Impact of Boundary Layer Processes on Simulated Tropical Rainfall

YOUNG-HWA BYUN and SONG-YOU HONG

Global Environment Laboratory, Department of Atmospheric Sciences, Yonsei University, Seoul, South Korea

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ABSTRACT

The impact of boundary layer (BL) processes on simulated tropical precipitation was studied using the National Centers for Environmental Prediction (NCEP) Medium-Range Forecast (MRF) Model. A new BL scheme, which is a nonlocal mixing concept of Noh et al. after Troen and Mahrt, was successfully incorporated into the MRF Model. In this study, simulations with 10-member ensembles were conducted for boreal summers of normal, El Niño, and La Niña years, respectively. In particular, the authors focused on the impact on tropical rainfall of the new BL scheme when two different convection schemes are utilized respectively in the model.

The new BL scheme improves simulated tropical precipitation overall and in particular reduces the simulated rainfall in the central and eastern equatorial Pacific Ocean. This reduction over the eastern Pacific is a direct effect of the new BL scheme resulting in less mixing of heat and moisture and is irrespective of the convection scheme. The effect of BL processes over the western Pacific, however, is indirectly related to the change of the Walker circulation and highly dependent on the convection scheme selected.

1. Introduction

Rainfall is a critical component of the role of the Tropics in the global hydrological cycle. Three-fourths of the atmosphere’s heat energy is derived from the release of latent heat by precipitation, an estimated two-thirds of which falls in the Tropics. Since differences in large-scale rainfall patterns and their associated energy release in the Tropics affect global circulation, proper treatments related to precipitation physics are important in modeling weather and climate processes that produce tropical rainfall.

Depending on the particular physical parameterization scheme utilized in a model, results vary considerably, as is revealed in numerous studies on general circulation model (GCM) intercomparisons (e.g., Sperber and Palmer 1996; Gates et al. 1999; Shukla et al. 2000; Kang et al. 2002a,b). For example, in their study on the predictability of five seasonal forecast models, Shukla et al. (2000) pointed out that the internal variability of models with different physics is extreme, even though these models have identical dynamic cores. In particular, Kang et al. (2002b) demonstrated the importance of convection schemes in GCMs. According to their analysis of 10 different models joined in intercomparison of atmospheric GCMs, two model groups were classified by characteristics of the simulated precipitation over the western Pacific. The authors indicated that the differences in performance of these two model groups should be related to the physical parameterizations and particularly to the convection schemes in the models.

The importance of the convective parameterization scheme in a GCM such as a seasonal prediction model has been pointed out (e.g., Sud et al. 1992; Kumar et al. 1996; Zhang et al. 1998; Kanamitsu et al. 2002a). Sud et al. (1992) showed that a cumulus parameterization scheme had a significant influence on simulated circulation and precipitation; the influence was conspicuous in a tropical region in a GCM. Kumar et al. (1996), Zhang et al. (1998), and Kanamitsu et al. (2002a) also showed that improvement of a simulated climate over the Tropics and subtropics was largely due to a change in convection scheme.

Recent studies suggest that the boundary layer (BL) scheme may be a sensitive factor that is particularly related to precipitation physics in global short- and medium-range forecast models, as well as in mesoscale models. For example, Hong and Pan (1996) showed that a slight change of parameters in BL formalism could significantly affect both the distribution and the amount of rainfall forecast in the National Centers for Environmental Prediction (NCEP) Medium-Range Forecast (MRF) Model. Basu et al. (2002) demonstrated systematic improvement of the simulation of a monsoon system over India by introducing a nonlocal vertical diffusion scheme for the mixed layer. Nevertheless, in GCMs that included seasonal prediction and climate models, few sensitivity studies have been carried out on the impact...
of the BL parameterization scheme [e.g., Martin et al. (2000) showed that the BL and cloud structure in the semipermanent stratocumulus regions over the eastern subtropical oceans were noticeably improved due to the change of the BL scheme].

The local-$K$, where $K$ is the eddy diffusivity coefficient, approach (Louis 1979) was popular among BL schemes in atmospheric models until the early 1990s. In recent years, because of its simplicity and ability to represent large-eddy turbulence within a well-mixed BL, the nonlocal scheme based on Troen and Mahrt (1986) has been successfully applied, with further generalization and reformulation, to GCMs and to numerical weather prediction models (Holtslag et al. 1990; Holtslag and Bouville 1993; Giorgi et al. 1993; Hong and Pan 1996). In particular, since Hong and Pan (1996) successfully used this concept with the NCEP MRF Model (hereafter referred to as the MRF BL scheme), this scheme has been widely implemented in both mesoscale and global models (e.g., Bright and Mullen 2002; Basu et al. 2002; Kanamitsu et al. 2002a,b). The MRF BL scheme is preferable because, compared with the local-$K$ approach through the inclusion of countergradient flux terms, it enables the realistic development of a well-mixed layer. However, there appear to be some deficiencies in the MRF BL scheme: recent studies report excessive mixing. For example, in referring to their respective simulations of Hurricanes Bob (1991) and Diana (1984), Braun and Tao (2000) and Davis and Bosart (2002) showed that the MRF BL scheme produced a significantly weaker storm than the other BL scheme; this was due to the excessively deep mixing and consequent drying of the lower BL. In addition, in his study on analytic solutions of a $K$-profile model with nonlocal fluxes, Stevens (2000) showed that the $K$-profile shape and the coefficients for countergradient mixing in Troen and Mahrt (1986) could be too large to account for a well-mixed BL.

Recently, a new nonlocal diffusion concept to improve deficiencies in Troen and Mahrt (1986) was proposed by Noh et al. (2003) and implemented within a numerical model by Hong et al. (2003), after further generalization and reformulation. The key to the new BL scheme is that the entrainment processes above the minimum flux level are expressed explicitly in the turbulence diffusion equations for the prognostic variables. According to the preliminary results of Hong et al. (2003) in experiments using a one-dimensional simple model with idealized surface boundary forcing, the new BL scheme effectively reduces mixing in the MRF BL scheme. Evaluation of daily forecast runs in the Weather and Research Forecast (WRF) Model revealed some improvement in the simulation of BL structure and precipitation forecasts (available online at http://wrf-model.org).

In the present study, the authors investigated changes in tropical precipitation due to using, in a GCM, the new BL scheme by Hong et al. (2003) and Noh et al. (2003) through a comparison with the current MRF BL scheme (Hong and Pan 1996). In particular, the authors focused on the effect on the simulation of tropical rainfall of the BL scheme’s dependence on two different cumulus parameterization schemes. Section 2 describes the global model and the BL scheme utilized in this study, as well as the experimental designs for three boreal summers and their characteristics for each year. Section 3 examines the impact of the new BL scheme on tropical rainfall and large-scale circulation. A summary and concluding remarks are discussed in section 4.

2. The model and experiment designs

a. The NCEP MRF Model

The NCEP MRF Model is a global spectral model (Sela 1980). Subsequent developments of this model are described in Kanamitsu (1989), Kanamitsu et al. (1991), Hong and Pan (1996), Caplan et al. (1997), and Kanamitsu et al. (2002a). The model used in this study is a version of the NCEP MRF Model with the physics package that was operational as of January 2000. Model physics include long- and shortwave radiation, cloud–radiation interaction, planetary BL processes, deep and shallow convection, large-scale condensation, gravity wave drag, enhanced topography, simple hydrology, and vertical and horizontal diffusions. The MRF Model utilizes a two-layer soil model of Pan and Mahrt (1987), which includes both soil thermodynamics and soil hydrology, modeled as diffusion processes. The evaporation process in the surface energy balance is modeled by direct evaporation from the bare soil surface, transpiration through the leaf stomata, and reevaporation of precipitation intercepted by the leaf canopy. Boundary layer physics employs a nonlocal diffusion scheme (Hong and Pan 1996), which is described in detail below.

For precipitation physics, both large-scale condensation and a deep-convection parameterization scheme are employed in the model. The large-scale precipitation algorithm checks for supersaturation in the predicted specific humidity. Latent heat is released as the specific humidity and temperature are adjusted to saturation values. The cumulus parameterization for deep convection uses the simplified Arakawa–Schubert (SAS) scheme in the current MRF Model. This follows Pan and Wu (1995), which is further based on Arakawa and Schubert (1974), as simplified by Grell (1993) with a saturated downdraft.

Another convection scheme selected for this study is the relaxed Arakawa–Schubert (RAS) scheme based on Moorthi and Suarez (1992), which was implemented in the recent NCEP seasonal forecast model (Kanamitsu et al. 2002a). According to an idealized experiment to examine the capability of the convection scheme, Kanamitsu et al. (2002a) found that the RAS parameterization scheme performs better than the SAS in terms of the Pacific–North America (PNA) pattern in response
to an idealized sea surface temperature (SST) forcing over the equatorial Pacific. For short- and medium-range forecasts, the SAS scheme had good skill in predicting tropical precipitation (Kalnay et al. 1996). The main differences between the SAS and RAS schemes have to do with two components: clouds and the treatment of downdrafts. The SAS scheme allows only one type of cloud, while the RAS scheme allows clouds with different tops. The SAS scheme considers saturated downdrafts based on empirical formulation; the RAS scheme incorporates no downdraft mechanisms. These variances result in different vertical heating and moistening profiles and change the tropical precipitation.

b. The BL parameterization scheme

The MRF BL scheme was implemented by Hong and Pan (1996) based on Troen and Mahrt (1986). In this scheme, turbulent diffusivity coefficients are calculated from a prescribed profile shape as a function of BL heights and scale parameters derived from similarity requirements. Above the mixed layer, the local diffusion approach is applied to account for free atmospheric diffusion.

The turbulence diffusion equation for the prognostic variables $C(u, v, \theta, q)$ can be expressed for the mixed layer as follows:

$$\frac{\partial C}{\partial t} = \frac{\partial}{\partial z} \left[ K_c \left( \frac{\partial C}{\partial z} - \gamma_c \right) \right].$$

(1)

Here, $K_c$ is the eddy diffusivity coefficient, and $\gamma_c$ is a correction to the local gradient that which incorporates the contribution of the large-scale eddies to the total flux. As shown in Hong and Pan (1996), because the turbulent diffusivity coefficient on the right side of Eq. (1) is calculated from a profile shape as a function of BL height, the determination of the BL height is critical to the representation of nonlocal mixing. The BL height $h$ is given by

$$h = \text{Rib}_c \left[ \frac{\theta_\text{v}}{g[\theta(h) - \theta_s]} \right].$$

(2)

where Rib$_c$ is the critical bulk Richardson number; $U(h)$ is the horizontal wind speed at $h$; $\theta_v(h)$ and $\theta_s$ are the virtual potential temperature at $h$ and the lowest model level, respectively; and $\gamma_c$ is the appropriate temperature near the surface. The temperature near the surface is defined as

$$\theta_s = \theta_v + \theta_v \left[ \frac{-b(w' \theta_v')_h}{w_v} \right],$$

(3)

where $\theta_v$ is the scaled virtual temperature excess near the surface, $w_v$ is the mixed-layer velocity scale, $(w' \theta_v')_h$ is the virtual heat flux from the surface, and the proportionality factor $b$ is set as 7.8.

Despite the importance of determining the BL height, several uncertainties exist in establishing $h$. Because $\theta_{v}$ in Eq. (2) could become too large when a weak surface wind results in an unrealistically large $h$, Hong and Pan (1996) put a maximum limit of $\theta_{v}$ as 3 K. On the other hand, $h$ could be too large with an excessively strong wind speed at a level $z$, as is pointed out by Noh et al. (2003) and Mass et al. (2002). One apparent reason for this is the characteristics of the bulk Richardson number at a level $z$, which is given by

$$\text{Rib}(z) = \frac{g[\theta(z) - \theta_s]z}{\theta_s u_U(z)^2}.$$  

(4)

In Eq. (4), it can be seen that the thermal excess $[\theta(z) - \theta_s]$ due to the nonzero Rib$_c$ (currently 0.5) becomes larger as the wind speed at $z$ is stronger, since computed Rib is reduced when the wind speed is weak. The excess can be as large as 6.1 K when the wind speed is 20 m s$^{-1}$. The combined effects of thermal excess due to surface flux in Eq. (3) and the bulk Richardson number in Eq. (4) can be unrealistically large. Furthermore, Eq. (4) provides the information that the estimated Rib increases as $z$ increases given the same temperature perturbation and wind speed, which implies more mixing when $h$ is smaller.

Recently, Noh et al. (2003) proposed some modifications, based on the large-eddy simulation data, of the previous $K$-profile model of the BL reported by Troen and Mahrt (1986). The modifications include three parts. First, the heat flux from the entrainment at the inversion layer is incorporated into the heat and momentum profiles. Second, vertically varying parameters are proposed, in contrast to constant values in the previous model. Finally, nonlocal mixing of momentum is included. According to their results, the new BL model improves the predictability of the BL heights because it resolves the problem of excessive mixing with strong wind shear and produces more realistic heat and momentum profiles. These modifications were successfully implemented within the WRF Model by Hong et al. (2003).

A brief description of the new BL parameterization scheme of Hong et al. (2003) appears below. The turbulence diffusion equation can be expressed for the mixed layer as follows:

$$\frac{\partial C}{\partial t} = \frac{\partial}{\partial z} \left[ K_c \left( \frac{\partial C}{\partial z} - \gamma_c \right) - (w'C')_h \left( \frac{\gamma_c}{h} \right) \right].$$

(5)

Here, $K_c$ and $\gamma_c$ are the same as in Eq. (1), and $(w'C')_h$ is the flux at the inversion. Compared with Eq. (1), the formula maintains the basic premise of Troen and Mahrt (1986) but includes an asymptotic flux term at the inversion layer; that is, the last term inside the square bracket of Eq. (5) explains entrainment processes at the BL top. Also, the new scheme specifies the BL height $h$ differently from the MRF BL scheme. The definition of the BL height in the new scheme is a level in which minimum heat flux exists, whereas the BL
height of the MRF BL scheme is the level that the heat flux becomes zero. In the new scheme, $h$ is determined to be the first neutral level considering the temperature perturbation due to surface buoyancy flux, which can be expressed by

$$\theta_v(h) = \theta_v + \theta_u \left[ \frac{a}{w_s} \right].$$

Here, $\theta_v(h)$ is the virtual potential temperature at $h$. This formula is the same as Eq. (3), but the proportionality factor $a$ and mixed-layer velocity scale $w_s$ have different values from the MRF BL scheme. Hong et al. (2003) examined the impact of this new BL scheme in a one-dimensional simple model. In this study, the new scheme effectively reduces excessive mixing in the MRF BL scheme. A detailed description of the new BL scheme, including equation sets, is presented in Hong et al. (2003) and Noh et al. (2003).

c. Experiment designs

To investigate the effect of the BL scheme’s dependence on different parameterization schemes for deep convection, the authors performed four experiments, as seen in Table 1. All the experiments were designed by combining two BL schemes (the MRF BL and the new BL) and two convective parameterization options (SAS and RAS). In Table 1, the first three letters refer to the BL scheme utilized; hereafter, the control scheme denotes the MRF BL scheme. The last letter indicates the convection scheme.

The MRF Model used in this study employs a resolution of T62L28 (triangular truncation at wave number 62 in the horizontal and 28 terrain-following sigma layers in the vertical). To avoid introducing uncertainties with the initial data, 10-member ensemble runs for each experiment were performed with an approximate 4-week lead time for the boreal summers (June–July–August) of 1996, 1997, and 1999. Hereafter, the authors refer to these periods by their respective years only. These three years were selected to examine a response of the BL scheme to different SST anomalies as a sur-

Table 1. Summary of the numerical experiments designed in this study.

<table>
<thead>
<tr>
<th>Experiments</th>
<th>BL scheme</th>
<th>Convection scheme</th>
</tr>
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<tbody>
<tr>
<td>CNT_S</td>
<td>MRF BL</td>
<td>SAS</td>
</tr>
<tr>
<td>NEW_S</td>
<td>New BL</td>
<td>SAS</td>
</tr>
<tr>
<td>CNT_R</td>
<td>MRF BL</td>
<td>RAS</td>
</tr>
<tr>
<td>NEW_R</td>
<td>New BL</td>
<td>RAS</td>
</tr>
</tbody>
</table>

Fig. 1. Observed SST anomaly (°C) and precipitation (mm month⁻¹) for boreal summers of each year. (a), (b), (c) SST anomaly with 0.5°C contour intervals. Dark (light) gray shading designates the anomalies greater (smaller) than ±0.5°C (±0.5°C). (d), (e), (f) Mean summer precipitation with contour intervals at 100 mm starting from 100 mm, and shading designates the precipitation amount greater than 200 mm.
face boundary forcing. The initial data for each ensemble was taken from the NCEP–Department of Energy (DOE) reanalysis II data (Kanamitsu et al. 2002b), starting from 0000 UTC of the 1st day of May to the 10th day with a 24-hr interval. In this study, the authors chose 3 yr in which to examine the effect of the BL scheme on different SST forcing. Observed SST data were used with a resolution of $1^\circ$ (Reynolds and Smith 1994) during the simulation period. Observed precipitation was taken from the Climate Prediction Center (CPC) Merged Analysis Monthly Precipitation (CMAP) data (Xie and Arkin 1997). The SST in 1996 shows a normal condition over the equatorial Pacific (Fig. 1a); the SST anomaly distribution in 1997 and 1999 reveals warm and cold El Niño–Southern Oscillation (ENSO) SST characteristics, respectively (Figs. 1b,c). In 1997 particularly, the El Niño signal as a warm SST anomaly over the eastern equatorial Pacific is so strong that an SST anomaly more than approximately $+4^\circ$C is shown near the western coast of Peru. This increased precipitation more than $+200$ mm when compared with normal precipitation over the eastern tropical Pacific (Figs. 1d,e). Strong precipitation activity near the western Pacific warm pool region also moved eastward in 1997. On the other hand, in 1999 the intensity of a cold SST anomaly over the eastern Pacific is not strong, yet a negative SST anomaly appears widely over the central Pacific (Fig. 1c). Tropical precipitation near the western Pacific warm pool region is shown to increase a little in comparison with a normal year, and this reduces precipitation over the central equatorial Pacific considerably (Fig. 1f).

3. Results and discussion

In this section, the authors focus on mean patterns of 10-member ensemble simulations for the boreal summer of 1996—first, because the authors are interested only in the mean feature of the tropical rainfall and atmospheric structure without uncertainties related to the initial data and, second, because a significant change of atmospheric structure due to a change to the BL scheme in 1997 and 1999 is considerably similar to that of 1996, except for the changes of location or intensity of response of the BL scheme due to different SST forcing. Therefore, most of the figures in this section show the effect of the BL scheme as mean fields.

In a seasonal forecast framework, climatology anom-
Fig. 3. Pattern correlation coefficients of monthly (Jun, Jul, and Aug) and seasonal (JJA) mean precipitation between the observation and simulated results for each year (1996, 1997, and 1999) with coefficients for (a)–(c) the SAS scheme and (d)–(f) the RAS cases. Each coefficient is averaged over the area with 50°S–60°N in lat and 0°–360° in lon.

Table 2. Global mean precipitation (mm day$^{-1}$) of summer obtained by each experiment.

<table>
<thead>
<tr>
<th>Experiments</th>
<th>1996</th>
<th>1997</th>
<th>1999</th>
</tr>
</thead>
<tbody>
<tr>
<td>CMAP</td>
<td>2.71</td>
<td>2.73</td>
<td>2.55</td>
</tr>
<tr>
<td>CNT_S</td>
<td>3.64</td>
<td>3.68</td>
<td>3.64</td>
</tr>
<tr>
<td>NEW_S</td>
<td>3.37</td>
<td>3.42</td>
<td>3.38</td>
</tr>
<tr>
<td>CNT_R</td>
<td>3.23</td>
<td>3.22</td>
<td>3.25</td>
</tr>
<tr>
<td>NEW_R</td>
<td>2.96</td>
<td>2.97</td>
<td>2.97</td>
</tr>
</tbody>
</table>

a. Tropical precipitation

In Fig. 2, both the CNT_S and the NEW_S experiments overall produce a tropical rainfall pattern comparable to the observation (cf. Fig. 1d) showing the main rainbelt along the intertropical convergence zone (ITCZ). However, it is apparent that the CNT_S experiment has some discernible defects, including excessive precipitation over the trade wind region north of the equator and underestimated precipitation over the western equatorial Pacific near the Maritime Continent. This aspect commonly appears in 1997 and 1999 (not shown). On the other hand, the new BL scheme improves the overall tropical precipitation pattern due to a considerable decrease in excessive rainfall over the eastern Pacific and an increase in insufficient rainfall over the western Pacific north of the equator (Fig. 2c). Such an improvement is well demonstrated, in comparison to pattern correlation coefficients (Figs. 3a–c). In this figure, the correlation coefficients of the model results with the new scheme are higher than those of the control case, except in June 1996. In addition to the precipitation pattern, the simulation with the new BL scheme improves global mean precipitation in terms of its amount (Table 2). In this table, global mean rainfall from the NEW_S experiment decreases approximately 10% as compared
with the CNT$_S$ case, which is closer to what was observed.

In the RAS convection scheme, it is also clear from Table 2 that the new scheme reduces the global mean precipitation by nearly 10% (cf. CNT$_R$ versus NEW$_R$); in turn, mean precipitation of the NEW$_R$ case comes close to the CMAP observation. According to Figs. 2d–f, the CNT$_R$ experiment shows a relatively well-organized rainfall pattern simulation when compared with the observation, except for an overestimate in rainfall near the Indian Ocean. In particular, over the eastern equatorial Pacific, the CNT$_R$ experiment shows a better pattern than the CNT$_S$ case in a quantitative respect by an overall decrease in equatorial precipitation along the ITCZ. The overall pattern does not, however, always improve, due to an excessive decrease in precipitation over the eastern Pacific region. This aspect can be easily seen in the pattern correlation coefficients of the experiments for the RAS convection scheme (Figs. 3d–f). In Fig. 3, it can also be seen that the RAS convection scheme generally outperforms the SAS scheme when the control BL scheme is employed, but the improvement due to the new scheme is not distinct in the RAS scheme.

In the case of the anomaly, the effect of the BL scheme on tropical rainfall is investigated. For the El Niño year, the observed precipitation anomaly shows the typical pattern for the warm ENSO SST forcing in the eastern Pacific, which is an increase of tropical precipitation over the central and eastern equatorial Pacific and a decrease over the western equatorial Pacific (Fig. 4a). The MRF Model reproduces this typical anomaly pattern (Figs. 4b–e) fairly well when compared with the observation. The model with the new scheme weakens the precipitation anomalies over the Tropics, irrespective of the convection scheme. The correlation coefficients due to the new BL scheme are also improved for both convection schemes selected, but to a greater extent in the SAS convection scheme (Fig. 6a).

In the La Niña case, the observed anomaly represents a decrease over a wide region in the western Pacific extending from north of Australia to the southern part of the Korean peninsula (Fig. 5a). Negative anomalies exist near the Indian Ocean and over the central Pacific. The simulated results reveal, however, randomly distributed anomalies, and the statistically significant areas are not distinct. Pattern correlations of precipitation anomalies are poor, irrespective of the convection scheme, but correlation due to the new BL scheme is better in the SAS scheme than in the RAS scheme (Fig. 6b). It is interesting to note that the model with the SAS convection scheme performs better than the RAS
scheme in terms of precipitation anomaly, even though the RAS scheme shows a better skill than the SAS scheme in terms of the mean pattern. The impact due to the new BL is clearer when the SAS convection scheme is employed, rather than the RAS scheme.

b. Large-scale circulation

Analyses of model results in section 3a generated the following questions: Why does the precipitation decrease overall? Why does the impact of the new BL scheme appear to be different in each convection scheme in the western Pacific and the eastern Pacific? What causes the varying responses of the new scheme to the SAS and RAS convection schemes? To answer these questions, the authors analyzed the differences in large-scale patterns between the control and the new scheme, focusing on 1996.

Figure 7 represents the distinctions in zonal mean temperature and specific humidity between the CNT_S
and NEW_S experiments. Variations in thermodynamical features due to the change of BL processes for the RAS convection scheme follow the analyses of SAS runs (not shown). In Fig. 7, it is apparent that temperature and moisture in the lowest layer below 950 hPa become colder and wetter due to the new BL scheme. Above it, warmer and drier air generally appears between 950 and 850 hPa. Such a feature results from less mixing due to the new scheme. According to Holtslag and Bouville (1993), the nonlocal BL scheme usually has a deep BL, especially in strong wind regions such as the Southern Hemisphere storm track and the eastern trade wind region, and the greatest BL height reaches about 2 km. The new scheme suppresses such excessive mixing by introducing explicit treatment of entrainment of the fluxes at the mixed-layer top (see Noh et al. 2003; Hong et al. 2003). Consequently, the model with the new BL scheme results in smaller convective available potential energy (CAPE) than the control scheme due to less mixing of heat and moisture and, in turn, reduces rainfall over the eastern Pacific, where strong trade wind convergence exists.

The effect of the new scheme over the western Pacific varies, depending on the convection scheme selected. First, a comparison of Figs. 8a,b and Figs. 2c,f indicates that the precipitation difference due to the BL scheme is primarily due to parameterized precipitation physics in the cumulus convection scheme. In Figs. 8a,b, an area of major decrease in rainfall appears along the ITCZ of the eastern Pacific in both simulations, but the amount is greater in the simulation with the SAS scheme than with the RAS scheme. Moreover, in the 10°–20°N area of the western Pacific, an increase in tropical precipitation is shown in the SAS case, whereas a decrease occurs in the RAS case. These spatial differences due to the convection scheme are clearly shown in the low-level moisture flux fields and their divergence fields (see Figs. 8c,d). As for the difference between the NEW_S and the CNT_S experiments, a divergence region appears over the central Pacific connected with enhanced easterlies over the central Pacific in the NEW_S experiment. Over the western Pacific, the NEW_S case shows that a wide convergence area due to intensified westerlies is located near 10°–20°N. In contrast, the experiment using the RAS scheme indicates a large area of divergence in the central Pacific near 5°N, whereas a convergence region is forced near the central Pacific at 5°–10°S. The NEW_R experiment shows, moreover, strong divergence in the 10°–20°N area of the western Pacific.

Consequently, it is apparent that SAS and RAS schemes have different influences on the large-scale flow with different sensitivities for change of the BL scheme and thus result in spatial differences. As mentioned in the previous section, the SAS scheme is different from the RAS scheme as to clouds and treatment of downdraft. The SAS scheme also has different triggers from the RAS scheme. According to Hong and Pan (1998), the SAS scheme in the MRF Model is modified with a convective trigger function that explicitly couples BL and convective precipitation processes, whereas the RAS scheme considers relative humidity only for triggering convection (Moorthi and Suarez 1992). Therefore, it is assumed that the differences between the SAS and RAS schemes in treatment of clouds and triggering processes affect the onset and development of the con-
Figure 9 shows vertical profiles of equivalent potential temperature \( \theta_e \) and saturation equivalent potential temperature \( \theta_{es} \) that were investigated to see the change of the CAPE related to the large-scale convergence. According to Yu et al. (1998), the distribution of moisture available for precipitation in the presence of large-scale convergence is spatially inhomogeneous and the effectiveness in producing precipitation per unit large-scale convergence is greater over the western Pacific Ocean than over the eastern Pacific. The vertical profiles of observed equivalent potential temperature (cf. thick lines marked as “ANAL”) in Fig. 9 simply represent the difference in moisture amount in two regions; that is, the \( \theta_e \) profile shows moister air over the western Pacific than the eastern Pacific, which is similar in aspect to the annual mean \( \theta_e \) climatologies from radiosonde observations in Folkins and Braun (2003).

Like the observed cases, these simulated results show moister profiles over the western Pacific than the eastern Pacific. For the SAS case in Figs. 9a,b, \( \theta_e \) and \( \theta_{es} \) profiles for the NEW_S and the CNT_S experiments are similar to the observed ones except for a \( \theta_e \) profile over the western Pacific. The NEW_S experiment indicates that \( \theta_e \) of the eastern Pacific moves to the drier side in the lower layer than in the layer from the surface to about the 700-hPa level; thus, the CAPE becomes smaller. Over the western Pacific, however, the difference in \( \theta_e \) profiles between the NEW_S and the CNT_S cases is unclear. Rather, the NEW_S experiment shows wetter conditions in the lowest model level (near 1000 hPa), resulting in an increase of the CAPE. In the RAS case (Figs. 9c,d), simulated profiles are quite different from the observed ones, perhaps because the convective parameterization of NCEP–DOE reanalysis data is based on the SAS scheme considering the deepest cloud; that is, the RAS scheme with a concept of multiple clouds.
tends to moisten the middle troposphere compared with the SAS because it allows detrainments by clouds with different tops. Despite the difference in vertical structures, $\theta_v$ and $\theta_w$ profiles of the RAS case clearly show the moisture content difference in two regions. However, the NEW_R experiment makes the $\theta_v$ profile in the low layer drier all over the western and eastern Pacific. The NEW_R experiment does not lead to an increase of the CAPE over the western Pacific that is different from the SAS case. Accordingly, the change of the CAPE due to the BL scheme over the western Pacific depends on the convection scheme; this response is closely related to the large-scale convergence.

To further clarify the change in large-scale patterns, the vertical motion and the streamlines in the zonal and vertical directions were plotted in Fig. 10. In the SAS case, it is apparent that the NEW_S experiment intensifies rising motion in the western Pacific and weakens sinking motion in the eastern Pacific, when compared with the control scheme (cf. Figs. 10a,c). This means that the new BL scheme plays a role in strengthening the Walker circulation (hereafter WC), resulting in a large amount of low-level moisture advection toward the western Pacific due to the intensified trade wind. Consequently, the enhanced WC in the NEW_S experiment leads to an increase in tropical precipitation over the western Pacific with active precipitation processes due to the large CAPE. However, the effect of the new BL scheme in the simulation with the RAS convection scheme is less distinct than in the SAS case (cf. Figs. 10b,d). In the figure, an intensification of the WC is not clear, although strengthened rising motion appears near the central Pacific near 170°E–150°W. Rather, the WC in the NEW_R case is weakened, leading to an overall weakness of the low-level easterly wind.

The change in the WC is clearly detected by the change in low sea level pressure (Fig. 11). The WC is associated with low sea level pressure in the west and high pressure in the east, and the basin-wide pressure gradient is the main driving force for the low-level zonal wind of the WC. The NEW_S experiment indicates higher sea level pressure over the central and eastern Pacific compared with the CNT_S case with lower pressure over

**Fig. 10.** (a), (b) Streamline for longitude–height section of wind differences and (c), (d) vertical velocity difference (hPa s$^{-1}$). All the variables are averaged from 20°S to 30°N for summer of 1996. (a) and (c) indicate the differences between the new and the control schemes for the SAS convection scheme, and (b) and (d) are the difference for the RAS convection scheme. Shaded areas in (c) and (d) indicate upward motion.

**Fig. 11.** The sea level pressure differences (hPa) between simulations with the new and the control schemes for summer of 1996. Sea level pressure is averaged over 20°S–30°N, and its deviations from the zonal mean are plotted.
the western Pacific (see solid dark line). The increase (decrease) of sea level pressure in the descending (ascending) branches of the WC indicates that the zonal pressure gradient between the western and eastern Pacific becomes stronger, leading to enhanced low-level easterly wind. Overall, the difference between the NEW_R and the CNT_R (thin line with closed circles) shows a positive value over the western Pacific and a negative value over the eastern Pacific. This zonal difference due to the new scheme in sea level pressure means that the driving force of the low-level easterly wind is weakened, as shown in Fig. 10b. Consequently, the NEW_R experiment reduces the rising (sinking) motion in the ascending (descending) branch of the WC, which in turn leads to a weakening of the WC.

4. Conclusions

In this study, the authors investigated the impact of BL processes on simulated tropical precipitation using the NCEP MRF Model. A new nonlocal BL process based on the concept of Noh et al. (2003) was successfully incorporated into the MRF Model, and simulations of tropical precipitation using the new BL scheme were compared with those using the MRF BL scheme. Simulations with 10-member ensembles were conducted for boreal summers of normal (1996), El Niño (1997), and La Niña (1999) years, respectively. The authors focused on the impact of using the new BL scheme with two different convection schemes on tropical rainfall.

Overall, the new scheme improves the precipitation over the Tropics by reducing rainfall in the central and eastern equatorial Pacific Ocean. This reduction is a direct effect of the new BL scheme’s resulting in less mixing of heat and moisture, which in turn produces smaller CAPE. This effect is common, irrespective of the convection scheme. Meanwhile, the effect of BL processes over the western Pacific is indirect and highly dependent on the convection scheme selected. The model with the SAS convection scheme improves precipitation over the western Pacific by increasing rainfall when the new scheme is used, whereas the RAS scheme suppresses rainfall activity in that region. As a result, improvement due to the new BL processes is clearer when the SAS convection scheme is employed than with the RAS scheme.

However, the response over the western Pacific is rather indirect and is closely related to the change of large-scale circulation induced by BL processes. A change in sea level pressure over the eastern Pacific induced by the new BL scheme results in a change in the zonal pressure gradient between the eastern and western Pacific, which is a main driving force of the low-level zonal wind. A change in low-level easterlies due to a change in the zonal pressure gradient has a great influence on the rising motion in the western Pacific, together with the low-level moisture content advected toward this area. This means that the new scheme induces the change of the WC. Furthermore, our results show that a change of the WC due to the new BL scheme highly depends on the convection scheme selected.

The results of this study demonstrate that the simulation of tropical precipitation in GCMs can be significantly improved by a change of BL processes. This implies that, in seasonal predictions as well as short-range forecasts, BL processes in GCMs are as important as precipitation processes. Moreover, these results suggest that balance between BL processes and convection processes should be considered essential to improving precipitation simulation.

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