Evaluating Low-Cloud Simulation from an Upgraded Multiscale Modeling Framework Model. Part I: Sensitivity to Spatial Resolution and Climatology

KUAN-MAN XU
Climate Science Branch, NASA Langley Research Center, Hampton, Virginia

ANNING CHENG

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ABSTRACT

The multiscale modeling framework, which replaces traditional cloud parameterizations with a 2D cloud-resolving model (CRM) in each atmospheric column, is a promising approach to climate modeling. The CRM component contains an advanced third-order turbulence closure, helping it to better simulate low-level clouds. In this study, two simulations are performed with 1.9 × 2.5 grid spacing but they differ in the vertical resolution. The number of model layers below 700 hPa increases from 6 in one simulation (IP-6L) to 12 in another (IP-12L) to better resolve the boundary layer. The low-cloud horizontal distribution and vertical structures in IP-12L are more realistic and its global mean is higher than in IP-6L and closer to that of CloudSat/Cloud–Aerosol Lidar and Infrared Pathfinder Satellite Observations (CALIPSO) observations. The spatial patterns of tropical precipitation are significantly improved; for example, a single intertropical convergence zone (ITCZ) in the Pacific, instead of double ITCZs in an earlier study that used coarser horizontal resolution and a different dynamical core in its host general circulation model (GCM), and the intensity of the South Pacific convergence zone (SPCZ), and the ITCZ in the Atlantic is more realistic. Many aspects of the global seasonal climatology agree well with observations except for excessive precipitation in the tropics. In terms of spatial correlations and patterns in the tropical/subtropical regions, most surface/vertically integrated properties show greater improvement over the earlier simulation than that with lower vertical resolution. The relationships between low-cloud amount and several large-scale properties are consistent with those observed in five low-cloud regions. There is an imbalance in the surface energy budget, which is an aspect of the model that needs to be improved in the future.

1. Introduction

A multiscale modeling framework (MMF) model uses a two-dimensional (2D) cloud-resolving model (CRM) in each atmospheric grid column of a general circulation model (GCM) to represent cloud dynamics and cloud physical processes in place of cloud parameterizations in a conventional GCM (Grabowski 2001; Khairoutdinov and Randall 2001, hereafter KR01). The MMF approach has demonstrated the ability to simulate modes of convective variability such as intraseasonal oscillations, higher-frequency tropical waves, and diurnal cycles of precipitation (KR01; Khairoutdinov et al. 2005, hereafter KRD05; Tao et al. 2009). However, the embedded CRM can only partially resolve small cloud-scale circulations, causing an inadequate simulation of low-level clouds (Cheng and Xu 2008, 2011; Khairoutdinov et al. 2008, hereafter KRD08; DeMott et al. 2010, hereafter DRK10). This is because low-level clouds are associated with large turbulent eddies whose circulations are not resolved by the embedded CRM with horizontal grid sizes on the order of a few kilometers. These eddies are typically a few hundred meters in size, too small for an embedded CRM to explicitly simulate. As a result, this MMF, known as the superparameterized Community Atmosphere Model (SPCAM) as a result of its use of the Community Atmosphere Model (CAM; Collins et al. 2006) as its host GCM, does not simulate the global climatology of low clouds well; that is, it produces too little in the cool subtropical ocean regions in which stratocumulus clouds prevail (KRD08; Marchand and Ackerman 2010; Cheng

Corresponding author address: Dr. Kuan-Man Xu, NASA Langley Research Center, Climate Science Branch, Mail Stop 420, Hampton, VA 23681.
E-mail: kuan-man.xu@nasa.gov

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and Xu 2011, hereafter CX11). When SPCAM was coupled to an ocean model, this deficiency led to warm biases in sea surface temperature (SST) in the eastern subtropical oceans (Stan et al. 2010). These biases were also seen in conventional coupled ocean–atmosphere GCMs (Ma et al. 1996; Yu and Mechoso 1999). The primary reason for the SST biases is the excessive amount of solar radiation reaching the ocean surface resulting from the lack of low-level clouds.

To produce more realistic simulations of low clouds, two parallel strategies have been proposed to improve the MMF approach. One of them is to drastically increase both the horizontal and vertical resolutions of the embedded CRM. Because of 200-fold increase\(^1\) in the computing time of an MMF over a conventional GCM at comparable horizontal and vertical resolutions, increasing the resolution of the embedded CRMs to fully resolve low-level clouds will be too costly for long-term climate simulations. Marchand and Ackerman (2010) modestly improved several aspects of the MMF low-level clouds by increasing both horizontal (from 4 km to 1 km) and vertical resolutions simultaneously. Another strategy is to improve the CRM with a more sophisticated subgrid turbulence scheme, which has shown promise in previous CRM and single-column model studies (e.g., Bougeault 1981; Lappen and Randall 2001; Golaz et al. 2002; Cheng and Xu 2006, 2008). CX11 recently adopted this approach to upgrade the embedded CRM with an advanced third-order turbulence closure [intermediately prognostic higher-order closure (IPHOC); Cheng et al. 2004; Cheng and Xu 2006]. They showed a greatly improved simulation of low-level clouds over the tropical and subtropical oceans despite the very coarse horizontal resolution (at T21) used in the host GCM. The computational time of the upgraded MMF, SPCAM–IPHOC, is about double that of the standard SPCAM.

CX11 attributed the improved simulation of low-level clouds not only to the representation of subgrid-scale (SGS) condensation in the embedded CRM, but also to the increases in surface sensible heat fluxes, the lower-tropospheric stability (LTS, defined as the difference in potential temperature between 700 hPa and the surface; Klein and Hartmann 1993), and stronger longwave radiative cooling. The SGS condensation scheme in SPCAM–IPHOC considers the SGS variabilities in thermodynamic and dynamic variables within a CRM grid cell, which are ignored in SPCAM (KR01), to determine whether or not a fraction of the grid cell can form a cloud. Large turbulent eddies and their vertical transports are parameterized in the CRM with IPHOC while they depend largely on the CRM grid-scale circulations in SPCAM. That is, the cloud formation process shifts to smaller spatial scales in the upgraded CRM, which may feed back more realistically to large-scale circulations in CAM than the standard CRM does.

CX11 further examined the detailed characteristics of a vertical cross section representative of the stratocumulus to trade cumulus transition along the southeastern Pacific, that is, along 15°S off the coast of South America. The SPCAM–IPHOC simulation produced reasonable cloud amounts in the cumulus regions but still underestimated the shallower clouds, especially the stratocumulus clouds near the coast, compared to observations from the merged CloudSat (Stephens et al. 2002), Cloud–Aerosol Lidar and Infrared Pathfinder Satellite Observations (CALIPSO; Winker et al. 2010), Clouds and the Earth’s Radiant Energy System (CERES; Wielicki et al. 1996), and Moderate Resolution Imaging Spectroradiometer (MODIS; King et al. 1992) data product (C3M; Kato et al. 2011). The vertical structures of simulated clouds were still very different from those observed. These differences were mainly related to the insufficient model resolution used in SPCAM–IPHOC. The host GCM, the CAM version 3 (Collins et al. 2006), only has six model layers between the surface and 700 hPa in its standard configuration. It cannot simulate the strong boundary layer inversion associated with stratocumulus clouds, which, in turn, degrades the simulation of these clouds in the embedded CRM. In a study with a conventional GCM, Bushell and Martin (1999) concluded that the full benefit of an improved boundary layer scheme could not be realized if the boundary layer structure was constrained by the rather poor lower troposphere resolution of the GCM. Thus, the exploration of the full benefit of SPCAM–IPHOC with higher lower-troposphere resolution is a primary motivation for the present study.

The goal of this series of papers is to evaluate the SPCAM–IPHOC performance with regard to its low-cloud simulation using state-of-the-art observations. This study (Part I) will present results on the sensitivity of low-cloud simulation to spatial resolution and a global seasonal climatology for the high-resolution simulation. The relationships between low clouds and large-scale variables at selected ocean basins are also examined. Part II will present the seasonal variations of low clouds in the eastern Pacific and the mechanisms of their seasonal variations (Xu and Cheng 2013). Part III will compare the cloud regime transitions in the northeast Pacific.

\(^1\) The exact amount of increase in computational costs depends upon the horizontal grid spacing of the host GCM and the number of CRM columns in each GCM grid cell, as well as the complexity of cloud microphysics parameterization (Tao et al. 2009). The 200-fold increase results from T42 GCM resolution and 32 CRM columns (KR01).
simulated from SPCAM, SPCAM–IPHOC, and a conventional GCM in order to understand the similarities and differences in the simulated seasonal-mean and instantaneous transitions (Cheng and Xu 2013).

To further improve the simulation of low clouds, we double the number of the vertical layers in the host GCM and the embedded CRM from 6 to 12 below 700 hPa while keeping the same model layers in the middle and upper troposphere. Offline testing was performed to assess the improvement before this modification was adopted in the MMF. The horizontal resolution of the host GCM also increases by choosing the finite volume dynamical core available in CAM, version 3.5 (CAM3.5), instead of the semi-Lagrangian dynamical core used in KRD05 and CX11. As shown later, this change of dynamical core and horizontal resolution has a pronounced impact on the simulation of tropical precipitation, but does not impact the simulation of tropical and subtropical low clouds much. The present study has three specific objectives: to evaluate and compare the performance of model simulations against state-of-the-art observations, to provide a global seasonal climatology for the high-resolution simulation, and to examine the physical relationships between low clouds and large-scale variables. The global climatology can be compared with similar simulations (Collins et al. 2006; KRD05); DRK10 examined some relationships between low-cloud and large-scale variables. The global climatology can be compared with similar simulations (Collins et al. 2006; KRD05); DRK10 examined some relationships between low-cloud and large-scale variables simulated with the standard SPCAM. The same relationships simulated by SPCAM–IPHOC will be examined to better understand the controlling processes.

The rest of this paper is organized as follows. The SPCAM–IPHOC model will be briefly described in section 2, along with the offline testing of the sensitivity of the embedded CRM to vertical resolution. Results from the low- and high-resolution SPCAM–IPHOC simulations are compared in section 3. Section 4 presents the climatology of the high-resolution simulation and the relationships between low-cloud and large-scale variables. Section 5 provides the conclusions and discussion.

2. Models, experiment design, and offline testing

The SPCAM–IPHOC used in this study consists of CAM, version 3.5 (Collins et al. 2006), and 2D system for atmospheric modeling (SAM) CRM embedded in each GCM atmospheric grid column (KR01). Its CRM component has been upgraded with the intermediate prognostic higher-order turbulence closure in place of a lower-order turbulence closure (CX11). The large-scale atmospheric circulation is represented on the coarse-resolution grid of the GCM via the finite volume dynamical core (Lin 1997). In CX11, however, the semi-Lagrangian dynamical core at T21 resolution was used in simulations with and without the IPHOC upgrade. There were 26 layers in the vertical including 6 layers below 700 hPa. The physical processes such as convection and stratiform cloudiness, usually parameterized in a conventional GCM, are resolved explicitly (more precisely, “cloud permitting”) on the CRM fine grid cells. Cloud microphysics, turbulence, and radiation are parameterized at the CRM scale, and they interact with each other at much smaller spatial and temporal scales than in conventional GCMs. Tendencies of heat and moisture from the CRM scale are communicated to the large scale via the GCM. The details on the coupling between the CRM and CAM3.5 were discussed in KRD05.

The upgraded CRM embedded in the SPCAM–IPHOC was described in detail in CX11. IPHOC assumes a joint double-Gaussian distribution of liquid water potential temperature, total water, and vertical velocity (Cheng and Xu 2006). The distribution is inferred from the first-, second-, and third-order moments of the variables given above and is used to diagnose the cloud fraction and grid-mean liquid water mixing ratio, as well as the buoyancy terms and fourth-order terms in the equations describing the evolution of the second- and third-order moments. These higher-order moments, which are not available in a low-order closure, are used to formulate the SGS condensation in IPHOC. IPHOC uses the first-order moments predicted by the SAM CRM, and the prognostic equations for the second- and third-order moments are shown in detail in CX11. The upgraded turbulence scheme is applied to the entire troposphere.

The upgraded 2D CRM is used to perform offline testing of the sensitivity of low-cloud simulation to vertical resolution here. Large-eddy simulation (LES) is used as a benchmark using the 3D LES version of SAM with a horizontal grid spacing of 100 m and vertical grid spacings of 25–40 m (Cheng and Xu 2008). Three CRM experiments are performed, one with the identical vertical resolution (60 or 75 layers depending upon cloud regime) to that used in LES (LL) and the other two with 6 or 12 layers below 700 hPa (6L or 12L) and the same vertical grid as that used in CAM3.5 above 700 hPa. Three different low-cloud regimes, cumulus [Barbados Oceanographic and Meteorological Experiment (BOMEX)], transitional cumulus–stratocumulus [Atlantic Trade Wind Experiment (ATEX)], and stratocumulus [Atlantic Stratocumulus Transition Experiment (ASTEX)] are simulated. The details of the LES and experiment LL and their configurations are given in Cheng and Xu (2008), which examined the sensitivity of CRM simulations to horizontal grid spacing and compared the performance of the CRMs with the low- and third-order closures. Experiment LL corresponds to those
with IPHOC using the 4-km horizontal grid spacing in Cheng and Xu (2008). The new experiments, 6L and 12L, further degrade the vertical resolutions of the CRM in order to match those used in the standard and modified vertical coordinates of CAM3.5.

Figure 1 shows the time–height cross sections of simulated cloud fractions from the LES, LL, 12L, and 6L experiments for three cloud regimes. It can be seen that the shallow cumuli are fairly reasonably simulated when there are only six layers below 700 hPa (Fig. 1d). The slightly higher-resolution experiment (12L) produces slightly smaller cloud fractions (Fig. 1c) and agrees better with LES (Fig. 1a) than 6L, while further increased resolution in LL (75 layers) does not significantly improve the performance (Fig. 1b). These results suggest that vertical resolution is not critical to model performance due to the fact that the boundary layer is not highly stratified and fewer model layers are needed to characterize the thermodynamic structures associated with this cloud regime (Figs. 2a,b). Also, the vertical profiles of the total transports of heat and moisture are reasonably comparable to those from LES (not shown), which play a significant role in maintaining the thermodynamic profiles and thus the vertical cloud structures.

For the transitional cumulus–stratocumulus regime, it is more difficult for the CRM to reproduce the LES results (Fig. 1e). Roughly speaking, the CRM results are degraded in the sense that the thin stratocumulus clouds disappear at ~1.5 km and the maximum cloud center appears at lower heights (~1 km) when the number of vertical layers decreases from 75 (LL) to 12 to 6 (Figs. 1f–h) because the stratocumulus clouds are too thin to be resolved. It should be noted that the benefit of additional layers in 12L is somewhat pronounced. The more complicated thermodynamic structures for this cloud regime (e.g., relatively strong inversion at cloud top, well-mixed layer below cloud base, and a less-mixed cloud layer) are the reasons for the poorer performance (Figs. 2c,d). Note that diimensionality may also play a role in contributing to the
difference between the LES and CRM simulation results for this cloud regime.

For the ASTEX stratocumulus regime, 6L substantially underestimates the cloud amount within the cloud layer (Fig. 1l), while 12L does not, except for the top portion of the cloud layer (Fig. 1k). Therefore, the additional layers help the simulation of stratocumulus clouds, as expected, because the inversion is better resolved. The decreased magnitude of the inversion strength relative to LL (60 layers) is the primary reason for the underestimate of cloud amount at the top portion of the cloud layer. The primary effect of coarsening vertical grid

Fig. 2. Vertical profiles of (left) liquid water potential temperature and (right) total water mixing ratio for (a),(b) BOMEX; (c),(d) ATEX; and (e),(f) ASTEX averaged over last 2 h of each simulation. Results from four experiments are shown in each panel.
spacing is the smoothing of the mean thermodynamic profiles (Figs. 2e,f). Underestimate of inversion leads to more mixing, which decays clouds and results in an underestimate of the cloud amount for the stratocumulus regime. Therefore, it is expected that the additional model layers below 700 hPa should benefit most from the simulation of stratocumulus clouds with the MMF.

With the guidance of these offline testing results, the MMF vertical coordinate is modified with 12 layers below 700 hPa in this study, without changing the size of the time step used in the CRM. However, a further increase in vertical resolution would greatly increase the computational cost. The rest of the MMF experiment designs are identical to those described in CX11, including the initial conditions, the SSTs, and the CRM setup. Briefly, the SPCAM–IPHOC MMF was forced by specifying climatological SSTs and sea ice with monthly-mean annual cycles (i.e., no interannual variability in SST and sea ice) while coupled with the Community Land Model at the land gridpoints (Oleson et al. 2004). The horizontal grid size of the GCM is 1.9° × 2.5° in both the low- and high-resolution experiments. In the high-resolution experiment (IP-12L), there are 32 layers in the vertical direction with 12 layers below 700 hPa while the low-resolution experiment (IP-6L) uses 26 layers as in the standard CAM3.5 configuration. The embedded CRMs have the same number of vertical layers as the host GCM and use homogeneous 4-km horizontal grid spacing and 32 columns in a periodic horizontal domain (KRD05). Experiment IP-12L was integrated for 10 years and 3 months. The results of the last 9 years are analyzed and compared to the observed climatology. Experiment IP-6L was integrated for 2 years and 3 months, as in CX11. The last 2 years of this sensitivity experiment can be compared with the results of the SPCAM–IPHOC experiment (IP-T21) presented in CX11 and the first 2 years and 3 months of IP-12L. IP-T21 also ran for 27 months. IP-T21 and IP-6L differ in horizontal resolution and dynamical core. The sensitivity of results to both horizontal and vertical resolutions will be discussed in section 3.

3. Comparison between the low- and high-resolution experiments

In this section, the annual means of the 2-yr integrations for the three experiments described in section 2—IP-12L, IP-6L, and IP-T21—are compared for low-level clouds and surface precipitation. Also summarized are comparisons of a few surface/vertically integrated quantities.

a. Low-level clouds

The low-level clouds refer to the clouds below 700 hPa in both models and C3M observations, not just those with tops at pressures greater than that value. Since SPCAM–IPHOC provides cloud fraction in each CRM grid point from its IPHOC’s SGS distribution, a maximum overlap assumption is used to calculate the low-level cloud amount for each CRM column. The cloud amount for each GCM gridbox is the average of the cloud amount of the embedded CRM columns. This variable is more commonly called “cloud cover” because it represents the fraction of the sky obscured by clouds.

Figure 3 shows the global distributions of the annual mean low-cloud amounts simulated in the three SPCAM–IPHOC experiments and obtained from the C3M data product (Kato et al. 2011). The C3M data product merges CloudSat, CALIPSO, CERES, and MODIS observations along the CloudSat/CALIPSO track. The observations are averaged to 1.9° × 2.5° grids in space and from July 2006 to June 2010 in time. Note that optically thin clouds (optical depth < 0.3) in the C3M data are filtered out before averaging so that the observed cloud amounts are comparable to those from passive satellite sensors commonly used in model evaluation.

The three experiments simulate the tropical, subtropical, and midlatitude storm-track oceanic low clouds well. In particular, the strength and location of the low-cloud maxima in the northeastern (NE) and southeastern (SE) Pacific, SE Indian Ocean, NE and SE Atlantic and Southern Ocean, and northern Atlantic and Pacific midlatitudes were well reproduced. The major improvements in the high-resolution experiment (IP-12L) are that 1) the stratocumulus clouds in narrow regions off the west coasts of continents are more abundant than in IP-T21 (more readily seen in Fig. 4). 2) The cloud amounts in the tropical and open oceans slightly increase, in better agreement with the observations. 3) The global mean of low-cloud amount increases from 39.5% in IP-T21 to 42.3% in IP-6L and 49.1% in IP-12L. The latter is close to the observed global mean of 50.3% [49.3% from CloudSat/CALIPSO data stored at the National Center for Atmospheric Research (NCAR) CAM portal, which was used in CX11]. All of these improvements can be substantially attributed to the increase in the vertical resolution.

On the other hand, IP-12L still underestimates the global mean low-cloud amount by 5.2%, compared to 8%–10% underestimates by its low-resolution counterparts, IP-6L and IP-T21. The regions with the largest underestimates are located at the open oceans (e.g., Pacific and Indian Ocean basins) with abundant shallow cumulus clouds (e.g., Xu et al. 2008), midlatitude storm tracks, and over the land areas (e.g., South/Central America). Over land, the surface heterogeneity and its interaction with the atmosphere cannot be represented in the embedded CRM; that is, there is no variation in topography within the CRM domain and the land
surface processes operate on the GCM grid scale. As discussed in section 2, the sensitivity tests showed that the cumulus cloud amount decreases in the higher-resolution experiment (see also Cheng et al. 2010). But the simulated cloud amount in IP-12L is higher than that in IP-6L. This result implies that cumulus clouds over the open ocean occur more often in IP-12L than in IP-6L, but still less often than in the observations.

One of the deficiencies identified from the low-resolution SPCAM–IPHOC simulation (IP-T21) presented in CX11 was the underestimate of thin stratocumulus clouds right off the continents, compared to the C3M data (Kato et al. 2011). This result was primarily attributed to the insufficient vertical resolution. Figure 4 shows the vertical cross section of annual-mean cloud fraction and cloud liquid water content along 15°S from the coast of South America (~80°W) to 120°W from the three SPCAM–IPHOC experiments and the updated C3M observations (2-yr versus 4-yr averages). Similar cross sections were examined in other studies with high-resolution regional climate models (Wang et al. 2004a,b; Lauer et al. 2009). This cross section represents a (coastal) stratocumulus to trade cumulus transition.

The agreement with the C3M observations is much better for IP-12L (Fig. 4). For example, the increase in cloud fraction between 10% and 20% also appear west of 100°W, and the cloud layers are optically thicker than in either IP-6L or IP-T21. Both of these features result in a closer agreement with the observations. Most importantly, the axis of the maximum cloud fraction resembles the observed one more closely than that in either IP-6L or IP-T21, whereas the increase in horizontal resolution does not alter the orientation (Figs. 4a,b). Although there is no apparent overestimate of cloud-top height, which was not the case for the regional climate model simulation (Wang et al. 2004a,b), the cloud-base heights in IP-12L are appreciably higher west of 95°W. An overestimate in cloud liquid water content (by approximately a factor of 2) is present throughout the entire cross section. Possible reasons for this overestimate are 1) the wrong cloud type is simulated and 2) no SGS variability of microphysical processes is taken into account in the CRM. It is likely that cumulus clouds rather than stratocumulus clouds are simulated. This can be explained by the raised planetary boundary layer (PBL) height (thick black lines in Figs. 4a–c), as diagnosed from GCM grid variables based upon the bulk Richardson number (Holtslag and Boville 1993), because cumulus clouds are associated with environments with deep PBLs and high surface latent heat fluxes. The drier surface air that is primarily responsible for higher surface latent heat flux raises the lifting condensation level and thus the cloud-base height. It will be interesting to see whether or not
the drizzle rate is increased and liquid water content is reduced when an improved formulation of the autoconversion process that considers the SGS variability (Cheng and Xu 2009) is tested in the SPCAM–IPHOC in the near future.

The increase in the horizontal resolution, which cannot be separated from different dynamical cores, may also play a role in increasing the cloud amount near the coast. The maximum cloud amount in IP-6L also increases by 10%–20%, compared to IP-T21 (Figs. 4a,b). However, the cloud vertical structures do not change much except for clouds extending closer to the coastline. Therefore, the higher vertical resolution may play a more significant role in improving the simulation of stratocumulus clouds than the higher horizontal resolution does.

**b. Surface precipitation**

Surface precipitation is a major source and critical component of the hydrological cycle. The two high horizontal-resolution simulations (IP-12L and IP-6L; Figs. 5b,c) with the finite volume dynamical core duplicate the observed global patterns of annual-mean precipitation (Fig. 5d) from the Global Precipitation Climatology Project (GPCP; Adler et al. 2003) more realistically than the low horizontal-resolution simulation (IP-T21; Fig. 5a) with the semi-Lagrangian dynamical core, particularly, in the tropical regions. For example, IP-T21 produces the double intertropical convergence zones in the western Pacific Ocean, which are replaced by a single strong ITCZ centered slightly north of the equator in IP-12L and IP-6L (Figs. 5b,c). We suspect that the double ITCZ problem in the MMF is caused by the coarse horizontal resolution. There is evidence for supporting this claim. For example, double ITCZs were not pronounced in the higher-resolution T42 simulation with the semi-Lagrangian dynamical core (KRD05). They were also not produced in another MMF with the finite volume dynamical core and the same horizontal resolution.
as in IP-12L and IP-6L (Tao et al. 2009). This problem is, however, more complicated in conventional GCMs. It is most likely attributed to deficiencies in convective parameterization (e.g., Song and Zhang 2009).

The ITCZ in the Atlantic Ocean is also stronger in IP-12L and IP-6L, but the maximum center south of the equator in the Indian Ocean shifts from the eastern side of the ocean to the central, compared to the observations (Figs. 5b–d). The midlatitude storm-track precipitation patterns are more or less similar among the simulations. However, the magnitudes of precipitation in some tropical regions are overestimated compared to the GPCP observations, which partly contributes to the small overestimates in the global means (~7%). This amount is still within the relative bias error estimate of 9% (Adler et al. 2012).

Some regional patterns of precipitation are also improved. For example, the summertime [June–August (JJA)] precipitation in China is improved in IP-12L and IP-6L in terms of magnitude and spatial patterns (increasing the magnitude of precipitation from northwest to southeast; Fig. 6), except for the strong center in southwestern China and weak precipitation south of the Yangtze River. The reason for the unrealistically large amount of precipitation in southwestern China is related to the periodic lateral boundary condition used in the embedded 2D CRM. This boundary condition, coupled with the destabilized (i.e., favorable) environment produced from the particular topography in southwestern China allows convection to last longer. It is not possible to eliminate this unrealistic feature in the present configuration of the MMF (KRD05; Tao et al. 2009), but a variable CRM orientation (with wind shear and stratification) may be helpful in minimizing this bias. Testing this idea is beyond the scope of the present study.

c. Concise comparisons with observations

As in our earlier study, CX11, a Taylor diagram (Taylor 2001) is used to provide an assessment of the overall performance of SPCAM–IPHOC among the three experiments (Fig. 7). The Taylor diagram is based upon the spatial pattern correlations and the normalized standard deviations (by the observed variables). The variables chosen for comparison with observations are the annual-mean surface pressure, surface precipitation, shortwave cloud radiative forcing (SWCRF), longwave cloud radiative forcing (LWCRF), latent heat surface flux, sensible heat surface flux, and low-, mid- (700–400 hPa), and high-level (<400 hPa) cloud amounts. The region between 30°S and 30°N is chosen for this analysis because of the high spatial variability of cloud-related parameters. The high spatial variability is well known and confirmed by global distribution plots presented in this study (see section 4b). The reference (REF) point on the Taylor diagram (Fig. 7) is for the reanalysis data or observations. The reference data are European Center for Medium-Range

FIG. 6. Summer (JJA) precipitation (mm day$^{-1}$) from experiments (a) IP-T21, (b) IP-6L, (c) IP-12L, and (d) from the GPCP observations. Observations from surface gauge and satellites (Zhou et al. 2008) are similar to (d) and less similar to the model simulations.

FIG. 7. Taylor diagrams for annual-mean surface pressure (PS), surface precipitation (PRECT), SWCRF, LWCRF, surface latent heat flux (LHFLX), surface sensible heat flux (SHFLX), low-level clouds (CLDLOW), midlevel clouds (CLDMID), and high-level clouds (CLDHIGH) between 30°S and 30°N for experiments (a) IP-T21 and IP-6L and (b) IP-6L and IP-12L. Point REF denotes the reanalysis and observations.
will also be used later for comparison with the simulated climatology.

Comparing IP-6L with IP-T21, one can see that there are significant improvements in surface pressure, surface precipitation rate, and high cloud amount (Fig. 7a). For surface pressure, the normalized standard deviation increases from 0.78 to 0.99 for high cloud amount, which is likely related to an improvement in the ITCZ simulation discussed earlier. The normalized standard deviation increases from 0.72 to 0.83, due primarily to the improvement in precipitation rate, the correlation coefficient increases from 0.89 to 0.96. For the surface pressure, the normalized standard deviation increases from 0.78 to 0.96 while the correlation coefficient, but their spatial variabilities slightly exceed those of the reference data.

The improvement from the higher vertical resolution (IP-12L) is relatively small except for shortwave (SW) cloud radiative forcing (CRF), precipitation, low-level cloud amount, and surface latent heat flux (Fig. 7b). The differences in correlation and spatial variability of these four quantities are moderate, compared to those between IP-6L and IP-T21. These results suggest that the higher horizontal resolution (and perhaps the dynamical core) is important in improving these vertically integrated/surface quantities. As stated earlier, the vertical resolution has the greatest impact on the vertical structures of simulated low-level clouds (Fig. 4).

4. Climatology

In this section, the simulated climatology from the last 9 years of the IP-12L simulation is presented. Except for radiative balance analysis, the climatology from the December–February (DJF) and June–August (JJA) seasons will be shown for the global distributions and zonal means of a few selected variables. Also discussed are the annual low-cloud cycle and the relationships of low-cloud amounts with large-scale variables.

a. Radiative balance and other global-mean quantities

Radiative balance is an important requirement for climate model simulations. In conventional GCMs, the tuning of some parameters in radiation and cloud parameterization schemes is required to achieve this balance. However, it is very difficult to tune the MMF models to achieve the same balance because clouds are explicitly simulated and there is no simple relationship that can be altered to produce a desired amount of clouds and the corresponding radiative balance. Therefore, the SPCAM–IPHOC model was run without any tuning in the radiative or cloud microphysics parameterizations. Small imbalances at the top of the atmosphere (TOA) were noted in previous short-term integrations (KR05; Tao et al. 2009; CX11), which was an encouraging result. It is not clear whether or not radiative energy balance can be achieved in a longer-term integration at both the surface and TOA, as in experiment IP-12L.

The global imbalance at the TOA is minimal in IP-12L (0.21 W m$^{-2}$; Table 1), TOA outgoing longwave radiation of 240.4 W m$^{-2}$ (239.6 W m$^{-2}$ from CERES) and TOA absorbed solar radiation of 240.6 W m$^{-2}$ (240.2 W m$^{-2}$ from CERES). The caveat is that the incoming solar radiation in the model (with a solar constant of 1367 W m$^{-2}$) is 1.9 W m$^{-2}$ higher than in CERES. At the surface, IP-12L produces a large imbalance with the opposite sign (−7.51 W m$^{-2}$), compared to an estimate of 1.2 W m$^{-2}$ based upon CERES Energy Balanced and Filled (EBAF), version 2.6 (Loeb et al. 2009).

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<tr>
<th>Property</th>
<th>SPCAM–IPHOC</th>
<th>Observation</th>
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<tr>
<td>Annual mean energy budgets (W m$^{-2}$, + upward)</td>
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</tr>
<tr>
<td>TOA</td>
<td>0.21</td>
<td>0.57$^a$</td>
</tr>
<tr>
<td>Surface</td>
<td>−7.51</td>
<td>1.20$^{a,e}$</td>
</tr>
<tr>
<td>TOA outgoing longwave radiation (W m$^{-2}$, + upward)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>All sky</td>
<td>240.4</td>
<td>239.6$^e$</td>
</tr>
<tr>
<td>Clear sky</td>
<td>263.2</td>
<td>265.9$^a$</td>
</tr>
<tr>
<td>TOA absorbed solar radiation (W m$^{-2}$, + downward)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>All sky</td>
<td>240.6</td>
<td>240.2$^d$</td>
</tr>
<tr>
<td>Clear sky</td>
<td>290.9</td>
<td>287.4$^b$</td>
</tr>
<tr>
<td>Longwave cloud forcing (W m$^{-2}$)</td>
<td>22.9</td>
<td>26.3$^a$</td>
</tr>
<tr>
<td>Shortwave cloud forcing (W m$^{-2}$)</td>
<td>−50.3</td>
<td>−47.2$^b$</td>
</tr>
<tr>
<td>Cloud amount (%)</td>
<td></td>
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</tr>
<tr>
<td>Total</td>
<td>61.6</td>
<td>68.0$^b$</td>
</tr>
<tr>
<td>Low-level</td>
<td>45.0</td>
<td>50.3$^c$</td>
</tr>
<tr>
<td>Midlevel</td>
<td>17.2</td>
<td>28.4$^a$</td>
</tr>
<tr>
<td>High-level</td>
<td>29.3</td>
<td>30.7$^b$</td>
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<tr>
<td>Cloud liquid water path (mm)</td>
<td>0.098</td>
<td>0.112$^c$</td>
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<tr>
<td>Precipitable water (mm)</td>
<td>24.2</td>
<td>24.6$^d$</td>
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<tr>
<td>Latent heat flux (W m$^{-2}$)</td>
<td>88.3</td>
<td>87.9$^e$</td>
</tr>
<tr>
<td>Sensible heat flux (W m$^{-2}$)</td>
<td>23.5</td>
<td>19.4$^f$</td>
</tr>
<tr>
<td>Precipitation (mm day$^{-1}$)</td>
<td>2.85</td>
<td>2.67$^f$</td>
</tr>
<tr>
<td>Net surface longwave radiation (W m$^{-2}$, + upward)</td>
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</tr>
<tr>
<td>All sky</td>
<td>57.7</td>
<td>54.5$^e$</td>
</tr>
<tr>
<td>Clear sky</td>
<td>85.0</td>
<td>83.6$^e$</td>
</tr>
<tr>
<td>Net surface shortwave radiation (W m$^{-2}$, + downward)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>All sky</td>
<td>162.0</td>
<td>163.0$^e$</td>
</tr>
<tr>
<td>Clear sky</td>
<td>189.3</td>
<td>192.1$^a$</td>
</tr>
</tbody>
</table>

$^a$ CERES EBAF (Loeb et al. 2009)
$^b$ C3M
$^c$ MODIS
$^d$ NASA’s Water Vapor Project (Randel et al. 1996)
$^e$ JRA-25
$^f$ GPCP
et al. 2009), and JRA-25. The larger surface sensible heat flux (+4.1 W m$^{-2}$) compared to JRA-25 and the larger longwave (LW) flux (+3.2 W m$^{-2}$) compared to CERES are responsible for the large imbalance. The underestimate of clouds (low and midclouds, in particular) shown in Table 2 can increase the LW emission.

It is expected that the surface energy imbalance (Table 2) can cause model drifting. The time series of yearly mean global quantities presented in Fig. 8 clearly shows drifting trends and some oscillations. The trend is most pronounced in the land surface temperature, but that of the global-mean surface temperature is weaker. The surface energy components and TOA SW also show increasing deviations from the means although their magnitudes are relatively small compared to the imbalance (7.5 W m$^{-2}$). Compared to the twentieth century mean temperatures (http://www.ncdc.noaa.gov/cmb-faq/anomalies.php), the land surface temperature bias is small (+0.2 K), but that over the oceans is much larger (+1.2 K). This is consistent with the positive bias in surface net LW flux (+3.2 W m$^{-2}$). But the bias in surface sensible heat flux (+4.1 W m$^{-2}$) and smaller bias

<table>
<thead>
<tr>
<th></th>
<th>SW-sfc</th>
<th>LW-sfc</th>
<th>LH</th>
<th>SH</th>
<th>Imbalance</th>
</tr>
</thead>
<tbody>
<tr>
<td>IP-12L</td>
<td>161.98</td>
<td>57.66</td>
<td>88.31</td>
<td>23.52</td>
<td>-7.51</td>
</tr>
<tr>
<td>Observations</td>
<td>162.98</td>
<td>54.47</td>
<td>87.94</td>
<td>19.37</td>
<td>1.20</td>
</tr>
</tbody>
</table>

Table 2. Surface energy budget balance components and imbalance for 9-yr average for experiment IP-12L and observations. Surface SW and LW fluxes are from CERES EBAF. Latent and sensible heat fluxes are from JRA-25.

Fig. 8. Time series of selected global-averaged quantities from experiment IP-12L: (a) column temperature, surface temperature, and land surface temperature; (b) precipitable water; (c) deviations of surface energy components from the means shown in Table 2; and (d) TOA energy components.
in surface latent heat flux (+0.4 W m$^{-2}$) cannot be explained. It is possible that an inconsistency arises within the MMF due to the fact that surface turbulent fluxes are calculated on the GCM grid cell instead of the CRM grid cells. Confirming this and revealing other potential causes are important for further improvements of the MMF, but they are beyond the scope of this study.

Other global annual-average properties are shown in Table 1 for IP-12L and observations. Overall agreements between the simulation and state-of-the-art observations are acceptable. Compared to the standard SPCAM simulation and CAM3 (Collins et al. 2006; KRD08), there are a few improvements (e.g., low-cloud amount) and many similarities between SPCAM and SPCAM–IPHOC (e.g., precipitable water, cloud liquid water path, sensible heat flux, precipitation rate, and high cloud amount). TOA LW and SW, SW CRFs, and surface clear and all-sky LW fluxes differ by 2–3 W m$^{-2}$ between the different models. This amount is comparable to the difference between IP-12L and observations. But clear-sky surface SW fluxes from IP-12L differ from the CERES observations by 3 W m$^{-2}$, compared to 29–34 W m$^{-2}$ overestimates in SPCAM and CAM3. Surface latent heat flux from IP-12L is 7 W m$^{-2}$ higher than in SPCAM and CAM3 and nearly matches that of JRA-25. The low- and midcloud amounts from IP-12L are much larger than SPCAM and CAM3 (by 20% and 4%–7%, respectively), but they are still 5%–10% less than those of the active-sensor CSM observations.

b. Seasonal global climatology

Measures of the hydrological cycle (i.e., surface precipitation and precipitable water) are shown in Fig. 9. Compared to the annual-mean precipitation shown in Fig. 5c, seasonal-mean precipitation bands in the tropics [ITCZ and South Pacific convergence zone (SPCZ)] and midlatitudes are narrower and more intense. The intensity of the tropical precipitation bands is, however, too strong, especially in the western Pacific. This excessive precipitation is a common feature among the MMFs (KRD05; KRD08; Tao et al. 2009). The correlation between observed and simulated precipitation is very high (0.86) while the root-mean-square (rms) error is slightly more than half of the global means. Precipitable water is highly related to convection in the tropics. The high correlation (>0.98) and small rms error (~10% of the global means from the National Aeronautics and Space Administration (NASA)’s Water Vapor Project (NVAP); Randel et al. 1996) is expected for precipitable water because the meridional circulations determine its spatial distribution. There is small underestimates in the global means (0.2–0.8 mm) in both seasons by SPCAM–IPHOC, whereas SPCAM overestimates precipitable water (0.8–0.9 mm; KRD05).

Figures 10 and 11 show the DJF and JJA global distributions of the TOA and surface SW and LW cloud radiative effects (CREs) or forcings. The TOA SW and LW fluxes are not shown because of their excellent agreements with the observations (e.g., correlations greater than 0.97). In MMF, the radiative heating rates are calculated on the CRM grid with the radiation transfer scheme within CAM3.5. Partial cloudiness from IPHOC is used as an input parameter in the radiation transfer scheme for each CRM column of SPCAM–IPHOC. The CERES observations on Terra/Aqua used in this study is the EBAF data product (Loeb et al. 2009), covering the period of 2000–10. The EBAF product provides highly constrained radiative energy fluxes at the TOA and surface. The diurnal cycle is accounted for using data from Terra, Aqua, and five geostationary satellites, and the net TOA irradiances are adjusted with ocean heat storage data. The surface EBAF fluxes are calculated from a radiative transfer model but are constrained by the observed TOA CERES fluxes. Because pure clear-sky pixels (drier) are chosen to represent clear sky in EBAF, it is likely that the magnitudes of CRE, especially LW CRE, are higher than those obtained from the simulation where the clear-sky fluxes are calculated using grid-mean thermodynamic profiles.

A common feature appearing in Figs. 10 and 11 is that the agreement between the simulation and observations is slightly better in DJF than in JJA. This can be explained by the well-simulated southern extratropical storm activities in DJF and the overestimates of CRE magnitudes in the northern Pacific in JJA. The latter decreases the correlations, which are between 0.65 and 0.95, by 0.02 (in TOA LW and SW CREs) to 0.19 (surface LW CRE). The differences in the global means range from 1.4 W m$^{-2}$ (in JJA surface LW CRE) to 6.7 W m$^{-2}$ (in JJA surface SW CRE). The ITCZ, SPCZ, and extratropical storm-track regions, as well as subtropical low-cloud regions (except for LW CRE), contribute greatly to the high correlations and good agreements in the global means. There are, however, some significant differences over land regions, especially for surface LW CRE of the polar region (Figs. 10f,h), likely due to difficulties in satellite cloud property retrieval over the ice surface (Minnis et al. 2011).

The zonal mean properties shown in Figs. 8 and 9 and TOA SW and LW fluxes for DJF and JJA are given in Figs. 12 and 13, respectively. As expected, all six properties agree well with their respective observations. The agreement is also better, compared to the SPCAM simulation presented in KRD05, in particular with the tropical precipitation, precipitable water, and CREs in
Disagreements with the observations include underestimates of DJF extratropical precipitation and the associated LW CREs and overestimates of JJA tropical precipitation (~50%) and SW CRE (15–20 W m$^{-2}$). Additionally, the magnitude of SW CREs (LW CREs) are also overestimated (underestimated), which is related to deficiencies in simulated cloud vertical structures and liquid and ice partitioning that is not discussed in this study (see Cheng et al. 2012).
The last set of comparisons is with the zonal mean low-, mid-, and high-level cloud amounts and the total cloud amount between the simulation and the C3M observations (Fig. 14). These cloud amounts are obtained from a maximum overlapping of cloud fractions over preset CRM column spans before spatial and temporal averaging is taken, as in the C3M data product (Kato et al. 2011). The agreement between the simulation and observations is good in the tropics but less so in midlatitudes for low, high, and total cloud amounts.
The midlevel clouds are underestimated by 10%–20% at all latitudes. Inadequate vertical resolution near the freezing level is the most likely reason for the underestimates of midlevel clouds, but they may be overestimated in C3M because of the inability to exclude precipitating hydrometeor areas.

c. Annual cycle of selected low-cloud regions

The relationships between low clouds and large-scale variables are fundamentally important to the fidelity of the low-cloud simulation in the upgraded MMF. An analysis of these relationships is performed for the five
marine stratocumulus regions discussed in Klein and Hartmann (1993): Californian (20°–30°N, 120–130°W), Peruvian (10°–20°S, 80°–90°W), Namibian (10°–20°S, 0°–10°E), Australian (25°–35°S, 95°–105°E), and Canarian (15°–25°N, 25°–35°W). The chosen large-scale variables are LTS, relative humidity (RH) at 1000 hPa, and PBL height, as in DRK10, to facilitate a direct comparison with the relationships simulated by the standard MMF. The “observed” large-scale variables are from ERA-Interim (Dee et al. 2011) data, instead of 40-yr ECMWF
Re-Analysis (ERA-40) data (Uppala et al. 2005) used in DRK10. Because of the comparison with DRK10 results, the observed low-cloud amount is from the International Satellite Cloud Climatology Project (ISCCP; Rossow and Schiffer 1999) data that were used in DRK10, rather than the C3M data that have been used so far in this study. Unlike DRK10, the low-cloud amount in the present study also includes that of midlevel clouds due to a well-known artifact in ISCCP cloud retrieval, which assigns low-level clouds at the midlevel if thin upper-level clouds are present (Garay et al. 2008; Ghate et al. 2009; Wang et al. 2011). However, its impact is very small for the selected stratocumulus regions. The MMF and ERA-Interim diagnose the PBL height in a similar
The PBL height is assigned at the altitude where the bulk Richardson number exceeds a critical value (0.3 for MMF and 0.25 for ERA-Interim). This minor difference in the threshold can slightly shift the slopes between RH and PBL height, as shown later.

Figure 15 shows the annual cycle in LTS and low-cloud amount from IP-12L and observations for the five regions. One plus and minus standard deviations from annual-mean cloud amounts are shown. The annual cycles of these two quantities (LTS is underestimated in most regions) are closely related in both observation and simulation. However, correlations are very different from one region to another as a result of different time lags between the two variables, for example, in the Namibian and Australian regions. In the Californian, Peruvian, and Canarian regions, the correlations agree within 0.04 between observations and IP-12L, which is impressive. In the Australian region, the negative correlations (−0.35/−0.27) between the observed cloud amounts and ERA-40/ERA-Interim LTS are due either to the lack of pronounced seasonal variability in cloud amount or to the uncertainties in cloud retrieval. The simulated correlation is 0.97 in IP-12L, compared to 0.95 in DRK10 for the same region. In the Namibian region, the correlations from the MMFs are lower (0.55 in IP-12L and 0.09 in DRK10) than the observed (−0.8).

The relationships between the cloud amount and LTS can be further examined as functions of RH at 1000 hPa, as in DRK10. DRK10 identified RH and PBL height as the only large-scale variables that yielded consistent relationships with MMF low-cloud amount in all five regions. Figure 16 shows the scatter diagrams between RH and LTS with low-cloud amount color coded for the five regions. Each point corresponds to a monthly average. There are significantly strong correlations in every region between RH and LTS for both observations and the simulation. It is interesting to point out that the RH and LTS relationship resembles that resulting from
the quasi-equilibrium hypothesis for deep convective regions (e.g., Arakawa 1993; Xu 1994; Xu and Randall 1998), except that a normalized lapse rate is used as the stability measure, instead of LTS. Whether or not the quasi-equilibrium hypothesis is valid for stratocumulus regions is beyond the scope of this study.

It is apparent that the shapes of the scatter and the locations of highest and lowest cloud amounts associated with RH and LTS pairs are slightly different from one region to another, but the agreement between the simulation and observation is outstanding. For example, the highest cloud amounts are associated with the highest LTS and highest RH in the Californian, Peruvian, and Canarian regions. In the Namibian region, the simulated highest cloud amounts are mostly associated with lower RHs, instead of higher RHs in the observations. In the

FIG. 15. Mean annual cycle of LTS (solid lines) and low-level cloud amount (dashed lines, ×10) for (bottom) observations and (top) experiment IP-12L for the five regions described in Klein and Hartmann (1993). For observations, LTS is calculated from ERA-Interim; low-level cloud amount is from ISCCP. Correlation coefficient between LTS and low-level cloud amount is listed for each region. Interannual variability of low-level cloud amount is indicated by shading for one plus/minus standard deviation.

FIG. 16. Monthly-mean low-level cloud amount (%; color-coded, color bar at bottom) as a function of monthly mean RH at 1000 hPa and LTS from (top) experiment IP-12L and (bottom) observations for the five regions described in Klein and Hartmann (1993). For observations, monthly-mean low-level cloud amount is from ISCCP; RH and LTS are calculated from ERA-Interim data.
Australian region, the simulated highest cloud amounts are associated with the highest LTS and RH, but the observed cloud amounts have no such preference. Compared to the standard MMF (DRK10), these results are superior as a result of the significantly better agreement in low-cloud amounts with observations.

Lastly, the relationships among the low-cloud amount, RH, and PBL height are shown in Fig. 17. The relationship between RH and PBL height is linear with small scatter in both simulation and observations for all five regions. There are some differences in the diagnosed ranges of PBL height between the MMF and ERA-Interim, with ERA-Interim being higher by 100–200 m, due to slightly different critical Richardson numbers in diagnosing the PBL heights. Despite the different ranges of the scatter diagrams among the regions, both the simulation and observations show that the highest cloud amounts are associated with the highest RHs for a given PBL height, except for the Australian region, which does not have a preference. These results suggest that the relationships among low-cloud amount, RH, and PBL height are rather similar between simulation and observations. This was not the case in DRK10, because of the severe underestimate of low-cloud amount in the standard MMF. DRK10 explained the deficiencies in the simulation in terms of the ratio of buoyancy to shear production, which is also related to the terms in the turbulent kinetic energy budget. The budget is more realistically simulated with the IPHOC, which transports heat/moisture via subgrid-scale eddies, instead of larger, more viscous eddies that are not effective in the CRM with a low-order closure (Cheng et al. 2010).

5. Conclusions and discussion

Several studies have shown that MMF is a promising approach to climate modeling (KR01; KRD05; KRD08; Tao et al. 2009; DRK10). This approach represents convective processes and phenomena well but not boundary layer turbulence and cloud processes. An upgrade of the CRM component with an advanced third-order turbulence closure has been made in CX11. In the present study, two simulations were performed using the finite-volume dynamical core with a grid size of 1.98.2.58 (instead of the semi-Lagrangian dynamical core) but they differed in the vertical resolution. The number of model layers below 700 hPa increases from 6 in one simulation (experiment IP-6L) to 12 in another (experiment IP-12L) in both the host GCM and the embedded CRM to better resolve the vertically thin stratocumulus clouds. This vertical reconfiguration was first tested in the CRM embedded in the MMF using a large-eddy simulation as the benchmark. The testing confirmed that low-cloud simulation could be sufficiently improved with the doubling of the vertical resolution in the lower troposphere, with the greatest improvement occurring for vertically thin stratocumulus clouds. This vertical reconfiguration was first tested in the CRM embedded in the MMF using a large-eddy simulation as the benchmark. The testing confirmed that low-cloud simulation could be sufficiently improved with the doubling of the vertical resolution in the lower troposphere, with the greatest improvement occurring for vertically thin stratocumulus clouds. The primary effects of coarsening vertical grid spacing are the smoothing of the mean thermodynamic profiles and underestimating the inversion strength. This leads to more mixing, which delays clouds and results in an underestimate

![Fig. 17. As in Fig. 16, but as a function of PBL height and RH at 1000 hPa.](http://journals.ametsoc.org/jcli/article-pdf/26/16/5717/4008437/jcli-d-12-00200_1.pdf)
of stratocumulus cloud amount in the low-resolution MMF simulation.

Even though the increased vertical resolution is still not adequate to realistically resolve all boundary layer clouds in the MMF, IP-12L produces not only a global mean low-cloud amount that is 3.6% higher than IP-6L and is much closer to that of the C3M observations, but also the spatial distributions that are slightly more realistic in several ocean basins. The latter is most pronounced for the stratocumulus clouds right off the continents and for the cumulus clouds over the open oceans. A vertical cross section at 15°S clearly shows that the increase in cloud fraction is 10%–20% off the coast and farther west in the cumulus and stratocumulus regions. The improved vertical resolution results in a substantially better agreement in the vertical structures of clouds with the C3M observations, although cloud liquid water content may be overestimated in all SPCAM–IPHOC simulations relative to the C3M observations.

Another significantly improved aspect of both IP-12L and IP-6L experiments is in the spatial patterns of tropical precipitation, compared to a coarser horizontal resolution (IP-T21) experiment performed in CX11, which also had a different dynamical core in its host GCM. A single ITCZ in the Pacific is simulated, which is centered slightly north of the equator, instead of double ITCZs in IP-T21. The intensity of South Pacific convergence zone and the ITCZ in the Atlantic are also more realistic. There are some excessive precipitation areas in the tropical regions, as in other MMF simulations (KRD05; KRD08; Tao et al. 2009). In terms of spatial correlations and patterns for the 30°S–30°N region, most surface/vertically integrated properties show greater improvement over the simulation in CX11 (IP-T21) than that with lower vertical resolution. A simulation with coarser horizontal resolution and an identical GCM dynamical core is needed to confirm these conclusions because it is likely that the improvement may be due to the different dynamical cores.

The 10-yr-and-3-month IP-12L simulation achieves a nearly balanced TOA radiative energy budget, but the energy balance at surface is not satisfactory; its imbalance is 7.5 \( \text{W m}^{-2} \). There is no tuning of parameters in radiation and cloud microphysics parameterizations in the model. The larger surface sensible heat flux and LW flux are responsible for the large imbalance, compared to the observational estimate. Despite this problem, global annual average TOA LW and SW CREs, and surface clear and all-sky LW and SW fluxes differ only by 2–3 \( \text{W m}^{-2} \) from CERES EBAF observations. The differences in most of these properties between SPCAM (KRD08) and IP-12L are 2–3 \( \text{W m}^{-2} \), but the differences between SPCAM and the CERES EBAF observations can be larger. For example, the clear-sky surface SW is 29 \( \text{W m}^{-2} \) larger.

The global seasonal climatology of the selected properties agrees well with observations. The simulated precipitable water has a correlation of 0.98 with observations and the rms error has a magnitude of less than 10% of the mean. The precipitation bands in the ITCZ and SPCZ realistically move with season, but the JJA tropical precipitation is too strong. The correlation in TOA LW and SW CREs between observations and the simulation exceeds 0.88 in both DJF and JJA, with agreements being slightly better with observations in DJF than in JJA. This result can be explained by the well-simulated southern extratropical storm activities in DJF and the overestimates of CRE magnitudes in the northern Pacific in JJA. Although there are some significant differences over land regions such as surface LW CRE of the polar region, the CREs of ITCZ, SPCZ, extratropical storm-track regions, and subtropical low-cloud regions (except for LW CRE) contribute greatly to the overall agreement with CERES observations. Additionally, the zonal means of precipitation rate, precipitable water, TOA LW and SW fluxes, LW and SW CREs, and cloud amounts show satisfactory agreements with state-of-the-art observations in both DJF and JJA, except for overestimated precipitation and SW CRE in the tropics. Midlevel cloud amount is underestimated by 10%–20% in all latitudes, while surface precipitation and associated CREs show some underestimates in the midlatitudes.

Finally, the relationships between low-cloud amount and large-scale variables were analyzed to better understand the controlling processes simulated in this upgraded MMF. This model produced the most physically realistic simulation of geographic distribution and annual cycle of low clouds among the different versions of SPCAM. The relationships among low-cloud amount, surface relative humidity, LTS, and PBL height agree well with observations in five stratocumulus deck regions, although the degree of agreements varies from one region to another with the closest agreements being in the Californian, Peruvian, and Canarian regions. In these regions, the highest cloud amounts are associated with the highest LTS, highest surface RH, and the lowest PBL height. Such preferences in either model or observation are weaker in the Australian (observation) and Namibian (model) regions. Relative to SPCAM, these relationships are more consistent between observations and simulation because of the improvement in low-cloud simulation and more realistic turbulent kinetic energy budget in the embedded CRM.

In the future, it will be interesting to couple this MMF with an ocean model to see whether or not the realistic
low-cloud geographic distributions are maintained and SST biases are minimized in the eastern ocean basins. Fixing the surface energy budget of the model is critically important for performing any long-term integration.

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