Simulations of the Tropical General Circulation with a Multiscale Global Model

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ABSTRACT

Cloud processes play a central role in the dynamics of the tropical atmosphere, but for many years the shortcomings of cloud parameterizations have limited our ability to simulate and understand important tropical weather systems such as the Madden–Julian oscillation. Since about 2001, “superparameterization” has emerged as a new path forward, complementing but not replacing studies based on conventional parameterizations. This chapter provides an overview of work with superparameterization, including a discussion of the method itself and a summary of key results.

1. Introduction

The tropics strongly absorb solar radiation at all times of year. Much of that solar energy is used to evaporate water at Earth’s surface. The water vapor thus added to the tropical atmosphere can be lofted to form beautiful convective cloud systems, which are organized on scales ranging from a few kilometers to many thousands of kilometers. The resulting weather systems include squall lines, tropical cyclones, and continent-spanning monsoons. The intense cloudy convective updrafts occupy only a tiny fraction of the tropics, but they carry prodigious amounts of energy upward through the depth of the troposphere. The outflows from the tops of the convective towers feed divergent upper-tropospheric winds, which carry energy poleward, and so contribute to Earth’s global energy balance. This cloud-filled multiscale tropical circulation was the main subject of Michio Yanai’s research.

In atmospheric global circulation models (GCMs), the small-scale processes at work in convective cloud systems must be included through simplified submodels called parameterizations (Fig. 15-1). Cloud parameterizations have been under development since the 1960s. A lot of progress has been achieved, but major difficulties remain (Randall et al. 2003a).

In 1999, National Center for Atmospheric Research (NCAR) scientists Wojciech Grabowski and Piotr Smolarkiewicz created a “multiscale” GCM in which the physical processes associated with clouds were represented by implementing a simple “cloud-resolving” model (CRM) within each grid column of a low-resolution global model (Grabowski and Smolarkiewicz 1999; Grabowski 2001, 2004). In this approach, parameterizations of radiation, cloud microphysics, and turbulence (including small clouds) are still needed (Fig. 15-2), but larger clouds and some mesoscale processes are explicitly (though crudely) simulated.

In highly idealized experiments, Grabowski and Smolarkiewicz found that their multiscale model...
produced promising simulations of organized tropical convection. In particular, the simulations showed propagating convective systems that resembled the Madden–Julian oscillation\(^1\) (MJO; Madden and Julian 1971, 1972). The MJO is an eastward-propagating tropical disturbance that spans thousands of kilometers in the zonal direction, with an irregular period in the range 40–50 days. Despite its large spatial and temporal scales, and its powerful effects on tropical weather, the MJO has proven very difficult to simulate with GCMs (e.g., Lin et al. 2006; Kim et al. 2009).

Inspired by the results of Grabowski and Smolarkiewicz, Khairoutdinov and Randall (2001) created a multiscale version of the Community Atmosphere Model (CAM; Collins et al. 2006). They disabled the parameterizations of the CAM and replaced them with a simplified version of Khairoutdinov’s CRM (Khairoutdinov and Randall 2003). One copy of the CRM runs in each grid column of the CAM. The CRM is two-dimensional (one horizontal dimension, plus the vertical) and uses periodic lateral boundary conditions. In the study of Khairoutdinov and Randall (2001), the CRM had a horizontal domain 64 grid columns wide, with a horizontal grid spacing of 4 km.

Khairoutdinov and Randall dubbed the embedded CRM a “superparameterization.” The combination of a GCM with a superparameterization is now called a multiscale modeling framework (MMF), and the MMF based on the CAM is now called the SP-CAM. A second MMF was created by Tao et al. (2009), using a different GCM and a different CRM. As of spring 2014, three more MMFs are being tested, each based on a different GCM.

2. Coupling the CRM and the GCM

Figure 15-3 schematically illustrates the coupling of the embedded CRM with the GCM, as implemented in the SP-CAM. The GCM provides advective forcing to the CRM, much as observed advective forcing can be supplied to the “column physics” of a model that is being tested in simulations based on field data (e.g., Randall et al. 1996). The CRM provides heating and drying as feedback to the GCM, just as any conventional parameterization would do.

The GCM’s time step is typically tens of minutes, whereas the CRM’s time step is on the order of 10 seconds. It is assumed (and required) that the GCM’s time step is an integer multiple of the CRM’s time step. The CRM “subcycles” through a sequence of its short time steps, in order to determine the physical tendencies averaged over one of the longer GCM time steps. The CRM runs continuously throughout a simulation; it is not restarted between GCM time steps.

The coupling between the GCM and the CRM is very simple and has some nice properties. It was used by Khairoutdinov and Randall (2001) and in all subsequent work with the SP-CAM. Here is how it works: Let \( q \) be a generic scalar variable that is defined in both the GCM and CRM. We write

\[
\bar{q}_G^{n+1} = q^n_G + B_G \Delta t_G, \tag{15-1}
\]

so that

\[
B_G = \frac{\bar{q}_G^{n+1} - q^n_G}{\Delta t_G}. \tag{15-2}
\]
Here \( B_G \) is the adiabatic tendency of \( q \) due to non-CRM effects, as determined by the GCM; the subscript \( G \) denotes a GCM value; the superscripts \( n \) and \( n + 1 \) denote successive GCM time steps; and \( \Delta t_G \) is the size of the GCM time step. The tilde on \( q_{n+1}^G \) denotes a provisional value. Equation (15-1) represents a “partial time step” before the diabatic tendency of \( q \) is accounted for by the CRM.

The CRM variables, denoted by \( q_C \), are updated using CRM time steps of size \( \Delta t_C \):

\[
\frac{q_C^{m+1} - q_C^m}{\Delta t_C} = B_C + \left( \frac{q_{n+1}^G - \langle q_C \rangle^n}{\Delta t_G} \right) + S_C. \tag{15-3}
\]

Here the superscripts \( m \) and \( m + 1 \) denote successive CRM time steps. In (15-3), the term \( B_C \) represents the adiabatic tendency of \( q_C \) due to advection on the CRM’s grid. Pointy brackets denote a horizontal average over the CRM’s domain. The quantity \( \langle q_C \rangle^n \) is the CRM-domain average of \( q_C \) at the beginning of the GCM time step. The term \( \left( (q_{n+1}^G - \langle q_C \rangle^n) / \Delta t_G \right) \) is the mechanism through which the GCM’s advective tendencies are felt by the CRM; in other words, it represents the forcing of the CRM by the GCM. The forcing is held constant as the CRM subcycles through one GCM time step, and it is also independent of horizontal position on the CRM’s grid. The source of \( q \) due to the CRM physics is denoted by \( S_C \), which has no counterpart in (15-1) because the physics is computed only in the CRM, not in the GCM.

We take enough CRM time steps to span the longer GCM time step, finally arriving at GCM time level \( n + 1 \). Then we let the CRM feed back on the GCM. To implement the feedback, the GCM variables are updated using...
Here \( \langle q_c \rangle^{n+1} \) represents the horizontal average of \( q_c \) at GCM time level \( n + 1 \), that is, at the end of the sequence of CRM time steps based on (15-2).

Comparing the left-hand side of (15-4) with the third line on the right-hand side, we see that

\[
q_G^{n+1} = \langle q_c \rangle^{n+1}.
\]

(15-5)

Of course, we can also write

\[
q_G^n = \langle q_c \rangle^n.
\]

(15-6)

The implication is that, at both the beginning and end of the GCM’s time step, the GCM’s value of \( q_G \) is guaranteed to agree with the horizontal average of the CRM values. The two models cannot “drift apart.”

Use of (15-1) in (15-3) shows that the CRM variables evolve according to

\[
\frac{q_G^{n+1} - q_G^n}{\Delta t_G} = B_G + \langle q_c \rangle^{n+1} - \langle q_G \rangle^n = \langle q_c \rangle^{n+1} - \langle q_G \rangle^n = \frac{\langle q_c \rangle^{n+1} - q_G^n}{\Delta t_G}.
\]

(15-7)

This means that the CRM feels the GCM advection as a simple additive term. If we horizontally average (15-7) over the CRM’s grid, we get

\[
\langle q_G \rangle^{n+1} - \langle q_G \rangle^n = B_G + \langle q_c \rangle^{n+1} - \langle q_G \rangle^n = \langle q_c \rangle^{n+1} - \langle q_G \rangle^n = \frac{\langle q_G \rangle^{n+1} - q_G^n}{\Delta t_G}.
\]

(15-8)

The CRM’s periodic lateral boundary conditions imply that \( B_G \) involves only vertical transports. Finally, from (15-5), (15-6), and (15-8), we see that

\[
q_G^{n+1} - q_G^n = B_G + (B_c + \langle S_c \rangle),
\]

(15-9)

where the overbar on the right-hand side represents an average over the GCM time step. The interpretation of (15-9) is simple and clear.

The key points to take away from the discussion above are that the coupling strategy represented by (15-1)–(15-3) leads to the result in (15-9), and also guarantees (15-5) and (15-6). The coupling strategy therefore guarantees consistency between the GCM fields and the CRM fields, as they coevolve during a simulation. See Grabowski (2004) for further discussion.

The CRM uses the anelastic system of equations with height as the vertical coordinate (Khairoutdinov and Randall 2003), while GCM uses the quasi-hydrostatic system with the terrain-following sigma-pressure coordinate (Collins et al. 2006). For each GCM grid column, the time-varying GCM sounding is used as the “reference sounding” in the anelastic system of the CRM. When the CRM is called by the GCM, it receives as input the provisional temperature and vapor profiles that have been updated by the GCM dynamics, as well as the height and pressure profiles. The CRM adjusts its vertical grid to agree with the GCM’s, assuming that the reference-state pressures at midlevels and interface levels are the same as the GCM’s. The CRM computes the reference-state air density from hydrostatic balance.

The heights of the GCM’s grid levels are slightly different from those on the previous GCM time step, so there is a small spike at the beginning of each CRM subcycle, but because the grid height in the GCM evolves slowly, the spike does not significantly affect the CRM results.

At the beginning of a sequence of CRM time steps, the vertical integral of water vapor in the CRM is not the same as it was when the CRM finished the preceding sequence. This discrepancy is eliminated at the end of CRM call, however, because it is automatically corrected by the prescribed large-scale forcing. As a result, the vertical integral of total water in the CRM at the end of the GCM time step always equals the vertical integral of total water in the GCM after the provisional dynamics step, minus the precipitation that falls out during the CRM time steps. A similar result holds for the frozen static energy.

A complication is that water in the form of falling precipitation inside the grid columns of the CRM is not passed back to GCM at the end of the sequence of CRM time steps; some of the precipitation remains in the CRM grid, “hanging in the air.” Although such hanging precipitation is not accounted for by the GCM, it is conserved by the CRM and can eventually reach the surface and be accounted for as rainfall, or else return to vapor through evaporation or sublimation.

Because the CRM is two-dimensional, we do not permit momentum feedback to the GCM winds. This means that the model cannot simulate the effects of vertical momentum transport by convection and gravity waves that are resolved on the CRM’s grid (Khairoutdinov et al. 2005). We return to this point later.

Further discussion of the superparameterization and how it differs from conventional parameterizations is given near the end of this chapter. We now turn to an overview of results that have been produced through the use of the superparameterization.

3. Simulations of tropical variability

Many investigators have published the results of studies based on the SP-CAM, in more than 50 refereed journal
articles. A recurring theme of these studies has been modes of variability in the climate system. In order of increasing time scale, the types of variability considered so far include the diurnal cycle (Khairoutdinov et al. 2005; Pritchard and Somerville 2009a,b; Pritchard et al. 2011; Kooperman et al. 2013), easterly waves and tropical cyclones (McCrary 2012; Stan 2012), fluctuations of the Asian summer monsoon (DeMott et al. 2011, 2013) and the Madden–Julian oscillation (Benedict and Randall 2009, 2011; Thayer-Calder and Randall 2009; Kim et al. 2009; Andersen and Kuang 2012; Arnold et al. 2013), El Niño and the Southern Oscillation (Stan et al. 2010), and anthropogenic climate change (Wyant et al. 2006, 2012; Stan and Xu 2014; Arnold et al. 2014). Here we provide a value-added overview of some of these results, with an emphasis on the effects of coupling the SP-CAM to an ocean model.

\textit{a. The diurnal cycle of precipitation}

Solar heating and nocturnal cooling of the land surface are associated with a vigorous diurnal cycle of convective activity over the warm continents of the tropics and the summer hemisphere midlatitudes, with a maximum in late afternoon or early evening (e.g., Dai 2001). Many global models fail to simulate the observed afternoon maximum (Dai 2006; Dirmeyer et al. 2011). Khairoutdinov et al. (2005, their Fig. 13) pointed out that the SP-CAM produces fairly realistic diurnal cycles of precipitation over both the continents and the oceans. Pritchard and Somerville (2009a,b) Pritchard et al. (2011), and Kooperman et al. (2013) performed detailed analyses of the SP-CAM’s simulations of the diurnal cycle of precipitation over central North America in summer. They showed that the model is able to simulate the observed propagation of diurnally forced convective systems from near the Rocky Mountains in the afternoon to near Omaha after midnight. An example is shown in Fig. 15-4, which depicts rapid eastward propagation of diurnally excited convective disturbances in both the observations and the SP-CAM, but not in the CAM.

The afternoon precipitation maximum over land is driven by solar heating of the ground, which leads to increasing sensible and latent heat fluxes from early morning until midafternoon. These energy fluxes directly modify the thermodynamic properties of the boundary layer. All models include these basic physical processes. To produce a realistic diurnal cycle of precipitation, however, deep convective clouds must respond realistically to the changing thermodynamic properties of the boundary layer. The coupling of the parameterized boundary layer with the parameterized deep convection is therefore key to the ability of a model to simulate the diurnal cycle of precipitation over heated land. This coupling includes the lofting of boundary layer air to form cumulus clouds in the free atmosphere, and the production of cold pools by precipitation-driven downdrafts that penetrate into the boundary layer. In many existing GCMs, the boundary layer and cumulus parameterizations have been developed independently,
their coupling is perfunctory, and the processes mentioned above are either not explicitly formulated or missing altogether. Many models include rather arbitrary assumptions about “triggers” that enable the development of deep convection (e.g., the requirement of a minimum relative humidity in the boundary layer).

The situation is very different with SP-CAM, in which the “boundary layer parameterization” of the GCM is limited to a parameterization of the surface fluxes. Turbulent transport above the surface is computed on the CRM’s relatively high-resolution grid, and includes both resolved and parameterized components. The coupling of the boundary layer with the cumulus layer is explicitly resolved, and does not rely on parameterizations, perfunctory or otherwise. Cumulus clouds form when warm, humid air near the surface is lofted by the vertical motion resolved on the CRM’s grid. Downdrafts resolved on the CRM’s grid can produce cold pools in the boundary layer. These processes are poorly resolved because of the CRM’s 4-km horizontal grid spacing, but the results suggest that an explicit representation with poor resolution can be more successful than a current-generation parameterization.

b. The Great Red Spot

Early work with the SP-CAM was based on the use of prescribed seasonally varying sea surface temperatures (SSTs). As first pointed out by Khairoutdinov and Randall (2001), the SP-CAM produces an unrealistically active hydrologic cycle over the western North Pacific Ocean during the northern summer months (Fig. 15-5c). The precipitation rate is unrealistically high, the precipitable water is excessive, the surface pressure is lower than observed, and the cyclonic low-level winds are too strong. The problem appears as an unrealistic eastward extension of the Asian summer monsoon, which also produces excessively strong monsoon precipitation. It does not occur at other times of year. Because color plots of this phenomenon typically show strong red features over the western North Pacific Ocean, we call it the Great Red Spot (GRS). Interestingly, the MMF created by Tao et al. (2009) also produces a GRS, even though it is based on different GCM and a different CRM.

A comparison of Figs. 15-5b and 15-5d shows that the GRS disappears when the SP-CAM is coupled with an ocean model. This was discovered when Stan et al. (2010) coupled the SP-CAM to a low-resolution version of POP (the Parallel Ocean Program); we call this coupled model the SP-CCSM.2 As reported by Stan et al. (2010), and further analyzed by DeMott et al. (2011, 2013, 2014), the SP-CCSM gives a more realistic simulation of the atmospheric circulation than the SP-CAM “right out of the box,” without any tuning, a somewhat surprising result in view of earlier experiences of others (e.g., Sausen et al. 1988). Besides eliminating the GRS, the SP-CCSM also produces a more realistic simulation of the Asian summer monsoon. In addition, Stan et al. (2010) found that the coupled model gives a more realistic simulation of El Niño, La Niña, and the Southern Oscillation, relative to the same coupled model with conventional atmospheric parameterizations.

The price paid for these improved results is an error in the simulated SST distribution. The error takes the form of a cooling in many locations (not shown), with a root-mean-square (RMS) value of 2.08 K. This is actually slightly smaller than the RMS error of the SST in version 3 of the conventionally parameterized CCSM, which also tends to produce tropical SSTs that are cooler than observed.

In the uncoupled model, the prescribed SSTs represent an effectively infinite reservoir of sensible and latent heat, which is unaffected by what happens in the simulated atmosphere. The strong cyclonic winds of the GRS promote intense evaporation, which supplies energy to maintain the GRS, as discussed by Luo and Stephens (2006). In the coupled model, the surface winds cool the SSTs, and this negative feedback prevents the development of the GRS. We can interpret the GRS as an error that comes from failure to include air-sea interactions in the uncoupled model.

c. The MJO

The ability of the SP-CAM to simulate the MJO was first reported by Khairoutdinov and Randall (2001), although Grabowski (2001) found something similar in a simplified global model. An extensive and generally favorable comparison of the simulated MJO with observations was reported by Benedict and Randall (2009). Thayer-Calder and Randall (2009) analyzed the relationship between water vapor and precipitation rate in the SP-CAM, and compared with version 3 of the conventionally parameterized CAM. They showed that in the SP-CAM heavy rainfall is accompanied by a very humid troposphere, whereas in version 3 of the CAM the lower troposphere remains dry even during intense rainfall (Fig. 15-6). They argued that a prerequisite for a realistic simulation of the MJO is that water vapor must exhibit a realistically large dynamic range. As discussed by Derbyshire et al. (2004) and others, deep convection may be unable to form unless the middle troposphere becomes sufficiently moist. This moistening is created by the convection itself. There are thus two separate issues: On the one hand, the simulated middle troposphere

2 The CCSM is the Community Climate System Model, which has recently been renamed as the Community Earth System Model (CESM). The SP-CCSM is a version of the CCSM that uses SP-CAM as its atmospheric component.
must become humid enough to permit realistic deep convection. On the other hand, the parameterized convection must be capable of moistening the middle troposphere. We return to this point later.

The analysis of Benedict and Randall (2009) shows that it is possible to produce a reasonably successful simulation of the MJO without taking into account the effects of air–sea interactions. Nevertheless, it is
interesting to examine the effects of air–sea coupling on the MJO. Benedict and Randall (2011) used a slab ocean model (SOM) to study the effects of air–sea interactions on the MJO in a simplified framework. In the SOM, departures of the net surface energy flux from climatology can create SST anomalies, depending on a prescribed ocean mixed layer depth, but the SST anomalies are damped with a 50-day time scale to keep the SSTs of the SOM close to those of a control run. As shown in Fig. 15-7, the air–sea interactions promote coherent eastward propagation of the MJO. The air–sea interactions also strengthen the coupling between dynamics, as represented by the 850-hPa zonal wind, and moist processes, as represented by the precipitation rate at 90°E.

In an analysis of results from their fully coupled model, Stan et al. (2010) also found that the air–sea interactions improve the realism of the simulated MJO. Figure 15-8 shows the variance of precipitation in the range of frequencies and wavenumbers associated with the MJO, plotted as a function of longitude, from the western Indian Ocean to the central Pacific. In the uncoupled SP-CAM, the simulated variance takes large values not only west of the date line, as observed, but also considerably farther east. In the SP-CCSM, the simulated variance decreases realistically to the west of the date line. Further analysis of the results shows that the excessive variance in the SP-CAM results is found mainly in the region of the GRS. The elimination of the GRS in the coupled model leads to the more realistic variance distribution in SP-CCSM.

d. Yanai waves

Wheeler–Kiladis diagrams (Wheeler and Kiladis 1999) are ubiquitous now, but are most commonly shown only for modes that are symmetrical across the equator, such as the MJO and Kelvin waves. Figure 15-9 shows the antisymmetric power, which includes the westward-propagating Yanai waves. The SP-CCSM produces a realistic level of Yanai-wave activity, while the SP-CAM generates considerably less. This improvement in the coupled model is due to a more realistic “basic state,” relative to the uncoupled model, as discussed by DeMott et al. (2011).

e. Monsoons

The Asian summer monsoon and its subseasonal variability attracted the attention of Prof. Yanai and his students (Luo and Yanai 1983, 1984; He et al. 1987; Yanai and Li 1994; Li and Yanai 1996; Hung et al. 2004). Recently, DeMott et al. (2011, 2013) have investigated the ability of the SP-CAM and the SP-CCSM to simulate these important tropical weather systems. Figure 15-10 shows the northward propagation (and also the weaker southward propagation) of precipitation anomalies that occur during the Asian summer monsoon. Both the SP-CAM and the SP-CCSM produce somewhat realistic northward propagation. The mechanisms of this variability are discussed in detail by DeMott et al. (2013).

Recently, McCrary (2012) has investigated the ability of the SP-CAM and the SP-CCSM to simulate the African summer monsoon and the associated African easterly waves, which are precursors to Atlantic tropical cyclones. Figure 15-11 shows the phase relationship between the vertically varying water vapor mixing ratio and the precipitation maximum associated with composite African easterly waves. The SP-CCSM correctly produces a very moist troposphere at the time of maximum rainfall, but the conventionally parameterized CAM3 produces a very unrealistic dry layer in the middle troposphere. This is strongly reminiscent of the earlier result of Thayer-Calder and Randall (2009), shown in Fig. 15-6, and suggests that the ability of models to produce realistic variations of water

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3 Also called mixed Rossby–gravity waves.
vapor is important not only for the MJO, but also for higher-frequency tropical disturbances. McCrory (2012) also shows that the SP-CCSM produces a somewhat realistic modulation of the African easterly waves by the MJO.

4. Concluding remarks

a. Parameterizations forever!

The results summarized in this paper demonstrate that superparameterization can simulate a wide variety of tropical variability more realistically than today’s conventionally parameterized models. We are confident, however, that future parameterizations can and will perform as well as or better than today’s superparameterizations. Our hope is that work with superparameterizations can make that future happen sooner.

Apart from their practical use in simulations, parameterizations encapsulate our understanding of the way in which the parameterized physical processes interact with the large-scale circulation. Even with very successful simulations of tropical variability from global cloud-resolving models, we would need parameterizations, coupled with simplified models, to understand why the simulations succeeded.

b. Compared to what?

It is easy to criticize the superparameterization. The CRM is two-dimensional. The 4-km horizontal grid spacing that is typically used in the CRM is too coarse to accurately resolve even large cumulus clouds. The periodic boundary conditions are unrealistic. Given these weaknesses of the approach, why does superparameterization work so well?

To answer this question, it is useful to review the key differences between an MMF and a conventionally parameterized GCM. First of all, the CRM used in an MMF simulates cloud dynamics by explicitly solving the (two-dimensional) equation of motion, thus tying back to the “basic physics” developed by Isaac Newton and others. In contrast, conventional GCMs parameterize all of the dynamical processes associated with cloud growth and decay. For example, most cumulus parameterizations

4 Modelers have to be optimistic!
represent cumulus clouds as one-dimensional entraining plumes (e.g., Arakawa and Schubert 1974). While it is true that the periodic lateral boundary conditions of the CRM are unrealistic, conventional parameterizations are so crude that the subject of lateral boundary conditions never even comes up.

In an MMF, the CRM simulates convective entrainment crudely but directly through its subgrid turbulence parameterization in combination with the simulated wind field resolved on the CRM grid. In contrast, entraining plume models often assume that the fractional entrainment rate is constant with height, which is in conflict with recent studies (e.g., Lin and Arakawa 1997a, b), or else they rely on entrainment parameterizations (e.g., Gregory 2001; Chikira and Sugiyama 2010).

Because the CRM solves the equation of motion, an MMF has no need for closure assumptions (e.g., quasi-equilibrium), nor does it use assumptions about triggers that initiate cloud formation. In addition, the CRM explicitly simulates some aspects of mesoscale organization, such as aggregation or squall lines; conventional parameterizations omit these processes entirely.

The CRM relies on questionable parameterizations of microphysics, turbulence, and radiative transfer. Of course, conventional models must also parameterize these same processes. A key difference, however, is that the CRM provides relatively detailed information as input to...
the parameterizations. For example, the CRM explicitly simulates (relatively) small-scale vertical velocities that can be used by microphysics parameterizations, whereas conventional parameterizations can only estimate cloud-scale vertical velocities using the entraining plume model. Last but not least, the CRM explicitly simulates fractional cloudiness and cloud overlap, which are critical inputs to a radiation parameterization.

The CRM simulates small-scale shear and buoyancy fluctuations that are key input to turbulence parameterizations.

FIG. 15-10. June-August (JJA) lag correlation of 20–100-day filtered precipitation with itself in GPCP, SP-CCSM, SP-CAM, and CCSM. The base point time series is the filtered precipitation averaged over 10°S–5°N, 80°–100°E. The contour interval is 0.2, with positive (negative) values shown with solid (dashed) lines.

FIG. 15-11. Longitude–height cross section of the specific humidity anomaly (g kg\(^{-1}\)) in African easterly waves as (left) observed, (middle) simulated by the SP-CCSM, and (right) simulated by CCSM3. The latitude is 10°N. The heavy contour is zero, dashed contours are negative (lightly shaded), and solid contours are positive (darkly shaded). The contour interval for the SP-CCSM results is 0.05 g kg\(^{-1}\), while the contour interval for the observations and for CCSM is 0.02 g kg\(^{-1}\). The small plots at the top of each panel show anomalies of the outgoing longwave radiation (OLR; W m\(^{-2}\)). The anomalies of both water vapor and outgoing longwave radiation are relative to seasonal means. Based on the work of McCrary (2012).
In addition, the CRM is a nonlinear fluid-dynamical model that exhibits sensitive dependence on initial conditions; that is, it behaves chaotically, as does the real atmosphere (e.g., Hohenegger et al. 2006). As a result, the CRM-grid-averaged heating and drying rates that are passed to the GCM contain quasi-random components, which originate primarily on the smallest scales resolved by the CRM. In this sense, the superparameterization is a stochastic parameterization. Efforts are under way to develop stochastic conventional parameterizations, but the superparameterization generates stochastic heating and drying rates in a particularly natural way.

Finally, the SP-CAM is almost embarrassingly parallel, so that it can make efficient use of a very large number of processors. The reason is that the many copies of the CRM (one per CAM grid column) run independently, with no communication among themselves. As a result, for a given GCM grid spacing the SP-CAM can use many more processors than a conventionally parameterized model. Although the SP-CAM does hundreds of times more arithmetic per simulated day than a conventionally parameterized GCM with the same resolution, the ability of the MMF to utilize more processors than a conventional GCM means that the wall-clock time required to complete a given simulation with the superparameterization is only moderately longer than that required with a conventional parameterization. An example is shown in Fig. 15-12. The SP-CAM is orders of magnitude less expensive than a global cloud-resolving model (GCRM; e.g., Tomita et al. 2005).

c. Process models and global models

Over the past two decades, cloud-parameterization testing has become organized on an international scale, beginning with the National Aeronautics and Space Administration (NASA) First International Satellite Cloud Climatology Project (ISCCP) Regional Experiment (FIRE) program in the 1980s (Cox et al. 1987), and continuing in the 1990s and beyond with the U.S. Department of Energy (DOE) Atmospheric Radiation Measurement Program (ARM; Stokes and Schwartz 1994) and the GCSS\(^5\) activities (Randall et al. 2003b). One strategy for testing parameterizations is to drive both the parameterized column physics of a GCM and a high-resolution CRM with “forcing” databased on field observations, and then to intercompare the results of the two models with additional observations from the field (Randall et al. 1996). The column-physics is called a single-column model, and the high-resolution model is sometimes called a process model.

An important limitation of this strategy is that, because single-column models and process models represent only a small regional domain, they cannot “feed back” to the large-scale circulation; the effects of the large-scale circulation are simply prescribed and non-interactive. This limitation is now breaking down. Both MMFs and global cloud-resolving models are simultaneously global models and process models (Fig. 15-13).

d. Improving the MMF

There are many ways to improve the MMF by enhancing the CRM. Perhaps the highest priority is to incorporate better parameterizations of shallow clouds and turbulence that cannot be resolved on the CRM’s grid. Higher-order closure methods have been developed for the CRM and tested in the MMF (Cheng and Xu 2011, 2013; Xu and Cheng 2013a,b; Bogenschutz et al. 2012, 2013), with very promising results. These modern turbulence parameterizations predict (or diagnose) fractional cloudiness and the variability of temperature and water vapor at scales that are not resolved on the CRM’s grid. With appropriate attention to coupling the parameterizations, this subgrid information can be used by improved parameterizations of microphysics (e.g., Morrison and Grabowski 2007, 2008) and radiation (e.g., Pincus and Stevens 2009). In addition, more advanced numerical methods have been implemented in the CRM.

\(^5\) GCSS was the GEWEX Cloud Systems Study; GEWEX was the Global Energy and Water Experiment. The work of GCSS is now continuing under the more general name GEWEX Atmospheric System Studies (GASS).
These various improvements are incremental, but very important.

More sweeping revisions of the MMF are also under development. The chapter of this book by Arakawa and colleagues describes a radically different “second-generation” MMF, called the quasi-three-dimensional (Q3D) MMF. The Q3D MMF partially removes the periodic boundary conditions of the CRM and includes the dynamical effects of three-dimensionality in a simplified way, while still consuming much less computer time than a GCRM. We hope that, in the years to come, the Q3D MMF will open the door to exciting new studies of global and tropical dynamics.

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FIG. 15-13. In this Venn diagram, the circle on the left represents process models, and the circle on the right represents global atmospheric models. Until recently, these two classes of models did not overlap. Today, as shown in the figure, there is some intersection in the form of GCRMs and MMFs. This figure first appeared in Randall (2013).


