

—Original Paper—

JAMSTEC Model Intercomparison Project (JMIP)

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The JAMSTEC Model Intercomparison Project (JMIP) provides a first opportunity to systematically compare multiple global models developed and/or used in JAMSTEC with the aim of moving toward better weather and climate predictions. Here, we evaluate climate simulations obtained from atmospheric models (AFES and MIROC5), atmospheric model with slab ocean (NICAM.12), and fully coupled model (SINTEX-F1 and SINTEX-F2). In these simulations, the sea surface temperature is fixed (for AFES and MIROC5) or nudged (NICAM.12, SINTEX-F1, and SINTEX-F2) to the observed historical one. We focus on the climatology and variability of precipitation and its associated phenomena, including the basic state, the energy budget of the atmosphere, extratropical cyclones, teleconnection, and the Asian monsoon. We further discuss the possible causes of similarities and differences among the five JMIP models. Though some or most of the dynamical and physical packages in the JMIP models have been developed independently, common model biases are found among them. The AFES and MIROC5, and the SINTEX-F1 and SINTEX-F2, show strong similarities. In many respects, NICAM.12 shows unique characteristics, such as the distributions of precipitation, shortwave radiation, and explosive extratropical cyclones and the onset of the Asian summer monsoon. To some extent, the similarities and differences among the JMIP models overlap with those among the Coupled Model Intercomparison Project Phase-5 (CMIP5) models, suggesting that JMIP can be used as a simple and in-depth version of CMIP to investigate the mechanisms of model bias. We suggest that this JMIP framework could be expanded to an intercomparison of weekly-to-seasonal scale weather forecasting; here, more fruitful discussion is expected through intensive collaboration among modeling and observation groups.

Keywords : Atmospheric model intercomparison, climate simulation, precipitation, large-scale circulation, radiation, teleconnection

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1. Introduction

In cooperation with universities and other research institutes, JAMSTEC (Japan Agency for Marine-Earth Science and Technology) has operated the Earth Simulator, a supercomputer with a peak performance of 1.3 PFLOPS, and developed and employed multiple general circulation models (GCMs) for weather and climate studies. Each GCM has its strength in terms of computational efficiency, spatial resolution, forecast skill, scale interaction, interdisciplinary purpose, and so on. Among them, the atmospheric GCM (AGCM) for the Earth Simulator (AFES; Enomoto et al. 2008), an atmospheric component of coupled general circulation model (CGCM) for the Earth Simulator (CFES; Enomoto et al. 2008), has been highly optimized for the Earth Simulator at JAMSTEC and can be run at a horizontal resolution of $O(10\text{ km})$. The Model for Interdisciplinary Research on Climate (MIROC; Watanabe et al. 2010) is widely used in universities and research institutes, primarily for climate research. The MIROC group is one of two major modeling groups in Japan that contribute to the Coupled Model Intercomparison Project (CMIP). The Multiscale Simulator for the Geoenvironment (MSSG; Takahashi et al. 2008) is a non-hydrostatic global atmospheric model with a Yin-Yang grid system that can simulate multiscale phenomena from global to urban scales on the Earth Simulator. The Non-hydrostatic Icosahedral Atmospheric Model (NICAM; Satoh et al. 2014), a non-hydrostatic global atmospheric model with an icosahedral grid system, has been developed and used for high-resolution weather and climate simulations on multi-platform environments including the Earth Simulator and the K computer, a 10-PFLOPS supercomputer operated by RIKEN. The Scale Interaction Experiment-Frontier ver. 1 (SINTEX-F1) model (Luo et al. 2005b), a coupled atmosphere–ocean model, has improved seasonal predictions of El Niño/La Niña, the Indian Ocean Dipole (IOD), and their teleconnections. The SINTEX-F1 model has routinely provided real-time seasonal forecasts since 2005 (see <http://www.jamstec.go.jp/frcgc/research/d1/iod/e/seasonal/outlook.html>). Recently, a revised CGCM called SINTEX-F2, which is a high-resolution version with a dynamical sea-ice model, has been developed (Masson et al. 2012). Other than the above models, an operational numerical weather prediction model, the global spectral model (GSM) at the Japan Meteorological Agency (JMA), and an AGCM jointly developed at JMA

and the Meteorological Research Institute (MRI-AGCM) (Mizuta et al. 2012) are major global models developed and used in Japan. Also noteworthy is that JAMSTEC has been operating the Earth Simulator, as well as fostering model developers, technical staff, and users, since 2002. Thus, JAMSTEC is a major hub for global model development, simulation, and application in Japan.

The diversity of global models is a result of the diversity of research in this field, and it is clear that there is currently no global model that can be perfectly applied to all cutting-edge research. Although the research focus differs among modeling groups, some issues common to all the models do exist. Obviously, model bias is a central concern for all modeling groups; therefore, a shared approach to model improvement, which has not been typically published in the literature, would help enhance our ability to research cutting-edge themes. The CMIP is an international framework for comparing the results of global models and provide climate projection information for the Intergovernmental Panel on Climate Change (IPCC). However, users of model data have experienced difficulties in analyzing model data with a deep understanding of the nature of the individual models. Also, model developers have found difficulties receiving feedback from users of the model data. The same issue exists between observation experts and model data users and between observation experts and model developers. As such, JAMSTEC, where model developers, model data users, and observation experts work together, can be the ideal institute at which to bridge these gaps.

These observations motivated us to launch the JAMSTEC Model Intercomparison Project (JMIP). As an initial step, the results of climate simulations from five participating models (AFES, MIROC5, NICAM.12, and SINTEX-F1 and -F2; hereafter, the JMIP models) were collected on a single data server with the same data format. Except for MIROC5, these JMIP models have never been included in CMIP (although NICAM has been partially included). The collected datasets were evaluated and compared based on the wide variety of the expertise of the participants. We discussed the comparison results and exchanged details on the model configurations in bimonthly JMIP meetings.

In this paper, outcomes of the first phase of JMIP activity are summarized. In Section 2, the participating models are introduced. Experimental design and datasets are described in Section 3. Some key results

are shown in Section 4: precipitation and zonal mean basic state in Section 4.1, atmospheric radiation in Section 4.2, extratropical cyclones in the Northwestern Pacific in Section 4.3, interannual variability of El Niño/La Niña and IOD in Section 4.4, and the Asian summer monsoon in Section 4.5. A summary and future perspectives are presented in Section 5.

2. Models

Table 1 shows the configurations of the models used in this study. In this section, the features, main scientific targets, and milestone papers of each model are summarized.

2.1 AFES

AFES is an AGCM for the Earth Simulator (Ohfuchi et al. 2004; Enomoto et al. 2008; Kuwano-Yoshida et al. 2010a). Since 1998, AFES has been developed by Frontier Research System for Global Change, Research Organization for Information Science and Technology (RIST), Earth Simulator Center and Application Laboratory in JAMSTEC. The original code of AFES was adopted from version 5.4.02 of an AGCM developed jointly by the Center for Climate System Research (CCSR) of the University of Tokyo and the Japanese National Institute for Environmental Sciences (NIES) (Numaguti et al. 1997). The first version of AFES achieved 26.58 TFLOPS on the ES, for which it received the Gordon Bell Award for Peak Performance at Super Computing 2002, held in Baltimore, MD, USA, November 2002 (Shingu et al. 2002). Now, most of algorithms for effective computation on ES used in AFES are adopted to other AGCMs compared in this study. Enomoto et al. (2008) improved the accuracy and efficiency via the Legendre transform and physical performance via the introduction of a new radiation scheme (MstrnX; Sekiguchi and Nakajima 2008) and convection scheme (Emanuel 1991; Emanuel and Živković-Rothman 1999; Peng et al. 2004); this new version was termed AFES version 2. AFES version 3 (AFES3), as used in the present comparison, introduced an improved PDF cloud scheme (Kuwano-Yoshida et al. 2010a). AFES is mainly used for mid- and high-latitude atmosphere-ocean interactions (Minobe et al. 2008; Kuwano-Yoshida et al. 2010b; Ogawa et al. 2012; Okajima et al. 2018), ensemble-based data assimilation (Miyoshi and Yamane 2007; Miyoshi et al.

2007), and weather predictability (Kuwano-Yoshida and Enomoto 2013; Sato et al. 2017). In addition, AFES is expanded to other planets such as AFES-Venus (Sugimoto et al. 2017; Takagi et al. 2018). In the present comparison, AFES3 was integrated with modest horizontal resolution T239 (approximately 0.5°) and 48 vertical levels of sigma coordinate from the surface to about 3 hPa to understand roles of western boundary currents such as the Kuroshio and the Gulf Stream, which provide massive heat/moisture to atmosphere in midlatitudes. The National Oceanic and Atmospheric Administration (NOAA) 0.25° daily sea surface temperature (SST) data (Reynolds et al. 2007) were used for the bottom boundary conditions. The dataset was used as a control experiment in O'Reilly et al. (2016), O'Reilly et al. (2017), and Kuwano-Yoshida and Minobe (2017).

2.2 MIROC5

MIROC5 is a version of the global climate model MIROC, which was included in CMIP5 (Watanabe et al. 2010). Its atmospheric component has a resolution of T85 (1.4°) with 40 vertical levels up to 3 hPa. The physical parameterizations included in the atmospheric component can be found in Table 1 (see Watanabe et al. 2010 for more details). In this paper, results of the AMIP-type simulation, in which observed SSTs and sea ice concentrations were given as the bottom boundary conditions of the AGCM, are compared with the results obtained from other models.

MIROC5 has been developed to study a wide variety of climate-scale phenomena, from internal variations in the climate system like the Arctic Oscillation (AO), tropical intraseasonal variations, El Niño/Southern Oscillation (ENSO), and Pacific Decadal Oscillation (PDO), to climate sensitivity, anthropogenic climate change and paleo climates (e.g. Chikira and Sugiyama, 2013; Watanabe et al. 2012; Chikamoto et al. 2013; Watanabe et al. 2013; Kamae and Watanabe 2013). In this regard, MIROC5 includes a coupled aerosol module SPRINTARS (Takemura et al. 2005) and considers feedback processes among aerosols, clouds, precipitation, radiation, atmospheric circulations, and land surface conditions. This is a unique strength of MIROC5 among other models.

2.3 NICAM.12

NICAM is a non-hydrostatic icosahedral atmospheric model (Tomita and Satoh 2004; Satoh et

Table 1. Model configurations.

Model name	AFES	MIROC5	NICAM.12	SINTEX-F1	SINTEX-F2
Model description papers	Ohfuchi et al. (2004), Enomoto et al. (2008), Kuwano-Yoshida et al. (2010a)	Watanabe et al. (2010)	Tomita and Satoh (2004), Satoh et al. (2008, 2014)	Roeckner et al. (1996), Gualdi et al. (2003), Luo et al. (2005b)	Roeckner et al. (2003), Masson et al. (2012), Sasaki et al. (2013)
Dynamical core	Hydrostatic spectrum model, σ coordinate	Hydrostatic spectrum model, hybrid σ -p coordinate	Non-hydrostatic finite volume model, z^* (terrain-following) coordinate	Hydrostatic spectrum model, hybrid σ -p coordinate	Hydrostatic spectrum model, hybrid σ -p coordinate
Radiation	MstrnX (Sekiguchi and Nakajima 2008)	MstrnX (Sekiguchi and Nakajima 2008)	MstrnX (Sekiguchi and Nakajima 2008)	SW scheme (Fouquart and Bonnell 1980) and LW scheme (Morcrette 1991)	SW scheme (Fouquart and Bonnell 1980) and LW scheme (Mlawer et al. 1997)
Cumulus convection	Emanuel (1991), Emanuel and Živković-Rothman (1999)	Chikira and Sugiyama (2010)	Not used	Tiedtke (1989)	Tiedtke (1989), Nordeng (1994)
Large-scale condensation	Kuwano-Yoshida et al. (2010a)	Watanabe et al. (2009)	Not used	Roeckner et al. (1996)	Roeckner et al. (2003)
Cloud microphysics	None	Bulk scheme (Wilson and Ballard 1999)	Single-moment six-category bulk scheme (Tomita 2008)	None	None
Turbulence	MYNN Level-2 (Nakanishi and Niino 2004)	MYNN Level-2.5 (Nakanishi 2001; Nakanishi and Niino 2004)	Modified MYNN Level-2 (Nakanishi and Niino 2006; Noda et al. 2010)	Brinkop and Roeckner (1995)	Brinkop and Roeckner (1995)
Land model	MATSIRO (Takata et al. 2003)	MATSIRO (Takata et al. 2003)	MATSIRO (Takata et al. 2003)	Roeckner et al. (1996)	Roeckner et al. (2003)
Ocean model	Not used	Not used	Slab ocean model of 15 m depth, nudging with a relaxation time of 7 days	Full OGCM, nudging (damping rate: $-2400 \text{ W m}^{-2} \text{ K}^{-1}$)	Full OGCM, nudging (damping rate: $-2400 \text{ W m}^{-2} \text{ K}^{-1}$)

al. 2008, 2014). It adopts a finite volume method with an icosahedral grid system to perform ultra-high-resolution simulations using a massive parallel supercomputer. The main scientific target of NICAM is tropical meteorology (see review in Satoh et al. 2014), specifically, convection, Madden-Julian Oscillation (MJO), tropical cyclones, diurnal cycles of precipitation, and so on. Diabatic heating by latent heat release is crucial to these studies, and GCMs with a horizontal resolution of $O(100\text{ km})$ rely on the cumulus convection scheme, which is known to be a primary cause of uncertainties (Randall et al. 2003). NICAM realistically simulates global climatology without the cumulus convection scheme using a 3.5-km mesh (or even a 14-km mesh) and can eliminate the ambiguity associated with the choice of convection scheme. On the first generation of the Earth Simulator, MJO was simulated for the first time using 3.5-km mesh NICAM (Miura et al. 2007). Continuous model development and enhanced computational power have improved model performance and promoted grand challenge simulations. In terms of horizontal resolution, the first-ever global cloud resolving simulation was achieved with a mesh size of 870 m (Miyamoto et al. 2013) and was continued for two days (Yashiro et al. 2016a). Though horizontal resolution of at least 2 km is necessary to resolve convection cores with multiple grids (Miyamoto et al. 2013), the statistical features of the convection and clouds except for shallow convection can be practically simulated using 14 km mesh model (e.g. Noda et al. 2010, 2012; Pauluis and Garner 2006). Present and future climate simulations using 14-km mesh NICAM were performed (Kodama et al. 2015; Satoh et al. 2015) and analyzed in terms of intraseasonal variation (Kikuchi et al. 2017), tropical synoptic-scale disturbance (Fukutomi et al. 2015), tropical cyclones (Satoh et al. 2015; Yamada et al. 2017), and climate sensitivity (Chen et al. 2016). The performance of MJO prediction was statistically confirmed through 14-km mesh, 54-ensemble simulation (Miyakawa et al. 2014). An intensive effort is further devoted to testing massive ensemble simulations (Miyoshi et al. 2015; Yashiro et al. 2016b). A recently developed double moment bulk cloud microphysics scheme (Seiki and Nakajima 2014) shows better performance in simulating climatology in a more realistic manner (Seiki et al. 2015).

In this study, the output from the 30-yr climate simulation of NICAM.12 (the 2012 version), in which SST is nudged toward the observed historical one using slab ocean

model, was used for an analysis; see Kodama et al. (2015) for a detailed description of the experiment and its performance.

2.4 SINTEX-F1 and F2

2.4.1 SINTEX-F1

SINTEX-F1, a fully coupled GCM, has been previously used for the Application Laboratory (APL)/JAMSTEC seasonal prediction system (Luo et al. 2005a,b). The atmospheric component (ECHAM4) has a resolution of T106 (1.125°) with 19 vertical levels (Roeckner et al. 1996). The oceanic component (OPA8) has a relatively coarse resolution of $2^\circ \times 2^\circ$ Mercator horizontal mesh but with a tropical refinement up to 0.58° in the meridional direction (Madec et al. 1998). The coupling information is exchanged every two hours without correction by the Ocean Atmosphere Sea Ice Soil (OASIS) 2 coupler (Valcke et al. 2000). Sea ice cover is relaxed toward the observed monthly climatology. In this study, we analyzed the outputs generated using a relatively simple SST-nudging initialization scheme with the SINTEX-F1 CGCM (Luo et al. 2005a). Model SSTs are strongly nudged toward daily observations in a coupled model with a restoring time of 1-day using the weekly NOAA OISSTv2 (Reynolds et al. 2002). The seasonal prediction system based on SINTEX-F1 has so far demonstrated high prediction performance for the ENSO (Luo et al. 2005a, 2008a; Jin et al. 2008), IOD (Luo et al. 2007, 2008b), subtropical dipole modes (Yuan and Yamagata 2015), and Coastal Niño (Doi et al. 2013, 2015a,b).

2.4.2 SINTEX-F2

The SINTEX-F2 coupled model, a higher-resolution and upgraded version with a dynamical sea-ice model, has been developed to resolve several physical processes, particularly relatively small-scale phenomena in the ocean (Masson et al. 2012; Sasaki et al. 2013). The atmospheric component (ECHAM5) has a horizontal resolution of T106 (same as the SINTEX-F1 model) with 31 vertical levels (Roeckner et al. 2003). The oceanic component (OPA9) has the horizontal resolution of a $0.5^\circ \times 0.5^\circ$ tri-polar grid (known as the ORCA05 configuration) with 31 vertical levels, similar to the SINTEX-F1 system (Madec 2008). While no sea ice model is incorporated in the SINTEX-F1 system, the dynamical sea ice model of the Louvain-la-Neuve Sea Ice Model (LIM) version 2 (LIM2) (Fichefet and Morales Maqueda 1997) is embedded in the SINTEX-F2 system. Although the atmospheric and

Table 2. Experimental designs of AMIP simulation.

Model name	AFES	MIROC5	NICAM.12	SINTEX-F1	SINTEX-F2
Description papers	O'Reilly et al. (2016)	Watanabe et al. (2010)	Kodama et al. (2015)	Luo et al. (2005b)	Doi et al. (2016)
Horizontal resolution, number of grids	T239 (~ 0.5°), 720 × 360	T85 (~ 1.4°), 256 × 128	glevel – 9 (~ 14km), 10 × 4 ⁹ + 2	T106 (~ 1.125°), 320 × 160	T106 (~ 1.125°), 320 × 160
Number of vertical levels, model top height	L48, 3 hPa	L40, 3 hPa	L38, 40 km	L19, 10 hPa	L31, 10 hPa
Integration term	1 September 1981–31 August 2001	1 January 1979–31 December 2008	1 June 1978–31 December 2008	1 January 1982–present	1 January 1982–present
Sea surface temperature / Sea ice	NOAA 0.25° daily OISST (Reynolds et al. 2007)	CMIP5 AMIP2 1° monthly (Taylor et al. 2000)	HadISST1 1° monthly with Taylor correction (Rayner et al. 2003; Taylor et al. 2000)	NOAA 1° weekly OISST (Reynolds et al. 2002)	NOAA 1° weekly OISST (Reynolds et al. 2002)
Ozone	AMIP2 zonal mean monthly climatology (Liang et al. 1997)	See A2.2.1 of Eyring et al. (2013)	MRI-CTM output (Shibata et al. 2005) under CCMVal REF2 condition (Eyring et al. 2008)	Zonal mean monthly climatology (Fortuin and Kelder 1998)	Zonal mean monthly climatology (Fortuin and Kelder 1998)
CO ₂	Constant mixing ratio	CMIP5 (Hansen and Sato 2004)	CMIP5 (Hansen and Sato 2004)	Constant mixing ratio, 348 ppmv	Constant mixing ratio, 348 ppmv
Aerosol	Given, monthly climatology simulated by SPRINTARS	Calculated online with SPRINTARS	None	Given, monthly climatology simulated by Global Aerosol Data Set (GADS) (Koepke et al. 1997)	Given, monthly climatology simulated by GADS (Koepke et al. 1997)
Insolation	Constant	CMIP5 historical solar constant	Constant (1365 W m ⁻²)	Constant (1365 W m ⁻²)	Constant (1365 W m ⁻²)

oceanic fluxes are exchanged every two hours with no flux correction by means of the Ocean Atmosphere Sea Ice Soil, version 3 (OASIS3) coupler (Valcke et al. 2004), we here analyzed the outputs of the SST-nudging run; model SSTs are strongly nudged toward daily observations similar to SINTEX-F1. More details and an overview of the prediction skill based on retrospective seasonal forecast experiments are given by Doi et al. (2016) and Doi et al. (2017).

3. Experimental designs and datasets

The simulations were performed in general accordance with AMIP protocol (Gates 1992). The configuration of each model is summarized in Table 2. The horizontal resolution ranges by one digit (from around 150 km to 14 km). The model top is placed inside the stratosphere or around the stratopause, and the interval of the vertical levels is similar among the models (as compared with the large differences in horizontal resolution). The observed SST boundary conditions were directly imposed in AFES and MIROC5, whereas they were imposed via nudging techniques in NICAM.12, SINTEX-F1, and SINTEX-F2, as shown in Table 1. Unless specified in the text, the analysis period is 1983–2000, and the data are re-gridded to 2.5° in longitude and latitude for the analysis. Abbreviations and observational dataset references used in this study are listed in Table 3.

4. Results

4.1 Precipitation and zonal mean basic state

Precipitation and zonal mean basic state are fundamental metrics for evaluating the hydrological,

dynamical, and thermodynamical aspects of the simulated climate. Figures 1 and 2 show annual mean precipitation climatology for GPCP observations and the JMIP

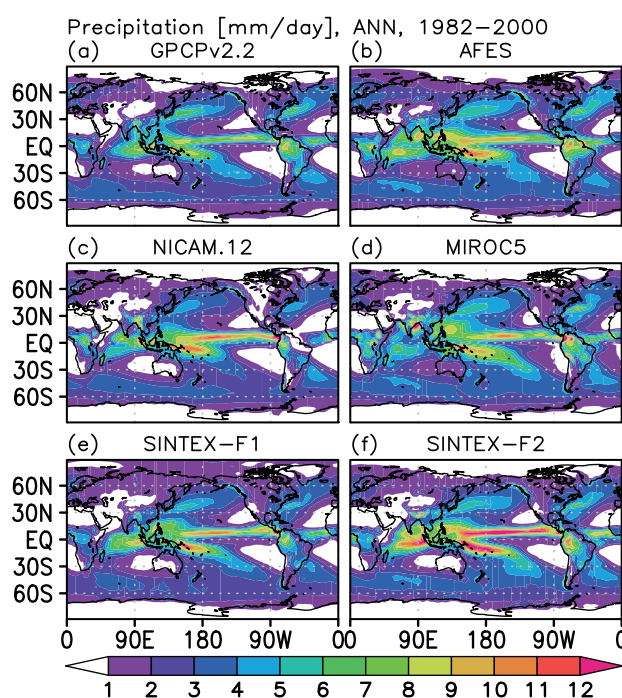


Fig. 1. Annual mean precipitation for GPCP version 2.2 (a) and JMIP models (b-f) [mm day^{-1}].

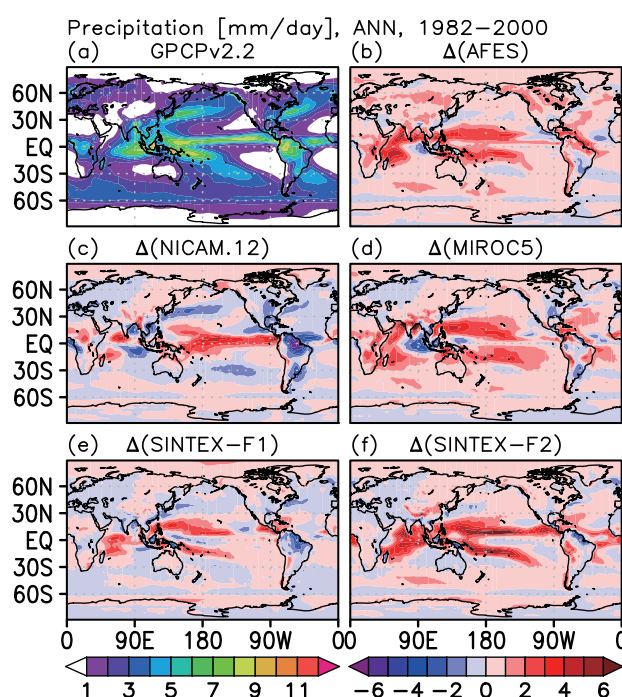


Fig. 2. Annual mean precipitation for GPCP version 2.2 (a, left color bar), and biases of the JMIP models from GPCP (b-f, right color bar).

Table 3. List of observational datasets.

Abbreviation	Full name	Reference
CERES	Clouds and Earth's Radiant Energy System (CERES) EBAF-TOA Ed2.8 and EBAF-Surface Ed2.8	Loeb et al. (2009)
GPCP	Global Precipitation Climatology Project	Adler et al. (2003)
JRA-55	Japanese 55-year reanalysis	Kobayashi et al. (2015)
JRA-55C	Japanese 55-year reanalysis assimilating conventional observations only	Kobayashi et al. (2014)
OISSTv2	Optimum interpolation sea surface temperature version 2	Reynolds et al. (2002)

models. Their global means and the bias and root mean square error from the GPCP observations are shown in Table 4. Precipitation concentrates around the intertropical convergence zone (ITCZ) and storm-track regions. Over the Pacific, precipitation has a southern branch, the so called South Pacific convergence zone (SPCZ). Dry regions spread in the subtropics, where Hadley circulation descends. These qualitative patterns of the observed precipitation are simulated by all the JMIP models. As highlighted in Figs. 2b-f, all the models present excessive precipitation around the ITCZ. Among the JMIP models, AFES and MIROC5 show many similarities over oceans, such as excesses of tropical North and South Pacific precipitation (which is also seen in SINTEX-F1 and SINTEX-F2) and the east-west dipole structure of a precipitation bias over the Indian Ocean (which is also seen in NICAM.12). The reasons for this strong similarity may not be explained easily because these two models incorporate different physical parameterizations, such as cumulus convection and stratiform clouds (Table 1), and run under different horizontal and vertical resolutions (Table 2). The dipole-type precipitation bias over the Indian Ocean is not seen in the coupled version MIROC5 (Watanabe et al. 2010) and may be related to the treatment of air-sea interactions. SINTEX-F1 shows a similar bias pattern to SINTEX-F2, although SINTEX-F2 produces much more global mean precipitation compared with SINTEX-F1. This is partly due to new longwave (LW) and shortwave (SW) radiation schemes (Lohmann et al. 2007). Among the JMIP models, NICAM.12 has a unique precipitation bias pattern; that is, an excess of precipitation over the Central to Eastern Pacific at the equator and a lack of precipitation over South America. NICAM.12 best simulates the amount of global mean precipitation (Table 4), although it includes strong compensating biases. The annual mean precipitation patterns in the JMIP models vary widely such that we can expect a certain degree of model uncertainty in other weather and climate phenomena; this is described in the following sections.

The zonal mean temperatures and zonal winds for December-January-February (DJF) for the JRA-55 reanalysis and the JMIP models are shown in Figs. 3 and 4. All the JMIP models capture the fundamental structures of the thermodynamical and dynamical fields. All the JMIP models (except for AFES) have cold biases around the extratropical tropopause regions, especially in the

Southern Hemisphere. Among the models, AFES, MIROC5, and SINTEX-F2 well simulate zonal mean temperatures and zonal winds in the troposphere, whereas NICAM.12 and SINTEX-F1 have a warm bias in the tropical upper troposphere. In relation to this warm bias in the tropical upper troposphere and cold biases in the extratropical tropopause regions, NICAM.12 overestimates the wind speed of subtropical jets at their high-latitude flank.

Overall differences in the precipitation patterns and zonal mean basic states among the JMIP models may originate from differences in the (i) choice of physics scheme, (ii) horizontal and vertical resolutions, and (iii) target of the intensive model tuning. For example,

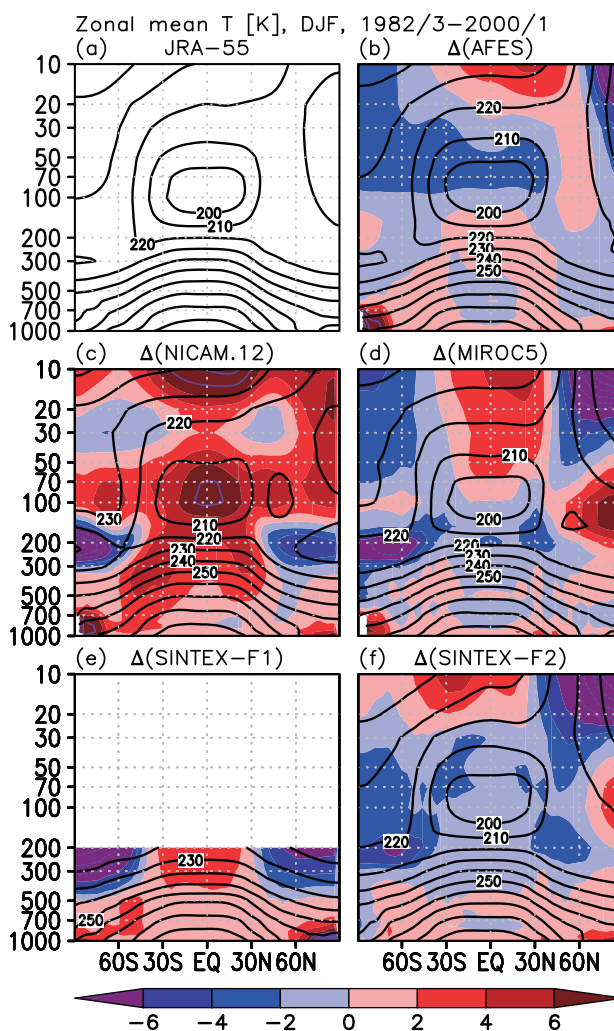


Fig. 3. Zonal mean temperature for December-January-February (DJF) during 1982/1983–2000/2001. (a) JRA-55 reanalysis. (b-f) JMIP models as black contours and their biases as shading and additional purple contours (± 8 K, ± 10 K, ...). Note that SINTEX-F1 did not output temperature above 200 hPa.

NICAM.12, the main focus of which is moist processes of the tropical meteorology, uses a shallow slab ocean with SST nudging; this leads to a better performance of MJO (Grabowski 2006) and an reduction of the double ITCZ bias as a result (Kodama et al. 2015). However, this also leads to a bias in the extratropics, as described in Section 4.3. Note that NICAM.12 does not use convection and orographic gravity wave drag schemes, and biases in the zonal mean basic state described in this study have been reduced significantly in the latest version by updating of the cloud microphysics scheme (Seiki et al. 2015) as well as the inclusion of an orographic

gravity wave drag scheme. This is one outcome of JMIP because all the other JMIP models use orographic gravity wave drag schemes to tune the mid-latitude jet.

These hydrological, thermodynamical, and dynamical aspects of the simulated climate are connected with other global and regional phenomena, as described in the following sections. Global mean precipitation is tightly connected with outgoing longwave radiation (OLR) (Section 4.2). The spatial precipitation pattern is related to local circulations including the Asian monsoon (Section 4.5) and storm-track (Section 4.3), whose interannual variabilities are affected by the ENSO, IOD, and their teleconnections (Section 4.4).

4.2 Atmospheric reflection and absorption of SW radiation

The energy balance of the radiation flux at the top of the atmosphere (TOA) is very important in simulations of Earth's climate. Table 5 shows temporally and globally averaged net fluxes of SW and LW radiation at the TOA, as calculated from the data of each model and observational products of the Clouds and Earth's Radiant Energy System (CERES) experiment (Wielicki et al. 1996). Here, we used the monthly averaged global data sets of CERES EBAF-TOA Ed2.8 (Loeb et al. 2009) as observation data. Note that the averaging period for the five models is 1982–2000 and for CERES is 2001–2015; the difference of the averaging period is due to data availability. The net LW fluxes in the JMIP models (except for SINTEX-F2) show similar values ($-236.0 \pm 0.4 \text{ W m}^{-2}$). In SINTEX-F2, the equatorial OLR is higher than that in CERES and the other models (not shown). This is due to a shortage of clouds in SINTEX-F2, as suggested by the negative bias of the atmospheric albedo analyzed below; that is, the strong LW radiation from the warm near-surface atmosphere is less blocked by the clouds and reaches the TOA.

In contrast to the similarity observed in the LW fluxes, the inter-model difference in the net SW fluxes is as large as 14.5 W m^{-2} . The fact that the inter-model difference in the net SW radiation flux is larger than that in the net LW radiation flux implies a large uncertainty in processes

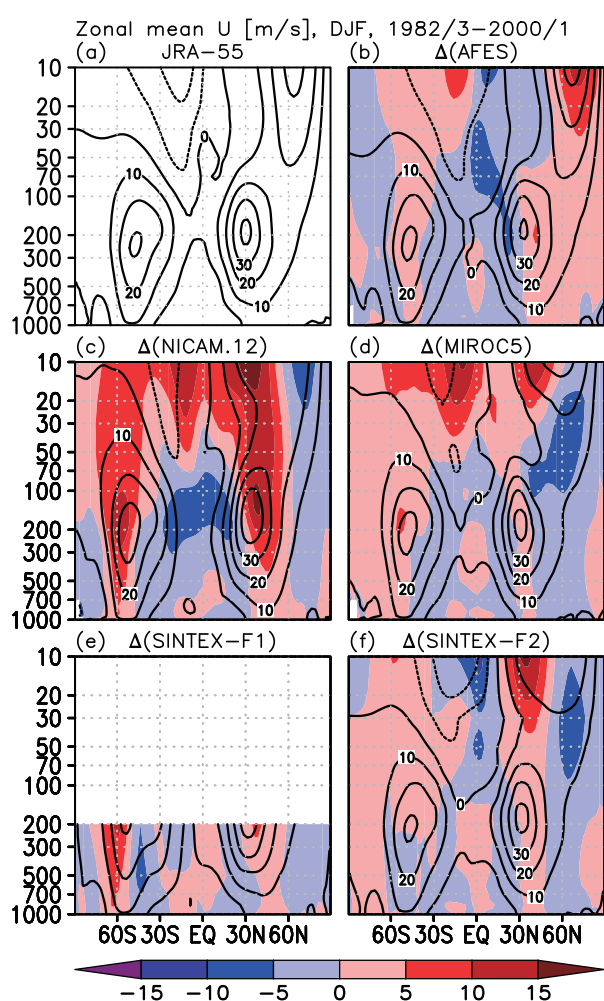


Fig. 4. Same as Fig. 3 but for zonal mean zonal wind.

Table 4. Statistics for annual mean precipitation (Figures 1 and 2). Global mean and bias and root mean square error from GPCP version 2.2 are shown in mm day^{-1} .

	GPCP	GAFES	NICAM.12	MIROC5	SINTEX-F1	SINTEX-F2
Global mean	2.68	3.31	2.67	3.24	2.76	3.24
Bias		+0.63	−0.01	+0.56	+0.08	+0.56
RMSE		1.19	1.22	1.36	0.94	1.49

related to SW radiation, such as reflection and absorption by the atmosphere and clouds. Donohoe and Battisti (2011) proposed a method to estimate the atmospheric albedo (i.e., reflection rate) and absorption rate for SW radiation from the upward and downward SW radiation fluxes at the TOA and the surface by assuming a single layer model of the atmosphere. Here, we adopt their methods and estimate the atmospheric albedo and absorption rate for the five models and the observational data (CERES EBAF-Surface Ed2.8 is used for the data at the surface).

Figures 5a and 5g show a temporally averaged atmospheric albedo map and its zonal mean, respectively, for SW radiation calculated from the CERES data. Because the reflection rate depends on the solar zenith angle, higher albedo is generally observed at higher latitudes. Horizontal variation of the atmospheric albedo at low- and mid-latitudes is mainly due to cloud; that is, cloudy regions show high albedo, whereas dry (non-cloudy) regions show low albedo. We confirm that the atmospheric albedo calculated from clear-sky values shows an almost zonally uniform distribution (not shown). Note that in recent years, a high albedo in China seems to be due to a large amount of SO₂ emission (Klimont et al. 2013) caused by industrial growth. Model biases, as compared with CERES data, are shown in Figs. 5b-f and h. The four models (except for SINTEX-F2) show higher albedo over the subtropical ocean where low albedo is observed. This implies that the albedo of the thin lower clouds above the ocean is overestimated in these models. On the other hand, there are negative biases in high albedo oceanic regions, such as the eastern part of Pacific in both hemispheres. This implies that the amounts of lower cloud in the cloudy regions are underestimated in the models. These results are consistent with Medeiros and Stevens (2011). They analyzed representations of lower clouds in four different GCMs and reported that the GCMs tend to overestimate the lower clouds due to the trade-wind inversion over the subtropical ocean and underestimate the marine stratocumulus in eastern part of Pacific (i.e., west coast). One of the most marked inter-model differences is

the negative albedo bias of NICAM.12 on land, which seems to be due to a negative bias in the cloud fraction (Kodama et al. 2015) and the lack of aerosols in NICAM.12. Such a negative bias over a wide area results in a positive bias in the net SW radiation flux at the TOA, as shown in Table 5. The bias over the Maritime Continent also shows significant inter-model differences: negative in AFES, NICAM.12 and SINTEX-F2, and positive in MIROC5 and SINTEX-F1. This may be relevant to the difficulty in simulating realistic amounts of both precipitation and clouds over the Maritime Continent, where convection, the boundary layer, large-scale condensation, radiation, and surface heat flux schemes would interact complexly with each other.

The atmospheric absorption rate for SW radiation is shown in Fig. 6. The SW absorption in the atmosphere is mainly due to water vapor. As such, a high absorption rate is observed above the ocean, especially in equatorial regions, and a low rate is observed in dry areas and high lands. Since the horizontal variation of water vapor at the synoptic scale and mesoscale is smaller than that of the clouds, and the absorption rate does not depend on the solar zenith angle, the horizontal distribution of the absorption rate is more uniform compared to that of the albedo. In polar regions, all the models show positive bias in the atmospheric SW absorption rate, possibly implying an overestimation of the water vapor amounts in these regions. Insufficient accuracy in modeling the latent and sensible surface heat fluxes in sea ice regions, which is a common problem in most global climate models, could be a reason for the overestimation of the water vapor amount in polar region. In fact, high temperature biases, which is equivalent to positive biases in saturated water vapor amount, are obtained in lower troposphere in polar region for all models as shown in Fig. 3b-f. Note that the absorption rate may be overestimated because of the positive bias in the surface albedo, which tends to be too high for snow and ice by tuning; however, this is not the case for at least the Antarctic Ocean. In most of the other regions, AFES, NICAM.12, and MIROC5 show negative bias, especially over the continents, whereas SINTEX-F1

Table 5. Globally and temporally averaged values of the net SW and LW radiation fluxes at the TOA, and those of the atmospheric albedo and absorption rate of SW radiation, for each model and observational dataset from CERES. The model-mean is also shown.

	CERES	AFES	NICAM.12	MIROC5	SINTEX-F1	SINTEX-F2	Model-mean
Net-SW [W/m ²]	240.6	240.2	249.4	235.3	238.0	245.8	241.7
Net-LW [W/m ²]	-239.7	-236.4	-236.1	-235.6	-236.0	-245.3	-237.9
Atmos. alb.	0.286	0.281	0.241	0.297	0.304	0.276	0.280
Atmos. abs.	0.212	0.202	0.197	0.207	0.252	0.213	0.214

shows a significantly positive bias. In SINTEX-F2, a positive bias is shown in the Eurasian Continent and northern part of the African Continent, and a negative bias is shown in other regions. Aerosols could be a reason for the observed biases over the continents; a relatively strong negative bias in NICAM.12 would be due to a lack of aerosols in the model, and a locally strong negative bias around China would be due to the difference in the averaging period between the CERES data and the models, which should reflect an increase of black carbon emission with industrial growth (Wang et al. 2012). The positive biases in SINTEX-F1 and SINTEX-F2 suggest that the amount of aerosols is overestimated or

the SW absorption rate of the aerosols is too high in these models.

The above analyses show that the inter-model differences in the net SW flux at the TOA would be caused by the differences in atmospheric albedo for SW radiation, which is mainly a result of the uncertainty in the cloud and radiative modeling as well as aerosol treatment.

4.3 Explosive extratropical cyclones in the Northwestern Pacific

Extratropical cyclones are the main phenomenon in the mid-latitude troposphere. In particular, explosive

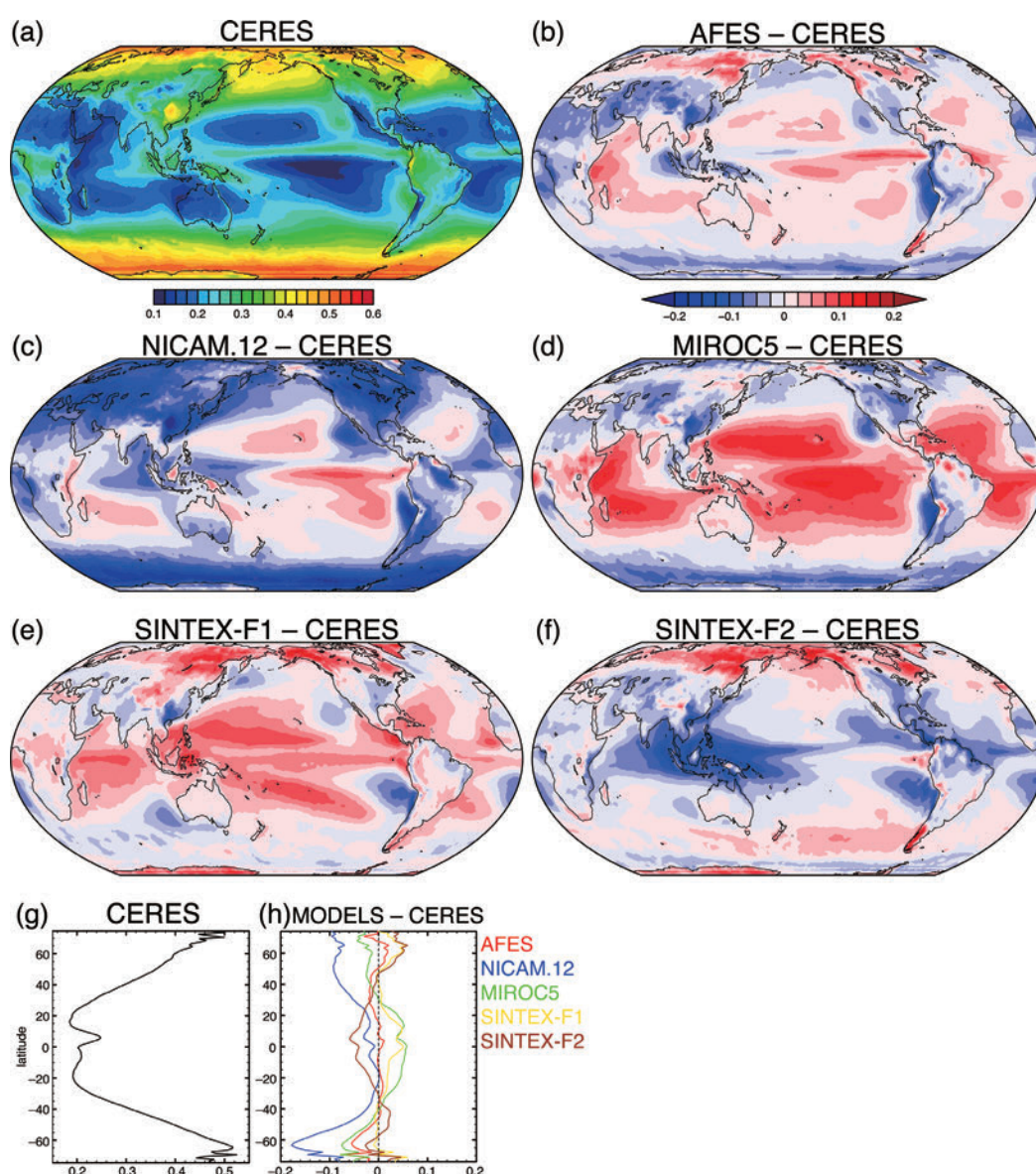


Fig. 5. (a) Atmospheric albedo map for SW radiation calculated from CERES data and (g) its zonal mean. Panel (h) shows the zonally averaged deviations of the models' values from CERES data. Other panels show the deviation maps for (b) AFES, (c) NICAM.12, (d) MIROC5, (e) SINTEX-F1, and (f) SINTEX-F2. Time-averaged data are used for each panel.

extratropical cyclones are active in winter and their activity controls precipitation and jet stream meandering over oceans. Therefore, in this section, the activity of explosive extratropical cyclones in January over the Northwestern Pacific in the five JMIP models is investigated. Observations have shown that the Northwestern Pacific is the most active region in the world for explosive extratropical cyclones and their activity reaches a maximum in January. To estimate the activity of the cyclones, the Local Deepening Rate for 24 hours (LDR24, Kuwano-Yoshida 2014) index is used. This index uses the surface pressure local tendency for 24 hours:

$$\text{LDR24} = -\frac{P_{sfc}(t+12h) - P_{sfc}(t-12h)}{24} \left| \frac{\sin 60^\circ}{\sin \theta} \right|,$$

where P_{sfc} is surface pressure, t is the time, and θ is the latitude. $\text{LDR24} \geq 1 \text{ hPa hr}^{-1}$ is defined as an explosive development. The monthly activity of explosive cyclones is estimated using the following expression;

$$\text{LDR24P1} = \frac{1}{n} \sum_{i=1}^n \sigma_1(t=i)$$

$$\sigma_1 = \begin{cases} \text{LDR24}, & \text{if } \text{LDR24} \geq 1 \text{ hPa h}^{-1} \\ 0, & \text{otherwise} \end{cases},$$

where n is the number of time steps in the month. In the present study, 6-hourly outputs of surface pressure are used. Unfortunately, SINTEX-F1 is not considered because it has no 6-hourly outputs. As a reference, LDR24P1 is also assessed using Japanese 55-year Reanalysis using

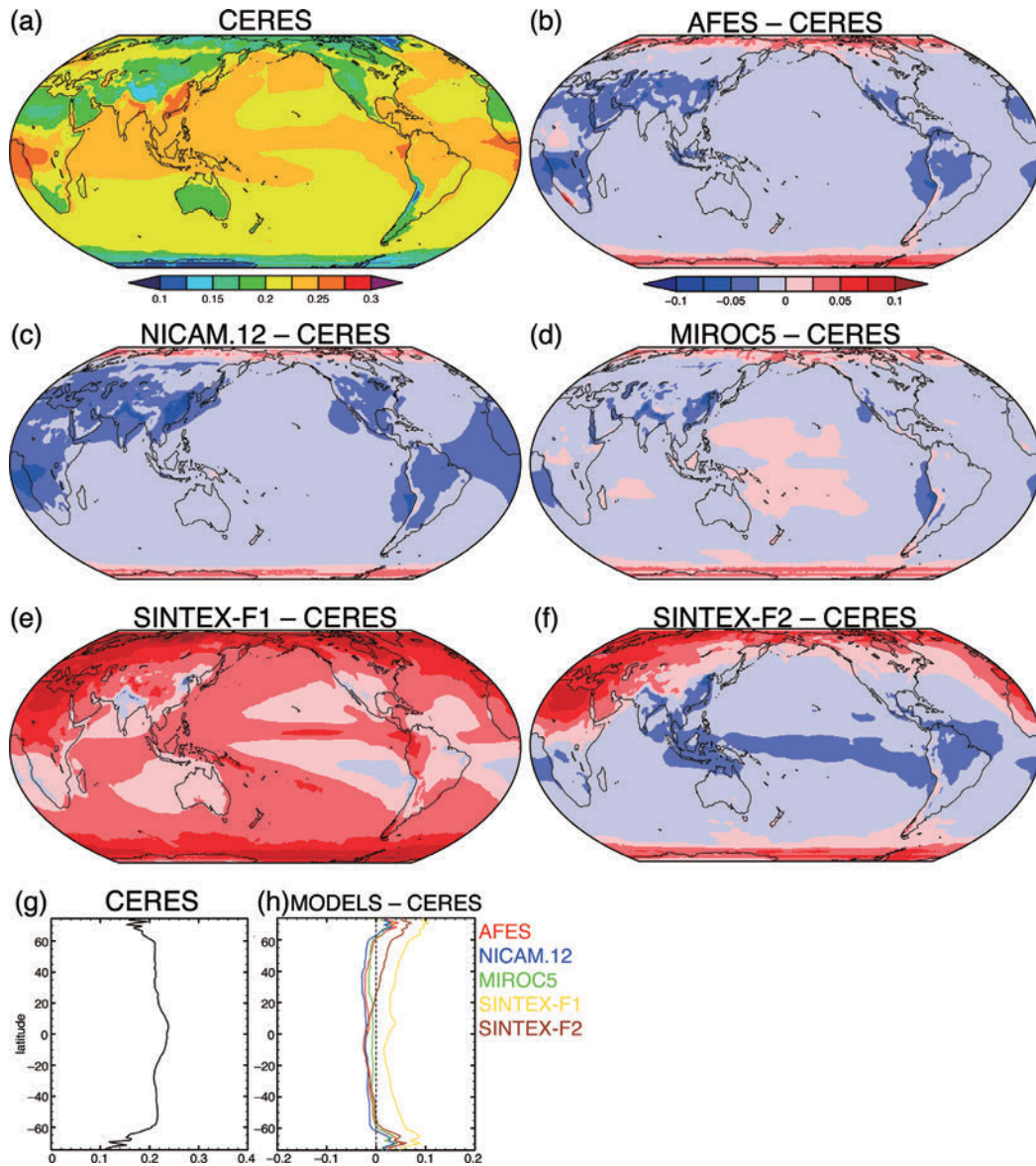


Fig. 6. Same as Fig. 5 but for atmospheric SW absorption rate.

conventional data only (JRA-55C, Kobayashi et al. 2014), which is more homogeneous dataset over a long period because it is unaffected by changes in historical satellite observing systems.

Figure 7 shows LDR24P1 in January averaged from 1982 to 2001. In JRA-55C, LDR24P1 has a peak around 165°E, 42°N with 3.5 hPa day⁻¹. AFES shows the strongest activity among the models and JRA-55C, although the distribution is similar to that of JRA-55C. MIROC5 shows the weakest activity among the models. SINTEX-F2 shows medium activity between AFES and MIROC5. These results are consistent with the results of Willison et al. (2013), i.e., that higher horizontal resolutions result in stronger explosive development. However, NICAM.12 shows weaker activity than SINTEX-F2 and AFES, despite NICAM.12 having the highest horizontal resolutions among the models. In addition, along the south coast of Japan, LDR24P1 shows weaker activity than the other models and JRA-55C.

The relative locations of the jet stream in the upper troposphere and the baroclinic zone in the lower troposphere can explain the activity differences among the models. Figure 8 shows zonal wind at 300 hPa and the horizontal gradient of equivalent potential temperature at 850 hPa. In JRA-55C, the upper jet and baroclinicity are strong over the East China Sea and the Kuroshio and Kuroshio Extension. The models (except for NICAM.12 in which the baroclinic zone is not represented) show similar baroclinic zone distributions to that of JRA-55C.

The cold SST bias in NICAM.12 (see upper-right panel of Fig. A1 in Kodama et al. 2015) causes weak baroclinicity along the southern coast of Japan. Figure 9 shows the total surface turbulent heat flux. In NICAM.12, the flux over the Kuroshio and Kuroshio Extension is the weakest among the models and JRA-55C because setting the slab ocean model with a 7-day nudging relaxation time and 15-m depth cannot represent the horizontal heat advection

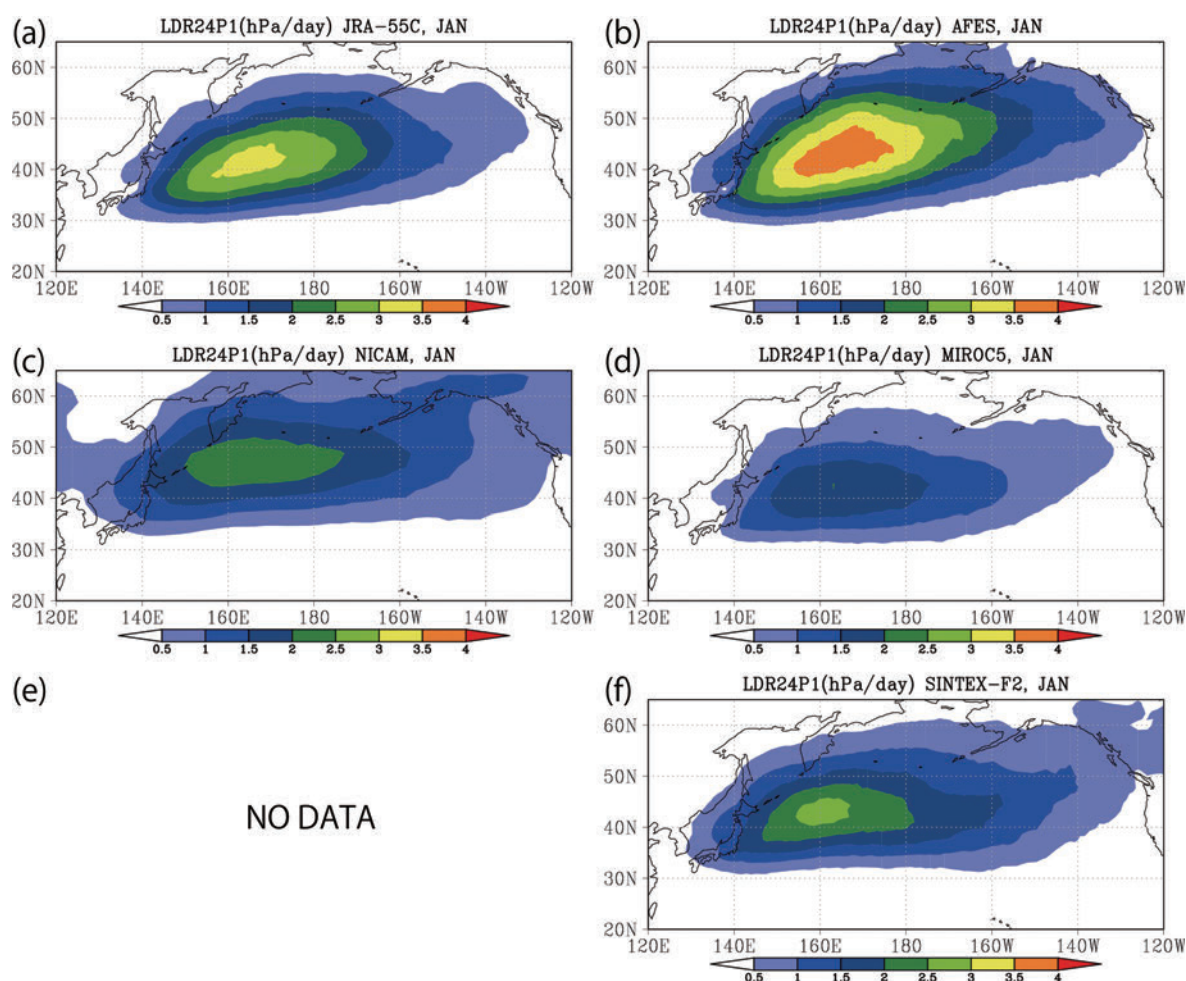


Fig. 7. Climatology of LDR24P1 in January (hPa day⁻¹). (a) JRA-55C, (b) AFES, (c) NICAM.12, (d) MIROC5, (e) SINTEX-F1 (no data), and (f) SINTEX-F2.

by the Kuroshio and Kuroshio Extension. SINTEX-F1 and SINTEX-F2, which use a shorter nudging relaxation time than does NICAM.12, present realistic distributions of SST and surface fluxes. These results are consistent with those of Kuwano-Yoshida and Minobe (2017), i.e., that the surface heat flux from the Kuroshio and Kuroshio Extension is important in representing the explosive cyclone activity in the Northwestern Pacific as well as the horizontal resolution.

4.4 Rainfall variability related to interannual climate variations

Unlike numerical weather prediction, numerical seasonal climate prediction presents challenges, particularly in the extratropics. The pioneering work by Bjerknes (1964; 1969) showed that the potential source of seasonal climate predictability is mostly related to air-sea coupled phenomena in the tropics and their teleconnection to the extratropics. It is well known that extratropical atmospheric variability

is not primarily driven by local oceanic variability (as it is in the tropics). Therefore, it is crucial to capture the atmospheric response of basin-scale air-sea coupled climate phenomena in the tropics such as the ENSO and IOD, both of which are dominant modes of interannual variability and have enormous impact on the global climate (see Yamagata et al. 2016 for a recent review).

4.4.1 ENSO

First, we focus on horizontal maps of the correlation coefficients between the interannual variations in rainfall and the Niño3.4 index. The Niño3.4 index is one of the most commonly used indices for ENSO and is defined as the SST anomalies averaged over the domain 170°E–120°W, 5°S–5°N. Figure 10 shows the correlation map in January, when the ENSO events are most mature. El Niño events bring heavy rainfall over the central/eastern tropical Pacific and less rainfall to the west of the Philippines

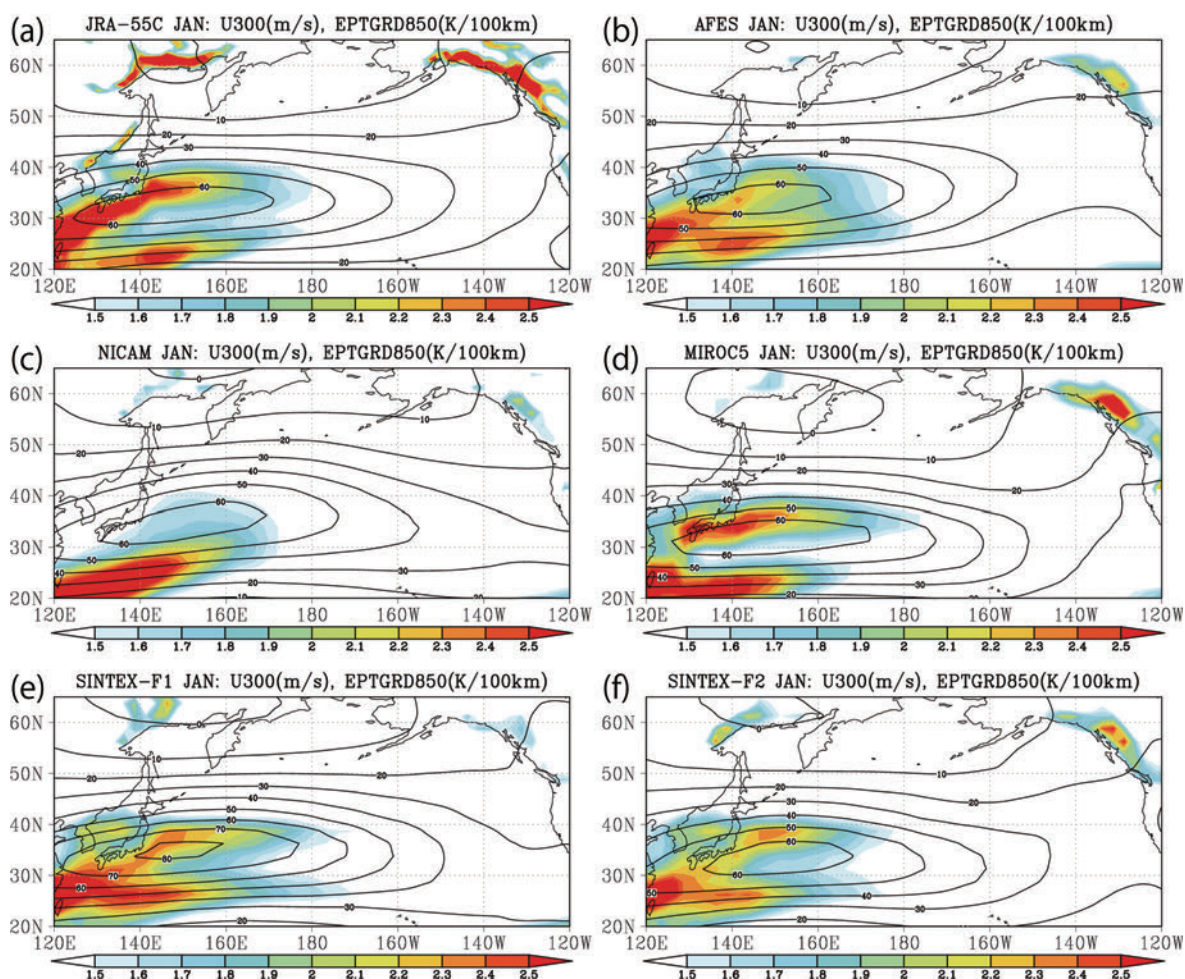


Fig. 8. Climatology of zonal wind at 300 hPa (m s^{-1} ; contours) and horizontal gradient magnitude of equivalent potential temperature at 850 hPa ($\text{K } 100 \text{ km}^{-1}$; shading) in January. (a) JRA-55C, (b) AFES, (c) NICAM.12, (d) MIROC5, (e) SINTEX-F1, and (f) SINTEX-F2.

and the Maritime Continent. As associated with El Niño (La Niña), the wetter (drier) than normal conditions over the central/eastern tropical Pacific are well captured by all models. The negative correlation over northern South America is also well captured by all models. These features of the JMIP models are consistent with the results of the AMIP runs from CMIP3 and CMIP5 (Langenbrunner and Neelin 2013). Although the observations show drier (wetter)

than normal conditions over the Maritime Continent, all models simulate the opposite condition over most of the area. This failure is again similar to the results of the AMIP runs from CMIP3 and CMIP5 (Langenbrunner and Neelin 2013). There is some discrepancy among the JMIP models in relation to convection off the east of the Philippines, which may be the origin of teleconnection patterns having a potential impact on the East Asian winter climate, including

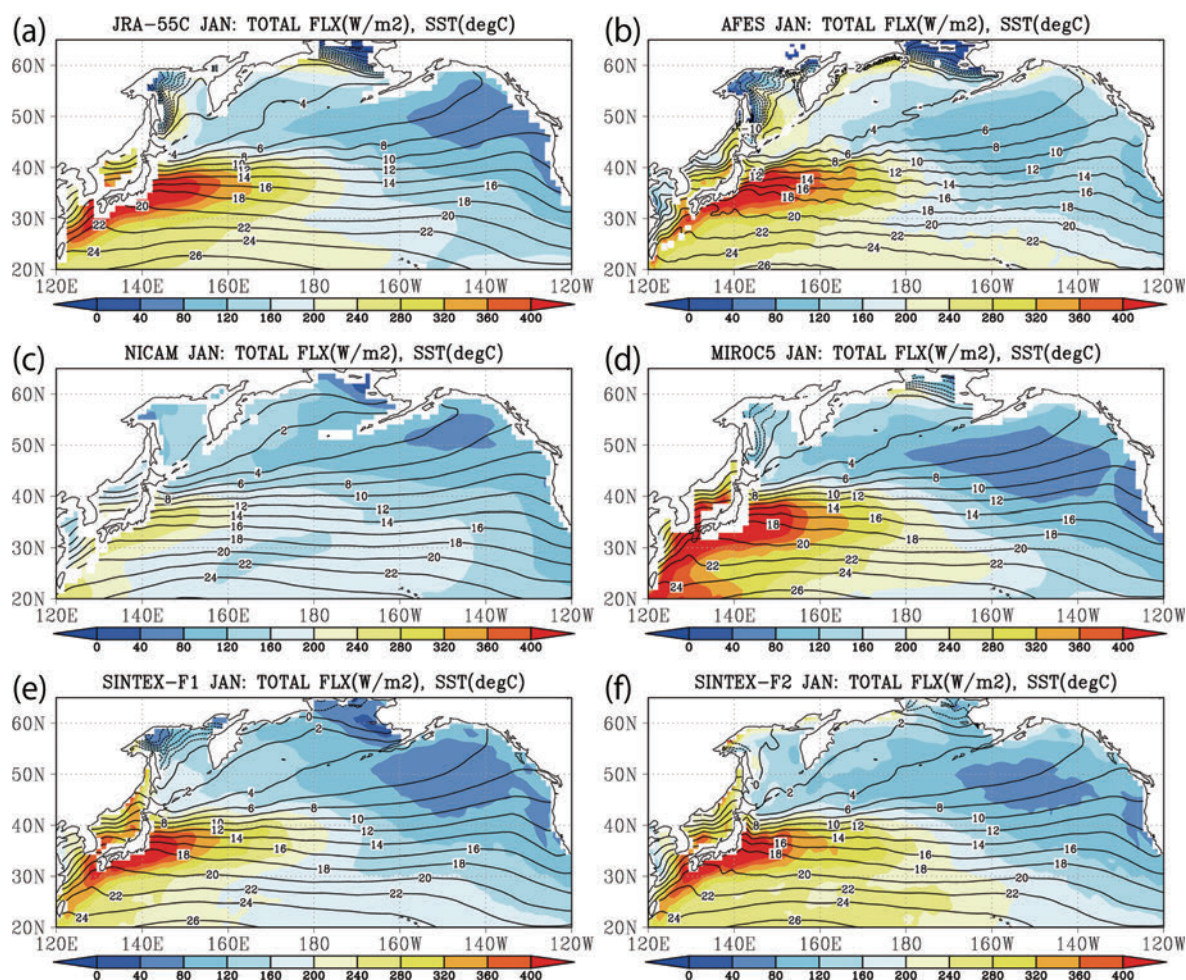


Fig. 9. Climatology of surface turbulent heat flux (W m^{-2} ; shading) and SST ($^{\circ}\text{C}$; contours). (a) JRA-55C, (b) AFES, (c) NICAM.12, (d) MIROC5, (e) SINTEX-F1, and (f) SINTEX-F2.

Table 6. (a) Regional average off the east of the Philippines (120° – 160°E , Eq. -20°N) of the correlation coefficients in January between the interannual variations in rainfall and the Niño3.4 index. (b, c, d) Same as (a), but for October with the DMI for (b) the southern part of the western pole of the IOD (50° – 70°E , 10°S –Eq.), (c) the Middle East (40° – 80°E , 20° – 40°N), and (d) eastern Australia (140° – 155°E , 40° – 10°S).

	(a) Off the east of the Philippines (120° – 160°E , Eq. -20°N)	(b) Southern part of the western pole of the IOD (50° – 70°E , 10°S –Eq.)	(c) Middle East (40° – 80°E , 20° – 40°N)	(d) Eastern Australia (140° – 155°E , 40° – 10°S)
Obs.	–0.54	0.46	0.41	–0.25
AFES	–0.32	0.28	–0.049	–0.23
NICAM.12	–0.43	–0.045	–0.061	–0.068
MIROC5	–0.18	0.091	0.24	–0.19
SINTEX-F1	–0.45	0.27	0.040	–0.051
SINTEX-F2	–0.45	0.38	0.31	–0.32

Japan (Wang et al. 2000; Watanabe and Jin 2003). Therefore, its skillful simulation is crucial in predicting boreal winter conditions in East Asia. As shown in Table 6a, the drier (wetter) than normal conditions to the east of the Philippines associated with El Niño (La Niña) are well captured by SINTEX-F1, SINTEX-F2, and NICAM.12, whereas they are slightly underestimated by AFES and MIROC5. This may be partly due to the similarities for excesses of tropical North and South Pacific precipitation in AFES and MIROC5 as highlighted in Fig. 2.

4.4.2 IOD

Next, we focus on the IOD mode. The IOD is normally characterized by dipole-type SST anomalies: cooling/warming of SSTs in the eastern/western tropical Indian Ocean as the positive event, and the reversed sign as the negative event. Associated with these SST anomalies, the convective region that is usually situated over the eastern Indian Ocean warm pool shifts to the west, leading to heavy rainfall over East Africa and severe droughts over the Indonesian region. Figure 11 is same as Fig. 10, but for October with the IOD Mode index (DMI), which is defined as the SST anomaly difference between

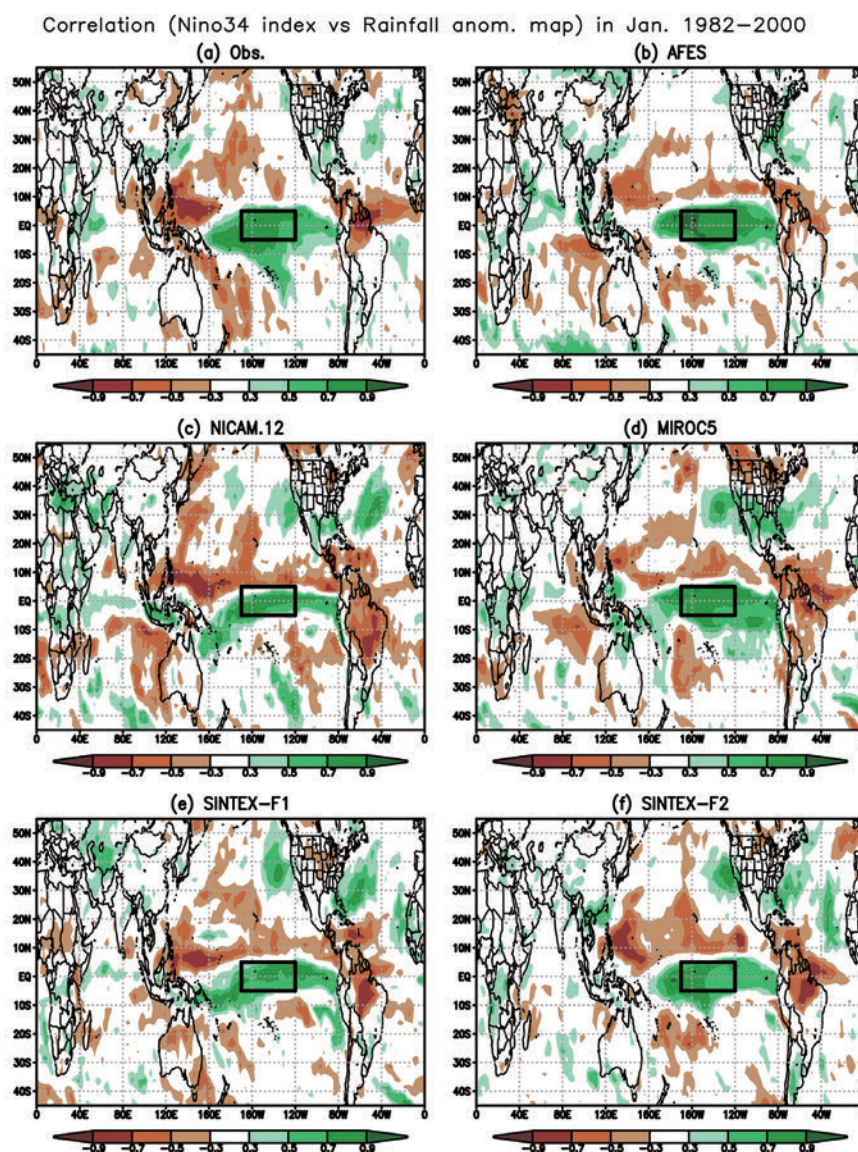


Fig. 10. (a) Horizontal maps of correlation coefficients between the Niño3.4 index and interannual variations of rainfall in January. The analysis period is 1983–2000. The Niño3.4 is from OISSTv2, and the rainfall is from GPCP (version 2.3). The Niño3.4 region is shown by a black rectangle. (b, c, d, e, f) Same as (a), but for rainfall simulated by AFES, NICAM.12, MIROC5, SINTEX-F1, and SINTEX-F2, respectively.

the western pole off East Africa (50°E–70°E, 10°S–10°N) and the eastern pole off Sumatra (90°E–110°E, 10°S–Eq) (Saji et al. 1999). Negative correlation coefficients for the eastern pole and the Maritime Continent, indicating the drier (wetter) than normal conditions associated with the anomalous cooling (warming) of SSTs, is well captured (but slightly underestimated) by all models. However, none of the models successfully simulate the broad domain of positive correlation centered on the western pole of the IOD. Namely, the wetter (drier) than normal conditions associated with the anomalous warming (cooling) of SSTs in the western pole during the positive (negative) event are underestimated

by all models. The positive correlation over the southern part of the western pole is better simulated by AFES, SINTEX-F1, and SINTEX-F2 relative to NICAM.12 and MIROC5 (Table 6b). The positive correlation coefficients over the Middle East are captured by SINTEX-F2 and MIROC5, whereas AFES, NICAM.12, and SINTEX-F1 underestimated these coefficients (Table 6c). The negative correlation over eastern Australia is well captured by SINTEX-F2 and AFES but not by the other models (Table 6d). This is very important for Australian winter wheat yields (Yuan and Yamagata 2015). Overall, SINTEX-F2 best captures the correlation distribution associated with

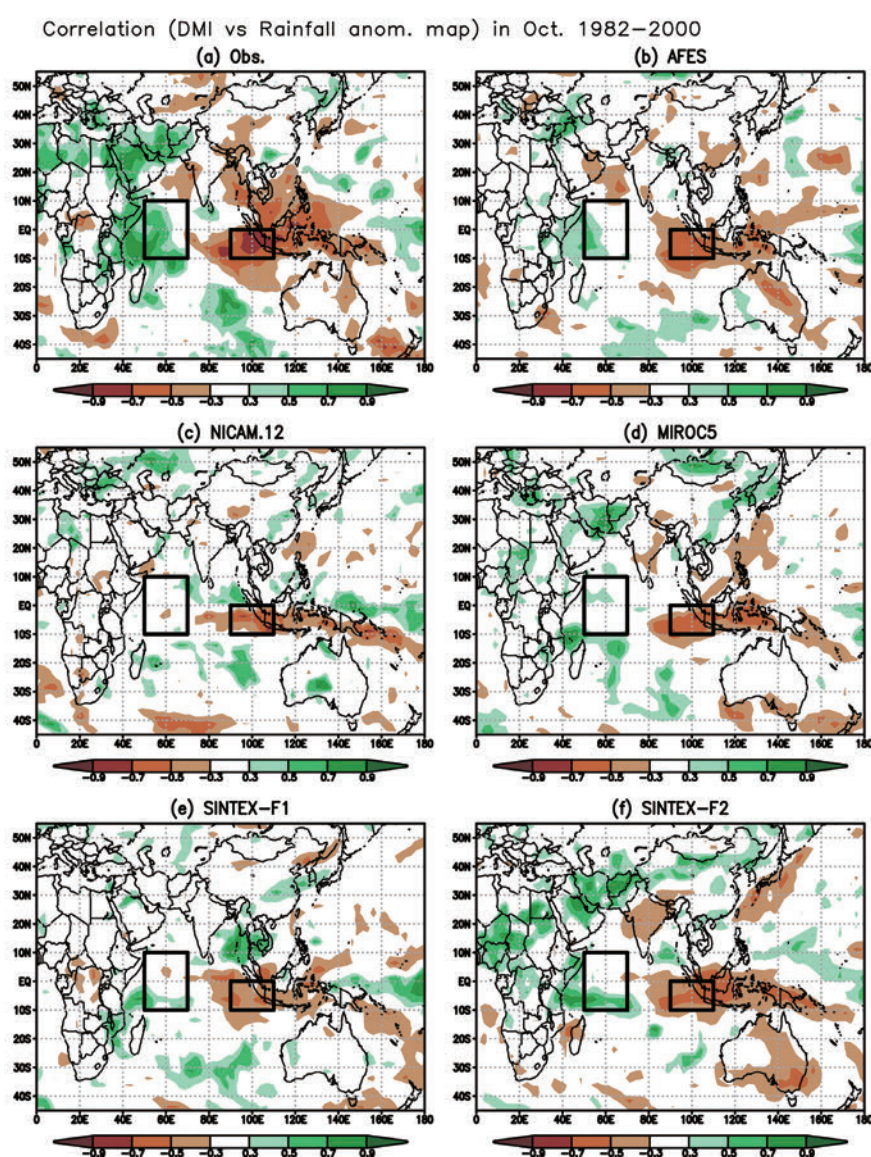


Fig. 11. Same as Fig. 10, but for the Indian Ocean Dipole Mode index (DMI), which is defined by Saji et al. (1999) as the SST anomaly difference between the western pole off East Africa (50°E–70°E, 10°S–10°N) and the eastern pole off Sumatra (90°E–110°E, 10°S–Eq). The analysis month is October.

the IOD. This may be partly due to that the east-west dipole structure of a precipitation bias over the Indian Ocean in the mean state is small with SINTEX-F2, although the precipitation over ocean is largely overestimated relative to the other models (Fig. 2).

In this subsection, we have shown that the tropical rainfall pattern associated with ENSO is relatively well captured by all models. The strong agreement among the models may provide a good predictor for ENSO teleconnection (Langenbrunner and Neelin 2013). However, simulating the IOD's response remains a challenge for all models. As far as we know, there is no study investigating IOD teleconnection using the AMIP runs from CMIP3 and CMIP5. More effort is necessary to improve the remote responses of the simulated convection to the IOD.

4.5 Asian monsoon

Monsoons are a basic variability on the Earth and are associated with seasonal cycles and land-ocean distribution; therefore, they are characterized by strong regionality. The Asian summer monsoon has the largest horizontal scale and most significant impacts on world weather and climate. In fact, it provides a basic environment for tropical cyclone genesis in the western North Pacific (Holland 1995). Because of its regionality coupling with multi-scales and its multi-process nature (e.g., air-sea-land interactions), the accurate simulation and prediction of monsoons are still challenging even for the latest climate models (Sperber et al. 2013). In this subsection, we examine the climatology and interannual variability of the Asian summer monsoons in the five JMIP models to understand the commonality and diversity across aspects that are reproduced both poorly and relatively well. Here, we use all the available data for each JMIP model: 1979–2008 for MIROC5, NICAM.12, and JRA-55, 1982–2001 for AFES, 1982–2013 for SINTEX-F1, and 1982–2014 for SINTEX-F2.

Figure 12 shows the June–September mean lower tropospheric zonal wind in the objective analysis (JRA-55) and in the simulations. The intensification of the westerlies along the southern edge of the warm air mass over the Eurasian continent and its eastward seasonal extension is a typical feature of the Asian summer monsoon. The westerlies exhibit a peak over the western Indian Ocean and extend southeastward along the southern edge of the Pacific high over the central Pacific. All the models successfully

simulate the westerlies over the tropical Indo-Pacific domain, although they have a tendency to overestimate the Pacific high. Close examination finds that the pattern of the wind field is better simulated in AFES and SINTEX-F2, whereas the strength of the peak westerlies over the western Indian Ocean are excessive in AFES, MIROC5, and SINTEX-F2 (Figs. 12a, b, d, and f). In NICAM.12 and MIROC5, the axis of the peak westerlies shows a northward bias over the Indian subcontinent and the Bay of Bengal (Fig. 12c, 12d). In SINTEX-F1, the westerlies intrude farthest over the western Pacific along approximately 15° N, with weak zonal contrast (Fig. 12e). This tendency is improved in SINTEX-F2 (Fig. 12f). NICAM.12 is distinct from the other models in that the meridional gradient of the Pacific high is very pronounced, yielding overestimation of trade wind (10°N–20°N) and westerly jet (40°N–60°N) over the central Pacific, and strong zonal confluence over the western North Pacific to the north of the Philippines (Fig. 12c).

Precipitation is another important parameter of the Asian summer monsoon. As a comparison with GPCP, Fig. 13 shows the June–September mean precipitation corresponding to Fig. 12. At a glance, the precipitation pattern is poorly reproduced compared with the flow pattern, implying difficulties in properly simulating moist processes, especially the nonlinear feedback between convection and circulation; intensification of vertical and horizontal circulation by latent heat release and invigoration of moist convection by the intensified circulation through enhancement of moisture convergence. This feedback should largely depend on the model physics, such as convective parameterizations. None of the models successfully capture the precipitation peak to the southeast of the Philippines (120°E–150°E, 5°N–20°N; Fig. 13a), which is closely related to tropical cyclogenesis in the western North Pacific. A common bias in the domain-mean overestimation of the precipitation amount is also noticeable. The tendencies (biases) in the individual models are generally consistent with the flow tendencies (biases) in Fig. 12. In AFES and MIROC5, precipitation peaks are formed over the Arabian Sea and the Bay of Bengal (Fig. 13b, 13d), reflecting the strong westerlies (Fig. 12b, 12d). Compared with GPCP, precipitation peaks over India are displaced northward in NICAM.12 and MIROC5 (Fig. 13a, 13c, 13d). SINTEX-F1 produces a large amount of precipitation to the east of the Philippines (Fig. 13e), which is collocated with the intrusion

of westerlies in that region (Fig. 12e). SINTEX-F2 produces a realistic precipitation pattern over the Indian Ocean and equatorial central Pacific, albeit with the greatest excess in terms of total amount (Fig. 13f). In all of the simulations, precipitation along the southern edge of the Pacific high partly accounts for the excessive precipitation amount over the domain.

Next, the seasonal march of the Asian monsoon is investigated by monsoon indices based on the 850-hPa zonal wind (Wang et al. 2001), i.e., the western North Pacific monsoon index (WNPMI), which is defined as the difference between the average for 100°E–130°E, 5°E–15°N and the average for 110°E–140°E, 20°N–30°N; and the Indian monsoon index (IMI), which is defined as the difference between the average for 40°E–80°E, 5°N–15°N

and the average for 70°E–90°E, 20°N–30°N (green boxes in Fig. 12). These indices represent the major sub-system of the Asian Monsoon. Figure 14 shows the annual cycle of the WNPMI and IMI as calculated from the climatological winds. The IMI sharply increases from April to May, its sign is reversed and it almost reaches its annual peak value in June. The IMI maintains its peak strength until August, and recedes thereafter (Fig. 14a). The sharp increase in early summer, namely, the onset of the monsoon, is successfully simulated in all the JMIP models. The difference in IMI strength among the models during the mature period (June–August) is attributable to the biases in the flow pattern, as seen in Fig. 12. The WNPMI shows a more gradual increase during April–August, reversing its sign in June and reaching its peak intensity in August. NICAM.12 simulates

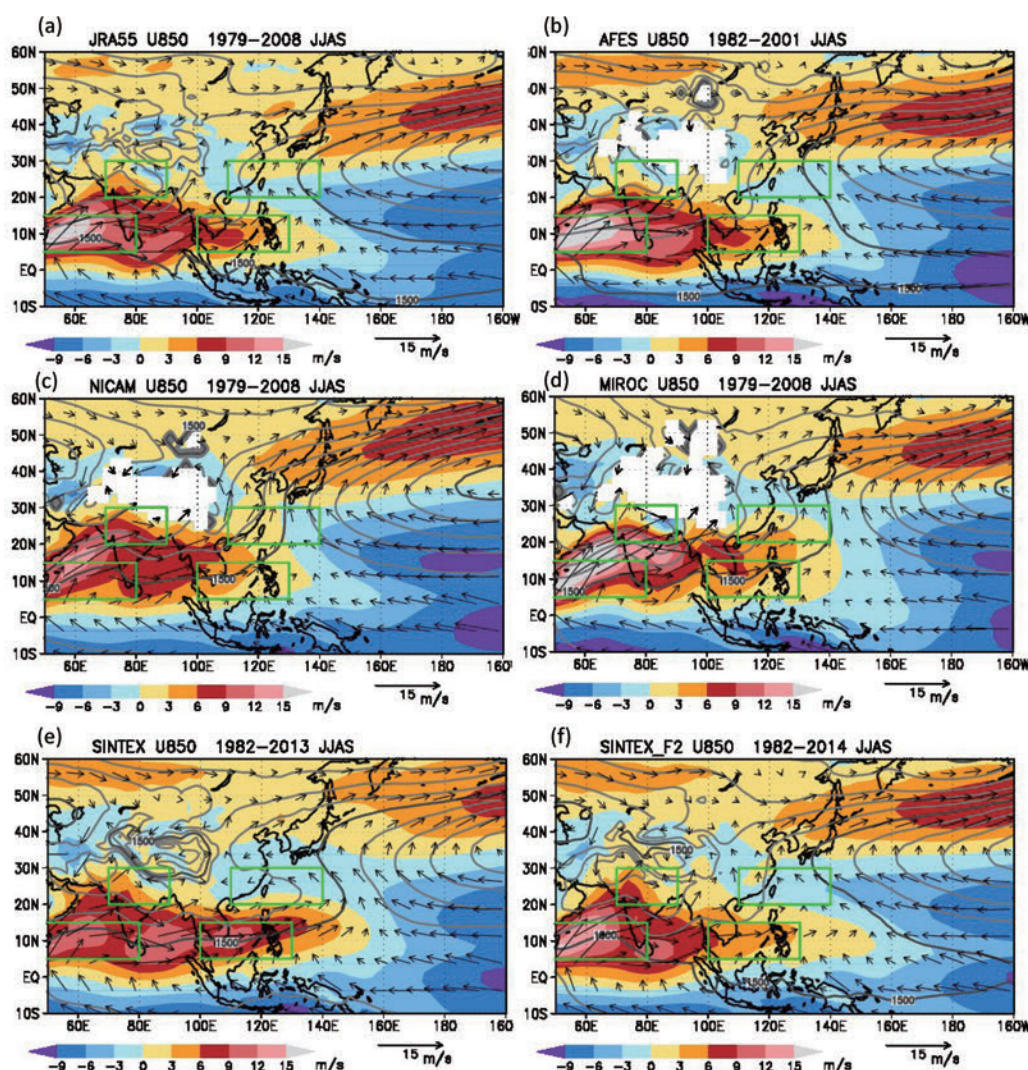


Fig. 12. Climatology of 850-hPa zonal wind (colors) and geopotential height (contours) averaged across June–September in (a) JRA-55, (b) AFES, (c) NICAM.12, (d) MIROC5, (e) SINTEX-F1, and (f) SINTEX-F2. The green boxes indicate the area considered for the monsoon indices.

the evolution of the WNPMI (from winter to summer) relatively well, with a deficit in peak intensity due to the northward displacement of the westerly axis (Fig. 14b; Kodama et al. 2015). AFES, SINTEX-F1, and SINTEX-F2 show earlier development of the western North Pacific monsoon in June, which is close to the onset of the Indian monsoon. The large WNPMI in SINTEX-F1 is associated with the robust westerlies in this domain (Fig. 12e), which become established earlier and are sustained longer than in the other models and in JRA-55. After June, this bias is reduced in SINTEX-F2, with the best performance occurring during the mature period of the summer monsoon (Fig. 14b). The strong performance of SINTEX-F2 is consistent with its good representation of the seasonal mean field (Fig. 12f). The failure of MIROC5 to simulate the seasonal cycle of WNPMI is attributed to the meridionally expanded pattern

of the westerlies (Fig. 12d). Across the JMIP models (except for SINTEX-F1) and JRA-55, the seasonal cycle of the average 850-hPa zonal wind over the western North Pacific (Fig. 14c) shows good agreement. In summary, the basic aspects of the climatology of the Asian summer monsoon are reasonably well simulated in all the models in terms of the lower tropospheric circulation, both in relation to flow pattern and temporal evolution; however, there also exist individual biases that should be handled cautiously. As to the moist convection associated with the monsoon, the biases are far more severe and diverse, presumably reflecting the differences in physical parameterizations.

Finally, we present a few remarks on interannual variability and discuss the impact of SST forcing on the Asian summer monsoon. Figure 15 shows the time series of the WNPMI anomalies in each model. The correlation

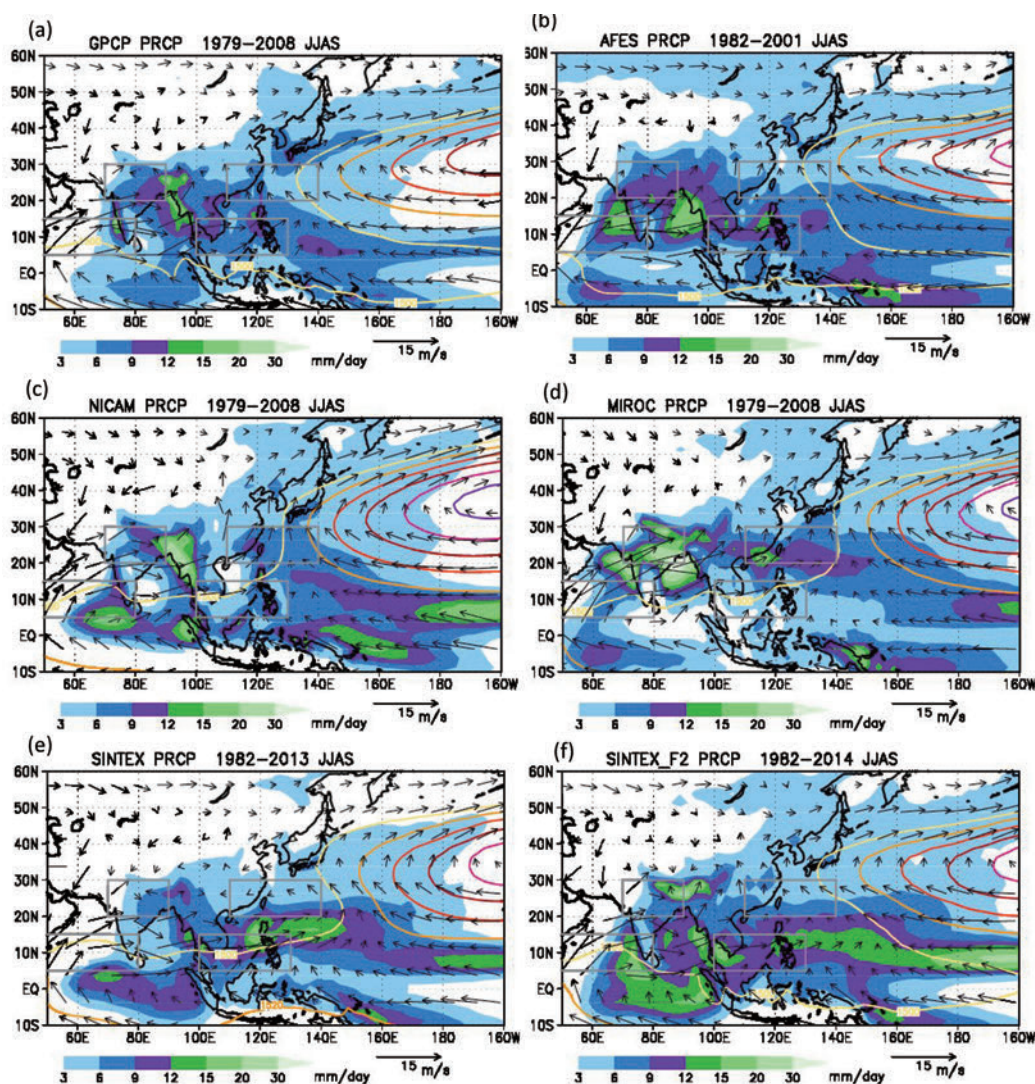


Fig. 13. Same as Fig. 12 but for precipitation rate (colors). GPCP precipitation is shown in (a).

with the time series in JRA-55 does not seem to be strong, especially after 2000. However, in the years of low WNPMI

(i.e., 1983, 1988, and 1998), almost all of the models show negative anomalies. These years correspond to the post El

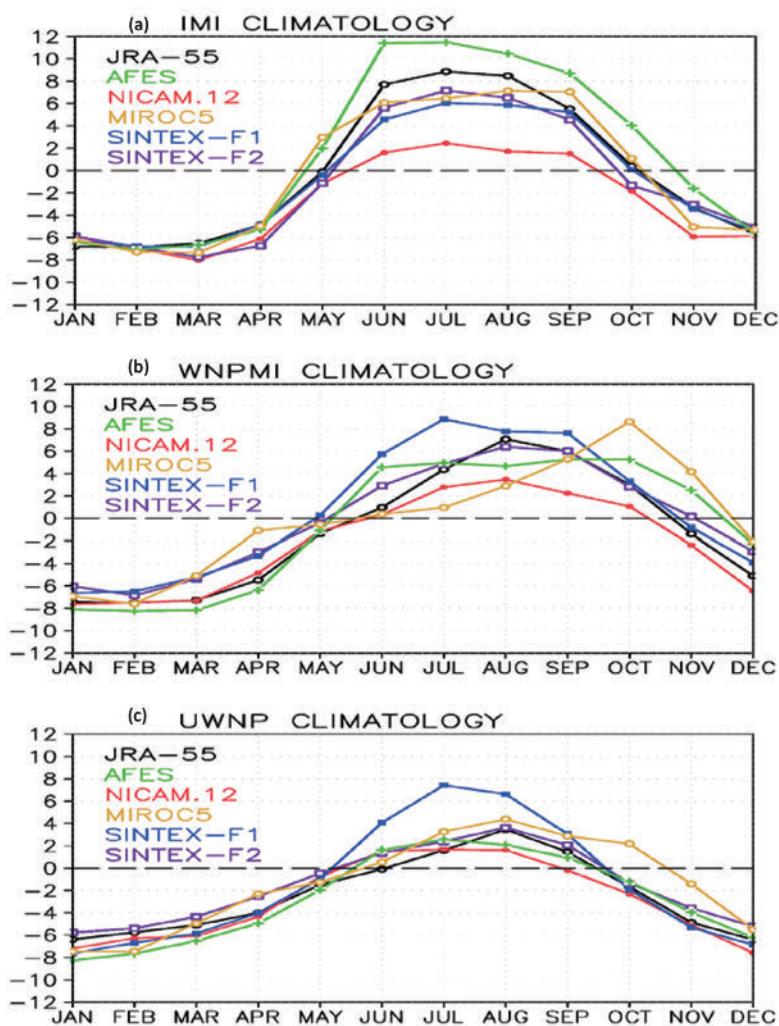


Fig. 14. Climatological annual cycle of the (a) Indian and (b) western North Pacific monsoon index (Wang et al. 2001), and (c) the 850-hPa zonal velocity averaged in the 110°E–140°E, 5°N–20°N domain.

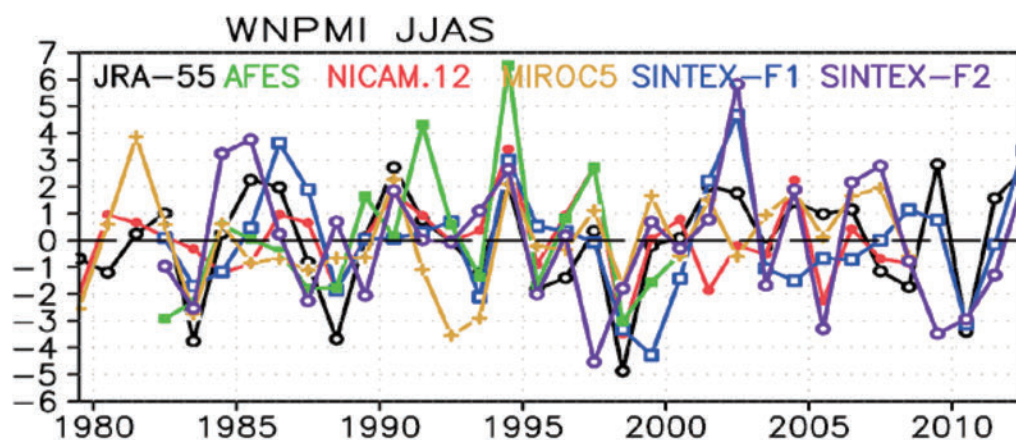


Fig. 15. Interannual variability of the western North Pacific monsoon index averaged over June–September. Anomalies from the climatology in each model are plotted.

Niño summer. It is generally known that in these years the Indian Ocean tended to be warm and anticyclonic circulation

anomalies appeared over the tropical western Pacific as an atmospheric response (Xie et al. 2009). In fact, the SST

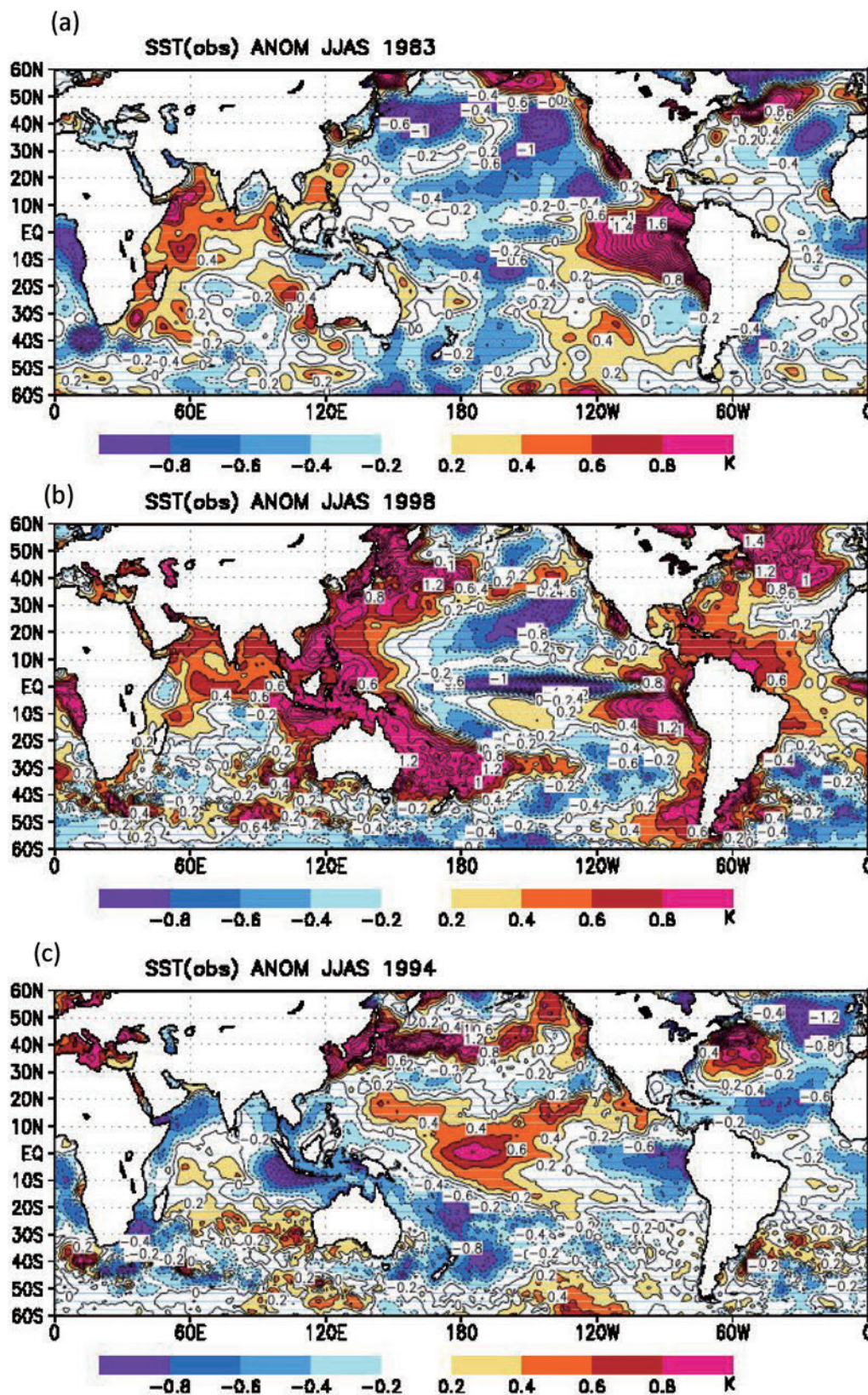


Fig. 16. Sea surface temperature anomalies averaged over June–September in NOAA OISST (1.0°) in (a) 1983, (b) 1998, and (c) 1994.

anomalies in these years (Fig. 16a, 16b) are positive over the northern Indian Ocean. In contrast, all of the models simulate the stronger WNPMI in 1994. In this year, the SST had a cold anomaly over the eastern Indian Ocean and a warm anomaly over the equatorial central Pacific (Fig. 16c), which may have enhanced cyclonic circulation over the tropical western Pacific under the influence of an extremely strong positive IOD event (Guan and Yamagata 2003; Saji et al. 1999) and positive Pacific meridional mode (Chiang and Vimont 2004). These results suggest control of local and remote SST forcing over the interannual variability of the monsoon circulation over the western North Pacific, but there is a fundamental limitation in the AMIP-type setup (Wang et al. 2005). Well-posed and systematic investigations are warranted to better understand and properly simulate the relationship between the Asian monsoon and SST forcing, where moist convection plays a key role.

5. Summary and future perspective

JMIP offers a first opportunity to systematically compare the climatology and variability simulated by the five AGCMs (AFES, MIROC5, NICAM.12, SINTEX-F1, and SINTEX-F2) developed and/or used in JAMSTEC. The outputs of the climate simulations are analyzed using subjectively chosen metrics, including climatological means of precipitation, the zonal mean basic state, radiation, explosive extratropical cyclones, and the Asian summer monsoon, and their variability associated with the ENSO and IOD mode. Although some or most of the dynamical and physical packages in the JMIP models have been developed independently, similar model biases are found among the models. In particular, AFES and MIROC5 have many similarities in terms of the simulated climate, e.g., precipitation pattern, zonal mean basic state, atmospheric shortwave absorption, Asian summer monsoon, and precipitation variability associated with the ENSO. This was noted despite the fact that the physics packages, i.e., cumulus convection, stratiform clouds, turbulent schemes, treatment of aerosols, and horizontal and vertical resolutions, are all different. Similarities are also found between SINTEX-F1 and SINTEX-F2. Although SINTEX-F2 systematically overestimates mean precipitation, it better simulates the variability in precipitation associated with the ENSO and IOD mode than does SINTEX-

F1. In relation to the performance of the climatology, NICAM.12 tends to show differences from the other models, e.g., in the precipitation, explosive extratropical cyclones, and the Asian summer monsoon onset. Our analysis revealed the challenges faced by all the models in simulating climatology and the variability of the regional precipitation patterns, especially in the Asian monsoon regions. These challenges may be linked to the difficulties in simulating the teleconnection pattern by the ENSO and IOD mode around the Maritime Continent. In addition, large differences in atmospheric SW reflection are found among the models, which are likely related to cloud characteristics (amount and optical thickness). Understanding and improving these model biases are key issues for all the modeling groups. Note that the observed SST boundary conditions were directly imposed in AFES and MIROC5, whereas they were imposed via nudging techniques in NICAM.12, SINTEX-F1, and SINTEX-F2 (Table 1). As discussed in Section 4.3, the large difference between the observed and modeled SST in NICAM.12 leads to weaker simulated explosive extratropical cyclones. Meanwhile, SST nudging improves the distribution of tropical precipitation in NICAM.12 (Kodama et al. 2015), suggesting the importance of understanding atmosphere-ocean coupling when reconsidering the methodologies for imposing SST boundary conditions in an AGCM.

To some extent, the similarities and differences among the JMIP models overlap with those among the CMIP5 models. Both the CMIP5 and JMIP models (note that MIROC5 belongs to both CMIP and JMIP) tend to underestimate the low-level cloud fraction (Wang and Su 2013) and simulate excessive tropical precipitation and a double ITCZ (Hirota and Takayabu 2013). They also struggle to simulate climatology and the seasonal march and interannual variability of the Asian summer monsoon (Sperber et al. 2013). In this context, JMIP can be considered a simple and in-depth version of CMIP with which to investigate the mechanisms of the model biases in a more process-oriented manner. For example, details of the cloud and precipitation process can be investigated using high-frequency, full vertical resolution JMIP data with the aid of satellite simulator (e.g. Suzuki et al. 2015 for warm rain process; Bodas-Salcedo et al. 2008 for global cloud). The results that AFES and SINTEX-F2 relatively better simulates climatology of Asian summer monsoon (Section 4.5) might give us a hint to improve climatology

in the other models, though further in-depth investigation is needed to understand the reasons for the difference of the simulated Asian summer monsoon among the models. We also point out the usefulness of the JMIP framework for exchanging information on model development and performance; this will aid the promotion of our overall scientific effort and assist in the potential future sharing some part of the physics packages; thus reducing the costs of multi-model development.

We would like to present this study as the first phase of JMIP (JMIP1), in which we have focused on the climatological aspects simulated by the atmospheric models. Another concern that modeling groups might share relates to model performance in weekly-to-seasonal scale weather forecasting. Recently, Nakano et al. (2017) conducted the Global 7 km mesh nonhydrostatic Model Intercomparison Project for improving TYphoon forecast (TYMIP-G7) to investigate the impact of using high resolution models on forecasting skill, in which NICAM, MSSG, DFSM (a double Fourier series model developed in MRI), and GSM were included. Considering that JAMSTEC not only possesses a wide variety of GCMs and the Earth Simulator, but also observation vessels, intensive collaboration between modeling and observation groups could also be an important future step for pushing our science forward. Such a collaboration has already been conducted in the Cooperative Indian Ocean Experiment on Intraseasonal Variability in the Year 2011 (CINDY2011)/Dynamics of the MJO (DYNAMO) intensive observation, in which NICAM was routinely performed to provide near real-time forecasting (Nasuno et al. 2017).

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