

Chapter 2

Studies on High-Resolution Atmospheric and Oceanic General Circulation Models

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2.1 Introduction

Atmospheric general circulation models (AGCMs) and oceanic general circulation models (OGCMs) are the most important components of a climate/earth system model. The development of higher resolution AGCMs and OGCMs is one of the most active areas of study in the model development field. At the Institute of Atmospheric Physics (IAP), Chinese Academy of Sciences (CAS), over 30 years have passed since the IAP's AGCM was first designed in 1982, during which time the model has been continually developed.

AGCMs can be divided into two parts, each solving different aspects of the problem at hand: the dynamical core and the physical processes parameterizations. The dynamical core solves the Navier–Stokes equations of the atmosphere using different numerical algorithms and physical processes parameterizations describe the statistical effect of unresolved subgrid-scale processes. The stability and time consumption of computation has become increasingly important to achieve a high-resolution AGCM. Wehner et al. (2008) reported that the dynamics and physics, respectively, consume 57 and 41 % of the computation time in the Community Atmosphere Model (CAM) with a horizontal resolution of 200 km. When the horizontal resolution is increased to 50 km, the percentages shift dramatically to 81 and 14 %. Therefore, designing a more efficient, stable, and accurate dynamical core is a critical aspect when developing high-resolution AGCMs.

On the other hand, in the climate system, radiative processes are fundamental, playing central roles in many proposed climate change mechanisms. Therefore, the parameterizations of radiation and related processes are the dominant physical processes in AGCMs. However, uncertainties in radiative transfer calculations represent a significant limitation on progress in climate modeling. IPCC (2013) reported that cloud–aerosol–radiation interaction is one of the most uncertain parts in current climate models, reducing the predictive abilities of models. How to objectively estimate the uncertainties caused by cloud, aerosol, or radiation

calculations, along with those describing their interaction and feedback mechanisms, has yet to be fully investigated. Therefore, a self-consistent cloud–aerosol–radiation (CAR) ensemble modeling system has been built (Liang and Zhang 2013; Zhang et al. 2013a). In the CAR system, all original radiation transfer schemes have been modified to decouple the treatments for cloud/aerosol physical and optical properties, including those for cloud subgrid-scale structures, and then consistent applications of those properties into different radiation transfer schemes have been carried out. The CAR system is a useful tool that enables us to further understand the roles of cloud/aerosol/radiation, especially under partly cloudy sky.

2.2 Objectives

In this subproject, three main targets of studies were:

- (1) To design a dynamical core of a high-resolution general circulation model (GCM) and to develop a high-resolution AGCM based on IAP AGCM4.0. The horizontal resolution would be about 50 km. To complete 30-year Atmospheric Model Intercomparison Project (AMIP) simulations with this high-resolution model and evaluate its performance.
- (2) To evaluate and improve the CAR ensemble modeling system and to incorporate the improved CAR system into the newly developed IAP high-resolution AGCM.
- (3) To develop a high-resolution GCM for the global ocean. The horizontal resolution would be up to 30–50 km.

2.3 Major Achievements

2.3.1 *Improvements of the Dynamical Core of the High-Resolution AGCM*

Version 4 of the IAP AGCM (IAP AGCM4.0) with a resolution $1.4^\circ \times 1.4^\circ$ has evolved from several earlier versions that have been documented in the literature (Zeng et al. 1989; Liang 1996; Zuo 2003; Zhang et al. 2013c). It has been used to simulate atmospheric circulations and climate, including climate change, monsoons, and palaeoclimatology (e.g., Dong et al. 2012; Wang and Wang 2013; Su et al. 2014).

The model uses a finite-difference scheme with a terrain-following σ coordinate (Phillips 1957). There are 26 vertical levels in IAP AGCM4.0, with the model top at about 2.2 hPa. A latitude–longitude grid with Arakawa’s C grid staggering is used in the horizontal discretization (Arakawa and Lamb 1977). A subtraction of the

standard stratification is used to reduce truncation errors, especially over regions of high terrain such as the Tibetan Plateau. The IAP transform is defined and applied to maintain the conservation of the mass and available energy. The nonlinear iterative time integration method described in Zuo et al. (2004) is used in the model. To develop the high-resolution ($0.5^\circ \times 0.5^\circ$) version of IAP AGCM4.0, some new methods are applied in the dynamical core, which we describe here.

The central difference scheme is used to discretize the baroclinic primitive equations with the Arakawa's C grid. Two calculations are applied in the model to alleviate the reduction of the physical grid spacing problem in high latitudes. Poleward of $\pm 70^\circ$, Fourier filtering is used on the tendencies of wind, temperature, and surface pressure to dampen the short longitudinal waves following Umscheid and Sankar-Rao (1971), which happen to be high-frequency waves as well. Between $\pm 30^\circ$ to $\pm 90^\circ$, the following flexible leaping-point scheme (referred to as FLP scheme) is used. In this scheme, $\partial F / \partial \lambda$ is discretized by

$$\left(\frac{\partial F}{\partial \lambda} \right)_{ij} = \alpha_j \cdot \frac{F_{i+\frac{1}{2}j} - F_{i-\frac{1}{2}j}}{\Delta \lambda} + \beta_j \cdot \frac{F_{i+\frac{3}{2}j} - F_{i-\frac{3}{2}j}}{3\Delta \lambda}, \quad (2.1)$$

where λ is the longitude and the subscripts i and j denote the i th and j th grid points for the longitudinal and latitudinal direction, respectively. α_j and β_j are coefficients dependent on latitude and satisfy $\alpha_j + \beta_j = 1$. β_j increases linearly from 0 at $\pm 30^\circ$ to 1 at $\pm 90^\circ$ (Zhang et al. 2009). Table 2.1 shows the time steps used in the dynamical core of IAP AGCM4.0 under the condition of applying the FLP scheme or Fourier filtering. We can see that if neither the FLP scheme nor the Fourier filter is used, the time step is very small at only 5 s. When only the FLP scheme is used, the time step increases to three times the original due to the three times larger zonal grid size. When both the FLP scheme and the Fourier filter are used, the time step can increase up to 1990s, which is 18 times the original, thus saving computing time.

The governing equation used in the AGCM can be divided into two parts: the advection term describing a slow process and the intergravity wave term describing a fast process. Therefore, the time-splitting method is applied to use different time steps in different terms. We define the equation

$$\Delta t_2 \equiv N \Delta t_1, \quad (2.2)$$

Table 2.1 The time step of the dynamical core in IAP AGCM4.0

Resolution	Coefficients		Filter	Time step (s)
	α_j	β_j		
$0.5^\circ \times 0.5^\circ$	1	0	No	5
	0	1	No	15
	0	1	Yes	90

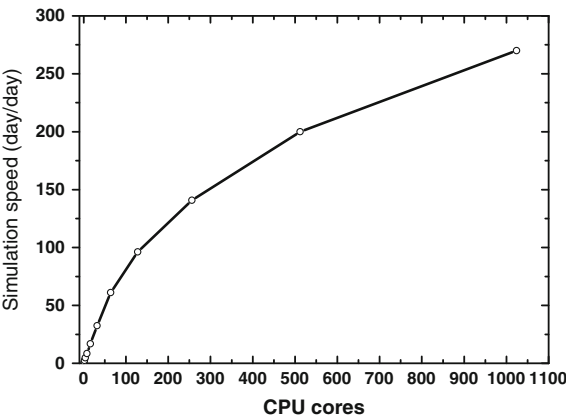
Table 2.2 Computing time of IAP AGCM4.0 using the time splitting method

$\Delta t_2/\Delta t_1$	Wall-clock time for 1 model day integration (s)	Time saving (%)
1	3270	–
5	2890	11.6
10	2550	22.0

where Δt_1 is the time step for the fast process and Δt_2 the time step for the slow process, and N is the proportional coefficient. We can see from the results presented in Table 2.2 that this method can save up to 22 % of computing time.

Parallel implementation is an important way to increase simulation speed, especially for high-resolution models. A 2D decomposition has been developed in the dynamical core of the high-resolution version of IAP AGCM4.0. In this parallel algorithm, a latitude–vertical decomposition is used for the main dynamical algorithms and a longitude–latitude decomposition is used for the semi-Lagrangian tracer transport. Halo updates are the primary multipoint interface (MPI) communications required in the computation for a given decomposition. IAP AGCM4.0’s physics is based on vertical columns, and computations are independent between columns. Thus, the parallel implementation of the physics is based on the assignment of columns to (MPI) processes. Transitions are needed from the dynamics grid decomposition to the physics grid decomposition. The model speed using this new parallel algorithm is shown in Fig. 2.1, which demonstrates that the model speed increases almost linearly as a function of core number for less than 100 cores. Above 100 cores, the model speed still increases, but acceleration slows down gradually. The parallel efficiency is about 20 % on 1024 cores and needs to be improved in the future.

Fig. 2.1 Model speed (units: model days per wall-clock days) for IAP AGCM4.0 with $0.5^\circ \times 0.5^\circ$ resolution as a function of Tianhe-1A supercomputer core numbers



2.3.2 *Sensitivity of Simulated Climate to Dynamical Cores*

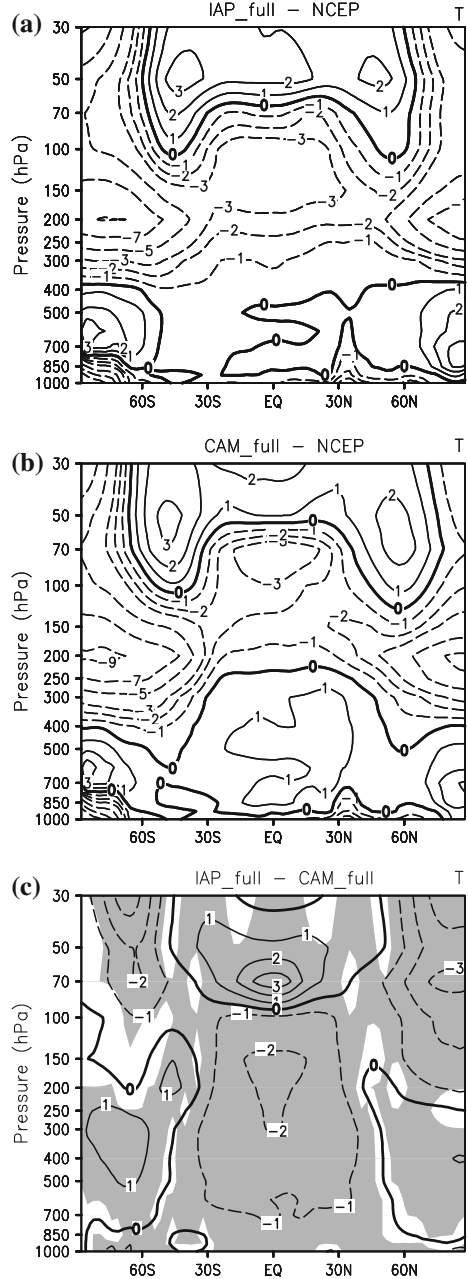
There are many different dynamical cores used to solve the same atmospheric governing equations in different AGCMs. There is therefore the need to understand and verify dynamical cores to assure the quality of model simulations (Williamson 2007). Several idealized cases have been designed to specifically test the numerical robustness of dynamical cores (Held and Suarez 1994; Boer and Denis 1997; Polvani et al. 2004; Jablonowski and Williamson 2006). These have been very valuable in understanding the sensitivity of numerical solutions to different schemes and resolutions. Most of these tests are formulated without moist physics. However, because numerical approximations are flow dependent, it is also important to understand the behavior of atmospheric dynamical cores in real settings of the atmosphere with moist physics. Accordingly, the objective of this section is to investigate the impact of the IAP AGCM dynamical core on simulated climate. We compared results from two models with the same physics package [the CAM3.1 physics package of Collins et al. (2004) and the dry physics of Held and Suarez (1994)], one using the IAP dynamical core (referred to as the IAP model), and the other using the Eulerian spectral transform dynamical core of the CAM3.1 (referred to as the CAM model). We report how the interactive physical parameterizations changed the sensitivities of simulated climate to the dynamical cores.

Two groups of experiments were carried out with IAP AGCM4.0 at a resolution of $1.4^\circ \times 1.4^\circ$ and CAM3.1 at a resolution of T85. One group was climate simulations with real moist physics and the other was Held and Suarez (1994) experiments (hereafter referred to as the H–S experiments) with idealized dry physics. In the climate simulations, IAP AGCM4.0 and CAM3.1 were both integrated for 17 years with the same physics package, and the boundary and initial conditions were interpolated from the same input data. Averages of the simulations over the last 15 years are presented. The SST and sea ice datasets were climatological monthly datasets from the global Hadley Centre Sea Ice and Sea Surface Temperature (HadISST) dataset (Rayner et al. 2003). The concentrations of greenhouse gases were held constant at their levels in 1990, with a CO_2 concentration of 355 ppmv. The primary sources of validation data were the National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) (Kalnay et al. 1996) and ERA-Interim (Dee et al. 2011) reanalyses. Both the model results and reanalysis data were truncated to T42 and 17 pressure levels for direct comparison.

In the H–S experiments, the thermal forcing was Newtonian relaxation to a specified temperature field of varying time scale with latitudes and pressure. The momentum forcing was a sink represented by Rayleigh friction in the lower troposphere. The atmosphere was integrated without moist processes, topography, land–sea contrast, and seasonal or diurnal cycle. Both models were integrated for 1220 days. By discarding 200 days, the remaining 1020 days were divided into 17 periods of 60 days. Independent 40-day samples were obtained by discarding 20 days in each period. The time–mean fields and statistics were taken from the 17 samples.

Figure 2.2a, b show the deviation of the zonally averaged temperature from the NCEP reanalysis in the models. The differences between the IAP model and the CAM3.1 simulations are shown in Fig. 2.2c. Shading indicates the differences judged statistically significant at the 0.05 level according to the Student's *t*-test

Fig. 2.2 The biases of zonally averaged 15-year annual mean temperature from **a** IAP and **b** CAM simulations in contrast to NCEP reanalysis data, and **c** their difference (IAP minus CAM). Contour intervals are 1 K between -3 K and 3 K, but 2 K beyond in **(a)** and **(b)**, and 1 K in **(c)**. Shaded areas are statistically significant at the 0.05 level according to the Student's *t*-test



based on 15 annual mean simulations. As in most AGCMs (Boer et al. 1992), a notable cold bias existed near the polar tropopause in both models. Moreover, CAM3.1 had a warm bias throughout most of the troposphere. This warm bias was systematically reduced in the model with the IAP dynamical core, especially in the tropics and midlatitudes, where the simulated temperature with the IAP core was colder than that in CAM3.1 by more than 2 K, and this difference was statistically significant. Poleward of 50°N and 50°S, the IAP model simulated a warmer troposphere than CAM3.1. Since the IAP model used polar filters, whereas the CAM model did not, it was difficult to trace the causes of the differences at high latitudes. We therefore focus on the simulation differences in the low and midlatitudes.

The observed climatological annual mean of zonally averaged transient eddy heat fluxes in Fig. 2.3c shows two peaks of poleward heat fluxes in the lower and upper troposphere at midlatitudes. Both simulations with the IAP model (Fig. 2.3a) and CAM3.1 (Fig. 2.3b) captured these two peak patterns well. Figure 2.3 indicates that the general pattern of the transient eddy heat fluxes was simulated well in both

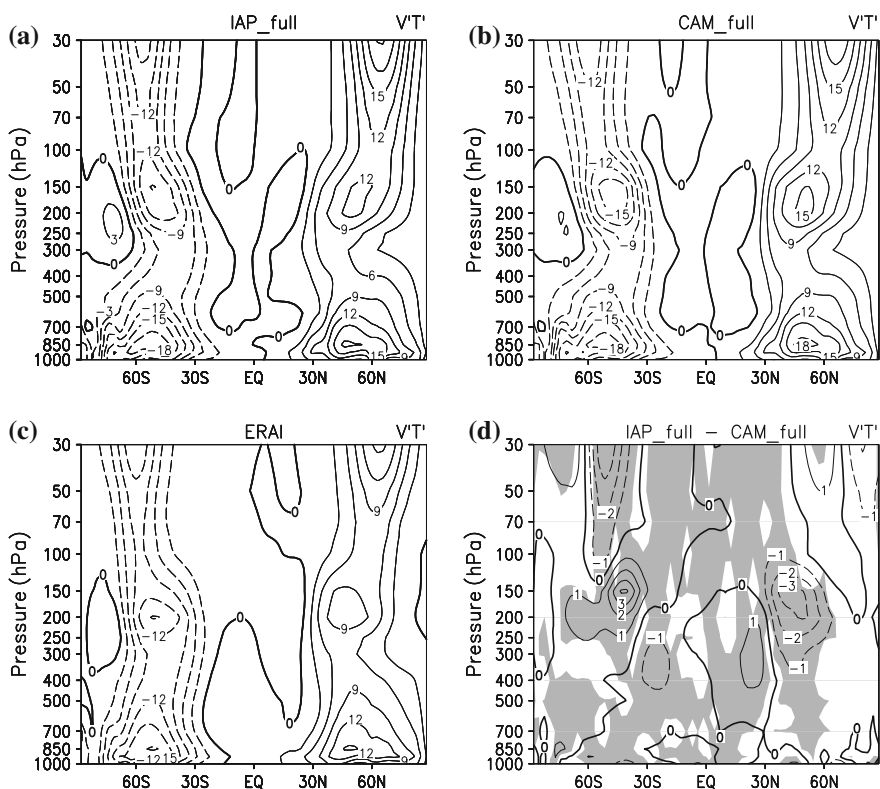
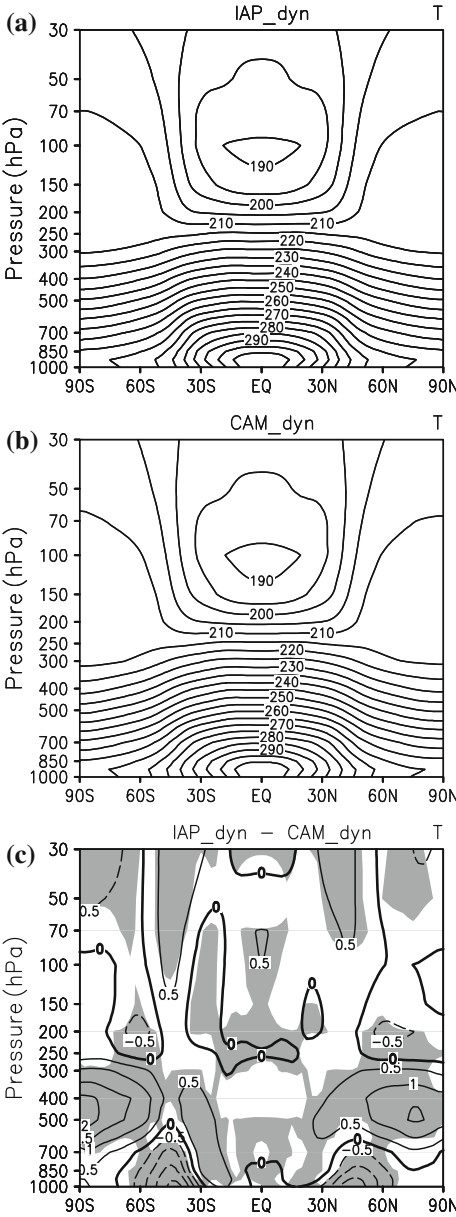


Fig. 2.3 Zonally averaged 15-year annual mean transient eddy heat flux from **a** IAP, **b** CAM, **c** ERA-Interim reanalysis data, and **d** the difference of the two simulations (IAP minus CAM). Contour intervals are 3 K m s⁻¹ in (a) and (b), 1 K m s⁻¹ in (c) and (d). Shaded areas are statistically significant at the 0.05 level according to the Student's *t*-test

the IAP model and CAM3.1, but the strength was too strong in the models. Relative to CAM3.1, Fig. 2.3d indicates weaker poleward heat transport above 500 hPa in the IAP model.

Figure 2.4 shows zonally averaged 17-period mean temperature from the H–S experiments with the IAP dynamical core (referred to as the IAP core) and that of

Fig. 2.4 Zonally averaged 17-period mean temperature from the H–S experiments with the **a** IAP and **b** CAM dynamical core, and **c** their difference (IAP minus CAM). Contour intervals are 5 K in **(a)** and **(b)**, and 0.5 K in **(c)**. Shaded areas are statistically significant at the 0.05 level according to the Student’s *t*-test



CAM3.1 (referred to as the CAM core) and the statistical significance of their difference according to the Student's t -test. The major features of the real atmosphere were reproduced with idealized forcing, as was expected. What was not expected, however, was that the simulated troposphere in the IAP core was generally warmer than that in the CAM core, which was opposite to the results found in the climate simulations shown in Fig. 2.2c.

The distribution of simulated eddy heat flux in the H-S experiments is presented in Fig. 2.5. The patterns of eddy heat flux in the H-S experiments were similar to the climate simulations, with two peaks located in the lower and upper troposphere, respectively. The difference between the two dynamical cores in Fig. 2.5c is similar to the climate simulations in Fig. 2.3d, with weaker heat transport in the IAP core. Therefore, the IAP model simulated weaker eddy activities in both models with full physics, and in the dry model, than those in CAM3.1, yet the IAP model simulated a colder troposphere in the full physics model but a warmer troposphere in the dry model than CAM3.1. With this sensitivity in mind, we next focus on why weaker eddies in the IAP model caused a colder troposphere in the model with full physics relative to CAM3.1, but a warmer troposphere in the dry model.

To aid the discussion, we write the zonally averaged atmospheric thermodynamic equation as

$$\frac{\partial \bar{\theta}}{\partial t} = -\frac{1}{a \cos \varphi} \frac{\partial \cos \varphi \bar{v} \bar{\theta}}{\partial \varphi} - \frac{\partial \bar{\omega} \bar{\theta}}{\partial p} + \bar{Q}, \quad (2.3)$$

in which φ is the latitude, v is the meridional wind, ω is the vertical pressure velocity, θ is the potential temperature, Q is the diabatic heating, and an overbar denotes zonal average. Upon vertical integration from the surface (p_s) to the tropopause (p_t), the vertical transport term on the right-hand side drops out. Using the quasi-geostrophic approximation of meridional transport as in Andrews et al. (1987), we have

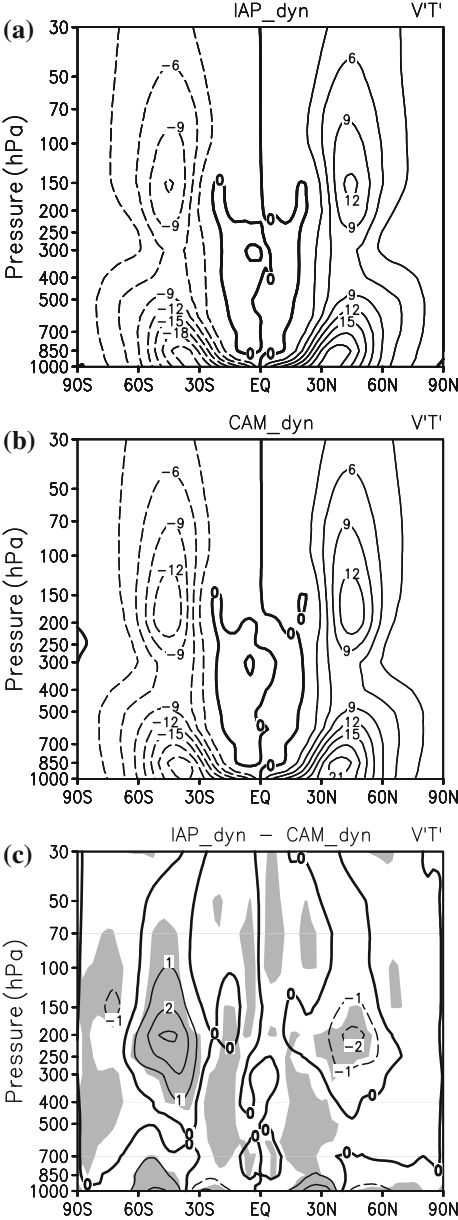
$$\int_{p_t}^{p_s} \bar{v} \bar{\theta} dp = \int_{p_t}^{p_s} \bar{v} \bar{\theta} dp + \int_{p_t}^{p_s} \bar{v}' \bar{\theta}' dp \approx \int_{p_t}^{p_s} \bar{v}_g \bar{\theta} dp + \int_{p_t}^{p_s} \bar{v}' \bar{\theta}' dp = \int_{p_t}^{p_s} \bar{v}' \bar{\theta}' dp \quad (2.4)$$

because the zonal average of the geostrophic velocity of meridional wind $\bar{v}_g$ is zero. For the dry model, Q is specified as a relaxation to the radiative equilibrium temperature $\bar{\theta}_e$, which can be schematically written as

$$\bar{Q} = -\alpha(\bar{\theta} - \bar{\theta}_e), \quad (2.5)$$

where α is a relaxation coefficient. Substituting Eqs. (2.4) and (2.5) into (2.3) and integrating over the tropics from the Equator to latitude φ_1 (say, 30°N), we get the

Fig. 2.5 As in Fig. 2.4, except for transient eddy heat flux. Contour intervals are 3 K m s^{-1} in (a) and (b), and 1 K m s^{-1} in (c)



following steady-state relationship between low-latitude temperature and eddy transport after some rearrangements:

$$\frac{1}{\cos \varphi_1} \iint_{\varphi=0, \varphi_1; p} (\bar{\alpha \theta}) a \cos \varphi d\varphi dp = \frac{1}{\cos \varphi_1} \iint_{\varphi=0, \varphi_1; p} (\bar{\alpha \theta}_e) a \cos \varphi d\varphi dp - \int_p v' \theta' |_{\varphi_1} dp. \quad (2.6)$$

In the IAP dry model, the last term on the right-hand side is smaller than that in CAM3.1 because of weaker eddy activities. This explains why the IAP core simulated a warmer troposphere in low latitudes relative to the CAM core.

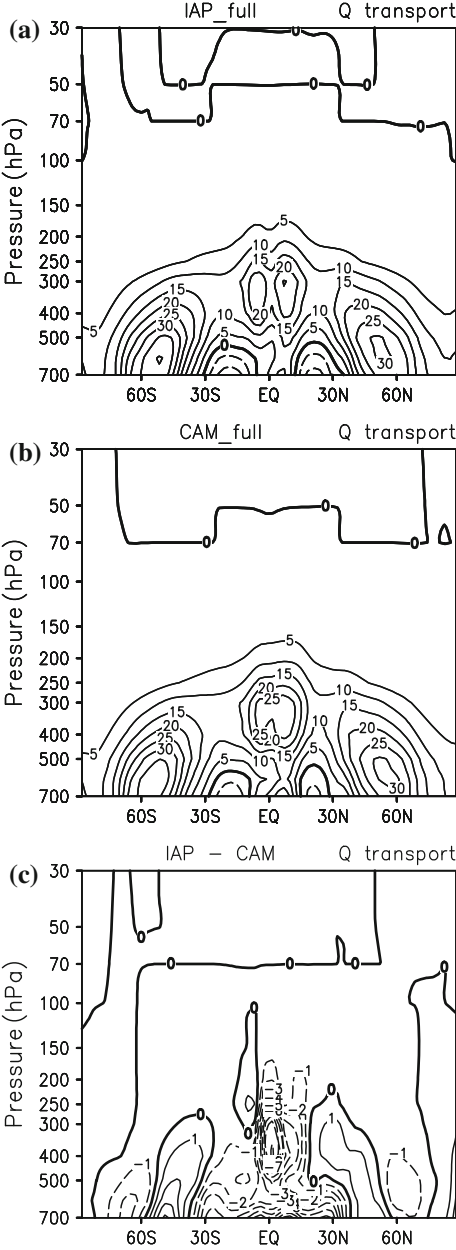
When moist physics is included with topography, the diabatic heating is no longer in the simple form of Eq. (2.5). The weak eddies with the IAP core can lead to weaker transport not only in heat, but also in moisture. Figure 2.6a, b show the 3D transport tendencies of water vapor in the two models. Because water vapor decreases exponentially with height, only values in the middle and upper troposphere are shown. Upward and poleward transport of water vapor were seen in both models. Figure 2.6c shows the difference between the two models. It is seen that the transport was systematically weaker in the model with the IAP dynamical core.

This difference led to a systematic difference in the relative humidity in the two models (Fig. 2.7), which in turn led to a difference in clouds, as shown in Fig. 2.8. The IAP dynamical core was drier than the CAM3.1 core, especially in the upper troposphere. The reduction of cloud amount in the upper troposphere reached about 20 % of the mean cloud amount. High clouds are greenhouse agents that trap infrared radiation in space. The smaller amount of high clouds in the IAP model corresponded to a cooling difference relative to the CAM model. This is shown in Fig. 2.9 for the total radiative cooling rate [sum of longwave (LW) radiative heating (qrl) and shortwave (SW) radiative heating (qrs)] in the IAP model (Fig. 2.9a), the CAM model (Fig. 2.9b), and the difference (Fig. 2.9c). There was larger upper-tropospheric radiative cooling in the IAP model than in the CAM model (Fig. 2.9c), which corresponded well with the smaller amount of high cloud in the IAP model shown in Fig. 2.8c.

2.3.3 Preliminary Results from the High-Resolution IAP AGCM4.0

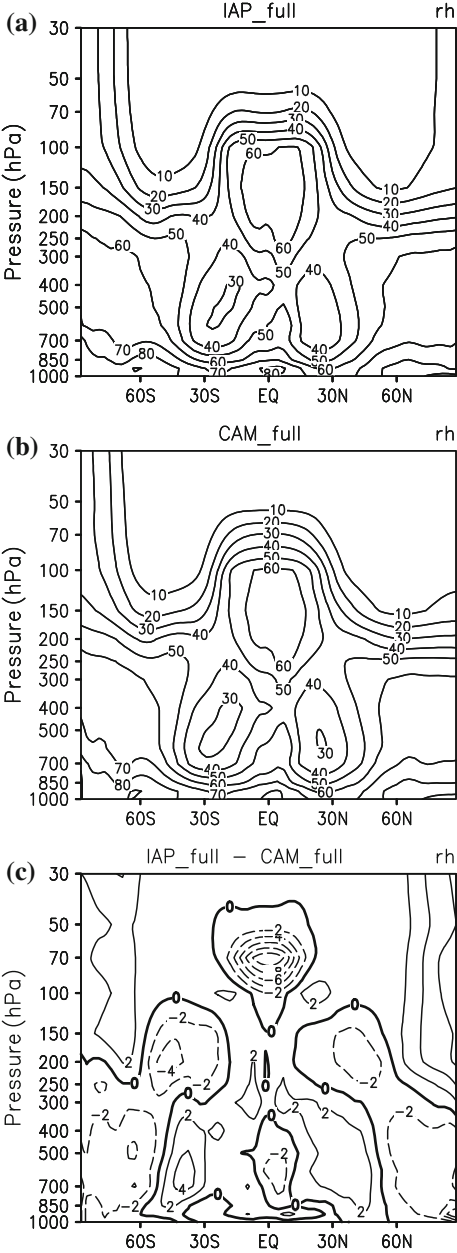
Thirty-year (1979–2008) AMIP integrations with the $0.5^\circ \times 0.5^\circ$ and $1.4^\circ \times 1.4^\circ$ version of IAP AGCM4.0 were completed and used for comparison. Three observational datasets were used to evaluate the model performance. The Global Precipitation Climatology Project (GPCP) monthly precipitation analysis V2 (Adler et al. 2003) and Global Precipitation Climatology Centre (GPCC) monthly precipitation dataset V6.0 with a horizontal resolution of $0.5^\circ \times 0.5^\circ$ (Schneider et al. 2011) were used to evaluate precipitation, and JRA-25 reanalysis data (Onogi et al. 2007) were used to evaluate sea-level pressure (SLP).

Fig. 2.6 Zonally averaged annual mean dynamical transport of moisture from **a** IAP and **b** CAM full physics experiments with realistic boundary conditions, and **c** their difference (IAP minus CAM). Contour values are $5 \times 10^{-2} \text{ g kg}^{-1} \text{ day}^{-1}$ in **(a)** and **(b)**, and 1×10^{-2} $\text{g kg}^{-1} \text{ day}^{-1}$ in **(c)**



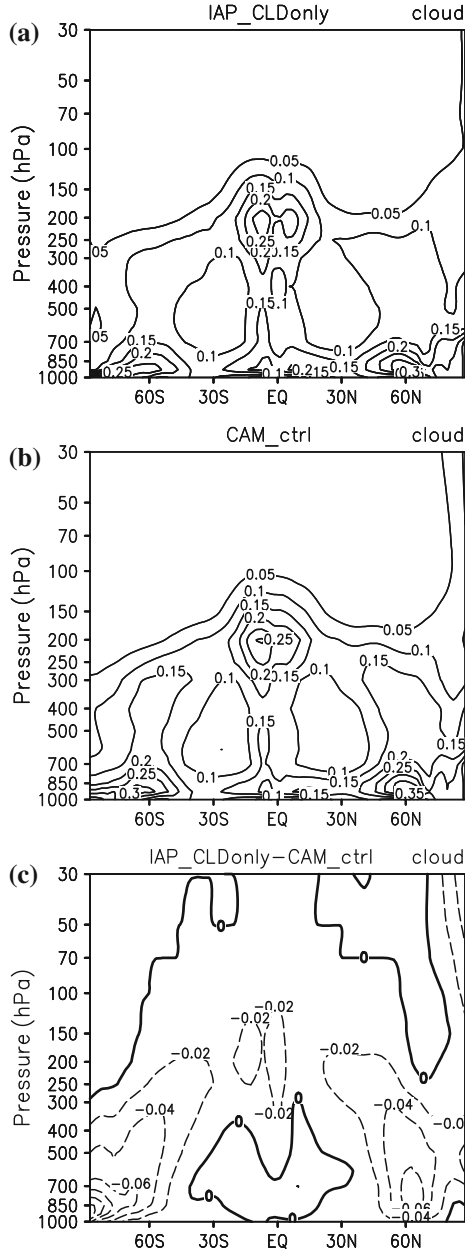
The SLP pattern is a useful indication of an AGCM’s ability to simulate the atmospheric circulation near the surface. As shown in Fig. 2.10, both the high-resolution and coarse-resolution version of IAP AGCM4.0 reproduced well the main patterns of the observed zonally averaged SLP in winter and summer, e.g.,

Fig. 2.7 As in Fig. 2.6, but for relative humidity. Contour values are 10 % in (a) and (b), and 2 % in (c)



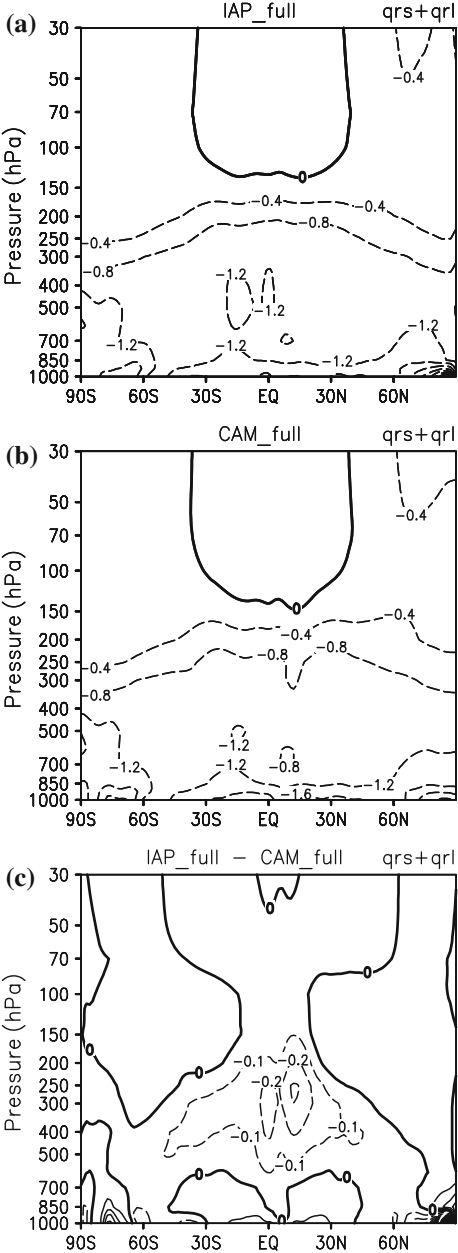
the high pressure over the subtropics and polar region, and low pressure over the tropics and midlatitudes. Over the Southern Hemisphere, the bias in the $0.5^\circ \times 0.5^\circ$ resolution version was reduced significantly in comparison to that in the $1.4^\circ \times 1.4^\circ$ resolution version, although the bias was slightly larger in the $0.5^\circ \times 0.5^\circ$ resolution version over the Northern Hemisphere.

Fig. 2.8 Cloud fraction used in experiment
a IAP_CLDonly and
b CAM_ctrl, and **c** their difference (IAP_CLDonly minus CAM_ctrl). Contour values are 0.05 in **(a)** and **(b)**, and 0.02 in **(c)**



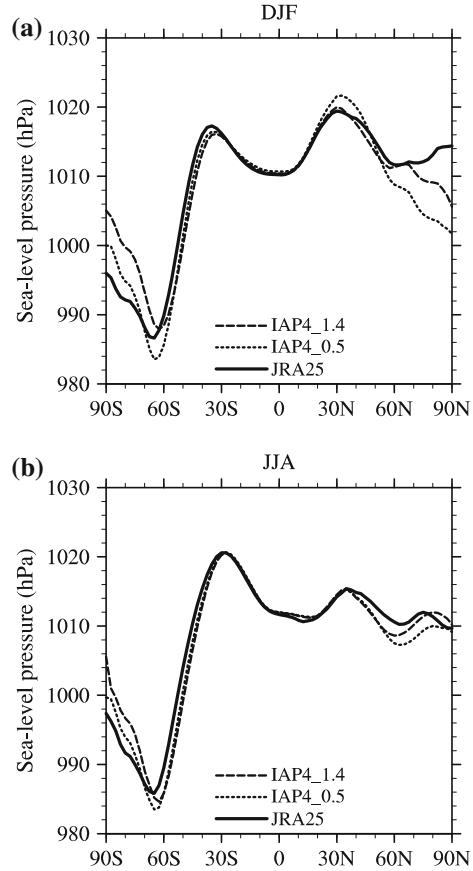
Precipitation is the result of links among moisture, thermodynamics and dynamics, and is thus difficult to simulate, especially over the monsoon region. Figure 2.11 presents the simulated and observed precipitation over land in South and East Asia in June. There were several maximum rainfall centers in the

Fig. 2.9 Zonally averaged annual mean radiative diabatic heating rate (qrs and qrl for SW and LW radiative heating, respectively) from **a** IAP and **b** CAM simulations, and **c** their difference (IAP minus CAM). Contour intervals are 0.4 K day⁻¹ in **(a)** and **(b)**, and 0.1 K day⁻¹ in **(c)**



observations, such as in the southeastern Tibetan Plateau, west coast of India, west coast of Burma, and Southeast China. Both versions of IAP AGCM4.0 simulations captured these major observed features, but nevertheless there were dry biases in

Fig. 2.10 The **a** December–January–February (DJF) and **b** June–July–August (JJA) averaged zonal mean SLP for IAP AGCM4.0 with resolutions of $0.5^\circ \times 0.5^\circ$ and $1.4^\circ \times 1.4^\circ$, and the JRA-25 reanalysis



the coarse-resolution simulation in the wet centers, especially over Southeast China. These biases in the $0.5^\circ \times 0.5^\circ$ resolution version of IAP AGCM4.0 were reduced by $2\text{--}10\text{ mm day}^{-1}$, which could be attributable to an increase in subscale precipitation in the high-resolution model.

There is a large seasonal variation of precipitation over the monsoon region. The south-to-north seasonal migration of the rainbelt is highly evident from the time–latitude section of the annual cycle of precipitation over East Asia, as shown in Fig. 2.12c. This migration feature was generally captured in the two model versions (Fig. 2.12a, b). However, the strength of the rainfall was much weaker near 30°N from June to September in the $1.4^\circ \times 1.4^\circ$ resolution version of IAP AGCM4.0. This weaker bias was improved in the high-resolution model, although it was still a little weaker than observed.

Fig. 2.11 Mean total precipitation in June from **a** IAP AGCM with a $0.5^{\circ} \times 0.5^{\circ}$ resolution, **b** IAP AGCM with a $1.4^{\circ} \times 1.4^{\circ}$ resolution, and **c** the GPCC

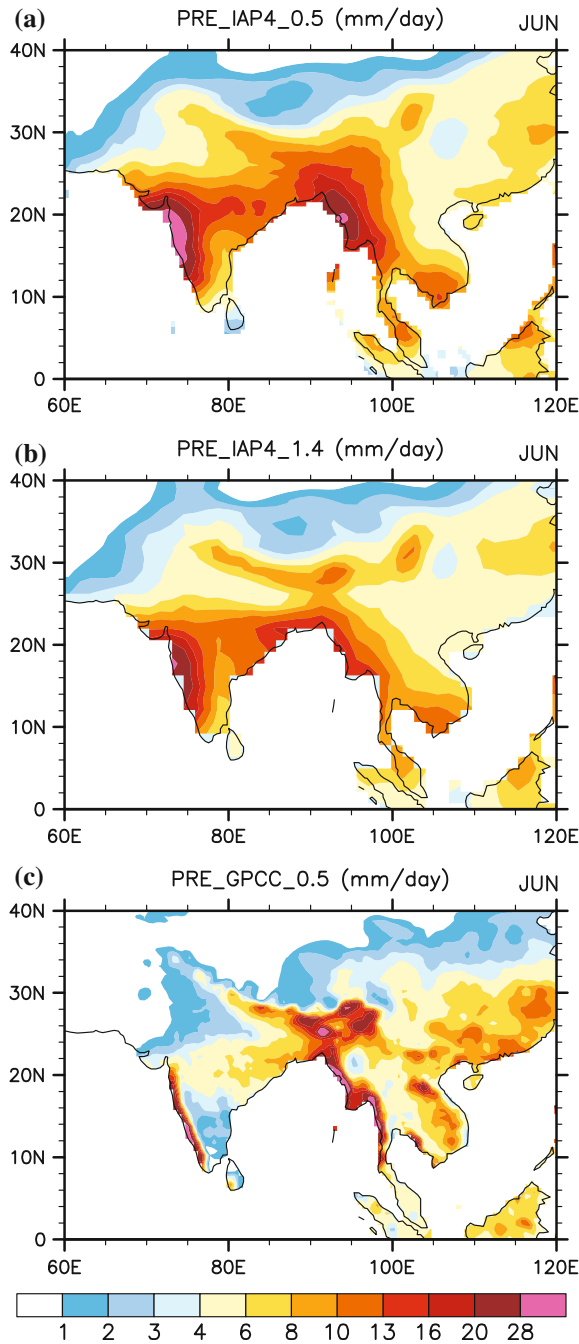
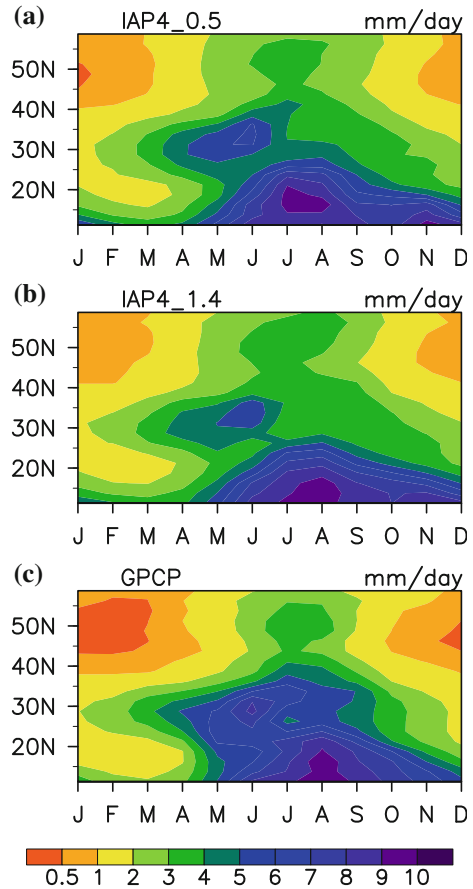


Fig. 2.12 The seasonal south–north migration of zonally averaged precipitation over 100° – 140° E from **a** IAP AGCM with a $0.5^{\circ} \times 0.5^{\circ}$ resolution, **b** IAP AGCM with a $1.4^{\circ} \times 1.4^{\circ}$ resolution, and **c** the GPCP



2.3.4 CAR Validation and Its Application to Further Improve the Performances of the Original Radiation Transfer Codes

2.3.4.1 Continual Intercomparison of Radiation Codes Experiments

Continual Intercomparison of Radiation Codes (CIRC) Phase I data (Oreopoulos et al. 2012; *cf.*, Table 1 in Zhang et al. 2013a) were used to drive the CAR system. Note that gas concentrations, aerosol single scattering properties, cloud water path and effective particle size, surface albedo (unweighted spectral), and emissivity are specified by the CIRC standard. The only freedom is the choice for cloud liquid optical property schemes. As shown in Fig. 2.13, there are eight and five schemes for cloud liquid SW (swls) and LW (lwls) optics (*cf.*, Table 1 in Zhang et al. 2013a), respectively. The outcomes were compared with the line-by-line (LBL) reference

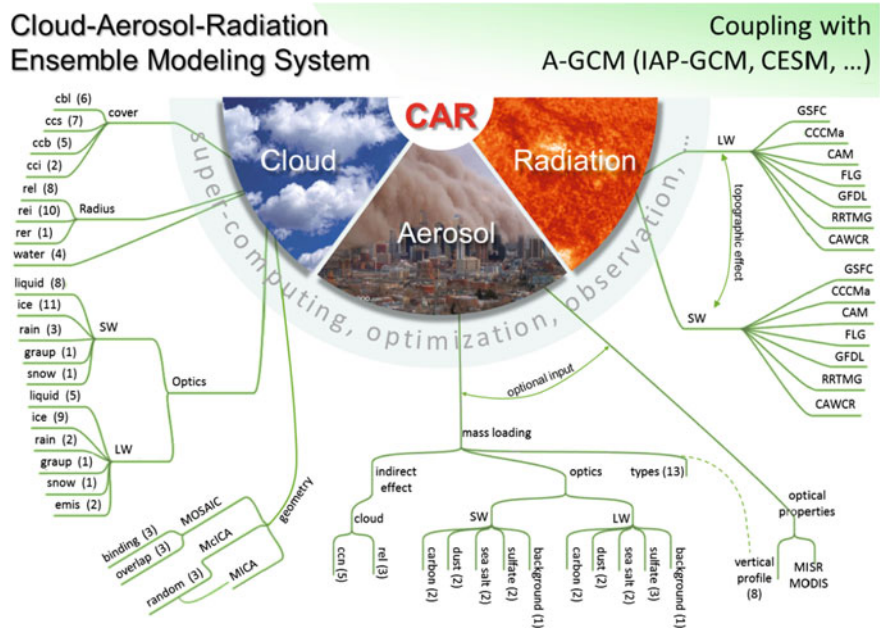


Fig. 2.13 Schematic diagram of the interactive CAR ensemble model (Liang and Zhang 2013; Zhang et al. 2013a), illustrating all key groups of parameterizations currently available (each with the number of schemes shown in parentheses) and their links with directional data flow by arrows. Acronyms/abbreviations are explained as follows: cbl, boundary cloud; ccs, stratiform cloud; ccb, deep convective cloud; cci, cirrus cloud; rei, cloud liquid effective radius; dei, cloud ice effective size; rer, cloud rain effective radius; graup, cloud graupel; bgd, background aerosol; sul, sulfate; ssa, sea salt; dst, dust; cab, carbon; ccn, cloud condensation nuclei; GSFC, CCCMA, CAM, FLG, GFDL, RRTMG, and CAWCR stand for the USA NASA Goddard, Canadian CCCma, USA NCAR, USA UCLA Fu-Liou-Gu, USA NOAA GFDL, USA AER, and Australian CAWCR radiation transfer codes, respectively

results available from CIRC to evaluate the accuracy of various radiation transfer schemes with different cloud liquid optical property parameterizations used.

The smallest errors were generally not identified with the default combination of radiation transfer and cloud optical property schemes in the original packages, and no single cloud optical property scheme worked best with all radiation transfer schemes. We defined the total percentage difference e_{tot} (%) from the LBL references (SW: LBLRTM-CHARTS line-by-line radiative transfer model) as: $e_{\text{tot}}(\%) = 0.5 \times (|e_{\text{SWDNS},\%}| + |e_{\text{SWUPT},\%}|)$, where $e_{\text{SWDNS},\%}$ and $e_{\text{SWUPT},\%}$ are the % differences of SWDNS [SW downward radiative fluxes at the ground surface (SFC)] and SWUPT (SW upward radiative fluxes at TOA) from the LBL references provided by CIRC. Figure 2.14 compares the $e_{\text{SWDNS},\%}$ and $e_{\text{SWUPT},\%}$ between the results of the original radiation codes (called ‘original’ here, blue bars) and the best results assuming the smallest e_{tot} (%) among the eight swl schemes (called “best,” red bars) for CIRC cases 6 and 7.

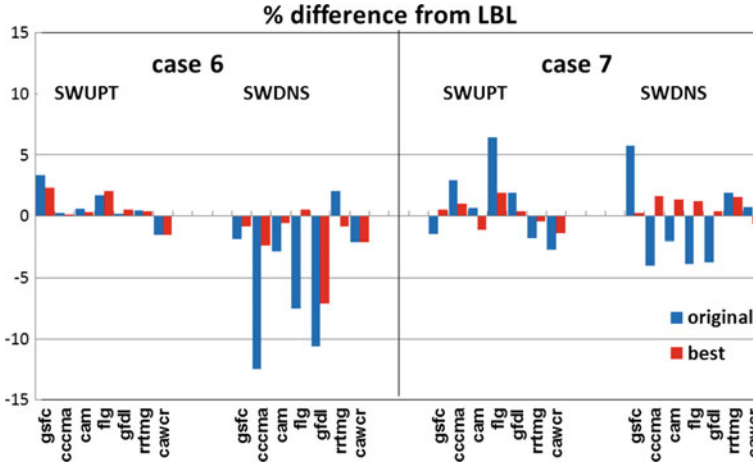


Fig. 2.14 The percentage difference (%) of SWUPT and SWDNS from the LBL reference (LBLRTM-CHARTS) for CIRC Phase I cases 6 and 7. Here, two results ('original' and 'best') from seven CAR major radiation codes are shown: the 'original' results (*blue bars*) are those with the original cloud optical property scheme used, while the 'best' results (*red bars*) assume the smallest e_{tot} (%) among the eight swls. Reprinted from Zhang et al. (2013a), with kind permission from Springer Science+Business Media

The $e_{\text{SWDNS},\%}$ and $e_{\text{SWUPT},\%}$ of the original radiation codes were within -12.5 to 6.5 %, which lay in the ranges disclosed by Oreopoulos et al. (2012). However, the obvious differences between the blue and red bars indicate that the original radiation codes did not always generate the best results. By using different cloud liquid optical property schemes, most percentage differences of SWUPT and SWDNS from LBL could be decreased by a factor of 1.22–23.41, depending on the case and radiation codes. The smallest percentage differences for these two CIRC cloudy cases were generally within ± 2.5 %, showing very good accuracies. Therefore, even when the same cloud droplet sizes and the same cloud water path are used, how to parameterize the cloud optical properties is also an uncertain factor for the solar radiation transfer calculations.

2.3.4.2 ERA-Interim Experiments

Six-hourly ERA-Interim observational analysis data (ERI) (Uppala et al. 2008) between January and July 2004 were used to drive the CAR system. The cloud water path was directly calculated from the input cloud water mass mixing ratio. MISR climatological mean aerosol optical properties over the period 2000–08 (Kahn et al. 2007) were used to account for the aerosol direct effects. Although only 2 months are evaluated here, the statistics over 60°S – 60°N included about 10^5 cloudy columns over the globe, which we assumed comprised sufficient cloud variety for the study.

Overall Accuracy and Further Improvement of Radiation Code Performance

To evaluate the CAR system’s overall accuracy, global simulations using the seven major CAR radiation schemes, each combined with the 64 cloud members listed in Table 4 in Zhang et al. (2013a), were designed. In total, 448 runs (LW and SW) were made.

Figure 2.15 illustrates boxplots of %bias of the 448 CAR members from different observations for July 2004 mean all-sky SWUPT, SWDNS, LWUPT (LW upward radiative fluxes at TOA), and LWDNS (LW downward radiative fluxes at SFC). The calculation was based on all 1.5° grids over (60°S–60°N). Despite the large spread among these 448 CAR members, the ensemble means of %bias compared to each of the chosen observations were within -0.67 to 8.59% (-0.6 to 8.2 W m^{-2}) for SWUPT, -5.34 to 0.06% (-10.4 to 0.1 W m^{-2}) for SWDNS, -2.40% to -0.82% (-5.7 to -2.0 W m^{-2}) for LWUPT, and -2.12 to -1.55% (-7.8 to -5.7 W m^{-2}) for LWDNS. The CAR ensemble means obtained very good overall accuracies for both SW and LW radiation calculations, establishing the credibility of future applications of CAR in climate models. Similar results were obtained for January 2004 (not shown). In addition, using ISCCP (International Satellite Cloud Climatology Project) as the reference usually generated the smallest %bias, while the largest %bias was from the CERES (cloud and the Earth’s radiant energy system), especially for the SW calculation. The ensemble means of %bias from ISCCP were only about -0.67% (-0.6 W m^{-2}), 0.06% (0.1 W m^{-2}), -0.82% (-2.0 W m^{-2}) and -2.12% (-7.8 W m^{-2}) for SWUPT, SWDNS, LWUPT and LWDNS, respectively. This strange phenomenon perhaps indicates that the schemes chosen in this study are not varied enough, that they are improperly tuned toward ISCCP, or the cloud water fields are closest to those from ISCCP.

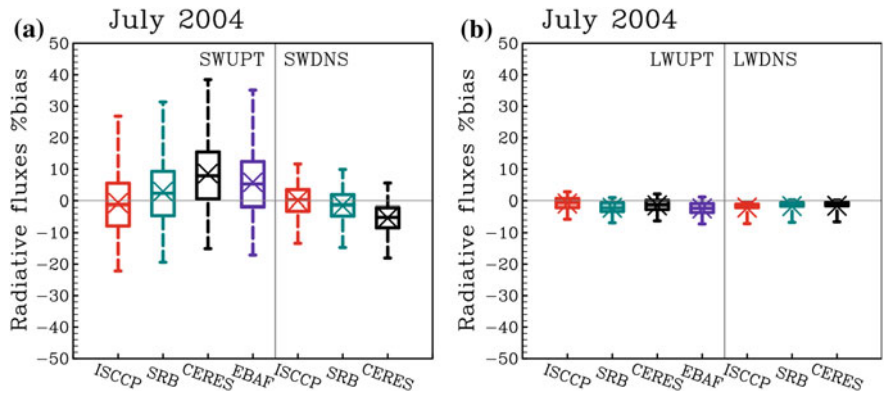


Fig. 2.15 Boxplots of %bias from different observations for July 2004 mean all-sky SWUPT, SWDNS, LWUPT and LWDNS for 448 CAR members. EBAF is CERES_EBAF. For each observation, the min, 25 % quartile, median, 75 % quartile, and max values among 448 CAR members are shown. The average among these 448 CAR members is also shown, marked ‘X’. ISCCP, SRB, CERES and EBAF refer to the different observations. Reprinted from Zhang et al. (2013a), with kind permission from Springer Science+Business Media

To better show the great potential of CAR to improve the radiation transfer calculation accuracies, the total $\overline{\%rmse}$ from ISCCP with those for the original radiation transfer codes (Table 3 in Zhang et al. 2013a) are shown in Fig. 2.16. Similar RMSE results were found for different observations, so only those from ISCCP are taken as an example here.

For all radiation transfer schemes except flg and cccma, generally $>25\%$ ($>16\%$) of CAR cloud members possessed smaller SW $\overline{\%rmse}$ than those for the original radiation codes, implying that although the original radiation transfer codes had quite good accuracies, some room for improvement still existed via the application of better cloud scheme combinations, emphasizing the large SW errors emerging from the treatments of cloud cover fractions, particle effective sizes, and optical

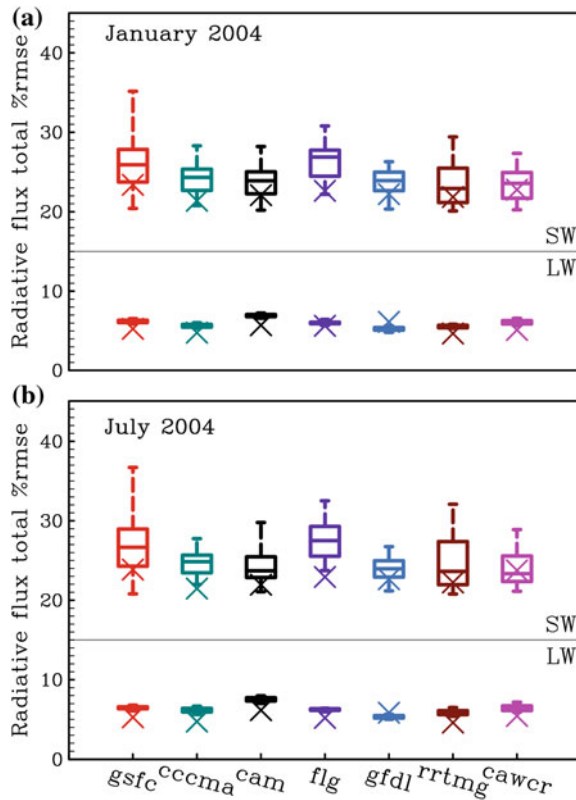


Fig. 2.16 Boxplots of total $\overline{\%rmse}$ from ISCCP for January and July 2004 mean all-sky SW and LW radiative fluxes for each radiation scheme with the min, 25 % quartile, median, 75 % quartile, and max values among 64 CAR cloud members shown. The 'X' depicts the result from the original radiation transfer code. Total $\overline{\%rmse} = 0.5 \times (\%rmse_{UPT} + \%rmse_{DNS})$, $\%rmse = RMSE/\bar{O}$, RMSE is the model spatial root mean square error, where $\bar{O}$ is the observed domain mean, and the calculation is based on all 1.5° grids over $60^\circ S-60^\circ N$. The subscripts UPT and DNS are the upward and downward fluxes at the TOA and SFC, respectively. Reprinted from Zhang et al. (2013a), with kind permission from Springer Science+Business Media

properties. Meanwhile, the better accuracies of the original flg and cccma SW calculations, and almost all better original LW calculations than those for the 64 CAR cloud members, may indicate an insufficient number of schemes chosen in this study. Undoubtedly, in the near future, we should include and investigate as many CAR cloud members as possible. Furthermore, to obtain better ensembles of CAR, the outliers should first be deleted, and then, based on what remains, we should try to identify the normal distribution of the problem at hand.

Effects of Cloud Subgrid Structures on Radiative Calculation Accuracies

The effects of the vertical overlap treatments of inhomogeneous clouds on radiative transfer calculations have also been investigated using the CAR. In the original radiation transfer codes, the maximum/random overlap among high/middle/low cloud blocks is adopted by gsfc and flg, while the maximum/random overlap among adjacent/nonadjacent layers is applied by cam and cawcr. The other three radiation transfer codes, i.e., cccma, gfdl, and rrtmg, employ McICA (Monte Carlo Independent column approximation) treatments with different maximum-random cloud generators.

Besides the results from the radiation schemes with original overlap treatments as mentioned above (called “original” here, blue bars), the results for those using the same McICA treatment with the maximum/random overlap among adjacent/nonadjacent layers of inhomogeneous clouds (called “McICA,” red bars) are also shown in Fig. 2.17. The major difference between “original” and “McICA” is the different treatment for the cloud subgrid-scale structures. Compared with the original treatments, the McICA treatment clearly reduced the RMSEs of TOA and

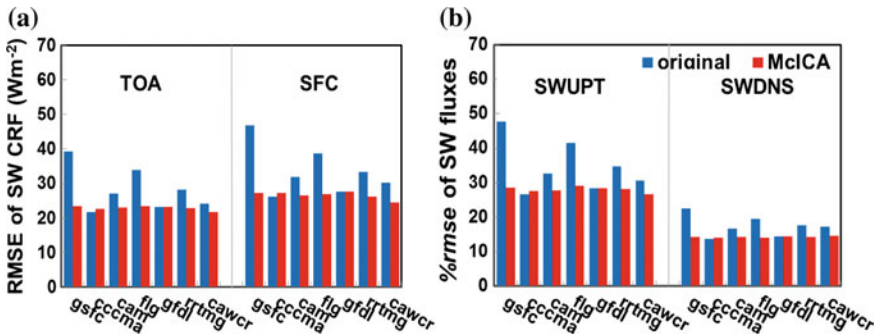


Fig. 2.17 a RMSEs of SW SFC and TOA CRFs and b %rmse for SWUPT and SWDNS under all-sky conditions for July 2004 over 60°S–60°N. Here, %rmse = $\text{RMSE}/\bar{O}$, where RMSE is the model spatial RMSE and $\bar{O}$ is the CERES data. Note that the same set of schemes for cloud cover fraction, particle effective size and optical properties was applied for each radiation transfer scheme. When different observations (ISCCP or SRB) were used, the same conclusions were reached (not shown). Reprinted from Zhang et al. (2013a), with kind permission from Springer Science+Business Media

SFC SW cloud radiative forcings (CRFs), especially for gsf. The RMSEs for gsf decreased by about 15 W m^{-2} from about 38 to 23 W m^{-2} and, accordingly, the corresponding %rmse of SWUPT and ~ 42 to $\sim 28 \%$ and from ~ 22 to $<18 \%$, respectively. For ccm, gfdl, and rrtmg, meanwhile, the relatively smaller differences between the original and McICA were mainly due to the different random cloud generator used. Hence, the vertical overlap treatments of inhomogeneous clouds should be considered as another major uncertainty factor influencing the current radiation transfer calculations.

2.3.5 *The Spread Related to Cloud and Radiation Calculations*

To demonstrate the seasonal and regional variations of cloud-radiation uncertainties, the member frequency distributions of the CAR domain-mean SW results over China (20° – 60°N , 75° – 135°E) and the U.S. (20° – 60°N , 140° – 55°W) are compared in Figs. 2.18 and 2.19, respectively.

For both domains, substantial spreads were seen among alternative members: 30 – 80 W m^{-2} in SW net fluxes at the top (SWUPT) or surface (NSWS). Similar uncertainty ranges could also be found for CRFs, implying that these model spreads were dominated by treatments for clouds, including those for cloud cover fractions, optical properties, and unresolved structures. The best available observational data also contained nontrivial uncertainties for the two domains, ranging between 5 and 15 W m^{-2} both for radiative fluxes and CRFs. The above domain ranges were similar to those of global means, i.e., 30 – 60 W m^{-2} in TOA/SFC LW/SW all-sky fluxes and in TOA/SFC CRFs (Liang and Zhang 2013). Furthermore, the disagreement in SW CRFs could be reduced to about 32 W m^{-2} when the constraint of the net TOA fluxes (LW + SW) within the observational ranges was applied (not shown). For July, the model spreads wholly included the observational ranges, with ensemble means located in the middle of such ranges. This indicates a good and reasonable performance of the CAR system, with very good accuracies of the CAR ensembles. It was also noted that, for January, almost all observations fell at the tails of the model distributions, which infers the possibility that the cloud/radiation representations in winter were not properly parameterized, the schemes chosen in this study were not varied enough, or the ERI cloud water fields had poor quality.

Moreover, for cloud-radiation uncertainties, quite large seasonal variations were found to exist: both the TOA and SFC SW CRF ranges changed from about 20 to 30 W m^{-2} in January to about 80 W m^{-2} in July. This indicates that larger uncertainties are associated with more cloud and more insolation.

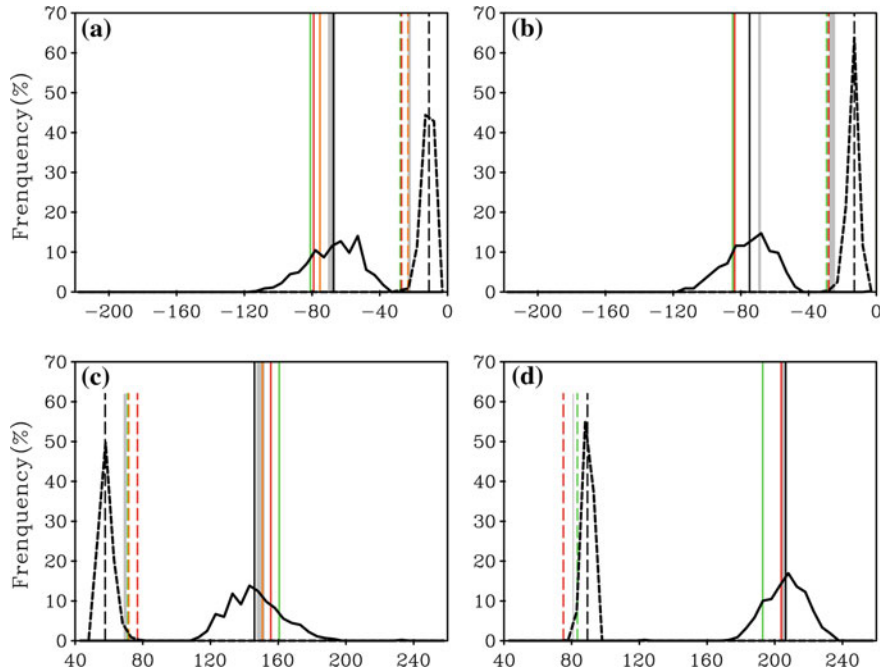


Fig. 2.18 Member frequency distributions (*solid lines* July; *dashed lines* January) of the CAR ensemble sensitivities in predicting the domain-mean **a, b** TOA and SFC SW CRF and **c, d** net radiative fluxes over China (20°–60°N, 75°–135°E) to a selective subset of 448 CAR members driven by 3D ERA-Interim data for the year 2004. *Black line* CAR ensemble mean. Also compared are the uncertainties in observational data from ISCCP (*red*), SRB (*green*), various versions of CERES (*gray shading*), and CERES_EBAF (*orange*). Reprinted from Zhang et al. (2013a), with kind permission from Springer Science+Business Media

2.3.6 Dominant Roles of Subgrid-Scale Cloud Structures in Model Differences of Cloud Radiative Effects

Model differences in estimated irradiances remain large, reducing our confidence in climate change projection. The individual roles of cloud cover fraction, optical properties, subgrid structure, and gaseous effects have been explored for the first time.

2.3.6.1 Origin and Reduction of Model Discrepancies in Cloud Radiative Effects

Figure 2.20 shows comparisons of zonal mean TOA/SFC SW cloud radiative effects (CREs) among the CAR’s seven major radiative transfer schemes for dcop_dovp1 (original) and scop_sovp1 (the same clouds used) (*cf.*, Table 3 in Zhang et al. 2013b). Clearly, for dcop_dovp1 (original), when different radiation

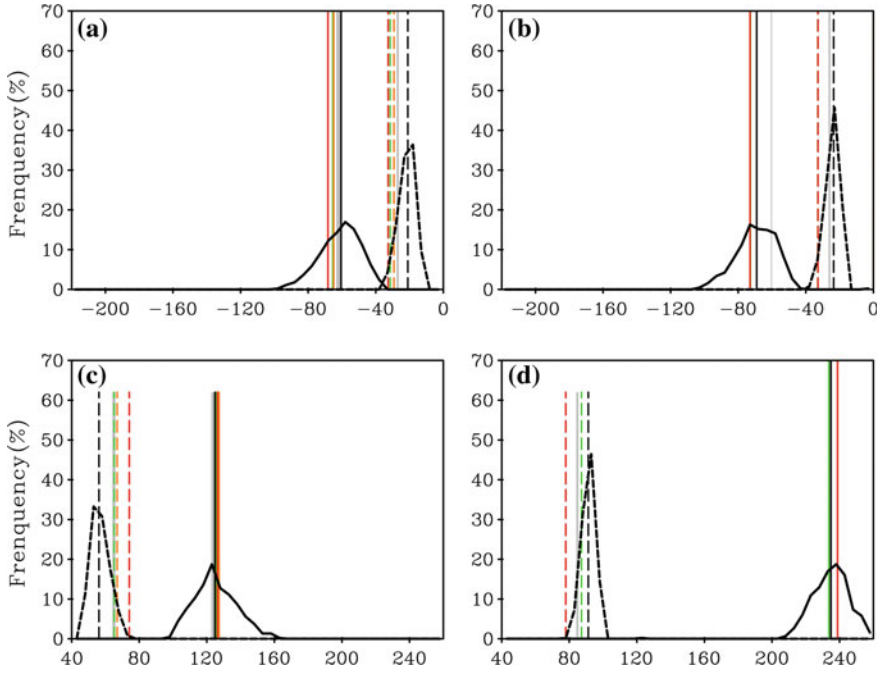


Fig. 2.19 As in Fig. 2.18, except for over the U.S. (20°–60°N, 140°–55°W). Reprinted from Zhang et al. (2013a), with kind permission from Springer Science+Business Media

algorithms employed different approaches for cloud optical properties and subgrid structures, although the same 3D distributions for cloud cover fraction and water path were used, large intermodel differences existed for both SW and LW CREs. For SW, spreads of 25–30 W m^{-2} were found in both the tropical and midlatitude areas. For LW, large differences were most remarkable for TOA CREs over the tropics with discrepancy of about 30 W m^{-2} ; and for SFC LW CREs, since their values of 10–40 W m^{-2} were quite small, the intermodel differences of about 5–10 W m^{-2} could not be ignored (not shown). In scop_sovp0 and scop_sovp1, by eliminating the greatest disagreement in cloud fields, the model discrepancies in dcop_dovp1 dramatically diminished. At most latitudes, model spreads were no more than 5 W m^{-2} for both SW and LW. Clearly, the large existing intermodel discrepancies in CREs primarily come from different cloud treatments that were originally adopted by different radiation codes.

2.3.6.2 Dominant Roles of Cloud Subgrid Structure in the Model Differences of Estimated Radiative Components

From Fig. 2.21a–e, it is clear that as cloud fields became similar, the intermodel discrepancies, as well as the min–max ranges among 14 cloud cover members

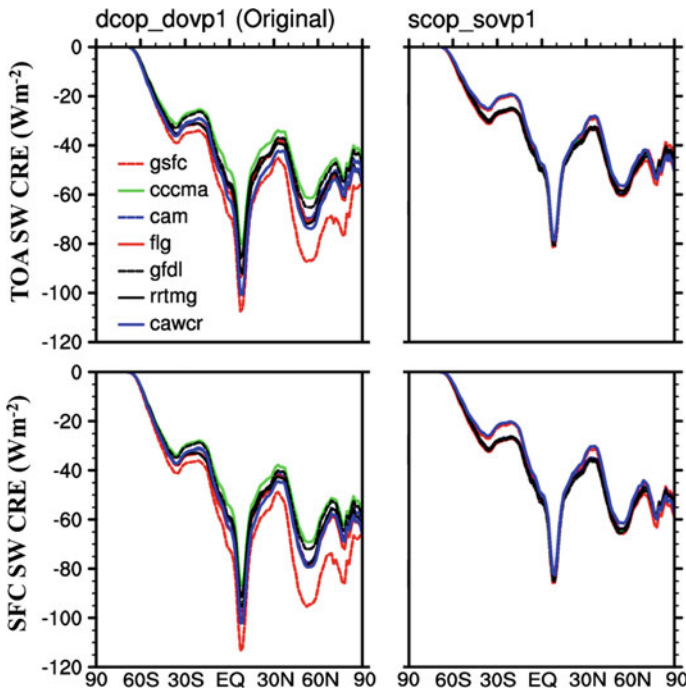


Fig. 2.20 Comparisons of zonal mean TOA/SFC SW CRE among the CAR's major radiative transfer schemes for dcop_dovp1 (i.e., the original cloud/radiation schemes used) and scop_sovp1 (i.e., the same treatments for clouds consistently applied by different radiative transfer schemes). Here, the CRE is defined as $CRE_{SW} = F_{SW}^{all} - F_{SW}^{clr}$, where F is the net (downward minus upward) flux, clr designates clear skies, all denotes a mixture of clear and cloudy skies; and ERA-Interim data in July 2004 were used. We took the combination of Xu and Randall (1996) for stratiform and Slingo (1987) for deep convective cloud cover fractions. Reproduced from Zhang et al. (2013b) by permission of John Wiley & Sons Ltd

(cld1–cld14), were significantly reduced, even to less than their corresponding interobservation ranges. In dcop_dovp1, based on the same 3D distributions for cloud condensate and cloud cover fraction, using the mixing overlap among adjacent/nonadjacent layers for inhomogeneous clouds generally produced the weakest SW CREs and smallest SWUPT, e.g., cccma and gfdl. Furthermore, both could be enhanced by removing the cloud inhomogeneity effects in dcop_dovp0, which then decreased the model spread in dcop_dovp0. Using similar cloud optical properties seemed to have no positive effect on the reduction of the SW model discrepancies. In scop_dovp0, the seven radiation codes formed two groups. Whether or not the conventional or McICA method was used, those using the same assumption of maximum/random overlap among adjacent and nonadjacent layers, i.e., cccma, cam, gfdl, rrtmg, and cawcr, converged together, while flg demonstrated similar performance as gsfc because of their adoption of the same mixing overlap among high/middle/low cloud blocks. In each group, the influence of cloud optical

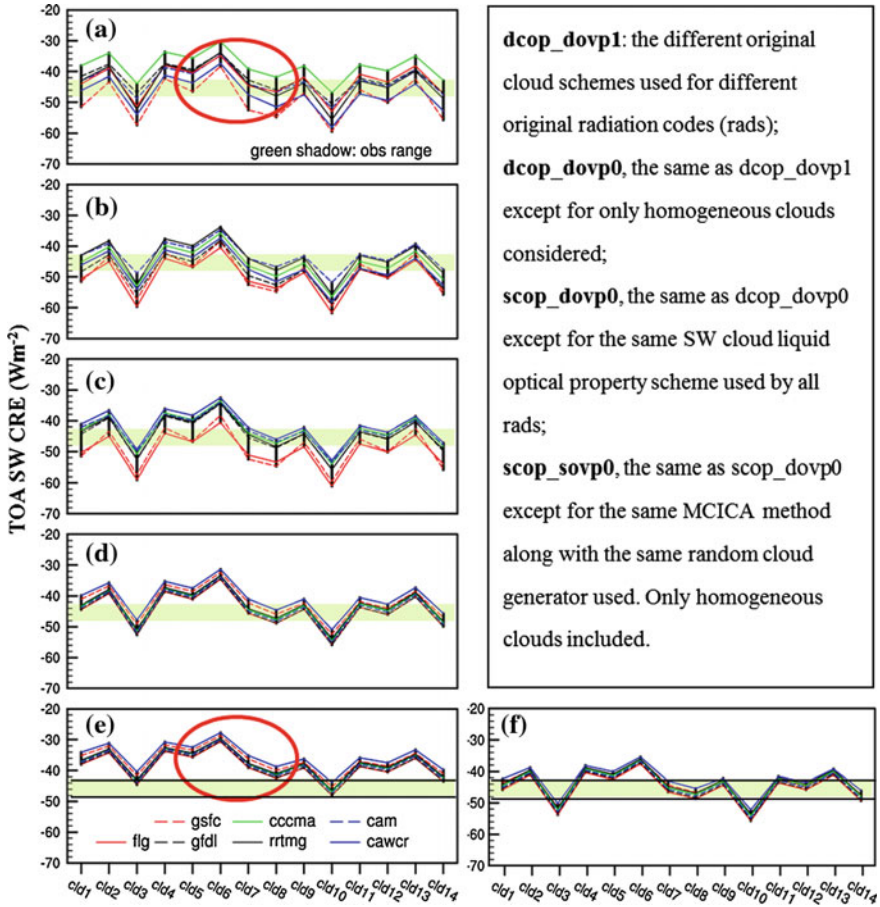


Fig. 2.21 Comparisons of global mean TOA SW CRE calculated using McICA_MRO (a–e) and McICA_Fgen1 (f) among 14 different cloud cover members (cld1–cld14) for different sets of experiments. ERA-Interim data for July 2004 were used. Green shading indicates the interobservation uncertainty among ISCCP, SRB, CERES and CERES_EBAF. scop_sovp1 is the same as scop_sovp0 except for the same cloud inhomogeneity effects are considered. McICA_MRO uses the McICA method along with the same implementation of the maximum-random overlap generator; McICA_Fgen1 uses the McICA method along with the same implementation of the generalized overlap generator (Hogan and Illingworth 2000), and the decorrelation length for overlapping fractional cloud, $L_\alpha = 2$ km, and the rank correlation decorrelation length, $L_\gamma = 1$ km. Reproduced from Zhang et al. (2013b) by permission of John Wiley & Sons Ltd

properties on model spread was large. However, the effects of using the same treatment of cloud optical properties were substantially limited by the different assumptions of cloud vertical overlap. Finally, when the same treatment of cloud vertical overlap was consistently applied in scop_sovp0, the model discrepancies were greatly reduced, even to less than their corresponding interobservation ranges.

Table 2.3 Percentages of model differences contributed by different treatments for different factors. Reproduced from Zhang et al. (2013b) by permission of John Wiley & Sons Ltd

Factor	TOA CRE		SFC CRE		UPT		DNS	
	SW	LW	SW	LW	SW	LW	SW	LW
Cloud subgrid variability	35.4	3.3	19.5	38.9	36.0	25.6	44.2	34.2
Cloud vertical overlap	44.4	36.9	44.4	27.9	50.6	0.1	8.4	13.5
Cloud optical properties (including particle effective size)	–	16.4	–	–	–	–	–	–
Trace gaseous effects (i.e., CO ₂ , O ₂ , CH ₄ and N ₂ O for SW; CH ₄ , N ₂ O, CFCs etc. for LW)	1.2	3.1	3.3	1.9	–	6.4	17.5	17.9

UPT upward fluxes at TOA; *DNS* downward fluxes at SFC; “–” means that no effective roles could be obtained, which was mainly due to the limitations from the roles of key factors, e.g., the variety in the existing assumptions of cloud vertical overlap largely obscuring the roles of different cloud optical properties

The percentages of the model differences contributed by different treatments for different factors are shown in Table 2.3. In summary, about 40–75 % of the model disagreement resulted from the different treatments of cloud subgrid structures, indicating a dominant role of cloud vertical overlap in the model differences. We have also discovered that the key factors are usually different for different radiative components, especially for LW. Generally speaking, using different assumptions of cloud vertical overlap is critical to the model spread of SW components and TOA LW CREs (>35 %), inferring the importance of both the position of cloud-top level and total cloud cover fractions; while different cloud inhomogeneity effects are key to those of SFC LW CREs and LWDNS (34–39 %), indicating roles of cloud optical properties. The difference in gas absorption has also been found to substantially contribute to model differences of SWDNS (~17.5 %) and LWDNS (~17.9 %). For SWDNS, this is most likely due to the long direct optical path to surface, as opposed to the TOA, where the differences due to gaseous absorption are smaller; while for LWDNS, gaseous absorption and emissions below clouds play important roles. In addition, the variety in existing assumptions of cloud vertical overlap largely limits the roles of different cloud optical properties.

From Fig. 2.21, we also note that in *dcop_dovp1*, *dcop_dovp0*, and *scop_dovp0*, due to the large intermodel diversity, all radiation codes could achieve quite good calculation accuracies by using the different combination of cloud schemes. The performance of each cloud scheme just depended on the host model, largely preventing us from reaching an accurate understanding of Earth’s radiation budget. However, after removing most of the disagreement in cloud fields in *scop_sovp0* and *scop_sovp1*, it became clearer as to how to further improve the estimated CREs and irradiances—one of the CAR’s major advances. As shown in Fig. 2.21e, in *scop_sovp1*, compared with observations, consistently weaker SW CREs were found for most clds. Although this is partly due to the prescribed surface emissivity and simplified surface albedo, the small model ranges also suggested at least three

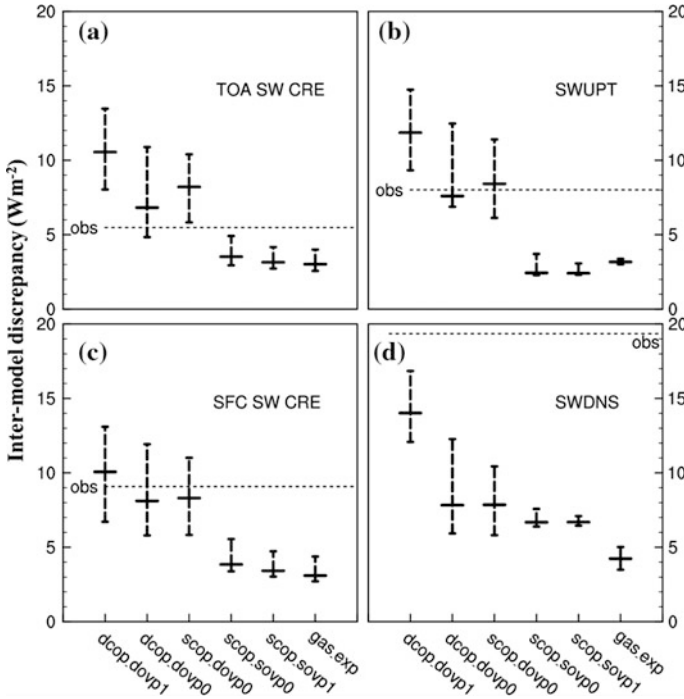


Fig. 2.22 Comparisons of the intermodel differences across the CAR's seven major radiation transfer codes in terms of the global mean **a** TOA SW CRE, **b** SWUPT, **c** SFC SW CRE, and **d** SWDNS among different sets of experiments. The medians and min–max ranges are among 14 intermodel differences generated by applying the same cloud cover members (cld1–cld14) to each intercomparison of the CAR's seven major radiation transfer codes. “obs” denotes the interobservation uncertainty among ISCCP, SRB, CERES and CERES_EBAF. The results are for July 2004. Reproduced from Zhang et al. (2013b) by permission of John Wiley & Sons Ltd

other consistent solutions to this problem: (1) increasing the cloud condensate field; (2) increasing total cloud cover fractions; and (3) increasing cloud optical depth for SW. Here, an example is shown to illustrate the improvement of model performances by using McICA_Fgen1 to increase the total cloud cover fractions (Fig. 2.21f). The generalized overlap generator produces larger total cloud cover fractions than the maximum-random overlap generator (Räisänen et al. 2004), improving the models' results.

2.3.6.3 Reduction in the Nonlinear Sensitivity of Model Spread to Cloud Fraction

As shown in Figs. 2.21 and 2.22, the complicated nonlinear nature that exists in the coupling of cloud/radiation processes is highlighted in this study. The current

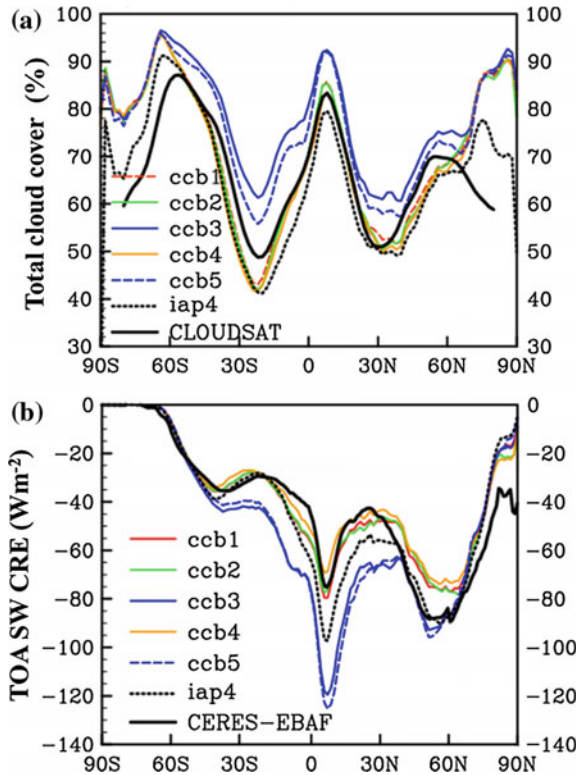
intermodel differences among different radiation codes vary strongly with different cloud cover members (clds), and the major reasons were investigated in this study. In general, using the different assumptions of cloud vertical overlap accounted for most of the sensitivity of the intermodel diversity to different clds. Different clds produced different vertical profiles of cloud cover fraction, which determined the degree to which the different assumptions of cloud vertical overlap mattered for radiation. Furthermore, the quite different ranges of clear-sky portions among different radiation codes then generated the large nonlinear sensitivity to different clds. In brief, the combination of the nonlinear roles of different cloud factors (cloud cover, optical properties, and subgrid-scale structures) in radiation transfer calculations is the major reason for such nonlinear variation. By removing the contributions from the different treatments of cloud vertical overlap mentioned above, the remaining nonlinear effects between cloud and radiation were largely reduced, as shown by the comparisons between scop_dovp0 and scop_sovp0 in Figs. 2.21 and 2.22.

2.3.7 Incorporation of the CAR System into the Physical Framework of IAP AGCM4

We have successfully incorporated the CAR system into the physical framework of IAP AGCM4 (called IAP AGCM4.5). Driven by the observed climatologic sea surface temperature data, five 6-year simulations (ccb1–ccb5) have been conducted using IAP AGCM4.5, along with one done by IAP AGCM4 (iap4). Figure 2.23 compares the zonal means of total cloud cover (a) and TOA SW CRE (b). Shown are the last 5-year means of model results. Here, ccb1–ccb5: five deep convective cloud cover schemes. Clearly, different CAR schemes produced quite different results, and assume some advantage to iap4 at some regions, especially over tropical areas. And generally speaking, the spread of results of IAP AGCM4.5 with different CAR members can fully comprise the observation, and hence using an optimized physics ensemble may have the ability to largely enhance the IAP AGCM4.5 performances as compared to observations. The above represent preliminary IAP AGCM4.5 results only; further analyses are ongoing.

Overall, the coupled IAP AGCM4/CAR system will certainly facilitate reliable probabilistic predictions of the climate, as defined by the likelihood and uncertainty of the physics ensemble forecasts, and thus could enhance the capability of IAP AGCM4 in simulating global cloud–aerosol–radiation interaction, and thus global climate change.

Fig. 2.23 Zonal means of total cloud cover (a) and TOA SW CRE (b). Shown are 5-year means of results from IAP AGCM4.5 (ccb1–ccb5) for July, along with observations (i.e., CLOUDSAT or CERES_EBAF) and simulation results from original IAP AGCM4 (iap4). Here, ccb1–ccb5: five deep convective cloud cover schemes



2.3.8 A High-Resolution Global Ocean General Circulation Model Based on the Hybrid Coordinate Ocean Model

The Hybrid Coordinate Ocean Model (HYCOM) (Bleck 2002) is a primitive equation ocean general circulation model developed based on the Miami Isopycnic Coordinate Ocean Model (MICOM) by combining three traditional vertical coordinates (z -level, terrain-following (σ), and isopycnic coordinates) for a proper representation of different ocean water characteristics.

In this subproject, a newer version of HYCOM, modified by the Nansen Environmental and Remote Sensing Center (NERSC-HYCOM 2.2), was used to set up a global ocean model with a resolution of at least 30 km. NERSC-HYCOM 2.2 is very different from the standard version of HYCOM, using different methods for nesting, the tidal boundary, ice modeling and thermodynamics and, the forcing field.

2.3.8.1 Model Setting

The hybrid coordinate is one that is isopycnal in the open, stratified ocean to retain sea water mass characteristics for centuries, but smoothly reverts to a terrain-following coordinate in shallow coastal regions to maintain sufficient vertical resolution, and to z -level coordinates in the mixed layer and/or unstratified seas, which need high vertical resolution to depict thermodynamical and biochemical processes (Wallcraft et al. 2009). In HYCOM, the reference pressure level should be chosen to define the target potential densities where a more accurate simulation is expected near the reference level and a relatively inaccurate one far away from the reference level. In this model, we chose two reference pressure levels at the surface and at 2000 m depth, separately, for different experiments to validate the model performance (denoted by EXP1 and EXP2 hereafter). We set 30 layers in the vertical direction, and Table 2.4 presents the target densities from 1000 kg m^{-3} for every layer for the two reference densities. Note that the densities in the upper five layers are less than any true seawater density to ensure that the top layers remain in z -coordinates because the model will decide which vertical coordinate (isopycnal or z -layer) is to be set, depending on how far away they are from the reference density.

A big problem in a grid point model that is discretized from a natural longitude–latitude orthogonal curve coordinate is that, at the polar regions, the computation grids in the model become very dense and converge to a point. A very short time step and special filter are needed to stabilize the computation. In order to avoid this problem, and save computing time, the horizontal coordinate transformation on a sphere was applied in NERSC-HYCOM by using conformal mapping, where the model's North Pole for the horizontal coordinate was placed in the center of the Eurasian Continent (50°N , 80°E), while the South Pole remained as normal (Bentsen et al. 1999). Therefore, computations are not needed in the model's polar regions where they are not oceans. There are a total of 1800×1200 grids in the horizontal direction, with a resolution of 7–30 km from the model's polar regions to the Equator (Fig. 2.24). The variables of the model are placed in Arakawa's C grid cells, and a periodic lateral boundary condition in the zonal direction is set. A barotropic and baroclinic mode discretization algorithm is used to increase the calculation efficiency. The time steps were set to 10 s for the barotropic mode and 600 s for the baroclinic mode in EXP1, and 9 s and 450 s in EXP2, respectively, after several tests.

There are several mixing algorithms supplied in HYCOM, e.g., the Kraus–Turner slab mixed layer model (Niiler and Kraus 1977), K-Profile-Parameterization (KPP) (Large et al. 1994, 1997), the Price–Weller–Pinkel dynamical instability model (Price et al. 1986), Mellor–Yamada level 2.5 turbulence closure (Mellor and Yamada 1982; Mellor 1998), and NASA Goddard Institute for Space Studies (GISS) level 2 turbulence closure (Canuto et al. 2001, 2002). In this model, the KPP scheme was chosen, which offers mixing from surface to bottom, efficiently matching the large surface boundary layer diffusivity/viscosity profiles to the weak diapycnal diffusivity/viscosity profiles of the interior ocean, and functions better on

Table 2.4 Target densities for the surface and 2000 m depth (units: kg m⁻³)

Layer	Depth	
	Surface	2000 m
1	00.10	20.10
2	00.20	20.20
3	00.30	20.30
4	00.40	20.40
5	00.50	20.50
6	21.00	30.83
7	21.50	31.11
8	22.00	31.73
9	22.50	32.19
10	23.00	32.68
11	23.50	33.36
12	24.00	33.87
13	24.45	34.22
14	24.90	34.66
15	25.30	35.07
16	25.65	35.50
17	26.00	35.83
18	26.30	36.07
19	26.60	36.25
20	26.84	36.38
21	27.03	36.47
22	27.22	36.65
23	27.38	36.82
24	27.52	36.93
25	27.64	37.01
26	27.74	37.08
27	27.82	37.15
28	27.88	37.19
29	27.94	37.21
30	27.96	37.24

a relatively coarse and unevenly spaced vertical grid (Halliwell 2004). It parameterizes the influence of a larger group of physical processes than other frequently used mixing schemes. The Critical Bulk Richardson Number is set to 0.45 and the Maximum Gradient Richardson Number is 0.7. The order of mixing is of 10⁻⁴.

The other algorithms or parameters applied in the model include a sea ice model comprising an Elastic–Viscous–Plastic dynamics coupled to the thermodynamics for the evolution of sea ice cover by Drange and Simonsen (1996), a bulk heat flux scheme in which the effect of cloud on radiation flux is considered, a relaxation to climatology for the surface salinity flux, and freshwater fluxes from 922 rivers.

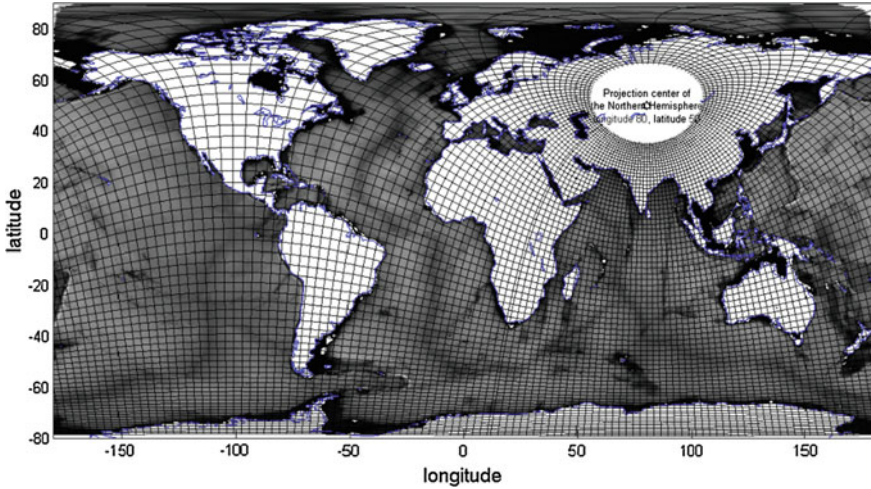


Fig. 2.24 The horizontal grid cells used in the model

The topography used in the model was created from 5-min Gridded Global Relief Data Collection (ETOPO5) with a minimum depth of 10 m, and no smoothing treatment was applied to the topography except for some small islands and coast regions. The forcing data used for driving the model are from monthly mean climatologies of ERA-40 (Uppala et al. 2005). The model was initiated from the climatology of temperature and salinity provided by Polar Hydrographic Climatology (PHC) data, which is a combination of World Ocean Atlas 1998 (WOA98) and AOA (the Environmental Working Group (EWG) Arctic-only climatology) and a static state.

Two types of parallel computation scheme, OpenMP and MPI, can be used separately or together to supply a very high-efficient parallelization. We divided the computation area into smaller ones—a maximum of 15 regions in longitude and 9 regions in latitude—and used 123 CPU cores to implement the parallelization.

2.3.8.2 Preliminary Results from the High-Resolution Global Ocean Model

Only a 42-year model integration was completed because of the model's high resolution and a lack of computing resources. Figure 2.25 shows the global mean of the modeled kinetic energy as a function of time in EXP1 and EXP2. The model showed a clear adjustment in the early years in both EXP1 and EXP2, reaching a relative balance after 35 years, albeit the integration was far away from the model equilibrium. This phenomenon was also found for the simulated SST (not shown). The simulated global mean SSTs in both EXP1 and EXP2 were in a slow state of change, but higher than the climatology of 18.17 °C from PHC data after a rapid

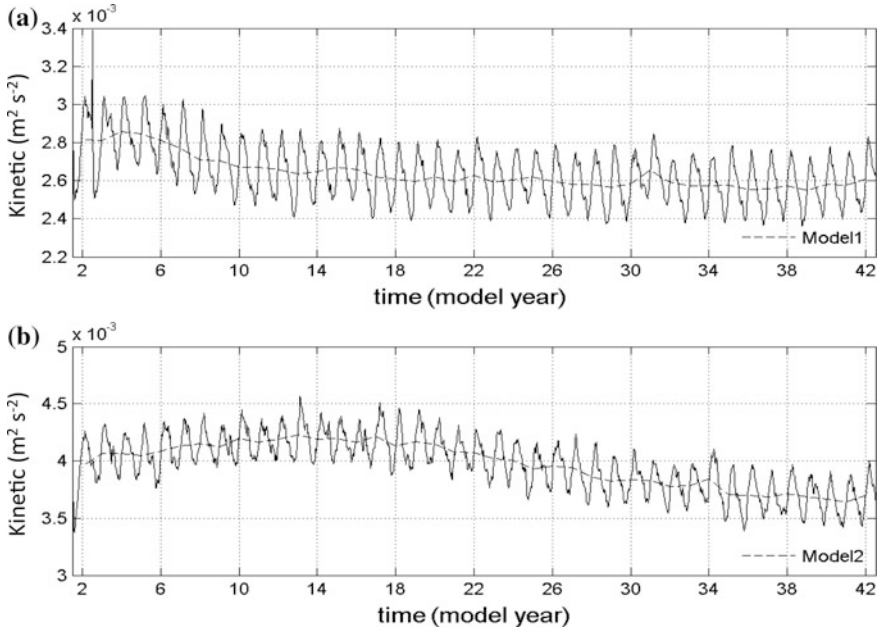


Fig. 2.25 The simulated global mean kinetic energy: **a** EXP1; **b** EXP2

adjustment at the beginning of the integrations. However, the simulated surface salinity decreased during the model integration, meaning that the model is still in the rapid adjustment to a equilibrium in both EXP1 and EXP2, especially in EXP2 where the reference density was at 2000 m depth and the model possibly adjusted slower at the surface than in EXP1.

Figure 2.26 shows the annual mean SST distributions for global ocean. The two simulations captured the main features of SST compared with the observation, e.g., the tropical warm pool in the western Pacific and the cold tongue in the eastern equatorial Pacific, the temperature front in the North Pacific, the Antarctic Circumpolar Current, and so on. However, some departures from the observation were also clear. The simulated cold tongues in both EXP1 and EXP2 were stronger and extended too far to the west—a discrepancy that is commonly found in many other models. Another discrepancy was that the simulated SSTs were higher in most areas than observed, especially in the strong current and coastal regions (maximums of 3 °C). The reason could be attributable to several aspects: One is that the model in both experiments was still under adjustment; the global mean SST showed a slow cooling trend in these two experiments. Another reason is that, in some cold current regions, the simulated currents were not strong enough to provide cold advection and upwelling. Finally, there may be some problems in the surface heat flux algorithm used in this model, suggesting it needs to be tuned more carefully.

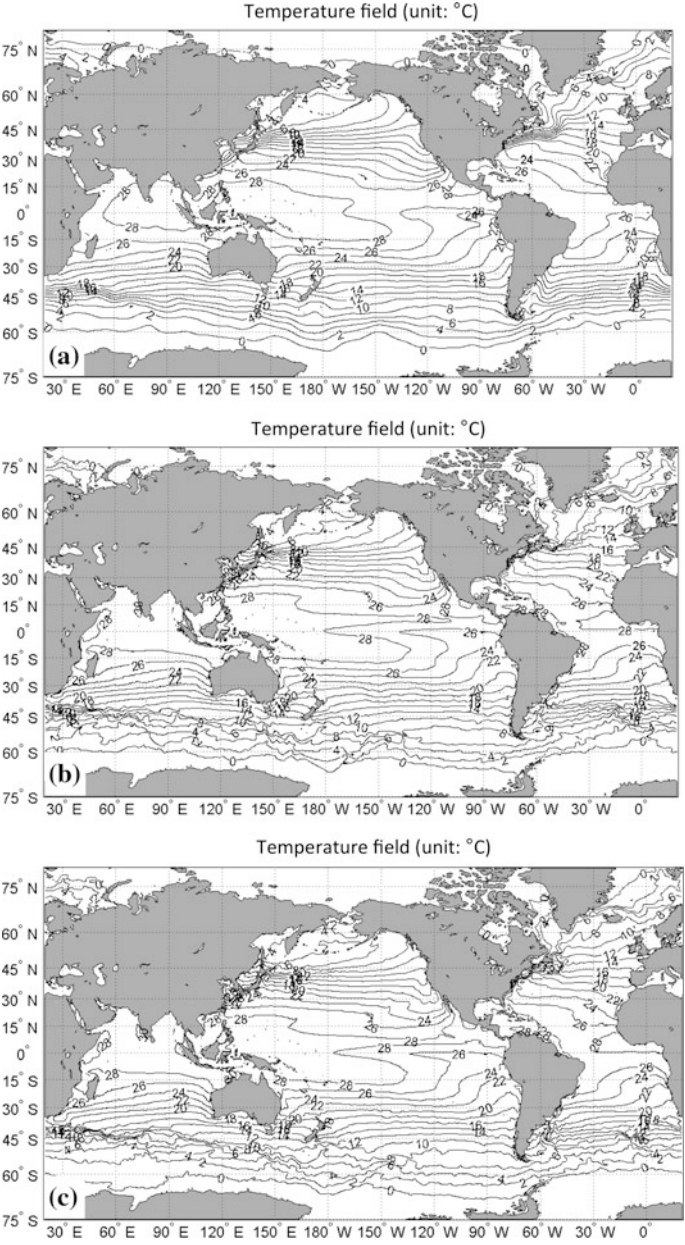
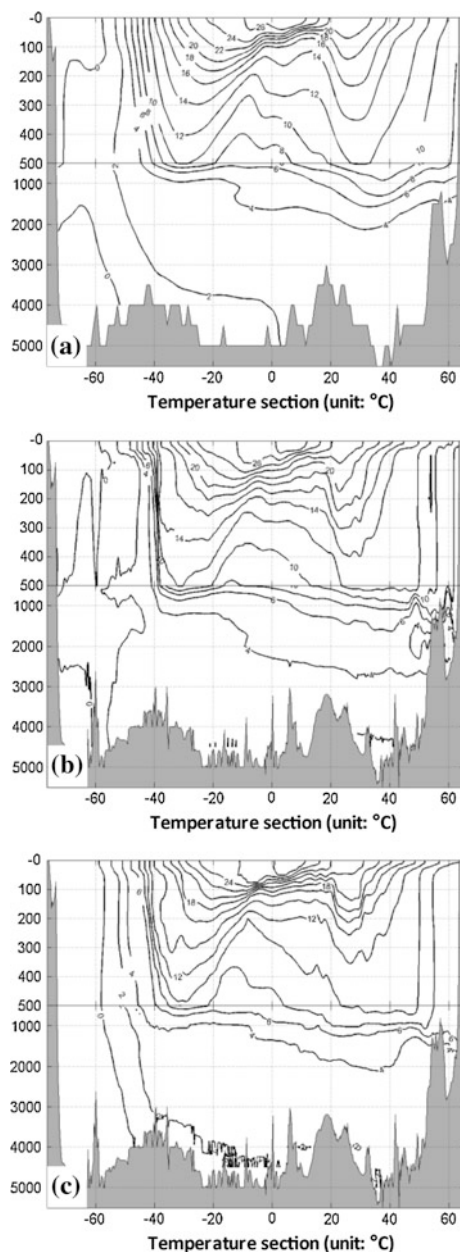


Fig. 2.26 Annual mean SST distributions for global ocean: **a** WOA98; **b** EXP1; **c** EXP2

Figure 2.27 shows the latitude–vertical section of annual mean temperature along 19.5°W. The two experiments were generally consistent with the observed patterns of vertical distribution. But it would have been difficult to give a

Fig. 2.27 Latitude–vertical sections of annual mean temperature along 19.5 °W:
a WOA98; **b** EXP1; **c** EXP2

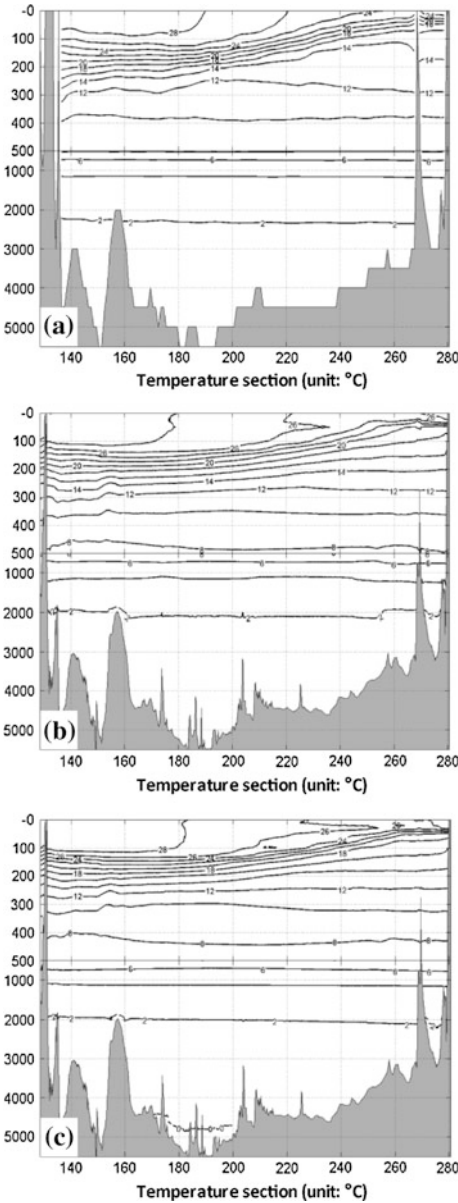


conclusion of achieving better results because the model is far away from reaching equilibrium in the middle and deep ocean, and the information from the initial conditions created from the PHC data, which were a combination of WOA98 and AOA, still dominates the simulated results. Nevertheless, there were also clear

differences between the experiments in their detail compared with the observation. In low latitudes, the position of 4 °C water in EXP2 was higher and better than in EXP1. In the Arctic region, EXP1 produced a better simulation in the upper ocean and EXP2 showed a better deep water representation extending from the Arctic to low latitudes. These differences can be explained from two aspects. First, as mentioned above, the model was expected to be more accurate near the reference density and less accurate as far away from the reference density. And second, the KPP algorithm was used, in which the Richardson Number (Ri) dominates the mixing rate, with a smaller Ri leading to stronger mixing. As the reference level is moved from the surface to 2000 m depth, the Ri will become larger and weaker mixing will occur in the deep ocean as a result. Therefore, greater similarities were observed in EXP2 compared with the observation, since more information from the initial conditions was retained in the very short integration.

In the upper ocean, the thermocline in the tropical Pacific is important for ocean variations related to interannual variability, e.g., ENSO. Figure 2.28 shows longitude–vertical sections of temperature along the Equator in the Pacific. The thermocline is characterized by a steep temperature gradient in the vertical direction, clearly located in the upper ocean with a slope from a depth of about 200 m in the west to only 50 m in the east. Above this, the mixed layer and temperatures are almost unique in the vertical direction. The two experiments reproduced this feature quite well in terms of the pattern, especially in the western Pacific. Nevertheless, some discrepancies were also apparent. The vertical gradient was too weak and the isotherms were too diffused in the eastern equatorial Pacific. The 14 °C isotherm did not rise from the central to the eastern Pacific, indicating that the upwelling along the east coast was not strong enough. This discrepancy in upwelling also led to warmer temperatures at the surface, as mentioned above. Another major weakness was that the slope of the thermocline in the central Pacific was flatter, meaning that the horizontal gradient of temperature/density in this region was weaker than observed. This led to a weaker undercurrent originated from the contrast of densities in the east and west, of which the center of maximum speed located in this region. We found one of the reasons for the above discrepancies could arise from the vertical coordinate settings in the model. Although we set the upper five layers to z -coordinates, they were confined within about 50 m depth. The characteristics of the water in the mixed layer in the tropical Pacific were uniform in the vertical direction, but the difference in density between the east and west was large. Meanwhile, the density became uniform just below the thermocline in the eastern equatorial Pacific, and this led to thicker isopycnal layers in these regions. Although the thermocline structure can now be reproduced to some extent through the use of isopycnal coordinate settings, the resolution in the vertical is not enough to greatly improve the result. The problem is that it is difficult to improve the simulation by increasing the number of density layers only because of the uniformity in the mixed layer and the large east–west contrast in density. As pointed out above, z -level coordinates are propitious to the mixed layer and/or

Fig. 2.28 Longitude–vertical sections of annual mean temperature along the Equator in the Pacific: **a** WOA98; **b** EXP1; **c** EXP2



unstratified seas; therefore, using z -coordinates in the whole thermocline section may be a good approach. To confirm this, we performed another experiment by setting z -coordinates of 10 levels in the upper 200 m in the model, and the primary result showed an obvious improvement in the simulations of both temperature and undercurrents in the tropical Pacific Ocean.

2.3.9 Other Achievements Related to IAP Model Performance

2.3.9.1 Simulation of Intraseasonal Variation of the East Asian Summer Monsoon

East Asia is characterized by a distinct monsoonal climate with a northerly in winter and a southerly in summer. Climatically, the East Asian summer monsoon (EASM) migrates northward in a stepwise fashion, with two northward jumps of the western Pacific subtropical high (WPSH), the first in mid-June and the second in late July (Ding 2007; Su and Xue 2010, 2011). Summer rainfall in China, therefore, can be divided into three distinct stages: the pre-flood period in southern China; the Mei-yu period over the Yangtze–Huihe River Basin; and the rainy season in northern China.

The complex intraseasonal variations of the EASM provide a rigorous standard for the evaluation of GCMs. Although the climatic features of the Asian monsoon including the EASM are basically simulated by the majority of models (Gadgil and Sajani 1998; Kang et al. 2002; Zhou et al. 2009), most fail to capture the intraseasonal features of the EASM. In general, the northward advance of the EASM and the associated rainbelt are not well reproduced in GCMs with resolutions lower than 200 km (Lau and Ploshay 2009). A similar problem also exists in the early GCMs developed at the IAP (Xue et al. 2001).

Recently, the fourth generation IAP AGCM (IAP AGCM4) was developed by Zhang (2009), with a much higher resolution compared with former IAP models. The new model shows much better performance in simulating the basic features of global atmospheric circulation. In order to evaluate the model's performances at different resolutions, model outputs with a resolution of $1.4^{\circ} \times 1.4^{\circ}$ during the period 1979–2008, forced by observed SST, were first used to evaluate the intraseasonal variation of the EASM (Su et al. 2014), because the integration with higher resolution $0.5^{\circ} \times 0.5^{\circ}$ had just been completed. Figure 2.29 shows time–latitude cross sections of observed and simulated pentad mean precipitation in East Asia. While the tropical rainbelt exhibits a clear intraseasonal oscillation, the subtropical rainbelt in East Asia migrates northward in a stepwise fashion, characterized by slow movements and two northward jumps (Fig. 2.29a). The model generally simulated the observed intraseasonal variation in the tropics and subtropics (Fig. 2.29b), except that rainfall was systemically underestimated.

Figure 2.30 shows the temporal evolution of various indices of the WPSH, which is closely related with the movement of rainbelts in East Asia. The simulated ridgeline index and northern borderline index were in agreement with the observation, except that both indices were located somewhat northward (Fig. 2.30a, c). The ridgeline index, in particular, was shifted northward by 5° of latitude. Similar to the rainbelt shown in Fig. 2.29, the simulated WPSH exhibited two jumps with a shorter interval than observed.

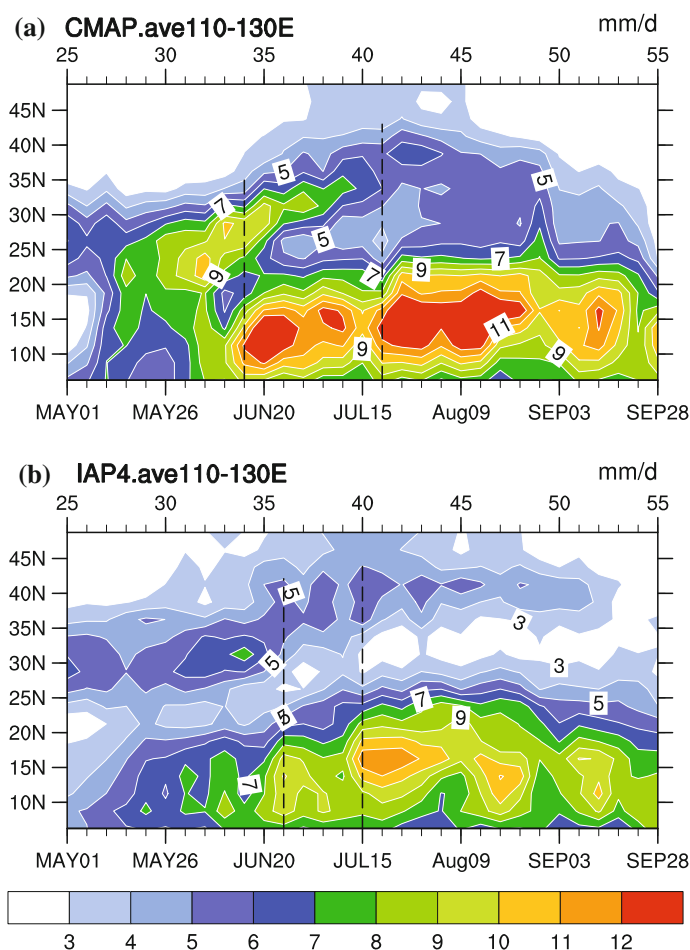


Fig. 2.29 Latitude–time cross sections of pentad mean precipitation averaged over 110°–130°E (units: mm day^{-1}): **a** observation; **b** simulation. The bold dashed lines indicate the jump stages in the observation and simulation. Reprinted from Su et al. (2014), with kind permission from Springer Science+Business Media

The simulated western extension index and area index were relatively poor (Fig. 2.30b, d). Although the variations before mid-July were generally simulated, the model failed to simulate a rapid eastward retreat and sharp reduction in intensity after mid-July. Compared with June and July, the observed WPSH in August retreats eastward with weak intensity. The simulated WPSH, however, tended to be located more northwestward with stronger intensity. As a result, a rainbelt still existed in August in East Asia (Fig. 2.29b).

During the northward jump of the WPSH, a remarkable change is found in the circulation pattern, which can be well represented by the interpentad difference

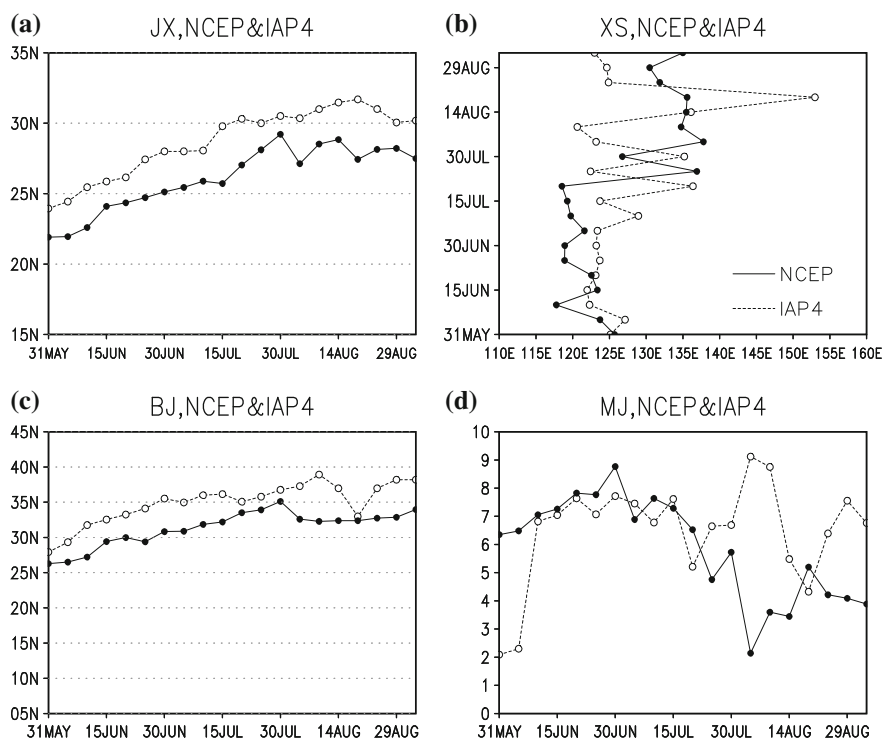


Fig. 2.30 Time series of various indices of the western Pacific subtropical high in the observation (solid circles, solid lines) and simulation (open circles, dashed lines): **a** ridgeline index (units: °N); **b** westward extension index (units: °E); **c** northern borderline index (units: °N); **d** area index (10^6 km^2). The dashed circles in (a) indicate the jump stages in the observation and simulation. Reprinted from Su et al. (2014), with kind permission from Springer Science+Business Media

between the jumping pentad and the previous pentad (Su and Xue 2010). Figure 2.31 shows the observed and simulated interpentad differences of 850 hPa wind and outgoing LW radiation (OLR) during the jump stages. Due to enhanced convection over the tropical western Pacific at P34 (15–19 June), a Rossby wavetrain appears, as represented by a cyclonic or anticyclonic anomaly alternating from the Philippines to the Bering Strait (Fig. 2.31a). This feature was generally simulated by the model, except that the propagation path was located more westward because the observed OLR anomaly near the Philippines shifted to the South China Sea in the model (Fig. 2.31b). Different from the first jump, the second one is not only related to the propagation of the Rossby wavetrain from the warm pool region, but also to the downstream propagation of Rossby waves in the high latitudes of Eurasia (Fig. 2.31c). The model simulated the second jump well due to a good representation of the phase-locking of the two factors (Fig. 2.31d). It was also noted that the second jump was not as evident as that in the observation.

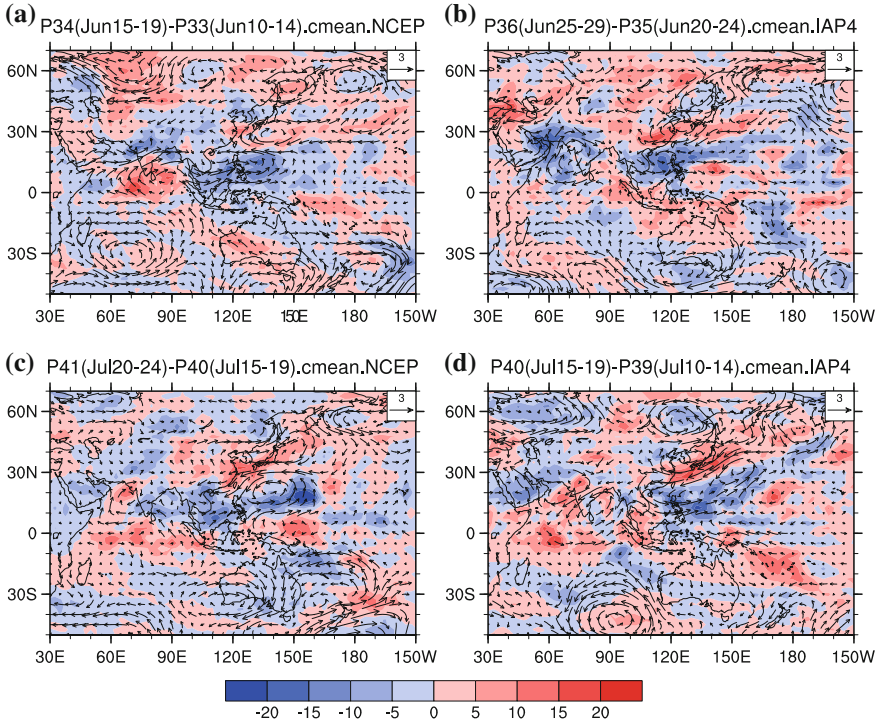


Fig. 2.31 The interpentad differences of 850 hPa wind (units: m s^{-1}) and OLR (units: W m^{-2}) during the jump stages in the observation (*left panels*) and simulation (*right panels*): **a** difference between P34 (15–19 June) and P33 (10–14 June)—the observed first jump; **b** difference between P36 (25–29 June) and P35 (20–24 June)—the simulated first jump; **c** difference between P41 (20–24 July) and P40 (15–19 July)—the observed second jump; **d** difference between P40 (15–19 July) and P39 (10–14 July)—the simulated second jump. Reprinted from Su et al. (2014), with kind permission from Springer Science+Business Media

The discrepancies in the northward jumps of the WPSH between the simulation and observation were possibly related to the climatological intraseasonal oscillation (CISO) over the warm pool region in the model. Figure 2.32 shows the observed and simulated OLR in the warm pool region from 1 April to 1 October, together with a power spectrum analysis. The observed OLR over the warm pool decreases from 260 W m^{-2} in April to 220 W m^{-2} in summer, with the lowest value in August. Although the model simulated this variation well, it overestimated OLR during the whole period (Fig. 2.32b), showing weaker convective activities in the tropical western Pacific to not only appear in summer, but also in other seasons. In particular, an OLR minimum appears during the jump stage of the WPSH (mid-June and late July), indicating that the jumps of the WPSH are associated with the CISO in the warm pool (Fig. 2.32c). A similar phenomenon was also found in the model, but with a shorter interval between the two minimums (Fig. 2.32d). It

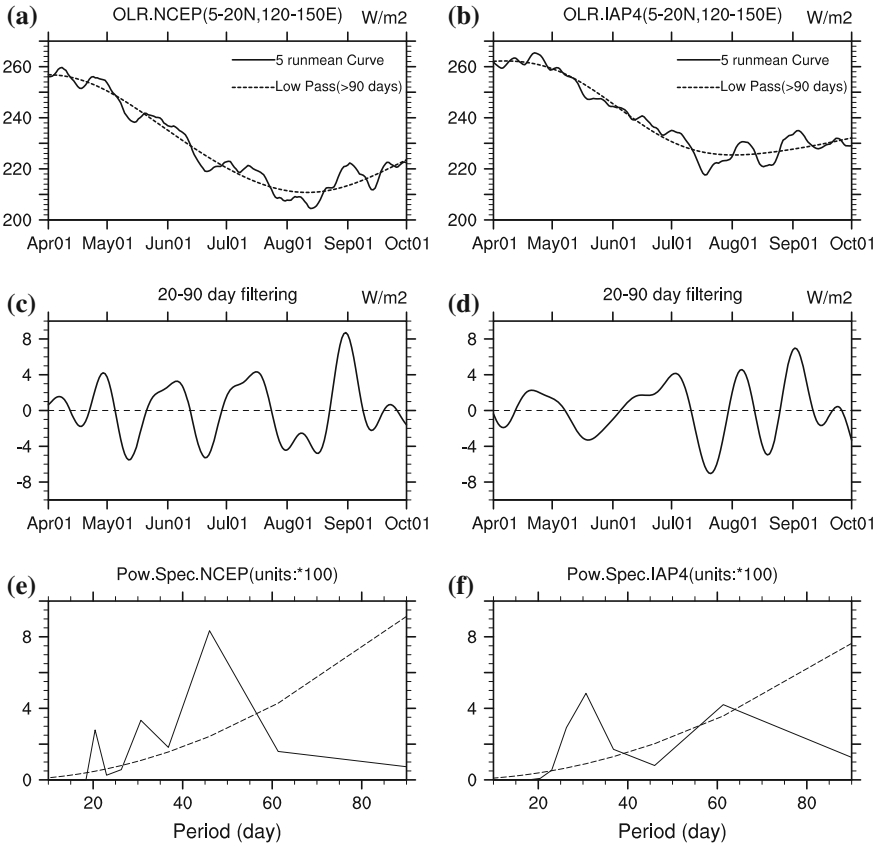


Fig. 2.32 **a, b** The time series of OLR over the warm pool region (5° – 20° N, 120° – 150° E) (units: W m^{-2}), in which the *solid* and *dashed* lines represent the original curve with a 5-day running mean and base annual cycle, respectively; **c, d** 20–90-day filtered time series of OLR over the region; **e, f** the corresponding power spectrum (*solid line*) and standard spectrum of red noise at the 5 % significance level. For the sake of clarity, the observed and simulated power spectra including the significance line are all scaled by 100. The observation and simulation are listed in the *left* and *right panels*, respectively. Triangles in (c) and (d) indicate the two jump stages in the observation and simulation, respectively. Reprinted from Su et al. (2014), with kind permission from Springer Science+Business Media

can be concluded that the phase disagreement of CISO in the warm pool region leads to the jump bias in the model.

Since the model simulates the northward propagating signals in the tropical western Pacific, the bias in the tropical northwestern Pacific intraseasonal oscillation could be attributable to the weak eastward propagation of Madden–Julian oscillation originating from the equatorial Indian Ocean. There are four possible reasons responsible for the weak eastward propagating signals in the model: the relatively lower horizontal resolution; the lack of air–sea interaction; weaker

westerlies from the Indochina Peninsula to the tropical western Pacific; and imperfect convective parameterization. The model's resolution, along with the land surface model and cumulus parameterization scheme, should further be improved in the near future in order to better simulate the intraseasonal variation of the EASM.

2.3.9.2 Simulation of Interdecadal Variation of the East Asian Summer Monsoon and Summer Rainfall in Eastern China

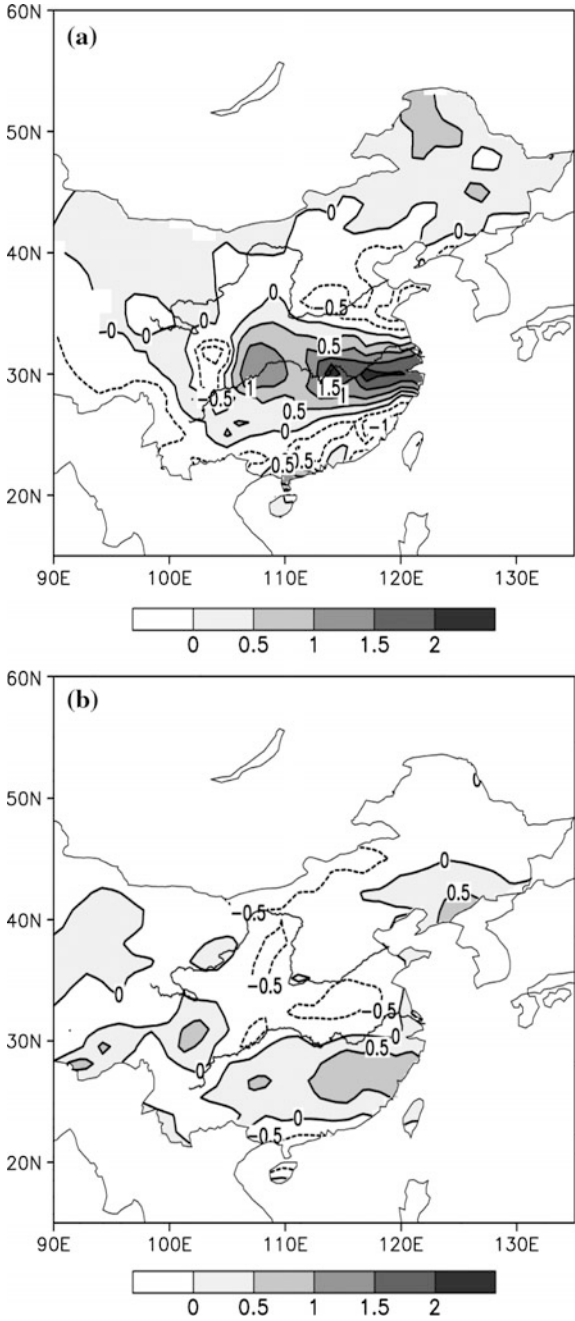
The EASM and summer rainfall in eastern China exhibits a significant decadal variation. The most significant decadal weakening of the EASM occurred in the late 1970s, characterized by a sharp reduction of the summer southwesterly and changes in the summer rainfall pattern in eastern China (Wang 2001; Yu et al. 2008). Observations show that this decadal variation is closely related with SST anomalies in the tropical Pacific and Indian Ocean (Xue 2001). Furthermore, it is also noted that the phase transition of the Pacific decadal oscillation (PDO) plays a role in the decadal variation (Yang et al. 2005). Forced by observed SST, most atmospheric models can simulate the decadal variation of EASM circulation, but they fail to simulate the summer rainfall changes (Zhou et al. 2009). Due to monsoon intensity being related to the thermal contrast between land and ocean, however, the influences of land surface processes have also been highlighted (Xu et al. 2007). Until now, it is unclear which factor plays the major role in this decadal variation. Besides, it is not well understood why most models fail to simulate the summer rainfall despite being able to simulate the EASM circulation.

Forced by observed SST from 1958 to 1999, IAP AGCM4 with a resolution of $1.4^{\circ} \times 1.4^{\circ}$ was used to simulate the decadal variation of the EASM and summer rainfall in eastern China during the late 1970s (Chen and Xue 2013). Based on previous studies, the difference between 1979–99 and 1958–78 was used to represent the decadal variation. Figure 2.33 shows the decadal variation of summer rainfall in eastern China. The observation shows that compared with 1958–78, a decadal variation of summer rainfall occurred in 1979–1999. While rainfall in the Yangtze River Basin tended to increase with the center over 2 mm day^{-1} , rainfall in North and South China tended to decrease. The above anomalous rainfall pattern was generally simulated by IAP AGCM4, except that the pattern was located somewhat southward than observed.

The decadal variation of summer rainfall pattern is closely related with EASM circulation. As shown in Fig. 2.34, a significant anomalous northerly appears along the coastal region of East Asia, reducing the southwesterly monsoon in summer. In this case, the moisture is not easily transported to North China. Besides, there is an anomalous cyclonic circulation from the Yangtze River Basin to Japan, which is favorable for an increase in summer rainfall in this region. The model simulated the anomalous northerly and cyclonic circulation, as shown in Fig. 2.34b.

To further understand the role of SST in the decadal variation of the EASM, singular value decomposition (SVD) was used to analyze the relationship between SST and EASM. Figure 2.35 shows the heterogeneous correlation of the first SVD

Fig. 2.33 The decadal variation of summer rainfall in eastern China (units: mm day^{-1}): **a** observation; **b** simulation. Reprinted from Chen and Xue (2013)



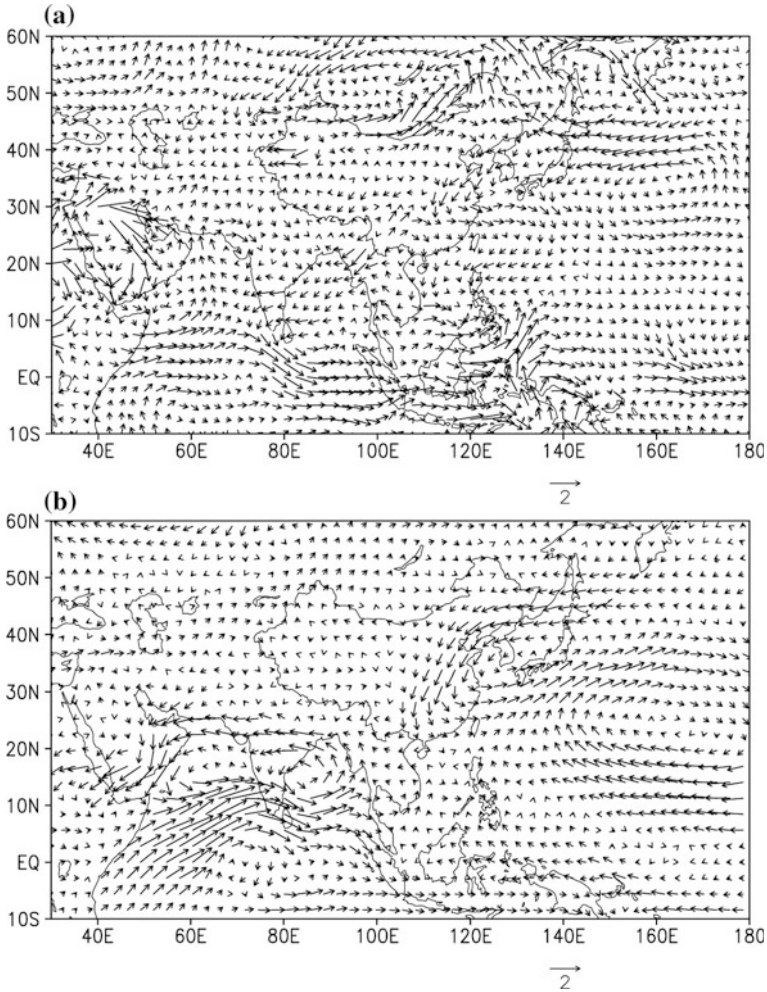


Fig. 2.34 The decadal variation of 850 hPa wind (units: m s^{-1}): **a** ECMWF reanalysis; **b** simulation. Reprinted from Chen and Xue (2013)

mode between summer rainfall in eastern China and SST. The correlation in summer rainfall indicates that there is a positive correlation to the south of the Yangtze River Basin and a negative correlation outside of this region (Fig. 2.35a), and the anomalous center is located southward compared to the observation. The SST correlation shows that, while there is a negative correlation in the North Pacific, there is a significant correlation in the tropical oceans, especially in the eastern Pacific and Indian oceans (Fig. 2.35b). The corresponding time series shows that the decadal variation occurred in the late 1970s (Fig. 2.35c). A further analysis revealed that the decadal variation of SST is related with the phase transition of the

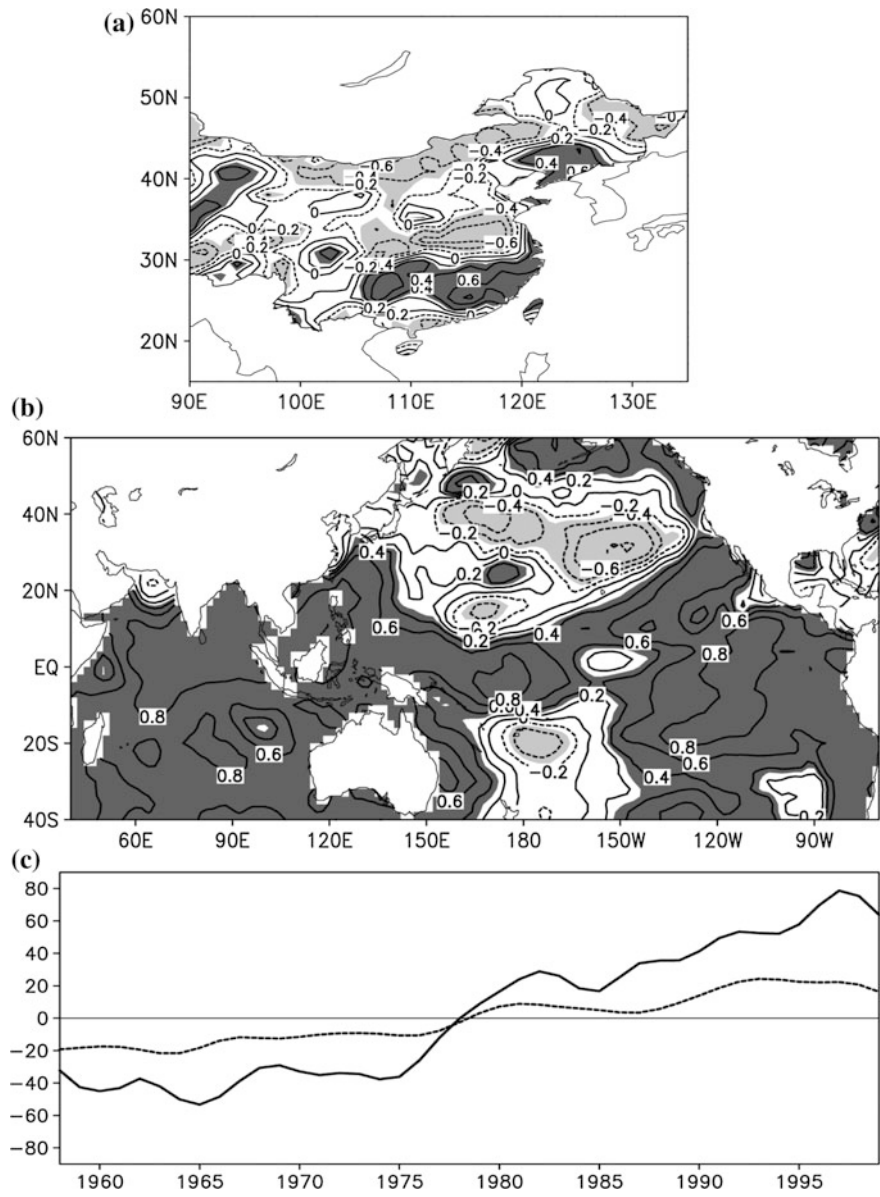


Fig. 2.35 The simulated heterogeneous correlation of the first SVD mode between summer rainfall in eastern China and SST, and the corresponding time series [regions above the 95 % significance level are shaded (units: dimensionless)]: **a** summer rainfall in eastern China; **b** SST; **c** the corresponding time series of SST (*solid line*) and rainfall (*dashed line*). Reprinted from Chen and Xue (2013)

PDO. After the late 1970s, the PDO shifted from a negative phase to a positive one, resulting in a warming in the tropical oceans and a cooling in the North Pacific.

Figure 2.36 shows the decadal variation of summer surface air temperature over Asia. After the late 1970s, some regions cooled, although global warming is evident in most regions (Fig. 2.36a). A significant cooling is found in the Yangtze River Basin, with a central value of over 0.5 K. Hence, the combined effects of a cooling

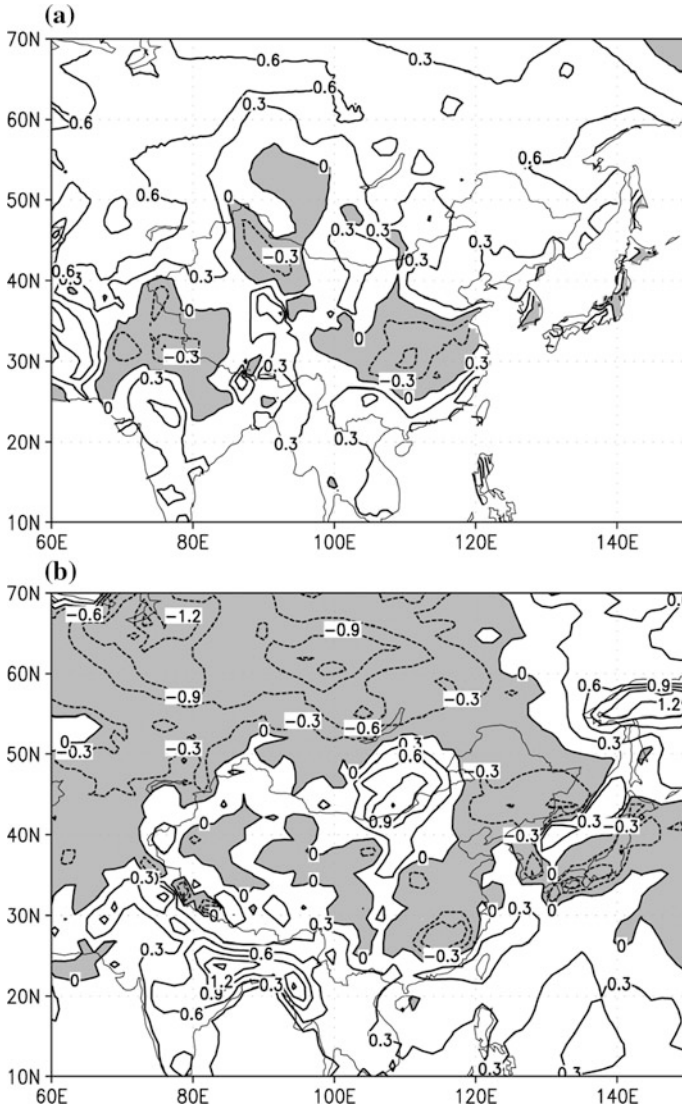


Fig. 2.36 The decadal variation of summer surface air temperature over Asia [regions with a negative value are shaded (units: K)]: **a** observation from the Climatic Research Unit; **b** simulation. Reprinted from Chen and Xue (2013)

over the continent and warming over the ocean leads to a reduced thermal contrast and a weakened EASM. Prescribed with observed SST, the model generally simulated the cooling trend in the Yangtze River Basin, but the center tended to be located slightly southward (Fig. 2.36b). Since cooling over the East Asian continent is also evident, the decadal variation of the EASM has been ascribed to cooling over the continent in some previous studies. The simulation by IAP AGCM4 indicated that the decadal variation in SST plays a major role in the EASM, and the cooling over the continent is a result of warming in the ocean.

On the other hand, the simulation produced opposite results to the observation in the high latitudes of Asia. While the observation shows a warming trend with a central value of over 0.6 K, a cooling trend was found in the simulation. It is widely accepted that the global warming signal associated with the increase in greenhouse gases is more evident at high latitudes, so it is reasonable that the model fails to simulate the warming trend at high latitudes with a constant level of greenhouse gases. In addition, the result indicates that the warming in the tropical oceans is essentially different from global warming. While global warming is related with increased greenhouse gases, the warming in the tropical oceans results from the phase shift of the PDO. It can be concluded that the PDO plays a primary role in the decadal variation of the EASM.

Previous studies have shown that, forced by observed SST, most atmospheric models can simulate the circulation changes associated with the decadal weakening of the EASM (e.g., the WPSH), but they fail to simulate the anomalous rainfall pattern in eastern China. This is possibly related to lower resolutions in early models. Although the large-scale circulation pattern was generally captured by these models, the rainfall pattern could not be simulated due to their lower resolutions. As for air–sea coupled models, performances are poorer than in atmospheric models. While most models are successful in simulating the spatial pattern of the PDO, few models can simulate the anomalous rainfall pattern. Some key features cannot be reproduced in these models, such as the cooling trend over the East Asian continent mentioned above. A large discrepancy among models is found, and simulation results are model-dependent to a large degree. This poses a great challenge for EASM prediction at the decadal timescale. When predicting the decadal change in the EASM, it is necessary to select models that demonstrate good performance.

2.4 Summary

IAP AGCM4.0, with its high resolution of $0.5^\circ \times 0.5^\circ$, was developed based on the previous version with a resolution of $1.4^\circ \times 1.4^\circ$. Several new methods, such as a flexible leaping-point scheme, time splitting methods, and a 2D parallel algorithm, were introduced to the dynamical core to make the high-resolution model more efficient and stable. Thirty-year (1979–2008) AMIP integrations of this high resolution model have been completed, and a preliminary evaluation of the results

implies that the simulated zonal mean and global distribution of precipitation and SLP agree well with observations. The seasonal migration of rainbelts over the East Asian monsoon area simulated with the high-resolution model is better than that with the coarse-resolution model, as compared with GPCP data. The sensitivity of the simulated climate to different dynamical cores has also been investigated.

With respect to physical process parameterizations, the CAR system has also been fully evaluated. The spread of radiation transfer calculations among different cloud/radiation scheme combinations was quantified with 448 selected CAR members. The discrepancies in the treatments of cloud fields that were originally adopted by different radiation codes were found to be the major reason for the current level of model spread, in which the dominant roles of cloud subgrid-scale structures were generally demonstrated. The CAR system has been incorporated into IAP AGCM4 and numerical simulations by selecting some of the parameterizations in CAR are in progress.

A global ocean model with a resolution of 7–30 km was set up based on HYCOM. Two experiments by setting different reference densities at the surface and at a depth of 2000 m for vertical coordinates were performed, and the results showed that the model reproduced the main features of global ocean circulations to some extent in both experiments. Further sensitivity experiments showed that the selection and setting of vertical coordinates in the model has a great effect on simulations.

The performances of the previous version of IAP AGCM4, with a resolution $1.4^\circ \times 1.4^\circ$, in simulations of the intraseasonal variation and decadal variation of the EASM were evaluated. The main features of these two phenomena were to some extent better simulated in comparison with previous coarser models, although some discrepancies were also clear compared with observations. These results are helpful for evaluating and understanding the performances of the high-resolution model.

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