

where F_i and A_i represent forecast and verification values, respectively. D_i is the deviation between the forecast and verification values in Eq. (A.1.1b), w_i is the weighting coefficient in Eq. (A.1.1c), and the subscripts x and y denote the differential with respect to x and y, as expressed by

$$\partial_x X = \frac{\partial X}{\partial x}, \quad \partial_y X = \frac{\partial X}{\partial y}. \quad (\text{A.1.10})$$

Lower S1 Scores indicate superior forecasts.

A.2 Verification Indices for Categorical Forecasts

Many meteorological phenomena can be regarded as simple binary events, and related forecasts or warnings are often issued as unqualified statements indicating that such events will or will not occur (Jolliffe and Stephenson 2003). In the verification of forecasts for binary events, outcomes for the targeted phenomenon are distinguished in terms of correspondence between forecasts and observations using a 2×2 contingency table as shown in Table A.2.1.

A.2.1 Contingency Table

In the contingency table, categorical forecasts for a binary event are divided into hits, false alarms, misses and correct rejections (or correct negatives) with numbers expressed as FO , FX , XO and XX , respectively. The total number of events is the sum of numbers for all outcomes, given by $N = FO + FX + XO + XX$. The numbers of observed events and non-observed events are $M = FO + XO$ and $X = FX + XX$, respectively.

A.2.2 Proportion Correct

Proportion Correct (PC) is the ratio of the number of correct events $FO + XX$ to the total number of events N , and is defined by

$$PC \equiv \frac{FO + XX}{N}, \quad (0 \leq PC \leq 1). \quad (\text{A.2.1})$$

Higher PC values indicate higher forecast accuracy.

A.2.3 False Alarm Ratio

The false alarm ratio (FAR) is the ratio of the number of false alarm events FX to the number of forecast events $FO + FX$, and is defined by

$$FAR \equiv \frac{FX}{FO + FX}, \quad (0 \leq FAR \leq 1). \quad (\text{A.2.2})$$

Lower FAR values indicate a lower number of false alarm events. In some cases, the total number N is used as the denominator in Eq. (A.2.2) instead of $FO + FX$.

A.2.4 Undetected Error Rate

The undetected error rate (Ur) is the ratio of the number of miss events XO to the number of observed events M , and is defined by

$$Ur \equiv \frac{XO}{M}, \quad (0 \leq Ur \leq 1). \quad (\text{A.2.3})$$

Lower Ur values indicate a lower number of miss events. In some cases, the total number N is used as the denominator in Eq. (A.2.3) instead of M .

A.2.5 Hit Rate

The hit rate (Hr) is the ratio of the number of hit events FO to the number of observed events M , and is defined by

$$Hr \equiv \frac{FO}{M}, \quad (0 \leq Hr \leq 1). \quad (\text{A.2.4})$$

Higher Hr values indicate a lower number of miss events. The hit rate is used to plot the ROC curve described in Subsection A.3.5.

A.2.6 False Alarm Rate

The false alarm rate (Fr) is the ratio of the number of false alarm events FX to the number of non-observed events X , and is defined by

$$Fr \equiv \frac{FX}{X}, \quad (0 \leq Fr \leq 1). \quad (\text{A.2.5})$$

Lower Fr values indicate a lower number of false alarm events and higher forecast accuracy. The denominator of the false alarm rate differs from that of the false alarm ratio (see Subsection A.2.3). The false alarm rate is also used to plot the ROC curve described in Subsection A.3.5.

A.2.7 Bias Score

The bias score (BI) is the ratio of the number of forecasted events $FO + FX$ to the number of observed events M , and is defined by

$$BI \equiv \frac{FO + FX}{M}, \quad (0 \leq BI). \quad (A.2.6)$$

If the number of forecasted events $FO + FX$ is equal to the number of observed events M , BI will be unity. If BI is larger than unity, the frequency of events is overestimated. Conversely, if BI is smaller than unity, the frequency of events is underestimated.

A.2.8 Climatological Relative Frequency

Climatological relative frequency (P_c) is the probability of occurrence of events estimated from samples, and is defined by

$$P_c \equiv \frac{M}{N}, \quad (A.2.7)$$

where M is the number of observed events occurring, and N is the total number of events. P_c is derived from the number of observed events, and is independent of forecast accuracy.

A.2.9 Threat Score

The threat score (TS) is an index value focused on hit events. It represents the ratio of the number of hit events FO to the number of events other than correct rejections $FO + FX + XO$, and is defined by

$$TS \equiv \frac{FO}{FO + FX + XO}, \quad (0 \leq TS \leq 1). \quad (A.2.8)$$

If the number of observed events is extremely small (i.e. $N \gg M$, and $XX \gg FO, FX$, or XO), the proportion correct (PC) value will be close to unity due to the the major contribution from the number of non-observed events. The TS is applicable to validation of forecasts accuracy without contribution from correct rejection events. Forecast accuracy rises as the TS value approaches the maximum value of unity. As TS values are often affected by climatological relative frequency, they are not applicable to comparison regarding the accuracy of forecasts validated under different conditions. To address this issue, equitable threat scores are often used for validation.

A.2.10 Equitable Threat Score

The equitable threat score (ETS) is similar to the threat score, but with the removal of contribution from hits by chance in *random forecasts*, and is defined by

$$ETS \equiv \frac{FO - S_f}{FO + FX + XO - S_f}, \quad \left(-\frac{1}{3} \leq ETS \leq 1\right), \quad (A.2.9)$$

and

$$S_f = P_c(FO + FX), \quad P_c = \frac{M}{N}, \quad (A.2.10)$$

where P_c is the climatological relative frequency and S_f is the number of hit events being forecast randomly $FO + FX$ times. Proximity to the maximum value of unity indicates higher forecast accuracy. For random forecasts, the ETS is zero. This metric has a minimum value of $-1/3$ if $FO = XX = 0$ and $FX = XO = N/2$.

A.2.11 Heidke Skill Score

The Heidke skill score (HSS) is used to remove the effects of issues in individual forecasts in consideration of the number of correct events in a random forecast estimated from climatological probability, and is defined by

$$\text{HSS} \equiv \frac{FO + XX - S}{N - S}, \quad (-1 \leq \text{HSS} \leq 1), \quad (\text{A.2.11})$$

where

$$S = P_c(FO + FX) + P_{x_c}(XO + XX), \quad (\text{A.2.12})$$

and

$$P_c = \frac{M}{N}, \quad P_{x_c} = \frac{X}{N} = 1 - P_c, \quad (\text{A.2.13})$$

where P_c and P_{x_c} are the climatological relative frequencies of observed and non-observed events in random forecasting, respectively. Proximity to the maximum value of unity indicates higher forecast accuracy. The Heidke skill score is zero in random forecasts and unity in perfect forecasts. The index has a minimum value of -1 if $FO = XX = 0$ and $FX = XO = N/2$.

A.2.12 Fractions Skill Score

The fractions skill score (FSS) is an index of how forecast skill varies with spatial scale. In other words, it is a measure to verify forecasted fractional event frequencies. The verification method (Roberts and Lean, 2008) is described here.

First, all model and observation data are projected onto the same verification grid. Suitable thresholds (q) are chosen and used to convert the observed (O) and forecast (F) fields into binary fields I_O and I_F . All grid squares exceeding the threshold have a value of 1 and all others a value of 0,

$$I_o = \begin{cases} 1 & (O \geq q) \\ 0 & (O < q) \end{cases} \quad \text{and} \quad I_F = \begin{cases} 1 & (F \geq q) \\ 0 & (F < q) \end{cases} \quad (\text{A.2.14})$$

Second, for every grid point in the binary fields obtained from Eq. (A.2.14), computation is performed to determine the fraction of surrounding points within a given square of length n that have a value of 1. These are described by

$$O(n)_{i,j} = \frac{1}{n^2} \sum_{k=1}^n \sum_{l=1}^n I_o \left[i + k - 1 - \frac{(n-1)}{2}, j + l - 1 - \frac{(n-1)}{2} \right], \quad (\text{A.2.15})$$

$$F(n)_{i,j} = \frac{1}{n^2} \sum_{k=1}^n \sum_{l=1}^n I_F \left[i + k - 1 - \frac{(n-1)}{2}, j + l - 1 - \frac{(n-1)}{2} \right]. \quad (\text{A.2.16})$$

Third, the mean square error (MSE) for the observed and forecast fractions from the neighborhood of length n is computed using

$$\text{MSE}(n) = \frac{1}{N_x N_y} \sum_{i=1}^{N_x} \sum_{j=1}^{N_y} [O(n)_{i,j} - F(n)_{i,j}]^2. \quad (\text{A.2.17})$$

Here i goes from 1 to N_x , where N_x is the number of columns in the domain, and j goes from 1 to N_y , where N_y is the number of rows. $O(n)_{i,j}$ is the resultant field of observed fractions for the square of length n and $F(n)_{i,j}$ is the resultant field of model forecast fractions. However, the MSE is not in itself very useful because it is highly dependent on the frequency of the event itself. The fractions skill score is defined by

$$\text{FSS}(n) = \frac{\text{MSE}(n) - \text{MSE}(n)_{ref}}{\text{MSE}(n)_{perfect} - \text{MSE}(n)_{ref}} = 1 - \frac{\text{MSE}(n)}{\text{MSE}(n)_{ref}} \quad (\text{A.2.18})$$

where $\text{MSE}(n)_{perfect}$ is the MSE of a perfect forecast for neighborhood length n and $\text{MSE}(n)_{ref}$ is the MSE of the reference.

A.3 Verification Indices for Probability Forecasts

A.3.1 Brier Score

The Brier score (BS) is a basic verification index for probability forecasts, and is defined by

$$\text{BS} \equiv \frac{1}{N} \sum_{i=1}^N (p_i - a_i)^2, \quad (0 \leq \text{BS} \leq 1), \quad (\text{A.3.1})$$

where p_i is the forecast probability of occurrence of an event ranging from 0 to 1 in probability forecasts, a_i indicates observation with binary values (1 for observed and 0 for not observed), and N is the number of samples. Smaller BS values indicate higher forecast accuracy. In a perfect forecast, the BS has a minimum value of 0.

The Brier score for *climatological forecasts* (BS_c), in which the climatological relative frequency $P_c = M/N$ is always used as the forecast probability p_i , is defined by

$$\text{BS}_c \equiv P_c(1 - P_c), \quad (\text{A.3.2})$$

Since the Brier score is influenced by the climatological frequency of events in the verification sample, it is not applicable to comparison of accuracy for forecasts with different sets of samples and/or different phenomena. For example, BS_c may differ with differing values of P_c even under the same forecast method (e.g., the climatological approach) because of its dependence on P_c . To reduce this effect, the Brier skill score is often used for verification with improvement from the climatological forecast (see Subsection A.3.2).

A.3.2 Brier Skill Score

The Brier skill score (BSS) is an index based on the Brier score. It indicates the degree of forecast improvement in reference to climatological forecasts, and is defined by

$$\text{BSS} \equiv \frac{\text{BS}_c - \text{BS}}{\text{BS}_c}, \quad (\text{BSS} \leq 1), \quad (\text{A.3.3})$$

where BS is the Brier score and BS_c is the Brier score for the climatological forecast. BSS is unity for a perfect forecast and zero for the climatological forecast. Its value is negative if the forecast error exceeds that of the climatological forecast.

A.3.3 Murphy's Decompositions

To provide deeper insight into the relationship between the Brier score (BS) and the properties of probability forecasts, Murphy (1973) decomposed the score into reliability, resolution and uncertainty terms (Eq. A.3.4a), referred to as Murphy's Decompositions.

Consider the probability of forecasts classified to L intervals. Let the sample number in the l th interval be N_l , and let the number of observed events in N_l be M_l . It follows that $N = \sum_{l=1}^L N_l$ and $M = \sum_{l=1}^L M_l$. The BS value can therefore be represented with Murphy's decompositions as follows:

$$\text{BS} = \text{Reliability} - \text{Resolution} + \text{Uncertainty}, \quad (\text{A.3.4a})$$

$$\text{Reliability} = \sum_{l=1}^L \left(p_l - \frac{M_l}{N_l} \right)^2 \frac{N_l}{N}, \quad (\text{A.3.4b})$$

$$\text{Resolution} = \sum_{l=1}^L \left(\frac{M}{N} - \frac{M_l}{N_l} \right)^2 \frac{N_l}{N}, \quad (\text{A.3.4c})$$

$$\text{Uncertainty} = \frac{M}{N} \left(1 - \frac{M}{N} \right), \quad (\text{A.3.4d})$$

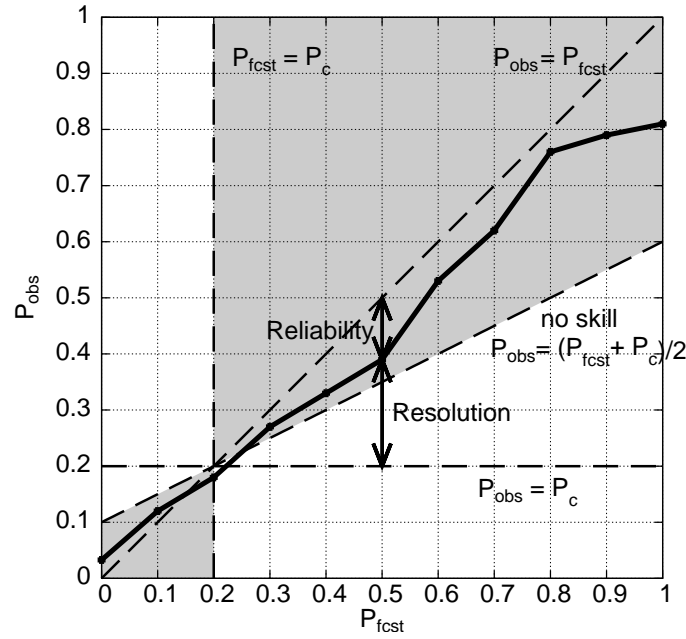


Figure A.3.1: Reliability diagram. The ordinate represents the relative frequencies of observed events P_{obs} , the abscissa is the probability of forecast event occurrence P_{fcst} , and the solid line is the reliability curve. Grey shading indicates positive contribution to the BSS.

where p_l is the representative value in the l th interval of predicted probability. Reliability is the minimum value of zero when p_l is equal to the relative frequency of the observed events M_l/N_l . If the distance between $M/N (= P_c)$ and M_l/N_l is longer, resolution will have a large value. Uncertainty depends on observed events regardless of forecast methods. When $P_c = 0.5$, Uncertainty will have the maximum value of 0.25. Uncertainty is equal to the Brier score for climatological forecasts (BS_c). In this regard, the Brier skill score (BSS) can be expressed as

$$\text{BSS} = \frac{\text{Resolution} - \text{Reliability}}{\text{Uncertainty}}. \quad (\text{A.3.5})$$

A.3.4 Reliability Diagram

Probability forecast performance is often evaluated using a reliability diagram, also called an attributes diagram. This is a chart detailing the relative frequencies of observed events P_{obs} as the ordinate and the probability of forecast event occurrence P_{fcst} as abscissa as shown in Figure A.3.1. The plots are generally displayed in the form of a reliability curve.

The properties of the curve can be related to the reliability and resolution terms of Murphy's decompositions. Contribution to reliability (or resolution) for each value of P_{fcst} is associated with the squared distance from a point on the reliability curve to the line $P_{\text{obs}} = P_{\text{fcst}}$ (or $P_{\text{obs}} = P_c$), and is derived from its weighted mean using the number of samples as weights. The contributions are the same for both reliability and resolution on the line $P_{\text{obs}} = (P_{\text{fcst}} + P_c)/2$, called the no-skill line, and contribution to the Brier score is zero on this line. The shading enclosed by the no-skill line, the line $P_{\text{fcst}} = P_c$ and the axes in Figure A.3.1 indicate the area of positive contribution to the BSS, since the contribution to reliability is larger than that to resolution. For further details of reliability diagrams, see Wilks (2006).

In climatological forecasting (see Subsection A.3.1) as a special case, the reliability curve corresponds to a point $(P_{\text{fcst}}, P_{\text{obs}}) = (P_c, P_c)$. Probability forecasts with the following properties will have higher accuracy.

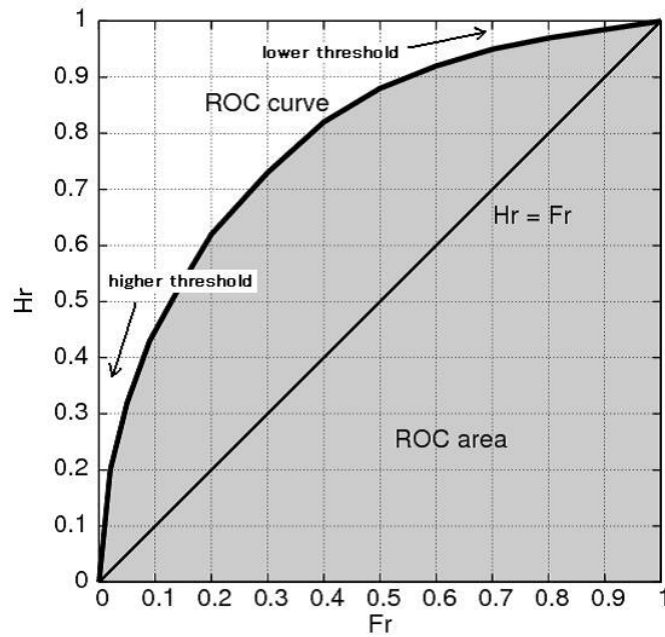


Figure A.3.2: Schematic diagram of an ROC curve. The ordinate is Hr and the abscissa is Fr. Gray shading indicates the ROC area.

- The reliability curve is close to the line $P_{\text{obs}} = P_{\text{fcst}}$ (reliability close to zero).
- Points with a large number of samples on the reliability curve are distributed away from the point of the climatological forecast $(P_{\text{fcst}}, P_{\text{obs}}) = (P_c, P_c)$ (around the lower left or the upper right of the reliability diagram) with higher resolution.

A.3.5 ROC Area Skill Score

If two alternatives in a decision problem, whether the event occurs or not, must be chosen on the basis of a probability forecast for a dichotomous variable, the choice will depend on the probability threshold. A relative operating characteristic (ROC) curve is often used to evaluate such decision problems. This involves the use of a schematic diagram whose ordinate and abscissa represent the hit rate (Hr) and the false alarm rate (Fr), respectively, and is made from contingency tables with variations of threshold values as shown in Figure A.3.2.

The threshold value is lower around the upper right of the diagram and higher around the lower left. Probability forecasting is more accurate when the curve is more convex to the top because the hit rate is higher than the false alarm rate; that is, $Hr > Fr$ around the upper left. The gray shaded area below the ROC curve, called the ROC area (ROCA), will be larger with higher values of information in probability forecasts. For further details of ROC curves, see Wilks (2006).

The ROC area skill score (ROCASS) is a validation index in reference to probability forecasts with no information values (i.e. $Hr = Fr$), and is defined by

$$\text{ROCASS} \equiv 2(\text{ROCA} - 0.5), \quad (-1 \leq \text{ROCASS} \leq 1). \quad (\text{A.3.6})$$

ROCASS is unity for a perfect forecast and zero for a forecast with no information values, such as one with a uniform probability as randomly sampled from the range $[1, 0]$.

References

- Abel, S. J. and I. A. Boutle, 2012: An improved representation of the raindrop size distribution for single-moment microphysics schemes. *Quart. J. Roy. Meteor. Soc.*, **138**, 2151–2162.
- Abel, S. J. and B. J. Shipway, 2007: A comparison of cloud-resolving model simulations of trade wind cumulus with aircraft observations taken during RICO. *Quart. J. Roy. Meteor. Soc.*, **133**, 781–794.
- Agresti, A., 2002: *Categorical Data Analysis, 2nd ed.* New York: Wiley, 734pp.
- An, H. S., 1977: A Numerical Experiment of the M2 Tide in the Yellow Sea. *J. Oceanogr.*, **33**, 103–110.
- Anderson, J. L., 2001: An Ensemble Adjustment Kalman Filter for Data Assimilation. *Mon. Wea. Rev.*, **129**, 2884–2903.
- Andersson, E. and H. Järvinen, 1999: Variational Quality Control. *Quart. J. Roy. Meteor. Soc.*, **125**, 697–722.
- Aoki, Te., Ta. Aoki, M. Fukabori, and T. Takao, 2002: Characteristics of UV-B Irradiance at Syowa Station, Antarctica: Analyses of the Measurements and Comparison with Numerical Simulations. *J. Meteor. Soc. Japan*, **80**, 161–170.
- Arakawa, A., 1972: Design of the UCLA general circulation model. Numerical simulation weather and climate, Tech. Rep. 7, Dept. of Meteorology, University of California, Los Angeles. 116pp.
- Arakawa, A. and V. R. Lamb, 1977: Computational design of the basic dynamical processes of the UCLA general circulation model. *Methods in Computational Physics*, **17**, 174–265, Academic Press.
- Arakawa, A. and W. H. Schubert, 1974: Interaction of a cumulus cloud ensemble with the large-scale environment, Part I. *J. Atmos. Sci.*, **31**, 674–701.
- Aranami, K. and J. Ishida, 2004: Implementation of two dimensional decomposition for JMA non-hydrostatic model. *CAS/JSC WGNE Res. Activ. Atmos. Oceanic Modell.*, **34**, 03.01–03.02.
- Aranami, K., T. Hara, Y. Ikuta, K. Kawano, K. Matsubayashi, H. Kusabiraki, T. Ito, T. Egawa, K. Yamashita, Y. Ota, Y. Ishikawa, T. Fujita, and J. Ishida, 2015: A new operational regional model for convection-permitting numerical weather prediction at JMA. *CAS/JSC WGNE Res. Activ. Atmos. Oceanic Modell.*, **45**, 05.05–05.06.
- Barnes, R. T. H., R. Hide, A. A. White, and C. A. Wilson, 1983: Atmospheric angular momentum fluctuations, length-of-day change and polar motion. *Proc. R. Soc.*, **A387**, 31–73.
- Bartholomé, E. and A. S. Belward, 2005: GLC2000: a new approach to global land cover mapping from Earth observation data. *International Journal of Remote Sensing*, **26**, 1959–1977.
- Beljaars, A. C. M., 1995: The parameterization of surface fluxes in large-scale models under free convection. *Quart. J. Roy. Meteor. Soc.*, **121**, 255–270.
- Beljaars, A. C. M. and A. A. M. Holtlag, 1991: Flux parameterization over land surfaces for atmospheric models. *J. Appl. Meteor.*, **30**, 327–341.
- Berger, M. J. and J. Olinger, 1984: Adaptive mesh refinement for hyperbolic partial differential equations. *J. Comp. Phys.*, **53**, 484–512.
- Berner, J., G. J. Shutts, M. Leutbecher, and T. N. Palmer, 2009: A Spectral Stochastic Kinetic Energy Backscatter Scheme and Its Impact on Flow-Dependent Predictability in the ECMWF Ensemble Prediction System. *J. Atmos. Sci.*, **66**, 603–626.
- Best, M. J., A. C. M. Beljaars, J. Polcher, and P. Viterbo, 2004: A proposed structure for coupling tiled surfaces with the planetary boundary layer. *J. Hydr. Meteorol.*, **5**, 1271–1278.
- Bishop, C. H., B. J. Etherton, and S. J. Majumdar, 2001: Adaptive Sampling with the Ensemble Transform Kalman Filter. Part I: Theoretical Aspects. *Mon. Wea. Rev.*, **129**, 420–436.
- Blackadar, A. K., 1962: The vertical distribution of wind and turbulent exchange in a neutral atmosphere. *J. Geophys. Res.*, **67**, 3095–3102.
- Bloom, S. C., L. L. Takacs, A. M. da Silva, and D. Ledvina, 1996: Data assimilation using incremental analysis updates. *Mon. Wea. Rev.*, **124**, 1256–1271.
- Bourke, W., 1974: A multi-level spectral model. I. Formulation and hemispheric integrations. *Mon. Wea. Rev.*, **102**, 687–701.
- Brasnett, B., 1999: A global analysis of snow depth for numerical weather prediction. *J. Appl. Meteor.*, **38**, 726–740.

- Briegleb, B. P., 1992: Delta-Eddington Approximation for Solar Radiation in the NCAR Community Climate Model. *J. Geophys. Res.*, **97**, 7603–7612.
- Briegleb, B. P., P. Minnis, V. Ramanathan, and E. Harrison, 1986: Comparison of Regional Clear-Sky Albedos Inferred from Satellite Observations and Model Computations. *Journal of Climate and Applied Meteorology*, **25**, 214–226.
- Brown, P. R. and P. N. Francis, 1995: Improved Measurements of the Ice Water Content in Cirrus Using a Total-Water Probe. *J. Atmos. and Oceanic Technol.*, **12**, 410–414.
- Broxton, P. D., X. Zeng, W. Scheftic, and P. A. Troch, 2014: A MODIS-Based 1 km Maximum Green Vegetation Fraction Dataset. *J. Appl. Meteor. Climat.*, **53**, 1996–2004.
- Buizza, R., 1998: Impact of Horizontal Diffusion on T21, T42, and T63 Singular Vectors. *J. Atmos. Sci.*, **55**, 1069–1083.
- Buizza, R., M. Miller, and T. N. Palmer, 1999: Stochastic representation of model uncertainties in the ECMWF Ensemble Prediction System. *Quart. J. Roy. Meteor. Soc.*, **125**, 2887–2908.
- Buizza, R. and T. N. Palmer, 1995: The Singular-Vector Structure of the Atmospheric Global Circulation. *J. Atmos. Sci.*, **52**, 1434–1456.
- Carter, W.P.L., 2000: Documentation of the SAPRC-99 Chemical Mechanism for VOC reactivity assesment. Final report to california air resources board.
- Cats, G., Van T. Vu, and L. Wolters, 2008: Overlapping communications with calculations. *Hirlam newsletter*, **54**, 189–201.
- Caumont, O., V. Ducrocq, É. Wattrelot, G. Jaubert, and S. Pradier-vabre, 2010: 1D+3DVar assimilation of radar reflectivity data: a proof of concept. *Tellus*, **62A**, 173–187.
- Cavaleri, L. and P. M. Rizzoli, 1981: Wind wave prediction in shallow water: Theory and applications. *J. Geophys. Res.*, **86**, 10961–10973.
- Charnock, H., 1955: Wind stress on a water surface. *Quart. J. Roy. Meteor. Soc.*, **81**, 639–640.
- Chikamoto, Y., H. Mukougawa, T. Kubota, H. Sato, A. Ito, and S. Maeda, 2007: Evidence of growing bred vector associated with the tropical intraseasonal oscillation. *Geophys. Res. Lett.*, **34**, L04806.
- Chou, M.-D., M. J. Suarez, X.-Z. Liang, and M. M.-H. Yan, 2001: A thermal infrared radiation parameterization for atmospheric studies. *Technical report series on global modeling and data assimilation, Vol. 19, NASA Goddard Space Flight Center*, 56pp.
- Coakley, J. A., Jr., R. D. Cess, and F. B. Yurevich, 1983: The effect of tropospheric aerosols on the Earth's radiation budget: A parameterization for climate models. *J. Atmos. Sci.*, **40**, 116–138.
- Collins, W. D., 2001: Parameterization of generalized cloud overlap for radiative calculations in general circulation models. *J. Atmos. Sci.*, **58**, 3224–3242.
- Collins, W. D., J. M. Lee-Taylor, D. P. Edwards, and G. L. Francis, 2006: Effects of increased near-infrared absorption by water vapor on the climate system. *J. Geophys. Res.*, **111**, D18 109.
- Courtier, P., J.-N. Thépaut, and A. Hollingsworth, 1994: A strategy for operational implementation of 4D-Var, using an incremental approach. *Quart. J. Roy. Meteor. Soc.*, **120**, 1367–1387.
- Culverwell, I. D., H. W. Lewis, D. Offiler, C. Marquardt, and C. P. Burrows, 2015: The Radio Occultation Processing Package, ROPP. *Atmos. Meas. Tech.*, **8**, 1887–1899.
- Deardorff, J. W., 1978: Efficient prediction of ground surface temperature and moisture, with inclusion of a layer of vegetation. *J. Geophys. Res.*, **83**, 1889–1903.
- Deardorff, J. W., 1980: Stratocumulus-capped mixed layers derived from a three-dimensional model. *Bound.-Layer Meteor.*, **18**, 495–527.
- Dee, D. P., 2004: Variational bias correction of radiance data in the ECMWF system. *Proceedings of the ECMWF workshop on assimilation of high spectral resolution sounders in NWP*, Reading, UK, 28 June - 1 July 2004, 97–112.
- DeFries, R. S., M. C. Hansen, J. R. G. Townshend, A. C. Janetos, and T. R. Loveland, 2000: A new global 1km data set of percent tree cover derived from remote sensing. *Global Chg. Biol.*, **6**, 247–254.
- Derber, J. and F. Bouttier, 1999: A reformulation of the background error covariance in the ECMWF global data assimilation system. *Tellus*, **51A**, 195–221.
- Derber, J. C. and W.-S Wu, 1998: The use of TOVS cloud-cleared radiances in the NCEP SSI analysis system.

- Mon. Wea. Rev.*, **126**, 2287–2299.
- Desroziers, G., L. Berre, B. Chapnik, and P. Poli, 2005: Diagnosis of observation, background and analysis-error statistics in observation space. *Quart. J. Roy. Meteor. Soc.*, **131**, 3385–3396.
- Deushi, M. and K. Shibata, 2011: Development of a Meteorological Research Institute Chemistry-Climate Model version 2 for the Study of Tropospheric and Stratospheric Chemistry. *Pap. Meteorol. Geophys.*, **62**, 1–46.
- Dobbie, J. S., J. Li, and P. Chýlek, 1999: Two- and four-stream optical properties for water clouds and solar wavelengths. *J. Geophys. Res.*, **104**, 2067–2079.
- Ebert, E. E. and J. A. Curry, 1992: A parameterization of ice cloud optical properties for climate models. *J. Geophys. Res.*, **97**, 3831–3836.
- ECMWF, 2014: Part IV: Physical Process, Chapter 7 Clouds and large-scale precipitation. *IFS Documentation—Cy40r1*, 91–108.
- Egbert, G. D. and S. Y. Erofeeva, 2002: Efficient Inverse Modeling of Barotropic Ocean Tides. *J. Atmos. Oceanic Technol.*, **19**, 183–204.
- Ehrendorfer, M., R. M. Errico, and K. D. Raeder, 1999: Singular-vector perturbation growth in a primitive equation model with moist physics. *J. Atmos. Sci.*, **56**, 1627–1648.
- Elliott, A. J., 1986: Shear diffusion and the spread of oil in the surface layers of the North Sea. *Dt. Hydrogr. Zeit.*, **39**, 113–137.
- Eyre, J. R. and W. P. Menzel, 1989: Retrieval of Cloud Parameters from Satellite Sounder Data: A Simulation Study. *J. Appl. Meteor.*, **28**, 267–275.
- Fairall, C. W., E. F. Bradley, J. E. Hare, A. A. Grachev, and J. B. Edson, 2003: Bulk parameterization of air–sea fluxes: Updates and verification for the COARE algorithm. *J. Clim.*, **16**, 571–591.
- FAO, IIASA, ISRIC, ISSCAS, and JRC, 2012: Harmonized World Soil Database (version 1.2). 42pp., URL http://webarchive.iiasa.ac.at/Research/LUC/External-World-soil-database/HWSD_Documentation.pdf.
- Field, P. R., A. J. Heymsfield, and A. Bansemmer, 2007: Snow Size Distribution Parameterization for Midlatitude and Tropical Ice Clouds. *J. Atmos. Sci.*, **64**, 4346–4365.
- Fingas, M. F., 2010: OIL SPILL SCIENCE and TECHNOLOGY. *Gulf Professional Publishing*, 201–242.
- Folland, C. K. and D. E. Parker, 1995: Correction of instrumental biases in historical sea surface temperature data. *Quart. J. Roy. Meteor. Soc.*, **121**, 319–367.
- Foster, D. J. and R. D. Davy, 1988: Global Snow Depth Climatology. Tech. Rep. USAF-ETAC/TN-88/006, Scott Air Force Base, Illinois. 48pp.
- Frank, W. M., 1977: The structure and energetics of the tropical cyclone I. Storm structure. *Mon. Wea. Rev.*, **105**, 1119–1135.
- Frank, W. M., 1984: A composite analysis of the core of a mature hurricane. *Mon. Wea. Rev.*, **112**, 2401–2420.
- Freidenreich, S. M. and V. Ramaswamy, 1999: A new multiple-band solar radiative parameterization for general circulation models. *J. Geophys. Res.*, **104**, 31 389–31 409.
- Fu, Q. and K. N. Liou, 1992: On the correlated k -distribution method for radiative transfer in nonhomogeneous atmospheres. *J. Atmos. Sci.*, **49**, 2139–2156.
- Fujii, Y. and M. Kamachi, 2003: A reconstruction of observed profiles in the sea east of Japan using vertical coupled temperature-salinity EOF modes. *J. Oceanogr.*, **59**, 173–186.
- Fujii, Y., M. Kamachi, S. Matsumoto, and S. Ishizaki, 2012: Barrier layer and relevant variability of the salinity field in the equatorial Pacific estimated in an ocean reanalysis experiment. *Pure Appl. Geophys.*, **169**, 579–594.
- Fujita, T., 1952: Pressure Distribution within Typhoon. *Geophys. Mag.*, **23**, 437–451.
- Gauthier, P. and J.-N. Thépaut, 2001: Impact of digital filter as a weak constraint in the preoperational 4DVAR assimilation system of Météo-France. *Mon. Wea. Rev.*, **129**, 2089–2102.
- Geleyn, J.-F. and A. Hollingsworth, 1979: An economical analytical method for the computation of the interaction between scattering and line absorption of radiation. *Beitr. Phys. Atmos.*, **52**, 1–16.
- Gent, P. R. and J. C. McWilliams, 1990: Isopycnal mixing in ocean circulation models. *J. Phys. Oceanogr.*, **20**,

- 150–155.
- Gifford, F. A., 1982: Horizontal diffusion in the atmosphere: a Lagrangian-dynamical theory. *Atmos. Env.*, **16**, 505–512.
- Giglio, L., J. T. Randerson, van der G. R. Werf, P. S. Kasibhatla, G. J. Collatz, D. C. Morton, and R. S. DeFries, 2010: Assessing variability and long-term trends in burned area by merging multiple satellite fire products. *Biogeosciences*, **7**, 1171–1186.
- Godske, C. L. et al., 1957: *Dynamic Meteorology and Weather Forecasting*, chap. 18. Amer. Met. Soc.
- Griffies, S. M. and R. W. Hallberg, 2000: Biharmonic friction with a Smagorinsky-like viscosity for use in large-scale eddy-permitting ocean models. *Mon. Wea. Rev.*, **128**, 2935–2946.
- Guenther, A., T. Karl, P. Harley, C. Wiedinmyer, P. I. Palmer, and C. Geron, 2006: Estimates of global terrestrial isoprene emissions using MEGAN (Model of Emissions of Gases and Aerosols from Nature). *Atmos. Chem. Phys.*, **6**, 3181–3210.
- Haimberger, L., C. Tavalato, and S. Sperka, 2008: Toward elimination of the warm bias in historic radiosonde temperature records—Some new results from a comprehensive intercomparison of upper-air data. *J. Climate*, **21**, 4587–4606.
- Haimberger, L., C. Tavalato, and S. Sperka, 2012: Homogenization of the global radiosonde temperature dataset through combined comparison with reanalysis background series and neighboring stations. *J. Climate*, **25**, 8108–8131.
- Hamrud, M., M. Bonavita, and L. Isaksen, 2015: EnKF and hybrid gain ensemble data assimilation part I: EnKF implementation. *Mon. Wea. Rev.*, **143**, 4847–4864.
- Han, J. and H.-L. Pan, 2011: Revision of Convection and Vertical Diffusion Schemes in the NCEP Global Forecast System. *Weather and Forecasting*, **26**, 520–533.
- Hara, T., 2015: Necessity of parameterizations for convective initiation in high resolution cloud-permitting models. *CAS/JSC WGNE Res. Activ. Atmos. Oceanic Modell.*, **45**, 04.06–04.07.
- Hara, T., K. Kawano, K. Aranami, Y. Kitamura, M. Sakamoto, H. Kusabiraki, C. Muroi, and J. Ishida, 2012: Development of Physics Library and its application to ASUCA. *CAS/JSC WGNE Res. Activ. Atmos. Oceanic Modell.*, **42**, 05.05–05.06.
- Hasegawa, H., N. Kohno, M. Higaki, and M. Itoh, 2017: Upgrade of JMA’s Storm Surge Prediction for the WMO Storm Surge Watch Scheme (SSWS). *RSMC Tokyo-Typhoon Center Technical Review*, **19**, 26–34.
- Hasegawa, Y., A. Sugai, Yo. Hayashi, Yu. Hayashi, S. Saito, and T. Shimbori, 2015: Improvements fo volcanic ash fall forecasts issued by the Japan Meteorological Agency. *J. Appl. Volcanol.*, **4**, 2.
- Hasselmann, K., T. P. Barnett, E. Bouws, H. Carlson, D. E. Cartwright, K. Enke, J. A. Ewing, H. Gienapp, D. E. Hasselmann, P. Kruseman, A. Meerburg, P. Muller, D. J. Olbers, K. Richter, W. Sell, and H. Walden, 1973: Measurements of wind-wave growth and swell decay during the Joint North Sea Wave Project (JONSWAP). *Dtsch. Hydrogr. Z.*
- Hasselmann, S., K. Hasselmann, J. H. Allender, and T. P. Barnett, 1985: Computations and parameterizations of the nonlinear energy transfer in a gravity-wave spectrum. Part II: Parameterizations of the nonlinear energy transfer for application in wave models. *J. Phys. Oceanogr.*, **15**, 1378–1391.
- Hasumi, H., 2006: CCSR Ocean Component Model (COCO) Version 4.0. CCSR Report 25, Center for Climate System Research, The University of Tokyo. 103pp.
- Hibler, W. D., 1979: A dynamic thermodynamic sea ice model. *J. Phys. Oceanogr.*, **9**, 815–846.
- Higaki, M., H. Hayashibara, and F. Nozaki, 2009: Outline of the storm surge prediction model at the Japan Meteorological Agency. *RSMC Tokyo-Typhoon Center Technical Review*, **11**, 25–38.
- Hirai, M., K. Miyaoka, H. Sato, H. Sugimoto, A. Minami, and C. Matsukawa, 2014: March 2014 upgrade of JMA’s One-month Ensemble Prediction System. *CAS/JSC WGNE Res. Activ. Atmos. Oceanic Modell.*, **44**, 6.09–6.10.
- Holmlund, K., 1998: The utilization of statistical properties of satellite-derived atmospheric motion vectors to derive quality indicators. *Wea. Forecasting*, **13**, 1093–1104.
- Holtslag, A. A. M. and B. A. Boville, 1993: Local versus nonlocal boundary-layer diffusion in a global climate model. *J. Climate*, **6**, 1825–1842.
- Honda, Y., M. Nishijima, K. Koizumi, Y. Ohta, K. Tamiya, T. Kawabata, and T. Tsuyuki, 2005: A pre-

- operational variational data assimilation system for a non-hydrostatic model at the Japan Meteorological Agency: Formulation and preliminary results. *Quart. J. Roy. Meteor. Soc.*, **131**, 3465–3475.
- Hortal, M., 2002: The development and testing of a new two-time-level semi-Lagrangian scheme (SETTLS) in the ECMWF forecast model. *Quart. J. Roy. Meteor. Soc.*, **128**, 1671–1687.
- Hoskins, B. J. and A. J. Simmons, 1975: A multi-layer spectral model and the semi-implicit method. *Quart. J. Roy. Meteor. Soc.*, **101**, 637–655.
- Hunke, E. C. and J. K. Dukowicz, 1997: An elastic-viscous-plastic model for sea ice dynamics. *J. Phys. Oceanogr.*, **27**, 1849–1867.
- Hunke, E. C. and W. H. Lipscomb, 2006: CICE. *The Los Alamos Sea Ice Model Documentation and Software User's Manual*, URL <https://github.com/CICE-Consortium/CICE/wiki>.
- Hunt, B. R., E. J. Kostelich, and I. Szunyogh, 2007: Efficient data assimilation for spatiotemporal chaos: A local ensemble transform Kalman filter. *Physica D*, **230**, 112–126.
- ICAO, 2018: *Annex 3 - Meteorological Service for International Air Navigation*. International Civil Aviation Organization.
- Ikawa, M. and K. Saito, 1991: Description of a nonhydrostatic model developed at the Forecast Research Department of the MRI. *Tech. Rep. MRI*, **28**, 238pp.
- Ikuta, Y. and Y. Honda, 2011: Development of 1D+4DVAR data assimilation of radar reflectivity in JNoVA. *CAS/JSC WGNE Res. Activ. Atmos. Oceanic Modell.*, **41**, 01.09–01.10.
- Ishida, J., 2007: Development of a hybrid terrain-following vertical coordinate for JMA Non-hydrostatic Model. *CAS/JSC WGNE Res. Activ. Atmos. Oceanic Modell.*, **37**, 03.09–03.10.
- Ishida, J., C. Muroi, and Y. Aikawa, 2009: Development of a new dynamical core for the nonhydrostatic model. *CAS/JSC WGNE Res. Activ. Atmos. Oceanic Modell.*, **39**, 05.09–05.10.
- Ishida, J., C. Muroi, K. Kawano, and Y. Kitamura, 2010: Development of a new nonhydrostatic model “ASUCA” at JMA. *CAS/JSC WGNE Res. Activ. Atmos. Oceanic Modell.*, **40**, 05.11–05.12.
- Ishihara, M., Y. Kato, T. Abo, K. Kobayashi, and Y. Izumikawa, 2006: Characteristics and Performance of the Operational Wind Profiler Network of the Japan Meteorological Agency. *J. Meteor. Soc. Japan*, **84**, 1085–1096.
- Ishii, M., A. Shouji, S. Sugimoto, and T. Matsumoto, 2005: Objective Analyses of Sea-Surface Temperature and Marine Meteorological Variables for the 20th Century using ICOADS and the Kobe Collection. *Int. J. Climatol.*, **25**, 865–879.
- Ishikawa, Y. and K. Koizumi, 2002: Meso-scale analysis. *Outline of the operational numerical weather prediction at the Japan Meteorological Agency. Appendix to WMO Technical Progress Report on the Global Data-processing and Forecasting System*, Tokyo, Japan, Japan Meteorological Agency, 26–31.
- Iwamura, K. and H. Kitagawa, 2008: An upgrade of the JMA Operational Global NWP Model. *CAS/JSC WGNE Res. Activ. Atmos. Oceanic Modell.*, **38**, 06.3–06.4.
- Iwasaki, T., T. Maki, and K. Katayama, 1998: Tracer transport model at Japan Meteorological Agency and its application to the ETEX data. *Atmos. Env.*, **32**, 4285–4295.
- Iwasaki, T., S. Yamada, and K. Tada, 1989: A parameterization scheme of orographic gravity wave drag with two different vertical partitionings, Part I: Impacts on medium-range forecasts. *J. Meteor. Soc. Japan*, **67**, 11–27.
- Jakob, C. and A. P. Siebesma, 2003: A new subcloud model for mass-flux convection schemes: Influence on triggering, updraft properties, and model climate. *Mon. Wea. Rev.*, **131**, 2765–2778.
- Janssen, P., 2004: *The Interaction of Ocean Waves and Wind*. Cambridge Univ. Press, 308pp.
- Japan Maritime Safety Agency, (Ed.) , 1983: *Nihonkinkai Kairyu Tokeizu (Statistical charts of ocean currents around Japan)*. Japan Maritime Safety Agency, (in Japanese).
- Japan Meteorological Agency, 2006: Characteristics of Global Sea Surface Temperature Analysis Data (COBE-SST) for Climate Use. *Monthly Report on Climate System Separated Volume*, **12**, 116pp.
- JMA, 1997: Objective Interpretation of NWP Products. *Outline of the operational numerical weather prediction at the Japan Meteorological Agency. Appendix to WMO Technical Progress Report on the Global Data-processing and Forecasting System*, Japan Meteorological Agency, Tokyo, Japan, 123–124.
- JMA, 2002: *Outline of the operational numerical weather prediction at the Japan Meteorological Agency*.

- Appendix to WMO Technical Progress Report on the Global Data-processing and Forecasting System.* Japan Meteorological Agency, Tokyo, Japan.
- JMA, 2007: *Outline of the operational numerical weather prediction at the Japan Meteorological Agency. Appendix to WMO Technical Progress Report on the Global Data-processing and Forecasting System and Numerical Weather Prediction Research.* Japan Meteorological Agency, Tokyo, Japan.
- JMA, 2013: *Outline of the operational numerical weather prediction at the Japan Meteorological Agency, Appendix to WMO Technical Progress Report on the Global Data-processing and Forecasting System (GDPFS) and Numerical Weather Prediction (NWP) Research.* Japan Meteorological Agency, Tokyo, Japan.
- JMA, 2017: Joint WMO Technical Progress Report on the Global Data Processing and Forecasting System and Numerical Weather Prediction Research Activities for 2016. 31–32.
- Jolliffe, I. T. and D. B. Stephenson, 2003: *Forecast Verification: A Practitioner's Guide in Atmospheric Science.* Wiley, 254pp.
- Joseph, J. H., W. J. Wiscombe, and J. A. Weinman, 1976: The delta-Eddington approximation for radiative flux transfer. *J. Atmos. Sci.*, **33**, 2452–2459.
- Juang, H.-M. H., 2004: A Reduced Spectral Transform for the NCEP Seasonal Forecast Global Spectral Atmospheric Model. *Mon. Wea. Rev.*, **132**, 1019–1035.
- Kadowaki, T., 2005: A 4-dimensional variational assimilation system for the JMA Global Spectrum Model. *CAS/JSC WGNE Res. Activ. Atmos. Oceanic Modell.*, **35**, 01.17–01.18.
- Kadowaki, T. and K. Yoshimoto, 2012: A new inner model with a higher horizontal resolution (TL319) in JMA's Global 4D-Var data assimilation system. *CAS/JSC WGNE Res. Activ. Atmos. Oceanic Modell.*, 01.09–01.10.
- Kain, J. S., 2004: The Kain-Fritsch convective parameterization: An update. *J. Appl. Meteor.*, **43**, 170–181.
- Kain, J. S. and J. M. Fritsch, 1990: A one-dimensional entraining/detraining plume model and its application in convective parameterization. *J. Atmos. Sci.*, **47**, 2784–2802.
- Kajino, M., Y. Inomata, K. Sato, H. Ueda, Z. Han, J. An, G. Katata, M. Deushi, T. Maki, N. Oshima, J. Kurokawa, T. Ohara, A. Takami, and S. Hatakeyama, 2012: Development of the RAQM2 aerosol chemical transport model and predictions of the Northeast Asian aerosol mass, size, chemistry and the mixing type. *Atmos. Chem. Phys.*, **12**, 11 833–11 856.
- Kawai, H., 2002: Forecast of sulfur dioxide flow from Miyake volcano with a high resolution regional transport model. *CAS/JSC WGNE Res. Activ. Atmos. Oceanic Modell.*, **32**, 05.24–05.25.
- Kawai, H., 2005: Improvement of a cloud ice fall scheme in GCM. *CAS/JSC WGNE Res. Activ. Atmos. Oceanic Modell.*, **35**, 04.11–04.12.
- Kawai, H., 2013: Improvement of a stratocumulus scheme for mid-latitude marine low clouds. *CAS/JSC WGNE Res. Act. Atmos. Ocean Model/WMO*, **42**, 4.03–04.
- Kawai, H. and T. Inoue, 2006: A Simple Parameterization Scheme for Subtropical Marine Stratocumulus. *SOLA*, **2**, 17–20.
- Kessler, E., 1969: *On the distribution and continuity of water substance in atmospheric circulation.* Meteorol. Monogr., American Meteorol. Soc., Boston, MA, 84pp.
- Kim, H., 2017: Global Soil Wetness Project Phase 3 Atmospheric Boundary Conditions (Experiment 1) [Data set]. *Data Integration and Analysis System (DIAS)*, URL <https://doi.org/10.20783/DIAS.501>.
- Kitada, T., 1994: Modelling of transport, reaction and deposition of acid rain. *Kishou Kenkyu Note*, **182**, 95–117, (in Japanese).
- Kitada, T., G. R. Carmichael, and L. K. Peters, 1986: Effects of dry deposition on the concentration-distributions of atmospheric pollutants within land- and sea-breeze circulations. *Atmos. Env.*, **20**, 1999–2010.
- Kitamura, Y. and J. Ito, 2016: Revisiting the bulk relation for heat flux in the free convection limit. *Bound.-Layer Meteor.*, **158**, 93–103.
- Klemp, J. B., W. C. Skamarock, and J. Dudhia, 2007: Conservative split-explicit time integration methods for the compressible nonhydrostatic equations. *Mon. Wea. Rev.*, **135**, 2897–2913.
- Kobayashi, S., Y. Ota, Y. Harada, A. Ebata, M. Moriya, H. Onoda, K. Onogi, H. Kamahori, C. Kobayashi, H. Endo, K. Miyaoka, and K. Takahashi, 2015: The JRA-55 reanalysis: General specifications and basic characteristics. *J. Meteor. Soc. Japan*, **93**, 5–48.

- Kohno, N., D. Miura, and K. Yoshita, 2012: The development of JMA wave data assimilation system. *Proceedings of the 12th International Workshop on Wave Hindcasting and Forecasting and Coastal Hazards Symposium*, Kohala Coast, Hawaii, 30 October - 4 November 2011, H2, (<http://www.waveworkshop.org>).
- Kohno, N., K. Murotani, H. Minematsu, and D. Miura, 2009: The improvement of JMA operational wave forecasting system. *Proceedings of the 11th International Workshop on Wave Hindcasting and Forecasting and Coastal Hazards Symposium*, Halifax, Canada, 18 - 23 October 2009, R3, (<http://www.waveworkshop.org>).
- Koizumi, K., Y. Ishikawa, and T. Tsuyuki, 2005: Assimilation of precipitation data to the JMA mesoscale model with a four-dimensional variational method and its impact on precipitation forecasts. *SOLA*, **1**, 45–48.
- Kondo, J., 1975: Air-sea bulk transfer coefficients in diabatic conditions. *Bound.-Layer Meteor.*, **9**, 91–112.
- Konishi, T., 1995: An experimental storm surge prediction for the western part of the Inland Sea with application to Typhoon 9119. *Pap. Meteor. Geophys.*, **46**, 9–17.
- Koren, B., 1993: A robust upwind discretization method for advection, diffusion and source terms. *CWI Technical Report NM-R 9308*, 1–22, URL <http://oai.cwi.nl/oai/asset/5293/05293D.pdf>.
- Kudo, A., 2011: Development of JMA's new turbulence index. *15th Conference on Aviation, Range, and Aerospace Meteorology*, Amer. Met. Soc., Los Angeles, CA, 1 - 4 August 2011.
- Kuragano, T., Y. Fujii, T. Toyoda, N. Usui, K. Ogawa, and M. Kamachi, 2014: Seasonal barotropic sea surface height fluctuation in relation to regional ocean mass variation. *J. Oceanogr.*, **70**, 45–62.
- Kuragano, T. and M. Kamachi, 2000: Global statistical space-time scales of oceanic variability estimated from the TOPEX/POSEIDON altimeter data. *J. Geophys. Res.*, **105**, 955–974.
- Kurihara, Y., T. Sakurai, and T. Kuragano, 2006: Global daily sea surface temperature analysis using data from satellite microwave radiometer, satellite infrared radiometer and *in-situ* observations. *Weather Bulletin, JMA*, **73**, s1–s18.
- Kyouda, M. and M. Higaki, 2015: Upgrade of JMAs Typhoon Ensemble Prediction System. **17**, 1–13.
- Lalaurette, F., 2003: Early detection of abnormal weather conditions using a probabilistic extreme forecast index. *Q. J. R. Meteorol. Soc.*, **129**, 3037–3057.
- Li, J., 2002: Accounting for unresolved clouds in a 1D infrared radiative transfer model. Part I: Solution for radiative transfer, including cloud scattering and overlap. *J. Atmos. Sci.*, **59**, 3302–3320.
- Li, J. and H. W. Barker, 2005: A radiation algorithm with correlated-*k* distribution. Part I: Local thermal equilibrium. *J. Atmos. Sci.*, **62**, 286–309.
- Lin, Y.-L., R. D. Farley, and H. D. Orville, 1983: Bulk Parameterization of the Snow Field in a Cloud Model. *J. Climate Appl. Meteor.*, **22**, 1065–1092.
- Lindner, T. H. and J. Li, 2000: Parameterization of the optical properties for water clouds in the infrared. *J. Climate*, **13**, 1797–1805.
- Liu, D. C. and J. Nocedal, 1989: On the limited memory BFGS method for large scale optimization. *Math. Programming*, **45**, 503–528.
- Lord, S. J. and A. Arakawa, 1980: Interaction of a cumulus cloud ensemble with the large-scale environment. Part II. *J. Atmos. Sci.*, **37**, 2677–2692.
- Lorenc, A. C. and F. Rawlins, 2005: Why does 4D-Var beat 3D-Var? *Q. J. R. Meteorol. Soc.*, **131**, 3247–3257.
- Louis, J.-F., M. Tiedtke, and J.-F. Geleyn, 1982: A short history of the operational PBL-parameterization at ECMWF. *Workshop on Planetary Boundary Layer Parameterization*, Reading, United Kingdom, 25 - 27 November 1981, ECMWF, 59–79.
- Loveland, T. R., B. C. Reed, J. F. Brown, D. O. Ohlen, Z. Zhu, L. Youing, and J. W. Merchant, 2000: Development of a global land cover characteristics database and IGBP DISCover from 1km AVHRR data. *Int. J. Remote Sensing*, 1303–1330.
- Luo, J.-J., S. Masson, E. Roeckner, G. Madec, and T. Yamagata, 2005: Reducing climatology bias in an ocean-atmosphere CGCM with improved coupling physics. *J. Clim.*, **18**, 2344–2360.
- Lyard, F., F. Lefèvre, T. Letellier, and O. Francis, 2006: Modelling the global ocean tides: Modern insights from FES2004. *Ocean Dyn.*, **56**, 394–415.
- Lynch, P., 1997: The Dolph-Chebyshev Window: A Simple Optimal Filter. *Mon. Wea. Rev.*, **125**, 655–660.

- Machenhauer, B., 1977: On the dynamics of gravity oscillations in a shallow water model, with application to normal mode initialization. *Contrib. Atmos. Phys.*, **50**, 253–271.
- Mahfouf, J.-F., 1999: Influence of physical process on the tangent-linear approximation. *Tellus*, **51A**, 147–166.
- Makihara, Y., 2000: Algorithms for precipitation nowcasting focused on detailed analysis using radar and raingauge data. *Technical Reports of the Meteorological Research Institute*, **39**, 63–111.
- Martin, G. M., D. W. Johnson, and A. Spice, 1994: The measurement and parameterization of effective radius of droplets in warm stratocumulus clouds. *J. Atmos. Sci.*, **51**, 1823–1842.
- Matsumoto, K., T. Takanezawa, and M. Ooe, 2000: Ocean Tide Models Developed by Assimilating TOPEX/POSEIDON Altimeter Data into Hydrodynamical Model: A Global Model and a Regional Model around Japan. *J. Oceanogr.*, **56**, 567–581.
- McFarquhar, G. M., S. Iacobellis, and R. C. J. Somerville, 2003: SCM Simulations of Tropical Ice Clouds Using Observationally Based Parameterizations of Microphysics. *J. Climate*, **16**, 1643–1664.
- McLandress, C. and J. F. Scinocca, 2005: The GCM response to current parameterizations of nonorographic gravity wave drag. *J. Atmos. Sci.*, **62**, 2394–2413.
- Meador, W. E. and W. R. Weaver, 1980: Two-stream approximations to radiative transfer in planetary atmospheres: A unified description of existing methods and a new improvement. *J. Atmos. Sci.*, **37**, 630–643.
- Mellor, G. L., 1977: The Gaussian cloud model relations. *J. Atmos. Sci.*, **34**, 356–358.
- Mellor, G. L. and L. Kantha, 1989: An ice-ocean coupled model. *J. Geophys. Res.*, **94**, 10 937–10 954.
- Mellor, G. L. and T. Yamada, 1974: A hierarchy of turbulence closure models for planetary boundary layers. *J. Atmos. Sci.*, **31**, 1791–1806.
- Mellor, G. L. and T. Yamada, 1982: Development of a turbulence closure model for geophysical fluid problems. *Rev. Geophys. Space Phys.*, **20**, 851–875.
- Mitsuyasu, H. and T. Honda, 1982: Wind-induced growth of water waves. *J. Fluid Mech.*, **123**, 425–442.
- Miyamoto, K., 2006: Introduction of the Reduced Gaussian Grid into the Operational Global NWP Model at JMA. *CAS/JSC WGNE Res. Activ. Atmos. Oceanic Modell.*, **36**, 06.09–06.10.
- Mizuta, R., H. Yoshimura, H. Murakami, M. Matsueda, H. Endo, T. Ose, K. Kamiguchi, M. Hosaka, M. Sugi, S. Yukimoto, S. Kusunoki, and A. Kitoh, 2012: Climate simulations using MRI-AGCM3.2 with 20-km grid. *J. Meteor. Soc. Japan*, **90A**, 233–258.
- Moorthi, S. and M. J. Suarez, 1992: Relaxed Arakawa-Schubert: A parameterization of moist convection for general circulation models. *Mon. Wea. Rev.*, **120**, 978–1002.
- Morton, B. R., G. Taylor, and J. S. Turner, 1956: Turbulent gravitational convection from maintained and instantaneous sources. *Proc. Roy. Soc. (London) A*, **234**, 1–23.
- Munk, W. H. and D. E. Cartwright, 1966: Tidal Spectroscopy and Prediction. *Phil. Trans. Roy. Soc. A*, **259**, 533–581.
- Murphy, A. H., 1973: A New Vector Partition of the Probability Score. *J. Appl. Meteor.*, **12**, 595–600.
- Myneni, R. B., S. Hoffman, Y. Knyazikhin, J. L. Privette, J. Glassy, Y. Tian, Y. Wang, X. Song, Y. Zhang, G. R. Smith, A. Lotsch, M. Friedl, J. T. Morisette, P. Votava, R. R. Nemani, and S. W. Running, 2002: Global products of vegetation leaf area and fraction absorbed PAR from year one of MODIS data. *Remote Sens. Environ.*, **83**, 214–231.
- Nagasawa, R., 2012: The problem of cloud overlap in the radiation process of JMA global NWP model. *CAS/JSC WGNE Res. Act. Atmos. Ocea. Model/WMO*, **42**, 4.15–4.16.
- Naito, I. and N. Kikuchi, 1992: Atmospheric contributions to non-seasonal variations in the length of day. *Geophys. Res. Lett.*, **19**, 1843–1846.
- Nakagawa, M., 2009: Outline of the High Resolution Global Model at the Japan Meteorological Agency. *RSMC Tokyo-Typhoon Center Technical Review*, **11**, 1–13.
- Nakamura, H., 1978: Dynamical effects of mountains on the general circulation of the atmosphere: I. Development of finite-difference schemes suitable for incorporating mountains. *J. Meteor. Soc. Japan*, **56**, 317–340.
- Nakanishi, M., 2001: Improvement of the Mellor-Yamada turbulence closure model based on large-eddy simulation data. *Bound.-Layer Meteor.*, **99**, 349–378.
- Nakanishi, M. and H. Niino, 2004: An improved Mellor-Yamada level-3 model with condensation physics: Its

- design and verification. *Bound.-Layer Meteor.*, **112**, 1–31.
- Nakanishi, M. and H. Niino, 2006: An improved Mellor-Yamada level-3 model: Its numerical stability and application to a regional prediction of advection fog. *Bound.-Layer Meteor.*, **119**, 397–407.
- Nakanishi, M. and H. Niino, 2009: Development of an Improved Turbulence Closure Model for the Atmospheric Boundary Layer. *J. Meteor. Soc. Japan*, **87**, 895–912.
- Nakano, H. and N. Sugimoto, 2002: Effects of bottom boundary layer parameterization on reproducing deep and bottom waters in a world ocean model. *J. Phys. Oceanogr.*, **32**, 1209–1227.
- Narui, A., 2006: Changing the Resolution of the Inner Loop of Global 4D-Var at JMA. *CAS/JSC WGNE Res. Activ. Atmos. Oceanic Modell.*, **36**, 01.23–01.24.
- Nocedal, J. and S. J. Wright, 2006: *Numerical Optimization*. 2d ed., Springer, 664pp.
- Noh, Y. and H. J. Kim, 1999: Simulations of temperature and turbulence structure of the oceanic boundary layer with the improved near-surface process. *J. Geophys. Res.*, **104**, 15 621–15 634.
- Ohara, T., H. Akimoto, J. Kurokawa, N. Horii, K. Yamaji, X. Yan, and T. Hayasaka, 2007: An Asian emission inventory of anthropogenic emission sources for the period 1980–2020. *Atmos. Chem. Phys.*, **7**, 4419–4444.
- Oleson, K. W., D. M. Lawrence, G. B. Bonan, M. G. Flanner, E. Kluzek, P. J. Lawrence, S. Levis, S. C. Swenson, P. E. Thornton, A. Dai, M. Decker, R. Dickinson, J. Feddema, C. L. Heald, F. Hoffman, J.-F. Lamarque, N. Mahowald, G.-Y. Niu, T. Qian, J. Randerson, S. Running, K. Sakaguchi, A. Slater, R. Stöckli, A. Wang, Z.-L. Yang, Xiaodong Zeng, and Xubin Zeng, 2010: Technical Description of version 4.0 of the Community Land Model (CLM). NCAR Technical Note 478, NCAR. 257pp.
- Onogi, K., 1998: A Data Quality Control Method Using Forecasted Horizontal Gradient and Tendency in a NWP System: Dynamic QC. *J. Meteor. Soc. Japan*, **76**, 497–516.
- Onogi, K., J. Tsutsui, H. Koide, M. Sakamoto, S. Kobayashi, H. Hatsushika, T. Matsumoto, N. Yamazaki, H. Kamahori, K. Takahashi, S. Kadokura, K. Wada, K. Kato, R. Oyama, T. Ose, N. Mannoji, and R. Taira, 2007: The JRA-25 Reanalysis. *J. Meteor. Soc. Japan*, **85**, 369–432.
- Ou, S. and K.-N. Liou, 1995: Ice microphysics and climatic temperature feedback. *Atmos. Res.*, **35**, 127–138.
- Oyama, R., 2010: Upgrade of atmospheric motion vector derivation algorithms at JMA/MSC. Meteorological Satellite Center Technical Note 54, Japan Meteorological Agency, Tokyo, Japan. 1–32. (<https://www.data.jma.go.jp/mscweb/en/product/library/note/index.html>).
- Palmer, T. N., G. J. Shutts, and R. Swinbank, 1986: Alleviation of a systematic westerly bias in general circulation and numerical weather prediction models through an orographic gravity wave drag parameterization. *Quart. J. Roy. Meteor. Soc.*, **112**, 1001–1039.
- Parrish, D. and J. C. Derber, 1992: The National Meteorological Center’s spectral statistical-interpolation analysis system. *Mon. Wea. Rev.*, **120**, 1747–1763.
- Persson, A. O., 1991: Kalman Filtering - A new approach to adaptive statistical interpretation of numerical meteorological forecasts. *WMO Technical Document*, **421**, XX27–XX32.
- Plant, W. J., 1982: A relationship between wind stress and wave slope. *J. Geophys. Res.*, **87**, 1961–1967.
- Prather, M. J., 1986: Numerical advection by conservation of second-order moments. *J. Geophys. Res.*, **91**, 6671–6681.
- Purser, R. J., W. S. Wu, D. F. Parrish, and N. M. Roberts, 2003: Numerical aspects of the application of recursive filters to variational statistical analysis. Part I: Spatially homogeneous and isotropic Gaussian covariances. *Mon. Wea. Rev.*, **131**, 1524–1535.
- Räisänen, P., 1998: Effective longwave cloud fraction and Maximum-Random overlap of clouds: A problem and a solution. *Mon. Wea. Rev.*, **126**, 3336–3340.
- Ramankutty, N., A. T. Evan, C. Monfreda, and J. A. Foley, 2008: Farming the planet: 1. Geographic distribution of global agricultural lands in the year 2000. *Global Biogeochemical Cycles*, **22**, GB1003.
- Randall, D. and D.-M. Pan, 1993: Implementation of the Arakawa-Schubert cumulus parameterization with a prognostic closure. *The representation of cumulus convection in numerical models, AMS Meteorological Monograph Series*, **46**, 137–144.
- Randel, W. J., F. Wu, J. M. Russell III, A. Roche, and J. W. Waters, 1998: Seasonal cycles and QBO variations in stratospheric CH₄ and H₂O observed in UARS HALOE data. *J. Atmos. Sci.*, **55**, 163–185.
- Ray, R. D., 1999: A Global Ocean Tide Model From TOPEX/POSEIDON Altimetry: GOT99.2. Technical

- Memorandum NASA/TM-1999-209478, Goddard Space Flight Center. 58pp.
- Reed, M., 1989: The physical fates component of the natural resource damage assessment model system. *Oil and Chemical Pollution*, **5**, 99–123.
- Reynolds, R. W., 1987: A real-time global sea surface temperature analysis. *J. Climate*, **1**, 75–86.
- Rothman, L. S., A. Barbe, D. C. Benner, L. R. Brown, C. Camy-Peyret, M. R. Carleer, K. Chance, C. Clerbaux, V. Dana, V. M. Devi, A. Fayt, J.-M. Flaud, R. R. Gamache, A. Goldman, D. Jacquemart, K. W. Jucks, W. J. Lafferty, J.-Y. Mandin, S. T. Massie, V. Nemtchinov, D. A. Newnham, A. Perrin, C. P. Rinsland, J. Schroeder, K. M. Smith, M. A. H. Smith, K. Tang, R. A. Toth, J. Vander Auwera, P. Varanasi, and K. Yoshino, 2003: The HITRAN molecular spectroscopic database: edition of 2000 including updates through 2001. *J. Quant. Spectrosc. Radiat. Transfer*, **82**, 5–44.
- Saito, K., J. Ishida, K. Aranami, T. Hara, T. Segawa, M. Narita, and Y. Honda, 2007: Nonhydrostatic Atmospheric Models and Operational Development at JMA. *J. Meteor. Soc. Japan*, **85B**, 271–304.
- Saito, K., T. Fujita, Y. Yamada, J. Ishida, Y. Kumagai, K. Aranami, S. Ohmori, R. Nagasawa, S. Kumagai, C. Muroi, T. Kato, H. Eito, and Y. Yamazaki, 2006: The Operational JMA Nonhydrostatic Mesoscale Model. *Mon. Wea. Rev.*, **134**, 1266–1298.
- Sato, N., P. J. Sellers, D. A. Randall, E. K. Schneider, J. Shukla, J. L. Kinter III, Y-T Hou, and E. Albertazzi, 1989: Effects of implementing the simple biosphere model (SiB) in a general circulation model. *J. Atmos. Sci.*, **46**, 2757–2782.
- Saunders, R., J. Hocking, P. Rayer, M. Matricardi, A. Geer, N. Bormann, P. Brunel, F. Karbou, and F. Aires, 2012: RTTOV-10 science and validation report. Tech. rep., EUMETSAT, 31 pp. (http://research.metoffice.gov.uk/research/interproj/nwpsaf/rtm/docs_rttov10/rttov10_svr_1.11.pdf).
- Schaaf, C. B., F. Gao, A. H. Strahler, W. Lucht, X. Li, T. Tsang, N. C. Strugnell, X. Zhang, Y. Jin, J. P. Muller, P. Lewis, M. Barnsley, P. Hobson, M. Disney, G. Roberts, M. Dunderdale, C. Doll, R. P. d’Entremont, B. Hu, S. Liang, J. L. Privette, and D. P. Roy, 2002: First operational BRDF, albedo nadir reflectance products from MODIS. *Remote Sens. Environ.*, **83**, 135–148.
- Scinocca, J. F., 2003: An accurate spectral nonorographic gravity wave drag parameterization for general circulation models. *J. Atmos. Sci.*, **60**, 667–682.
- Sellers, P. J., Y. Mintz, Y. C. Sud, and A. Dalcher, 1986: A simple biosphere model (SiB) for use within general circulation models. *J. Atmos. Sci.*, **43**, 505–531.
- Selwood, P., 2012: The Met Office Unified Model I/O Server. *ENES Workshop: Scalable IO in climate models*, available at: https://verc.enes.org/computing/hpc-collaborations/parallel-io/workshop-scalable-io-in-climate-models/presentations/Unified_Model_IO_Server_Paul_Selwood.ppt/view.
- Semtner, A. J., 1976: A model for the thermodynamic growth of sea ice in numerical investigation of climate. *J. Phys. Oceanogr.*, **6**, 379–389.
- Shimbori, T., Y. Aikawa, and N. Seino, 2009: Operational implementation of the tephra fall forecast with the JMA mesoscale tracer transport model. *CAS/JSC WGNE Res. Activ. Atmos. Oceanic Modell.*, **39**, 5.29–5.30.
- Shimokobe, A., 2012: Improvement of the Stratocumulus Parameterization Scheme in JMA’s Operational Global Spectral Model. *CAS/JSC WGNE Res. Activ. Atmos. Oceanic Modell.*, 4.17–4.18.
- Simmons, A. J. and D. M. Burridge, 1981: An energy and angular-momentum conserving vertical finite-difference scheme and hybrid vertical coordinates. *Mon. Wea. Rev.*, **109**, 758–766.
- Simonsen, C., 1991: Self adaptive model output statistics based on Kalman filtering. *WMO Technical Document*, **421**, XX33–XX37.
- Smagorinsky, J., 1963: General circulation experiments with the primitive equations: I. The basic experiment. *Mon. Wea. Rev.*, **91**, 99–164.
- Smith, R. N. B., 1990: A scheme for predicting layer clouds and their water content in a general circulation model. *Quart. J. Roy. Meteor. Soc.*, **116**, 435–460.
- Smith, S. D. and E. G. Banke, 1975: Variation of the sea surface drag coefficient with wind speed. *Quart. J. Roy. Meteor. Soc.*, **101**, 665–673.
- Sommeria, G. and J. W. Deardorff, 1977: Subgrid-scale condensation in models of nonprecipitating clouds. *J. Atmos. Sci.*, **34**, 344–355.
- Sugi, M., R. Kawamura, and N. Sato, 1997: A study of SST-forced variability and potential predictability of

- seasonal mean fields using the JMA global model. *J. Meteor. Soc. Japan*, **75**, 717–736.
- Sundqvist, H., 1978: A parameterization scheme for non-convective condensation including prediction of cloud water content. *Quart. J. Roy. Meteor. Soc.*, **104**, 677–690.
- Sundqvist, H., E. Berge, and J. E. Kristjánsson, 1989: Condensation and cloud parameterization studies with a mesoscale numerical weather prediction model. *Mon. Wea. Rev.*, **117**, 1641–1657.
- Suzuki, T., 1983: A theoretical model for dispersion of tephra. *Arc volcanism: Physics and Tectonics*, Shimozuru, D. and I. Yokoyama, Eds., TERRAPUB, Tokyo, 95–113.
- Takasa, S, M. Chikasawa, and H. Mori, 2011: A new method of tide prediction along the Japanese coastline and its accuracy. *Weather Bulletin, JMA*, **78**, S33–S42, (in Japanese).
- Takaya, Y., J.-R. Bidlot, A. Beljaars, and P. A. E. M. Janssen, 2010: Refinements to a prognostic scheme of skin sea surface temperature. *Geophys. Res. Lett.*, **115**, C06 009.
- Takaya, Y., S. Hirahara, T. Yasuda, S. Matsueda, T. Toyoda, Y. Fujii, H. Sugimoto, C. Matsukawa, I. Ishikawa, H. Mori, R. Nagasawa, Y. Kubo, N. Adachi, G. Yamanaka, T. Kuragano, A. Shimpo, S. Maeda, and T. Ose, 2018: Japan Meteorological Agency/Meteorological Research Institute-Coupled Prediction System version 2 (JMA/MRI-CPS2): atmosphere–land–ocean–sea ice coupled prediction system for operational seasonal forecasting. *Clim. Dyn.*, **50**, 751–765.
- Takeuchi, Y. and T. Kurino, 1997: Document of algorithm to derive rain rate and precipitation with SSM/I and AMSR. *Algorithm description of PIs for SSM/I and ADEOS-II/AMSR, 2nd AMSR Workshop*, Tokyo, Japan, 61.1–61.9.
- Takeuchi, Y. and T. Tsuyuki, 2002: The operational 3D-Var assimilation system of JMA for the Global Spectrum Model and the Typhoon Model. *CAS/JSC WGNE Res. Activ. Atmos. Oceanic Modell.*, **32**, 1.59–1.60.
- Tanaka, T. Y., K. Orito, T. T. Sekiyama, K. Shibata, and M. Chiba, 2003: MASINGAR, a global tropospheric aerosol chemical transport model coupled with MRI/JMA98 GCM: Model description. *Pap. Meteorol. Geophys.*, **53** (4), 119–138.
- Tanguay, M., E. Yakimiw, H. Ritchie, and A. Robert, 1992: Advantage of spatial averaging in semi-implicit semi-Lagrangian schemes. *Mon. Wea. Rev.*, **120**, 113–123.
- The WAMDI Group, 1988: The WAM model—A third generation ocean wave prediction model. *J. Phys. Oceanogr.*, **18**, 1775–1810.
- Tiedtke, M., 1993: Representation of Clouds in Large-Scale Models. *Mon. Wea. Rev.*, **121**, 3040–3061.
- Tokioka, T., K. Yamazaki, A. Kitoh, and T. Ose, 1988: The equatorial 30–60 day oscillation and the Arakawa-Schubert penetrative cumulus parameterization. *J. Meteor. Soc. Japan*, **66**, 883–901.
- Toth, Z. and E. Kalnay, 1993: Ensemble Forecasting at NMC: The Generation of Perturbations. *Bull. Amer. Meteor. Soc.*, **74**, 2317–2330.
- Toth, Z. and E. Kalnay, 1997: Ensemble Forecasting at NCEP and the Breeding Method. *Mon. Wea. Rev.*, **125**, 3297–3319.
- Toyoda, T., Y. Fujii, T. Yasuda, N. Usui, T. Iwao, T. Kuragano, and T. Kamachi, 2013: Improved analysis of the seasonal interannual fields by a global ocean data assimilation system. *Theoretical and Applied Mechanics Japan*, **61**, 31–48.
- Tsujino, H., T. Motoi, I. Ishikawa, M. Hirabara, H. Nakano, G. Yamanaka, T. Yasuda, and H. Ishizaki, 2010: Reference Manual for the Meteorological Research Institute Community Ocean Model (MRI.COM) Version 3. Technical Reports of the Meteorological Research Institute 59, Meteorological Research Institute of Japan Meteorological Agency, Ibaraki, Japan. 241pp. (http://www.mri-jma.go.jp/Publish/Technical/DATA/VOL_59/59_en.html).
- Tsujino, H., H. Nakano, K. Sakamoto, S. Urakawa, M. Hirabara, H. Ishizaki, and G. Yamanaka, 2017: Reference Manual for the Meteorological Research Institute Community Ocean Model Version 4 (MRI.COMv4). Technical Reports of the Meteorological Research Institute 80, Meteorological Research Institute of Japan Meteorological Agency, Ibaraki, Japan. 284pp. (http://www.mri-jma.go.jp/Publish/Technical/DATA/VOL_80/index_en.html).
- Udin, I. and A. Ullerstig, 1976: A numerical model for forecasting the ice motion in the Bay and Sea of Bosnia. *SMHI MMK*, **6**, 1–40.
- Ueno, K., 1998: An energy dissipation term in a numerical prediction wave model. *Sokkou-Jihou*, **65**, S181–

- S187, (in Japanese).
- Ueno, K., 2004: The development of the third generation wave model MRI-III for operational use. *Proceedings of the 8th International Workshop on Wave Hindcasting and Forecasting*, North Shore, Oahu, Hawaii, 14 - 19 November 2004, G2, (<http://www.waveworkshop.org>).
- Usui, N., Y. Fujii, K. Sakamoto, and M. Kamachi, 2015: Development of a four-dimensional variational assimilation system toward coastal data assimilation around Japan. *Mon. Wea. Rev.*, **143**, 3874–3892.
- Usui, N., S. Ishizaki, Y. Fujii, H. Tsujino, T. Yasuda, and M. Kamachi, 2006: Meteorological Research Institute multivariate ocean variational estimation (MOVE) system: Some early results. *Adv. Space Res.*, **37**, 806–822.
- van de Berg, L., J. Gustafsson, and A. Yildirim, 2002: Reprocessing of atmospheric motion vectors from Meteosat image data. ECMWF ERA-40 Project Report Series 3, European Centre for Medium-Range Weather Forecasts, Reading, UK. 159–168. (<https://www.ecmwf.int/en/publications>).
- van Vuuren, D. P., J. Edmonds, M. Kainuma, K. Riahi, A. Thomson, K. Hibbard, G. C. Hurtt, T. Kram, V. Krey, J.-F. Lamarque, T. Masui, M. Meinshausen, N. Nakicenovic, S. J. Smith, and S. K. Rose, 2011: The representative concentration pathways: an overview. *Climatic Change*, **109**, 5–31.
- Veersé, F., D. Auroux, and M. Fisher, 2000: Limited-memory BFGS diagonal preconditioners for a data assimilation problem in meteorology. *Optimization and Engineering*, **1**, 323–339.
- Wicker, L. J. and W. C. Skamarock, 2002: Time-splitting methods for elastic models using forward time schemes. *Mon. Wea. Rev.*, **130**, 2088–2097.
- Wilks, D. S., 2006: *Statistical Methods in the Atmospheric Sciences, Second Edition*, International Geophysics Series, Vol. 91. Academic Press, 648pp.
- WMO, 1993: *Guide on the Global Data-processing System*. WMO-No.305, World Meteorological Organization.
- WMO, 2010: Standardized Verification System (SVS) for Long-Range Forecasts (LRF), Attachment II-8. *Manual on the Global Data-processing and Forecasting System Volume I - Global Aspects, Updated in 2012*, World Meteorological Organization.
- WMO, 2017: *Manual on the Global Data-processing and Forecasting System: Annex IV to the WMO Technical Regulations, 2017 edition*. World Meteorological Organization.
- Wurtele, M. G., R. D. Sharman, and T. L. Keller, 1987: Analysis and simulations of a troposphere-stratosphere gravity wave model. Part I. *J. Atmos. Sci.*, **44**, 3269–3281.
- Wyser, K., 1998: The effective radius in ice clouds. *J. Climate*, **11**, 1793–1802.
- Xie, S. C. and M. H. Zhang, 2000: Impact of the convection triggering function on single-column model simulations. *J. Geophys. Res.*, **105**, 14 983–14 996.
- Yabu, S., 2013: Development of longwave radiation scheme with consideration of scattering by clouds in JMA global model. *CAS/JSC WGNE Res. Activ. Atmos. Oceanic Modell.*, **43**, 4.07–4.08.
- Yabu, S., T. Y. Tanaka, and N. Oshima, 2017: Development of a multi-species aerosol-radiation scheme in JMA's global model. *CAS/JSC WGNE Res. Activ. Atmos. Oceanic Modell.*, **47**, 4.15–4.16.
- Yamaguchi, H., M. Higaki, and M. Kyouda, 2014: Upgrade of JMA's One-Week Ensemble Prediction System. *CAS/JSC WGNE Res. Activ. Atmos. Oceanic Modell.*, **44**, 6.17–6.18.
- Yamaguchi, M., R. Sakai, M. Kyoda, T. Komori, and T. Kadowaki, 2009: Typhoon Ensemble Prediction System Developed at the Japan Meteorological Agency. *Mon. Wea. Rev.*, **137**, 2592–2604.
- Yanagino, K. and S. Takada, 1995: Quantitative Analysis and Application to Weather Prediction by Neural Networks. *Technical Report of IEICE*, **NC95-37**, 63–70, (in Japanese).
- Yano, S, Johan C. Winterwerp, A Tai, and T. Saita, 2010: Numerical Experiments on Features of Nonlinear Tide and Its Influences on Sediment Transport in the Ariake Sea and the Yatsushiro Sea. *J. of JSCE, Ser. B2 (Coastal Engineering)*, **66**, 341–345, (in Japanese).
- Yonehara, H. and M. Ujiie, 2011: A stochastic physics scheme for model uncertainties in the JMA one-week ensemble prediction system. *CAS/JSC WGNE Res. Act. Atmos. Ocean Model/WMO*, **41**, 6.09–6.10.
- Yonehara, H., M. Ujiie, T. Kanehama, R. Sekiguchi, and Y. Hayashi, 2014: Upgrade of JMA's Operational NWP Global Model. *CAS/JSC WGNE Res. Activ. Atmos. Oceanic Modell.*, 6.19–6.20.
- Yonehara, H., T. Tokuhiro, R. Nagasawa, M. Ujiie, A. Shimokobe, M. Nakagawa, R. Sekiguchi, T. Kanehama, H. Sato, and K. Saitou, 2017: Upgrade of parameterization schemes in JMA's operational global NWP

- model. *CAS/JSC WGNE Res. Activ. Atmos. Oceanic Modell.*, 4,17–4.18.
- Yonehara, H., R. Sekiguchi, T. Kanehama, K. Saitou, T. Kinami, A. Shimokobe, D. Hotta, R. Nagasawa, H. Sato, M. Ujiie, T. Kadowaki, S. Yabu, K. Yamada, M. Nakagawa, and T. Tokuhiro, 2018: Upgrade of JMA's operational global NWP system. *CAS/JSC WGNE Res. Activ. Atmos. Oceanic Modell.*, 6,15–6.16.
- Yoshimura, H., 2002: Development of a Semi-Implicit Semi-Lagrangian Global Model using Double Fourier Series. *The 4th International Workshop on Next Generation Climate Models for Advanced High Performance Computing Facilities.*, NCAR, Boulder, Colorado, 12 - 14 March 2002.
- Yoshimura, H. and T. Matsumura, 2003: A Semi-Lagrangian Scheme Conservative in the Vertical Direction. *CAS/JSC WGNE Res. Activ. Atmos. Oceanic Modell.*, **33**, 03.19–03.20.
- Yoshimura, H. and T. Matsumura, 2004: Semi-Lagrangian Toitsu model. *Report of Numerical Prediction Division (Suuchiyohouka Houkoku Bessatsu Houkoku)*, **50**, 51–60, (in Japanese).
- Yoshimura, H. and S. Yukimoto, 2008: Development of a Simple Coupler (Scup) for Earth System Modeling. *Pap. Meteor. Geophys.*, **59**, 19–29.
- Yukimoto, S., H. Yoshimura, M. Hosaka, T. Sakami, H. Tsujino, M. Hirabara, T. Y. Tanaka, M. Deushi, A. Obata, H. Nakano, Y. Adachi, E. Shindo, S. Yabu, T. Ose, and A. Kitoh, 2011: Meteorological Research Institute-Earth System Model Version 1 (MRI-ESM1) - Model Description -. Technical Reports of the Meteorological Research Institute 64, Meteorological Research Institute, 83 pp.
- Zeng, X. and A. Beljaars, 2005: A prognostic scheme of sea surface skin temperature for modeling and data assimilation. *Geophys. Res. Lett.*, **32**, L14 605.
- Zhang, H., T. Nakajima, G. Shi, T. Suzuki, and R. Imasu, 2003: An optimal approach to overlapping bands with correlated k distribution method and its application to radiative calculations. *J. Geophys. Res.*, **108**, D20, 4641.
- Zhong, W. and J. D. Haigh, 1995: Improved broadband emissivity parameterization for water vapor cooling rate calculations. *J. Atmos. Sci.*, **52**, 124–138.
- Zsótér, E., 2006: Recent developments in extreme weather forecasting. *ECMWF Newsletter*, **107**, 8–17.

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Outline of the Operational Numerical Weather Prediction at the Japan Meteorological Agency

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