Comparison Studies of Cloud- and Convection-Related Processes Simulated by the Canadian Regional Climate Model over the Pacific Ocean

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ABSTRACT

This paper presents results from the Canadian Regional Climate Model (CRCM) contribution to the Global Energy and Water Cycle Experiment (GEWEX) Pacific Cross-section Intercomparison Project. This experiment constitutes a simulation of stratocumulus, trade cumulus, and deep convective transitions along a cross section in the tropical Pacific. The simulated seasonal mean cloud and convection are compared between an original version of CRCM (CRCM4) and a modified version (CRCMM) with refined parameterizations. Results are further compared against available observations and reanalysis data. The specific parameterization refinements touch upon the triggering and closure of shallow convection, the cloud and updraft characteristics of deep convection, the parameterization of large-scale cloud fraction, the calculation of the eddy diffusivity in the boundary layer, and the evaporation of falling large-scale precipitation. CRCMM shows substantial improvement in many aspects of the simulated seasonal mean cloud, convection, and precipitation over the tropical Pacific, CRCMM-simulated total column water vapor, total cloud cover, and precipitation are in better agreement with observations than in the original CRCM4 model. The maximum frequency of the shallow convection shifts from the ITCZ region in CRCM4 to the subtropics in CRCMM; accordingly, excessive cloud in the shallow cumulus region in CRCM4 is greatly diminished. Finally, CRCMM better simulates the vertical structure of relative humidity, cloud cover, and vertical velocity, at least when compared to the 40-yr ECMWF Re-Analysis. Analyses of sensitivity experiments assessing specific effects of individual parameterization changes indicate that the modification to the eddy diffusivity in the boundary layer and changes to deep convection contribute most significantly to the overall model improvements.

1. Introduction

Clouds and cloud-related processes are among the most variable and complicated phenomena in the earth’s climate system. Through nonlinear interactions with dynamical and thermodynamic processes, clouds influence the energy balance, the hydrological cycle, and even the vertical distribution of momentum in the climate system (Cess et al. 1990; Houghton et al. 2001). Improving our understanding of and our ability to simulate clouds and their physical properties are two of the largest challenges in climate modeling. At present, cloud feedbacks remain the largest source of uncertainty in climate sensitivity estimates, as indicated in the recent Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4; IPCC 2007; also see Soden and Held 2006).

Among the variety of cloud types, stratocumulus, shallow cumulus, and cumulonimbus clouds are the most common in the subtropical and tropical oceans, with shallow cumulus clouds being the most frequently occurring (Bony et al. 2006) and, therefore, those most strongly influencing the surface and top-of-the-atmosphere (TOA) radiation balance. It is crucial that climate models accurately simulate these various cloud types and the transitions between them. It was with this goal in mind that the Global Energy and Water Cycle Experiment (GEWEX) Cloud System Study (GCSS) project was developed (Randall et al. 2003) and, in particular, the GCSS Pacific Cross-section Intercomparison (GPCI) experiment (Siebesma et al. 2004).

In both global and regional climate models (GCMs and RCMs, respectively), cloud processes and their collective influence on resolved scale variables are generally parameterized. This is because the scales of motion...
and the cloud particles themselves occur on time and space scales that are far smaller than the typical resolution of a climate model grid box. Despite great progress in the past few decades, it is generally recognized that a lack of understanding and inadequate parameterization of cloud processes still contribute importantly to uncertainties in climate and numerical weather prediction models (Gates et al. 1999; IPCC 2007). Recent studies, reported upon by Bony et al. (2006), suggest that low-level cloud processes in the tropical boundary layer are the most uncertain and poorly modeled, and as a result contribute most significantly to uncertainties in simulated cloud–radiation feedbacks in coupled GCMs.

The GPCI, inspired by the European Cloud Systems study (EUROCS; Siebesma et al. 2004), focuses on the complete simulation of coupled cloud, convection, turbulence, and radiation processes in fully three-dimensional climate models, including both GCMs and RCMs. Within the GPCI framework, GCMs and RCMs are integrated for the boreal summer season (May–August) of 1998 and 2003. A cross section of model grid boxes along a trajectory from the coast of California, through the trade cumulus regions of the subtropical Pacific, to the equatorial west Pacific intertropical convergence zone (ITCZ) was extracted from the model output and evaluated against satellite observations and reanalysis products. This cross section encompasses all three cloud and convective regimes (namely boundary layer stratocumulus, trade wind shallow cumulus, and deep precipitating convection, plus the transitions between these three regimes) active in the tropical ocean, allowing an efficient evaluation of model-simulated cloud, convection, and turbulence parameterization within a single framework. So far, over 20 modeling groups have contributed results to the GPCI initiative (information online at http://gcss-dime.giss.nasa.gov/gpci/modsim_gpci_models.html).

In this paper we present results from the Canadian Regional Climate Model (CRCM) GPCI simulation and, in particular, we analyze the impact of a number of parameterization modifications made in light of these initial results. These modifications include changes to the parameterization of shallow convection, deep convection, the calculation of eddy diffusivity in the boundary layer, and the formation of large-scale clouds. In addition, evaporation of falling precipitation is also now taken into account to improve the simulation of precipitation in the stratocumulus region. Comparison of results demonstrates that the modified version of CRCM has a number of encouraging improvements in simulating the first-order characteristics of subtropical and tropical clouds and precipitation systems.

The rest of this paper is arranged as follows: Section 2 briefly introduces the CRCM and its configuration for the GPCI experiment, along with the datasets used for model evaluation. Section 3 provides a description of the major modifications to the physical parameterizations in CRCM. Section 4 presents a comparison between the modified version of CRCM and its original counterpart. In section 5 the impacts of specific modifications on the simulated seasonal mean climate are assessed, through a set of sensitivity experiments that highlight the main parametric improvements leading to the results in the modified version CRCMM. Section 6 presents a summary and conclusions.

2. Model description, experiment configuration, and evaluation datasets

a. A brief description of the CRCM

Over the past decade, CRCM has provided a comprehensive regional atmospheric modeling system for use in the analysis and understanding of regional climate processes over North America. It has evolved into the standard tool for regional climate change assessment in Canada (Laprise et al. 1998; Laprise et al. 2003; Plummer et al. 2006).

The original version of CRCM was based on the dynamical kernel of the Mesoscale Compressible Community model and the physical parameterization package of the Canadian General Circulation Model (Caya and Laprise 1999). The dynamical formulation of CRCM is characterized by an advanced semi-implicit, semi-Lagrangian advection scheme, permitting relatively long time steps to be used at high resolution (CRCM currently uses a time step of 15 min at a resolution of 45 km). CRCM employs an Arakawa C-type staggered grid with uniform horizontal resolution (typically 45 km at 60°N) in a polar-stereographic projection. A terrain-following hybrid vertical coordinate (Gal-Chen and Somerville 1975) is used to stagger the momentum and thermodynamic variables in the vertical. CRCM employs a one-way nesting technique (Davies 1976) to provide large-scale lateral atmospheric boundary conditions, derived either from a forcing GCM or from reanalysis data, such as the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis (NNRA; Kalnay et al. 1996) or the 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis data (ERA-40; Uppala et al. 2005). At the lateral boundaries, Davies nudging (Davies 1976) is enforced only on the wind field, over a nine-gridpoint buffer zone.

The first- and second-generation CRCMs were used
to study current climate and climate change over western Canada (Laprise et al. 1998; Laprise et al. 2003). The third-generation CRCM was introduced in July 2003. Major improvements in the dynamic aspect of the model included the introduction of spectral nudging of the large-scale fields (Riette and Caya 2002). Changes to the model physics mainly included the implementation of the Bechtold–Kain–Fritsch mass-flux scheme (hereafter referred to as BKF; Bechtold et al. 2001) as the convective parameterization scheme for both shallow and deep convection, and an interactive mixed-layer lake model (Goyette et al. 2000).

The most ambitious set of model improvements occurred with the introduction of the fourth generation of the CRCM, hereafter CRCM4, which was released at the end of 2005. While the dynamics of CRCM4 is identical to its predecessor, this version was the product of a major effort to improve the physical representation of a wide range of key climate processes, including clouds, radiation, turbulence in the boundary layer, and the representation of land surface processes, characterized by the incorporation of the Canadian Land Surface Scheme (CLASS; Verseghy 1991; Verseghy et al. 1993). The majority of the physical parameterization schemes employed in CRCM4 are identical to those in the third generation of the Canadian Centre for Climate Modeling and Analysis (CCCma) Atmospheric Global Climate Model (AGCM3; Scinocca and McFarlane 2004), with only the BKF convective scheme differing from the parameterization package in AGCM3. Interested readers are referred to the relevant descriptions of AGCM3 for a detailed explanation of the remaining schemes (Scinocca and McFarlane 2004; McFarlane et al. 2006). CRCM4 represents the starting point for the GPCI integrations and parameterization changes described in this article.

b. Experimental configuration and evaluation data

The GPCI protocol requires modeling groups to extract simulated variables along a cross section of points traversing the subtropical to equatorial Pacific. The points along this cross section are identical to those used in the EUROCS project (Siebesma et al. 2004) and are illustrated in Fig. 1. The cross section is composed of 13 points in the Pacific Ocean, starting at (35°N, 235°E) then moving southward with steps of 3° latitude and 4° longitude until (1°S, 187°E), encompassing the main subtropical stratocumulus, trade wind shallow cumulus, and tropical deep convection regimes, allowing all three regimes to be studied within one framework.

The CRCM domain, as shown in Fig. 1, is centered on 17°N, 200°E with 115 × 75 polar stereographic grid points at 180-km (true at 60°N) horizontal resolution, in both zonal and meridional directions. This configuration puts the cross section at the center of the model domain and also covers the two-dimensional common area (a rectangle between 5°S–45°N and 160°–240°E) required by GPCI. In the vertical, the model is discretized into 29 uneven Gal-Chen levels, half of them being assigned in the lower troposphere below 3 km (Jiao and Caya 2006).

Both the initial and lateral boundary conditions are derived from the NNRA-2 data (Kanamitsu et al. 2002). Observed monthly mean sea surface temperature and sea ice concentrations, from the Atmospheric Model Intercomparison Project (AMIP), are prescribed as a lower boundary condition. The simulations were initialized on 21 May 1998 and 2003, respectively, and run in continuous climate mode for June–August (JJA) of each year. Model output was saved every 3 h as required by GPCI. In this way, model results are available for both monthly–seasonal mean comparison, as well as for studies of the diurnal cycle. In this paper, we focus on the seasonal mean evolution of cloud and convection as simulated along the GPCI cross section.

The observational datasets used in this study largely include monthly mean cloud cover from the International Satellite Cloud Climatology Project (ISCCP; Rossow and Duenas 2004), high-resolution precipitation from the Tropical Rainfall Measuring Mission (TRMM; Adler et al. 2003; Huffman et al. 2007), cloud liquid water path and total-column water vapor from the Special Sensor Microwave Imager (SSMI; Wentz and Spencer 1998), and shortwave radiation, at the surface and TOA, from the Clouds and the Earth’s Radiant Energy System (CERES; Wielicki et al. 1996). Finally, the ERA-40 data (Uppala et al. 2005), including vertical profiles of relative humidity, cloud fraction, and vertical velocity, are used as a proxy for observations, allowing an evaluation of the CRCM vertical structure along the cross section.

3. Modifications to the model physics

Based on the original CRCM4 GPCI results, a number of modifications to aspects of the cloud, convection, and turbulence parameterizations have been introduced into CRCM4. The modified model will hereafter be referred to as CRCMM.

In this section we detail the major changes introduced to the various parameterization schemes, with references to the original formulation included at the appropriate point. Section 4 compares seasonal mean statistics along the cross section between the original CRCM4 and CRCMM results. In section 5, we try to
isolate the major contribution of the individual parameterization changes to the overall improvements seen in section 4.

a. Modifications to shallow convection

The modifications to shallow convection focus on the trigger function and mass flux closure at cloud base.

1) Trigger function

The trigger function for shallow convection originates from Fritsch and Chappell (1980) with a slight modification in the BKF scheme (Bechtold et al. 2001). Essentially, it requires that a mixed layer of at least 50-hPa thickness is buoyant when lifted to the height of its lifting condensation level (LCL):
where $\theta_v^{\text{mix}}$ is the mean virtual potential temperature of a mixed updraft layer and $\bar{\theta}_v^{\text{env}}$ the environmental value, an overbar refers to a gridbox mean value. Here, $\Pi = (P/P_0) R_d / C_p a$ is the Exner function. The virtual temperature term $\Delta T_v^w$ crudely represents the effect of resolved-scale vertical motion in increasing the buoyancy of a parcel at the LCL through subgrid-scale enhanced motion. In the deep convection parameterization, this is related directly to the resolved vertical velocity in the model by $\Delta T_v^w = \text{sgn}(w_n) c_w |w_n|^{1/3}$, where sgn is the signum function and $c_w = 6 \text{ K m}^{-1/3} \text{s}^{-1/3}$; $w_n$ is the large-scale vertical velocity in a grid box normalized by a 25-km reference square. In the shallow convection scheme, this thermal perturbation is simply assigned a constant value of 0.2 K.

In CRCMM, a further perturbation is applied to the shallow convection trigger function once the grid-scale mean relative humidity is sufficiently high. The trigger function for shallow convection in CRCMM is expressed as

$$\left(\theta_v^{\text{mix}} - \bar{\theta}_v^{\text{env}}\right) + \frac{\Delta T_v^w}{\Pi} + \frac{\Delta T_v^{\text{RH}}}{\Pi} > 0,$$

the newly added perturbation term $\Delta T_v^{\text{RH}}$ aims to represent the contribution of subgrid-scale variations in the environmental moisture field to parcel buoyancy at the LCL (Jones and Sanchez 2002):

$$\Delta T_v^{\text{RH}} = \begin{cases} 0.2(\bar{\text{RH}}_{LCL} - 0.7)\theta_v^{\text{mix}} / \frac{\partial q_s}{\partial T} & \text{for } 0.7 \leq \bar{\text{RH}}_{LCL} \leq 0.9, \\ (1.0/\bar{\text{RH}}_{LCL} - 1.0)\theta_v^{\text{mix}} / \frac{\partial q_s}{\partial T} & \text{for } \bar{\text{RH}}_{LCL} > 0.9, \end{cases}$$

where $\bar{\text{RH}}_{LCL}$ is the mean relative humidity and $\bar{\theta}_v$ the saturated mixing ratio at the LCL; $\theta_v^{\text{mix}}$ represents the mean mixing ratio of the mixed updraft layer. The physical motivation behind this term is that when the grid-box mean relative humidity is above a certain value (e.g., 70%), there likely exist subgrid-scale regions already at or above saturation. Latent heat release in these regions will lead to subgrid-scale buoyancy and an increase in the parcel vertical velocity. In these regions, convection will be preferentially triggered. As the grid-box mean relative humidity tends toward 100%, this perturbation gradually reduces to zero as we now assume the entire grid box is homogeneously at saturation.

2) CLOSURE OF SHALLOW CONVECTIVE MASS FLUX AT CLOUD BASE

In BKF, the intensity of both shallow and deep convection is controlled by a convective available potential energy (CAPE) adjustment closure. This assumes that at least 90% of the mean vertical profile of CAPE is removed by overturning mass in convective updrafts and downdrafts and by mixing with the surrounding environment, within a prescribed time scale.

The idea behind CAPE adjustment is that the mass flux at cloud base is completely controlled by conditions in the upper cloud layer. This is a reasonable assumption for deep convection, which is largely driven by latent heat release in the cloud plume. However, it is not the case for shallow convection where the fundamental mechanisms controlling convective growth are quite different. Based on the observation that shallow cumulus clouds are often rooted in the subcloud layer, Grant (2001) proposed a relationship between mass flux at cloud base and turbulent kinetic energy (TKE) in the subcloud layer. Formally, he suggested using the free convective vertical velocity scale $w^*$ of the subcloud layer as a closure for the cloud-base mass flux in shallow convective clouds, specifically

$$M_b = 0.03 w^*, \quad \text{with } w^* = \left[ \frac{g z_b}{\bar{\theta}_v} \left( w^* \bar{\theta}_v \right) \right]^{1/3},$$

where $M_b$ represents mass flux at cloud base and $z_b$ the height of the convective boundary layer; $\bar{\theta}_v$ is the vertically averaged virtual potential temperature in the boundary layer and $\left( w^* \bar{\theta}_v \right)$ the buoyancy flux at the surface; and the constant 0.03 was derived from a number of large-eddy simulation (LES) integrations for several different shallow cumulus cases (Grant 2001). Neggers et al. (2004) have demonstrated that this type of closure provides the most accurate results in simulating the diurnal cycle of shallow cumulus in a number of LES case studies. In CRCMM, we replace the standard CAPE closure in the BKF shallow convection scheme with that presented by Grant (2001). The original CAPE closure is retained for BKF deep convection.
b. Modifications to deep convection

Following the work of Kain (2004), updates to the deep convection scheme touch upon the minimum cloud depth required to allow deep convection, the assumed cloud radius, and the calculation of CAPE used in the deep convective closure.

In CRCMM, the minimum cloud depth for deep convection is a function of the cloud-base temperature, as proposed by Kain (2004), and lies in the range 2000–4000 m, rather than specified as a constant of 3500 m, as in the original BKF scheme used in CRCM4. If a convective cloud depth is less than the threshold minimum, shallow convection can still be supported where the minimum cloud depth required is 300 m.

Cloud radius controls the maximum possible entrainment rate in the BKF deep convection, and was set to 1500 m in the original scheme. In CRCMM we now specify the cloud radius to lie in the range 1000–4000 m, as a function of the grid-box mean vertical velocity at the LCL as in Kain (2004). The stronger the upward vertical velocity at cloud base, the larger the assumed cloud radius.

The third modification to deep convection concerns the calculation of the CAPE value used in the closure. In the original Kain and Fritsch (1990) and BKF schemes, the calculation of the CAPE in the closure routine used an undilute updraft profile, which ignored the effects of entrainment. Kain (2004) suggested that a dilute updraft profile resulted in a more realistic calculation of CAPE, with systematically lower CAPE values. As a result, convective rainfall and stabilization of the atmosphere were reduced when a dilute profile was used in the CAPE closure. In CRCMM, we have therefore modified the CAPE calculation from the original undilute updraft to a dilute one. This approach takes into account the entrainment contribution from the surrounding environment:

\[
\text{DCAPE} = \int_{\text{LCL}}^{\text{ETL}} \left( \frac{\theta_e^{up} - \theta_e^{env}}{\theta_e^{env}} \right) g \, dz,
\]

\[
(\theta_e^{up})_k = f_k (\theta_e^{env})_k + (1 - f_k)(\theta_e^{up})_{k-1},
\]

with

\[
f_k = \frac{M_k^{\prime}}{M_k^{up} - M_k^{\prime}}.
\]

where DCAPE is the dilute CAPE; \(\theta^{up}\) and \(\theta^{env}\) are the equivalent potential temperature in the updraft and environment, respectively; \(g\) is the gravity constant; LCL represents the lifting condensation level and ETL the equilibrium temperature level where the buoyancy of the updraft becomes zero; \(M^{up}\) represents mass flux in the parameterized updraft, while \(M^{\prime}\) and \(M^{\prime\prime}\) represent the mass of air entrained into, and detrained from, the convective updraft; and the subscript \(k\) indicates the model vertical level (see a schematic illustration in Fig. 2).

Finally, the original BKF scheme potentially allowed shallow and deep convection to be triggered simultaneously at the same grid point at the same model time step. This frequently occurred in the CRCM4 simulations. In CRCMM, a switch has been added to disallow the possibility of shallow convection if deep convection has already been triggered at the same grid point and time step (i.e., deep convection is tested for first, and shallow convection is only tested for if deep convection is not simulated at a given model grid box and time step).

c. Modifications to the vertical diffusion scheme

Boundary layer processes are primarily responsible for the occurrence of low-level clouds such as stratocumulus and are of fundamental importance in determining the thermodynamic structure of the boundary layer and, by association, the characteristics of near-surface air parcels that ultimately support convection. An accurate representation of boundary layer processes in climate models is therefore essential for simulating large-scale quantities such as cloud cover and precipitation.

In CRCM4, vertical subgrid-scale fluxes in the atmosphere were parameterized using a first-order closure, stability-dependent eddy diffusivity formulation. The diffusion coefficients were defined on the basis of a mixing length hypothesis, modified to account for the
effects of vertical stable stratification (see McFarlane et al. 1992 for a detailed description).

In CRCMM, the calculation of vertical diffusivity has been replaced by a nonlocal mixing scheme, initially employed by Troen and Mahrt (1986) and subsequently developed for use in the ECMWF Integrated Forecast System [IFS; ECMWF (2007), where a detailed description of the scheme can be found]. The important difference relative to the CRCM4 diffusion scheme is that eddy diffusivities are calculated in a nonlocal manner for unstable conditions, as defined by the surface buoyancy flux. The vertical diffusion coefficients are specified as a function of height over a diagnosed mixed-layer depth. In stable situations, a neutral mixing length approach modified by a bulk Richardson number is utilized. Finally, the new scheme includes an explicit parameterization of entrainment at the boundary layer top.

d. Parameterization of resolved cloud fraction

The large-scale cloud fraction in CRCM4 was originally calculated as solely a function of relative humidity (e.g., Sundqvist 1978). In CRCMM, this has been replaced by the scheme proposed by Xu and Randall (1996). The cloud fraction $f_c$ diagnosed from this scheme depends not only on the large-scale relative humidity $\bar{R}H$, but also on the saturation water vapor mixing ratio $\bar{q}_{\text{sat}}$ and the total cloud water mixing ratio $\bar{q}_l$, providing a cloud fraction estimate that is potentially more responsive to changes in the large-scale thermodynamic structure of the atmosphere. Specifically,

$$f_c = \begin{cases} \bar{R}H^p \left(1 - \exp\left\{-\frac{-\alpha q_l}{\left(1 - \bar{R}H/\bar{q}_{\text{sat}}\right)\gamma}\right\}\right), & \text{if } \bar{R}H < 1, \\ 1.0, & \text{if } \bar{R}H \geq 1; \end{cases}$$

(8)

where $p = 0.25$, $\alpha = 100$, and $\gamma = 0.49$ are empirical constants suggested by Xu and Randall (1996) based on a variety of cloud-resolving model simulations.

e. Evaporation of falling large-scale precipitation

In CRCM4, precipitation produced by the large-scale condensation scheme was assumed to immediately fall to the ground, without experiencing evaporation during its fall path. It is well established that a significant fraction of precipitation falling through a subsaturated atmosphere evaporates, reducing the total precipitation flux reaching the ground (Oliver and Holzworth 1953). Cooling and moistening associated with this evaporation can play an important role in boundary layer thermodynamic processes (Katzfey and Ryan 1997). To include this process in CRCMM, we introduce the parameterization of Kessler (1969), where precipitation evaporation is assumed to be proportional to the degree of subsaturation in a model grid box. Starting from the uppermost model layer, the evaporation of falling precipitation is calculated as described in the ECMWF IFS documentation (ECMWF 2007):

$$E = 5.44 \times 10^{-4}(1 - f_c)(1.0 - \bar{R}H)(\bar{q}_{\text{sat}} - \bar{q})$$

$$\times \left[\left(\frac{P}{P_s}\right)^{1/2} \frac{P_t}{5.9 \times 10^{-3}}\right]^{0.577},$$

(9)

where $P_t$ is the large-scale precipitation falling from the layer directly above the vertical layer in question, $P$ is the pressure of a given model layer, and $P_s$ is the surface pressure.

4. Comparison of the CRCM4 and CRCMM results

Both CRCM4 and the modified version CRCMM have been integrated for the GPCI cases of 1998 and 2003. In this section, we present simulated seasonal mean results related to the cloud and convection processes for 1998, since most of the observations and/or reanalysis data are available only for this year. A limited evaluation of the JJA 2003 results indicates similar findings.

a. Evaluation of vertical integrated quantities and precipitation

In this section, we focus on a few important column-integrated quantities, such as total column water vapor (TCWV), cloud liquid water path, and integrated total cloud cover. Simulated precipitation will also be discussed in this section.

Water vapor is the most abundant of the greenhouse gases in the atmosphere and plays a fundamental role in the hydrological cycle. It is important that climate models accurately simulate TCWV in order to get reasonable cloud and precipitation statistics, along with accurate clear-sky radiation fluxes. Figure 3 presents the seasonal mean distribution of TCWV over the Pacific Ocean as simulated both by CRCM4 and CRCMM, together with observations from SSM/I. Generally speaking, both models simulate a reasonable pattern of TCWV except that CRCM4 (Fig. 3a) is slightly drier than the observations (Fig. 3c). Looking at the TCWV values along the GPCI cross section (Fig. 3d), one can clearly see that the TCWV values in CRCMM are increased (and improved) relative to CRCM4, both in the
stratocumulus and trade cumulus regions (15°–30°N) as well as in the ITCZ (5°–10°N).

In the original EUROCS cross-section analysis by Siebesma et al. (2004), the simulation of cloud cover along the trajectory proved to be one of the most poorly simulated variables, with the stratocumulus cloud fraction typically underestimated by 30%–50% and conversely, trade cumulus overestimated by 20%. These errors led to commensurate errors in the surface solar radiation flux, an extremely important term in coupled climate simulations. This problem largely remains in the new GPCI results (J. Teixeira 2007, personal communication). Figures 4a–c present spatial distributions of the seasonal mean (JJA 1998) cloud cover simulated by CRCM4 and CRCMM, together with ISCCP observations over the GPCI 2D domain. Corresponding to the basic characteristics of the large-scale circulation over the tropical Pacific, cloud cover can generally be separated into three distinct regimes (Fig. 4c): (i) low-level stratocumulus clouds in the north and northeastern Pacific Ocean under the descending branch of the Hadley circulation, and coincident with northeast trade winds and relatively cold ocean surface temperatures; (ii) shallow trade wind cumulus over the warmer subtropical oceans between 15° and 30°N, with total cloud cover usually less than 50%; and (iii) cumulonimbus in the ITCZ, corresponding to the ascending branch of the Hadley circulation and identifiable in Fig. 4c by the cloud band of 50%–60%, paralleling 8°N. CRCM4 reasonably simulates stratocumulus amounts in the northeastern Pacific, along the west coast of North America, although the large cloud fraction extends too far into the subtropics, where shallow cumuli are observed to be dominant. This results in a positive cloud cover bias of ~20%–30% in CRCM4 in this region. The high cloud band in the ITCZ, with cloud amounts greater than 50%, is confined to the western half of the domain in CRCM4, while in the observations this cloud band stretches zonally across the entire domain at ~8°N, corresponding to the seasonal mean location of the ITCZ in JJA.

These key features in cloud cover are substantially improved in the CRCMM simulation (Fig. 4b). For example, the ITCZ cloud band around 8°N is clearly more visible and extends across the entire domain. The southeast extension of the stratocumulus region is also reduced in CRCMM, with a clear minimum in the cloud fraction in the subtropics, associated with the dominant region of shallow convective activity. Figure 4d further compares the cloud cover variation, together with the
ERA-40 cloud fraction, along the GPCI cross section. The ISCCP total cloud amount is more than 80% in the stratocumulus region, drops to ~50% in the shallow cumulus region, and then increases back to ~70% in the ITCZ. As previously noted, the major deficiency in CRCM4 is that the model dramatically overpredicts the cloud cover in the shallow cumulus region by 20%–35%; this is clearly improved in CRCMM. Several parameterization changes combined to produce this improvement. The revised calculation of eddy diffusivity helped improve the thermodynamic structure of the boundary layer, particularly in the trade cumulus region, providing improved conditions for the triggering of shallow convection. Inclusion of cloud liquid water in the formation of large-scale cloud led to improved cloud diagnostics in the stratocumulus region. Finally, the modifications to shallow convection, including both the triggering and the changed closure calculation, described in sections 3a(1) and 3a(2), significantly improved the performance of the shallow convection scheme in CRCMM. A more quantitative assessment of which specific parameterization change led to the improvements seen in Fig. 4d is deferred to section 5, where we present a number of controlled numerical experiments aimed at isolating the role of specific parameterization changes.

Liquid water path (LWP) was another variable poorly simulated in the EUROCS study. Most participant models underestimated LWP in the stratocumulus region and overestimated LWP in the equatorial regions, leading to positive biases, both for net shortwave radiation at the TOA and downward shortwave radiation at the surface in the stratocumulus regions and negative biases elsewhere (Siebesma et al. 2004). Figure 5a presents the seasonal mean LWP simulated both by CRCM4 and CRCMM, together with observational estimates from SSM/I. CRCM4 simulates less than 40 g m\(^{-2}\) LWP in the stratocumulus region, while SSM/I observations suggest a value of ~100 g m\(^{-2}\) in this region. Coincident with the LWP and cloud biases, CRCM4 overestimates the surface downwelling shortwave radiation (DSWR) in the stratocumulus region by ~60 W m\(^{-2}\) (Fig. 5b). Conversely, in the shallow cumulus region (~12°–20°N), the large overestimate of cloud fraction causes an underestimate of surface DSWR of ~60 W m\(^{-2}\). The CRCMM-simulated LWP is somewhat improved in the stratocumulus region although still underestimated; this results in a reduction of the positive DSWR bias in this region by ~20 W m\(^{-2}\). In the shallow cumulus region, the improvements seen in the cloud fraction amounts in CRCMM (Fig. 4d) are not realized in the surface DSWR, which remains biased low. This is due to a positive bias in the simulated LWP in CRCMM and reiterates the impor-
tance of an accurate simulation of both the cloud fraction and LWP in order to achieve realistic surface radiation fluxes.

Figure 6 shows the frequency of occurrence of deep or shallow convective activity in the BKF scheme along the GPCI cross section. Frequency of occurrence is defined as the percentage of times that the BKF scheme is activated out of 736 three-hourly outputs from the model at each location along the cross section. The observed cloud amounts in Figs. 4c and 4d suggest shallow convection is most frequent in the latitude band 11°–22°N along the cross section, where a relative minimum in cloud fraction is seen. This is coincident with the acknowledged climatological zone of trade wind cumulus (Riehl 1954). In CRCM4, shallow convection is relatively infrequently triggered in this region, being far more prevalent in the ITCZ with a 75% frequency of occurrence. In CRCMM, shallow convective activity is greatly reduced in the ITCZ and its maximum activity, ~80% of the total simulation time, shifts to the 14°–17°N zone. This increase in shallow convective activity acts to deepen the boundary layer in this region (refer to the height of 50% contour in Fig. 10) and as a result, the cloud fraction is reduced by 20%–25% (Fig. 4d). A small improvement in the TOA and surface solar radiation, accompanying this reduction in cloud fraction, can be seen in Figs. 5b and 5c. However, the total reflectivity of the shallow cumulus zone remains too high, mainly due to an overestimate of LWP (Fig. 5a).

Comparison of the apparent heating rate \( Q_1 = (\partial T/\partial t) \) and the moistening rate \( Q_2 = (L/C_p)(\partial q/\partial t) \) can quantitatively illustrate the contribution of the subgrid-scale parameterized process (e.g., convection) to the total model heat and moisture budget. Figure 7 illustrates the seasonal mean contribution of the parameter-
ized shallow convection to Q1 and Q2. Consistent with the frequency of occurrence of shallow convection illustrated in Fig. 6, only one Q1 maximum, located in the ITCZ and peaked at 850 hPa, is produced by the CRCM4 shallow convection (Fig. 7a). CRCMM simulates double maxima in Q1 at 5°N and 15°N, respectively (Fig. 7b). Both peaks are located around 900 hPa but the heating source in the north is much stronger and wider. Corresponding with the Q1 changes, Q2 in CRCMM also splits into two maxima with the northward peak clearly being the more intense (Fig. 7d). The Q1 and Q2 profiles clearly show the action of shallow convection is to dry the boundary layer though an upward flux of moisture, which subsequently condenses and warms the atmosphere around the 900-hPa level. One important feature to note in the CRCMM Q2 results is the moistening due to shallow convection at the top of the boundary layer (~875 hPa). We assume this is due to shallow convective overshooting and subsequent evaporation of detrained liquid water, which is an important moisture source for the subtropical midtroposphere (von Salzen et al. 2005). This feature is largely absent in CRCM4 due to the shallow convection preferentially detraining into the much moister equatorial region (2°–8°N).

A comparison of the seasonal mean precipitation along the cross section is given in Fig. 8, with TRMM 3B satellite estimates (Adler et al. 2003; Huffman et al. 2007) used as observations. TRMM suggests close to zero precipitation in the stratocumulus and shallow cumulus regions, and a peak in precipitation of around 11 mm day$^{-1}$ in the ITCZ at 8°N. CRCM4 overestimates precipitation by 1–2 mm day$^{-1}$ along the cross section in the stratocumulus and shallow cumulus regions and underestimates precipitation in the ITCZ by 4 mm day$^{-1}$. Figure 8b shows the contribution of just the large-scale precipitation scheme to the total precipitation. The majority of the erroneous precipitation in the zone 15°–30°N clearly emanates from the large-scale precipitation in CRCM4. Inclusion of the parameterization of precipitation evaporation in CRCMM is the main cause of the reduction in large-scale precipitation in this zone, with total precipitation rates now being closer to the TRMM observations. There is a large increase in the convective precipitation in the ITCZ in CRCMM relative to CRCM4, leading to a positive bias in ITCZ precipitation. This increase does not appear to be linked to an increase in the frequency of deep convection in the ITCZ, which is actually marginally reduced in CRCMM (see Fig. 6). Figure 9 compares the

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**Fig. 7.** The apparent (a), (b) heating (Q1) and (c), (d) moistening (Q2) rates contributed only by shallow convection in (a), (c) CRCM4 and (b), (d) CRCMM (K day$^{-1}$).
seasonal mean vertical profiles of the convective mass fluxes, $Q_1$ and $Q_2$, at $8^\circ$N in CRCM4 and CRCMM. The mean convective mass flux and $Q_1$ terms are clearly more intense in CRCMM, indicating that when deep convection occurs in CRCMM, it does so at a much higher intensity with the mean mass flux profile penetrating to $\sim 100$ hPa compared to 200 hPa in CRCM4. There is also an indication of convective over-shooting at the tropopause in CRCMM, with associated cooling and moistening resulting from evaporation of detrained convective ice (Derbyshire et al. 2004); this feature is absent in CRCM4. As will be seen later (in Fig. 12), upward-directed vertical velocities increase dramatically in CRCMM, being much closer to those in ERA-40. The specific parameterization modifications causing this large change in convective activity will be discussed in section 5.

### b. Vertical profiles along the GPCI cross section

In this section, we present latitude–pressure plots of relative humidity, cloud fraction, and vertical velocity along the GPCI cross section to better understand the changes in cloud and convection seen earlier. The
model-simulated results are compared with each other and against the ERA-40 reanalysis, which is used as a quasi-observational reference.

Figure 10 displays vertical profiles of relative humidity along the cross section. While the broad features are similar, the CRCM4 results differ from those of the ERA-40 in several important respects. The model simulates a relatively flat and shallow boundary layer below 900 hPa, from the stratocumulus region to the tropical ITCZ, while ERA-40 shows a clear slope of the boundary layer top (defined here by the height of the 50% relative humidity contour) in the subtropics, as the underlying sea surface temperature gets warmer and the boundary layer becomes increasingly unstable. CRCM4 is also drier in the midtroposphere in the deep convective ITCZ region, which can be seen by comparing the height of the 50% contour lines at 8°N between the CRCM4 simulation (Fig. 10a) and the ERA-40 reanalysis (Fig. 10c). The CRCMM-simulated relative humidity profile is improved compared to CRCM4 if one uses ERA-40 as a reference. CRCMM better simulates the increasing depth of the boundary layer southward along the cross section in the subtropics. Furthermore, CRCMM improves the dry bias seen in CRCM4 in the midtroposphere of the ITCZ (Fig. 10b). However, some deficiencies still exist: for example, the boundary layer simulated by CRCMM is shallower and moister than that with the ERA-40 in the stratocumulus and shallow cumulus regions; both of the models simulate a sharper gradient from the moist boundary layer to the drier midtroposphere than in ERA-40. In this respect, there is no difference between the CRCMM and CRCM4 results.

A comparison of the seasonal mean vertical profile of cloud fraction simulated by both of the models is shown in Fig. 11, along with the ERA-40 reanalysis. Corresponding to the distributions of relative humidity, both models simulate the basic structures of cloud cover along the cross section, with a relative maximum in low-level clouds in the northern part of the cross section and a maximum in upper-level cloud in the ITCZ. In detail, the simulation of CRCMM agrees more favorably with the ERA-40 reanalysis than does CRCM4. CRCM4 incorrectly simulates a maximum in cloud fraction in the subtropical boundary layer (around 14°–20°N) with less cloud in the stratocumulus region to the north (20°–25°N). This feature is clearly improved in CRCMM, with stratocumulus cloud amounts reaching 40% at 950 hPa reducing to a minimum in cloud fraction around 900 hPa in the band 11°–15°N, coincident with the minimum in ERA-40 clouds, although at a slightly lower elevation. Both CRCM4 and CRCMM simulate a reasonable amount of upper-tropospheric clouds in the ITCZ, although both have the cloud maximum at slightly lower levels than in ERA-40. Additionally, CRCMM also simulates cloud cover in the mid- to upper troposphere near 25°N. While the observation of optically thin cirrus clouds is difficult, ISCCP observa-
tions suggest a 5%–10% high cloud fraction in this region; another observational dataset, the High-Resolution Infrared Sounder (HIRS; Wylie and Menzel 1999), reports slightly more high cloud than ISCCP over this region. We therefore feel the CRCMM cloud amount simulated in the subtropical upper troposphere may be reasonable.

Figure 12 presents the seasonal mean cross section of the resolved vertical velocity. Compared to ERA-40, CRCM4 simulates a relatively weak vertical circulation, including both the ascending branch in the ITCZ and the descending branch over the subtropics. CRCMM more closely matches the vertical circulation strength
seen in ERA-40, reproducing the narrow ascending branch located between 5° and 10°N, with a maximum vertical velocity of −0.1 Pa s⁻¹ at 850 hPa. It also more correctly simulates the large-scale subsidence motion, which usually reaches its maximum in the subtropics around 35°N with a vertical velocity over 0.1 Pa s⁻¹. While acknowledging that the ERA-40 values are a mixed model–assimilation product and potentially an overestimate, related to the known excess precipitation in the ERA-40 ITCZ (Hagemann et al. 2005; also see Fig. 8a), we feel the changes in vertical velocity going from CRCM4 to CRCMM do suggest a significant improvement in the vertical structure of the convective forcing in the CRCMM atmosphere.

5. Sensitivity experiments

In the previous section, some basic improvements in the CRCMM results were documented for the total package of parameterization modifications. In discussing these results, certain aspects of the improvements were qualitatively related to specific changes in a model parameterization (e.g., consideration of the evaporation of falling large-scale precipitation in reducing the total surface precipitation rate in the subtropics). In this section, a more quantitative approach of identifying the contribution of the various parameterization changes to the seasonal mean climate is made through a series of sensitivity experiments.

In section 3, we described the modifications to the model shallow convection, deep convection, turbulence, large-scale cloud formation, and large-scale precipitation, respectively. Corresponding to these modifications, we have performed five sensitivity experiments, named NSHL, NDEP, NPBL, NCLD, and NEVP, respectively. Each experiment explores the effects of one specific category of modifications by excluding it from the physics of the CRCMM. The sensitivity experiments, therefore, have the same model physics as CRCMM except that one group of modifications has been set back to the original CRCM4 approach. For instance, experiment NSHL means that modifications related to shallow convection, as summarized in section 3a, have not been included in the CRCMM physics, so that the only difference between NSHL and CRCMM is the parameterization of shallow convection. Similarly, experiment NCLD means that no modification to the large-scale cloud formation has been made and so on for the other experiments. Differences between a given sensitivity experiment and CRCMM can more easily be attributed to one specific set of parameterization changes, although a complete separation of the contribution of each individual parameterization to the combined response is impossible due to the highly nonlinear way in which the individual changes interact (Stein and Alpert 1993). In general, the larger the difference between an experiment and CRCMM, the bigger the impact a given change has on the simulated seasonal mean values.

Figure 13 displays the total precipitation, total cloud cover, and frequency of shallow cumulus convection simulated by the sensitivity experiments along the GPCI cross section.
experiments, together with the observations and the results from CRCM4 and CRCMM. Compared with the CRCM4 results, all sensitivity experiments reduce the excessive stratiform precipitation over the stratocumulus and shallow cumulus regions in the subtropics; all experiments also intensify precipitation in the ITCZ (Fig. 13a). We therefore conclude that all the parameterization changes, to varying degrees, have contributed to these two changes in CRCMM.

For total cloud cover, there is a general pattern followed by most sensitivity experiments with the excessive cloud in the shallow cumulus region simulated by CRCM4 being corrected. However, experiment NPBL, which calculates eddy diffusivity as in the original CRCM4, simulates the worst cloud cover, with more than 90% cloud amount in the shallow and deep convection regions. This seems to imply that the modifications to the boundary layer scheme have a significant impact on the improved cloud fraction in the shallow cumulus region. Figure 14 further displays the vertical structures of the vertical velocity, relative humidity, and cloud cover simulated by experiment NPBL. As can be seen in Fig. 14c, NPBL simulates a quite weak Hadley circulation, including both the ascending branch in the ITCZ and the descending branch in the subtropics; moreover, NPBL simulates the flattest, thinnest and thus wettest boundary layer among all experiments. In this experiment, moisture is unrealistically trapped near the surface, below 900 hPa (Fig. 14a), with little vertical transport through the boundary layer (Fig. 13c indicates that nearly no shallow convection has been triggered in experiment NPBL because of the extreme stable conditions in the boundary layer). The direct result of this is excessive stratus cloud amounts in the boundary layer and a deficiency of cloud in the upper troposphere (Fig. 14b). Figure 15 further compares the seasonal mean eddy diffusivity simulated by CRCM4 and CRCMM. CRCMM simulates much stronger turbulent mixing, resulting in a deeper boundary layer along the cross section than is seen in CRCM4. Compared with diagnosed eddy diffusivities from LES model simulations of a stratocumulus case during the First ISCCP Regional Experiment (FIRE-I; De Roode 2007), the magnitude and the profile of the eddy diffusivity simulated by CRCMM are closer to the LES results, where the maximum eddy diffusivity for moisture could reach 550 m² s⁻¹ between 300–400 m above the sea surface.

Although the boundary layer is even flatter, the NPBL relative humidity profile looks somewhat like that in CRCM4. This is understandable since both versions use the original boundary layer scheme. However, the cloud fraction diagnosed from NPBL has changed enormously from CRCM4. NPBL uses the Xu–Randall cloud scheme, while CRCM4 uses the relative humidity–based scheme. It would appear that the CRCM4 cloud parameterization has been tuned to give reason-
able cloud amounts based on an incorrect relative humidity input. Introduction of a new cloud scheme, using largely the same relative humidity values, then leads to a significant degradation in the simulated cloud amounts. This illustrates that different schemes interact and strongly depend on each other. If one scheme (e.g., the cloud fraction scheme) of a model is changed, it is important to critically assess the relevant fields that impact this scheme (e.g., relative humidity); otherwise, an improvement in only one scheme may lead to significantly worse overall results. The logical conclusion from the analysis of the NPBL results, compared to all other experiments, is that changes to the boundary layer eddy diffusivity calculation were crucial in the development of a deeper boundary layer in CRCMM and subsequent improved triggering of shallow convection and subtropical cloud amounts.

The importance of the modification to the cloud scheme can be further illustrated by comparing the total cloud cover simulated by NCLD, CRCM4, and CRCMM, as displayed in Fig. 13b. Clearly, the original relative humidity–based cloud scheme does not work well in the CRCMM physics. The total cloud fraction simulated by NCLD is even worse than CRCM4, with only 50%–60% cloud cover in the stratocumulus regions, 30% lower than in the observations. It seems that compensating errors, between a too moist boundary layer and the cloud parameterization scheme in CRCM4, led to reasonable stratocumulus cloud amounts for the wrong thermodynamic structure. Improving the boundary layer structure while not updating the cloud scheme (NCLD) results in a worse simulation of cloud fraction, even with an improved thermodynamic structure, again emphasizing the need to carefully consider interactions between various parameterization schemes in a model.

The effect of modifications to the shallow cumulus scheme do not appear to be overly significant when we compare the frequencies of the shallow cumulus convection activated in experiments NSHL, CRCM4, and CRCMM, as shown in Fig. 13c. The shallow convection frequency in NSHL is quite similar to that in CRCMM, indicating that modifications to the shallow cumulus convection are not the fundamental reason for the improvements in shallow cumulus activity seen between CRCM4 and CRCMM. On the other hand, nearly no shallow convection is triggered in experiment NPBL, implying that changes to the boundary layer scheme have a significant impact on the triggering of shallow convection in the model. An additional experiment, which excluded both modifications to shallow cumulus convection and boundary layer schemes, produced very similar results to CRCM4 with respect to shallow convective activity (not shown), indicating the tight coupling between the shallow convection and boundary layer mixing schemes.

The comparison between NDEP, CRCM4, and CRCMM demonstrates the impact of the modifications to deep convection. Figure 16 displays the vertical profiles of the relative humidity, cloud cover, and vertical velocity simulated by NDEP; all of which are quite similar to CRCMM in the stratocumulus and shallow cumulus regions. In contrast, the vertical profiles in the ITCZ region shift back to be like CRCM4, since experiment NDEP uses the original deep convection with an undilute updraft in the CAPE closure. In summary, the large sensitivities displayed by NDEP indicate that updates to the deep convection scheme from CRCM4 to CRCMM are primarily responsible for the changes seen in the ITCZ in CRCMM.
have been refined to improve the overall model simulation over the tropical Pacific. The modifications touch upon shallow and deep convection, large-scale cloud formation, vertical eddy diffusivity, and large-scale precipitation. The main modification set includes (i) the addition of a relative humidity–related perturbation to the trigger function for shallow convection; (ii) the use of a convective vertical velocity scale in the boundary layer to close the shallow cumulus-cloud-base mass flux; (iii) the replacement of the original undilute updraft by a dilute updraft in the calculation of the CAPE closure in deep convection; (iv) the introduction of a nonlocal eddy diffusivity term that helps to improve the turbulence and moisture structures in the boundary layer, thus providing realistic triggering conditions for deep and shallow convection; (v) the utilization of a more advanced cloud fraction parameterization, including cloud water in the calculation of cloud fraction, leading to improved large-scale cloud amounts in the stratocumulus and shallow cumulus regions; and (vi) the consideration of evaporation of falling precipitation, which reduces the amount of precipitation erroneously reaching the surface over the subtropical oceans.

The updated model, CRCMM, shows substantial improvements in many aspects of the simulated seasonal mean climate compared to the original CRCM4. The maximum frequency of shallow convection shifts from the ITCZ in CRCM4 to the subtropics, near the Hawaiian Islands, in CRCMM; accordingly, excessive cloud in the shallow cumulus region in CRCM4 is significantly reduced. Finally, CRCMM-simulated vertical structures of relative humidity, cloud cover, and vertical velocity are in better agreement with the ERA-40 reanalysis than those of CRCM4.

A number of sensitivity experiments were performed in order to isolate the main modifications leading to the improved seasonal mean results seen in CRCMM. The large sensitivity displayed by experiment NPBL showed that the subgrid-scale boundary layer mixing scheme is of fundamental importance in setting the correct thermodynamic structure in the boundary layer, from which many schemes (e.g., shallow and deep convection, cloud fraction parameterization) subsequently are influenced. The main finding from these sensitivity tests is that changes to one parameterization scheme often do not directly lead to an overall improvement in model results unless potential interactive deficiencies in other schemes are also addressed at the same time.

Although significant improvement in the simulation of the seasonal mean climate has been achieved by this set of modifications, some deficiencies still remain. For example, CRCMM precipitation appears to be exces-

6. Summary and conclusions

This paper presents seasonal mean results from the Canadian Regional Climate Model GPCI integration for JJA 1998. Several cloud-related parameterizations

![Fig. 16. Seasonal mean vertical profiles of (a) relative humidity (%), (b) cloud cover (%), and (c) vertical velocity (Pa s⁻¹) simulated by sensitivity experiment NDEP.](http://journals.ametsoc.org/mwr/article-pdf/136/11/4168/4239646/mwr2494_1.pdf)
sive in the ITCZ region (Fig. 8a). CRCMM also overestimates LWP in the shallow cumulus region, coincident with an underestimate of the downwelling surface solar radiation. Finally, CRCMM simulates a shallower and moister boundary layer than ERA-40 in the strato-cumulus and shallow cumulus regions, and for that reason, the model simulates a sharper gradient between the moist PBL and the dry midtroposphere above (Fig. 10). Further investigation is required to isolate the causes of these deficiencies before targeted improvements can be investigated.

It should be pointed out that because of the limitation of available observational data, our comparison has been focused on 1998. Limited evaluation for JJA 2003 suggests the major improvements seen in 1998 are also applicable in other years. We are presently evaluating the impact of the CRCMM updates on the regional climate over North America, where an interactive land surface scheme can now respond to changes in surface fluxes and precipitation. We also plan in the near future to explore the transferability of these modifications to other nonnative model domains such as those outlined in the Intercontinental Transferability Study (ITCS; Takle et al. 2007).

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