High-resolution simulation of an extreme heavy rainfall event in Shanghai using WRF: Sensitivity to planetary boundary layer parameterization

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ABSTRACT

In this study, the extreme rainfall event on 25 May 2018 over Shanghai and its nearby area was simulated using the Weather Research and Forecasting model (WRF), with the focus on the effects of planetary boundary layer (PBL) physics using double nesting with large grid ratios (15:1 and 9:1). The sensitivity of precipitation forecast was examined through three PBL schemes, i.e., the Yonsei University Scheme (YSU), the Mellor-Yamada-Nakanishi Niino Level 2.5 (MYNN) and the Mellor-Yamada-Janjic (MYJ) scheme. The PBL effects on boundary layer structures, convective thermodynamic and large-scale forcings were investigated to explain the model differences in extreme rainfall distributions and hourly variations. The results indicated that in single coarser grids (15 km and 9 km), the extreme rainfall amount was largely underestimated with all three PBL schemes. In the inner 1-km grid, the underestimated intensity was improved; however, using the MYNN scheme for the 1-km grid domain with explicitly resolved convection and nested within the 9-km grid using the Kain-Fritsch cumulus scheme, significant advantages over the other PBL schemes are revealed in predicting the extreme rainfall distribution and the time of primary peak rainfall. MYNN, with the weakest vertical mixing, produced the shallowest and most humid inversion layer with the lowest lifting condensation level (LCL), but stronger wind fields and upward motions from the top of boundary layer to upper levels. These factors all facilitate the development of deep convection and moisture transport for intense precipitation, and result in its most realistic prediction of the primary rainfall peak.
Key words: PBL parameterization, extreme rainfall, high-resolution

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Article Highlights:

- For a 1-km grid nested in the outer 9-km grid, the MYNN scheme most realistically simulates extreme rainfall distribution and primary hourly peak.

- Diurnal rainfall simulation is influenced by the use of PBL schemes and the PBL sensitivities vary as model resolution is increased.

- The MYNN scheme with the weakest vertical mixing more facilitates the development of deep convection and intense precipitation.
1. Introduction

Accurate and quantitative forecast of heavy rainfall events in warm season over East China is a longstanding challenge for regional climate and/or weather prediction models. Difficulty arises not only from the complicated multi-scale interactions between convective activities and different weather systems (Sun, 2010; Fu, 2017), but also from the limitations of the model itself such as initial state, model resolutions and large uncertainties in representing the related precipitation physical processes (Qiao and Liang 2016; Srinivas et al. 2018). Among these processes, planetary boundary layer (PBL) is directly affected by the underlying surface and responds to its changes through vertical mixing and turbulent eddies that transport heat, momentum, and moisture to the atmosphere above. The vertical eddy transport in the PBL determines the profiles of low-level moisture, temperature, and winds (Beljaars and Viterbo 1998), which have an impact on convection and precipitation development by surface fluxes exchanges, thermodynamic instability and low-level forcings (Roebber et al., 2004; Hu et al., 2010; Coniglio et al., 2013; Holtslag et al., 2013; Clark et al., 2015; Cohen et al., 2015).

However, large uncertainties exist in representing the PBL processes in current mesoscale models, because different assumptions and various kinds of tuning parameters are required in PBL parameterization schemes to describe the vertical subgrid-scale (SGS) fluxes due to eddy transport (Wyngaard, 2004). To date, different kinds of PBL schemes have been proposed, but no consensus yet exists with respect to their applications in model predictions.
Previous studies have focused on the development of PBL schemes and their impacts on the simulations of boundary layer structures under various atmospheric conditions (Shin and Hong, 2011; Efstathiou et al., 2013; Dong et al., 2019). Generally, there are two categories for the PBL schemes in the WRF model based on their turbulence closure assumptions. The first category is the nonlocal closure, such as the Yonsei University (YSU, Hong et al., 2006) and the Asymmetric Convection Model 2 (ACM2) scheme (Pleim, 2007a; Pleim, 2007b). These conventional nonlocal closure PBL schemes determine the nonlocal SGS transport through a mass-flux term (Pleim, 2007a; Pleim, 2007b) or a gradient-adjustment gamma term (Hong et al., 2006) and adopt multiple vertical levels to simulate the effect of larger eddies in the convective boundary layers (Stensrud, 2007). They are typically characterized by overly vigorous vertical mixing and tend to simulate deep, dry and warm boundary layers (Coniglio et al., 2013; Burlingame et al., 2017). The second is local closure, such as the 1.5-order Bougeault-Lacarrere (Bougeault and Lacarrere, 1989), Mellor-Yamada-Janic (MYJ, Janjic and Zavisa, 1994) and the Mellor–Yamada Nakanishi Niino (MYNN) Level 2.5 and 3.0 (Mellor and Yamada, 1982; Nakanishi and Niino, 2006; Nakanishi and Niino, 2009; Olson et al., 2019). They are generally turbulent kinetic energy (TKE) closure schemes and adopt only adjacent vertical levels to estimate the turbulent fluxes through full range of atmospheric turbulent regimes. These local closure schemes such as MYNN and MYJ simulate more likely the shallow, cool and moist boundary layers due to relatively weak vertical mixing (Burlingame et al., 2017). However, controversy still exists regarding their application under convectively unstable conditions. For instance,
Shin and Hong (2011) presented that the nonlocal closure PBL schemes such as YSU and ACM2 are favourable in unstable conditions, while Burlingame et al. (2017) suggested that local closure PBL schemes such as MYJ and Quasi-normal Scale Elimination (QNSE; Sukoriansky et al., 2005) showed advantages over the nonlocal closure schemes in detecting convection initiation.

Many studies have likewise demonstrated the model sensitivity to the choice of PBL schemes in the simulation of heavy precipitation related to tropical cyclones (Dong et al., 2019) and severe convections in mid-latitude (Shin and Hong 2011; Efstathiou et al., 2013). For instance, Wang et al. (2013) suggested that the YSU scheme predicted the intensity, track and associated precipitation more realistically for a weak typhoon Muifa (2011) than the MRF scheme. But Liu et al. (2017) argued that the MYJ scheme generated better simulation of the rapid intensification process of Hurricane Katrina (2005) offshore before its landfall than that of the YSU scheme. Studies of Dong et al. (2019) showed that the YSU and MYNN2.5 scheme outperformed the MYJ and QNSE scheme in producing precipitation after landfall of Typhoon Fitow (2013). For the severe convection and heavy precipitation in the mid-latitudes, Efstathiou et al. (2013) suggested that the YSU scheme had enhanced vertical mixing and moisture transport, and more realistically simulated the strong convection and intense precipitation over Chalkidiki peninsula in northern Greece. However, Srinivas et al. (2018) examined an extreme heavy rainfall event in December over Indian and found that in 1-km cloud-resolving grid, the MYNN2.5 scheme showed advantages over the YSU and MYJ schemes in simulating the strong convection and associated heavy rainfall. Therefore,
large uncertainty still exists with the effect of PBL schemes on heavy rainfall predictions. In particular, as the model resolution is increased to explicitly represent the cumulus convection, the effect of PBL schemes becomes as important as microphysics parameterizations (Braun and Tao 2000; Li and Pu 2008). Therefore, this study attempts to explore the sensitivity of PBL schemes in high-resolution WRF simulations based on an extreme rainfall event over Shanghai and its nearby area during 25 May 2018. It involved with the interaction of multi-scale systems from the synoptic circulation, mesoscale low-level jet (LLJ) and low-level shear line to local convective activities, which provides a great test bed for examining the boundary layer processes and their interaction with large-scale environment and mesoscale convections.

The objectives of this study are two-fold: 1) to investigate the different sensitivities of PBL schemes in extreme rainfall simulations over the mesoscale (15-, 9-km) and high-resolution grid (1-km) with different means of treating convection; 2) to identify the model differences and to explore the underlying mechanisms that affect the interaction of boundary layer processes with the large-scale forcings and convection development.

The paper is organized as follows. Section 2 briefly describes the extreme rainfall event in Shanghai. Section 3 describes the WRF model experimental designs with physics setting and three PBL parameterization schemes (YSU, MYJ, MYNN2.5) used in this study. Section 4 compares the spatiotemporal characteristics of heavy rainfall simulations using different PBLs over varying grid spacings. Section 5 focuses on
expanding the physical mechanisms for model differences using three PBL schemes in the inner 1-km grid simulations. Section 6 concludes the results.

2 Observed synoptic conditions and mesoscale features

Figure 1 presents the double nesting grids for WRF experiments and the geographic distribution of daily rainfall observations from the Automated Weather Stations (AWS) network on 25 May 2018. There are 5207 sites distributed in the inner domain D2. The daily rainfall was accumulated from 00 BJT to 23 BJT. (Hereafter BJT denotes Beijing time which is 8-hours earlier than the universal time). Of the 5207 sites, 379 stations registered a daily rainfall accumulation exceeding 100 mm. Among them, daily rainfall for two stations of Shanghai, Chongming (173.8 mm) and Baoshan, (132.6 mm) even broke the extreme record of 50-year history in May shown in Fig. 1c. The rain band covered by these 379 AWS sites is the focus of our study and referred to below as the core rain band, which was roughly divided into two parts (east and west) by the longitudinal line 120° E (outlined by the red boxes in Fig. 1b).

To determine the dividing line, we fully examined the observed hourly variations of extreme rainfall over the stations within the core rain band. As is shown in Fig. 2, the coastal stations east of 120° E basically had three rainfall peaks, whereas the stations in the western inland had two peaks without the early-morning secondary one. The core rain band had three rainfall peaks with the primary peak at 12 BJT and two secondary peaks at 07 BJT and 18 BJT. The primary and late-afternoon peaks were contributed by both east and west parts, while the early morning peak was primarily attributable to
the east part. This implies that the extreme rainfall in these two sub regions was caused by different precipitation systems or mechanisms, which will be further analysed.

Figure 3 presents the evolution of 3-hourly AWS accumulated rainfall and wind field, moisture convergence as well as geopotential heights at 850 hPa from the National Centers for Environmental Prediction (NCEP) Final (FNL) Operational Global Analysis data at 1° grids at 08 BJT, 14 BJT, 20 BJT of 25 May and 02 BJT 26 May. Hereinafter, LLJ is defined by wind speeds greater than 12 m/s at 850 hPa. At 08 BJT, the southwesterly LLJ was strong and transported a great amount of water vapor to the Anhui-Shanghai region, producing intense moisture convergence and a heavy rainfall center with the presence of a low-level warm shear line along the east coast region. Warm shear line is a discontinuous line between southerly and easterly winds with the cyclonic shear in the lower troposphere (Yan and Yao 2019). This explains the early morning rainfall peak in the east rain band. Thereafter, LLJ gradually weakened due to its diurnal variation (Bonner, 1970), but the moisture convergence and uplifting motion associated with the low-level warm shear lines were still conducive to the development of mesoscale convection in the east coastal region. At the same time, a short-wave trough at 500 hPa was propagating eastward (Figure not shown), and the low-pressure vortex at 850 hPa ahead of the trough began to affect the west rain band. The strong convergence at the back of low vortex intensified the uplifting motion and mesoscale convection, resulting in the primary rainfall peak at 12 BJT for the west rain band. At 14 BJT, the LLJ and associated low-level moisture convergence decreased, and then until 20 BJT, the nocturnal LLJ intensity began to increase but its northern terminus
moved south, the moisture convergence and rainfall at the back of low vortex were intensified. Therefore, the diurnal variation of LLJ and associated moisture advection contributed more to the secondary peaks at early morning 07 BJT and late-afternoon 18 BJT, and the strong uplifting motion caused by the low-level warm shear line (east rain band) and the convergence at the back of the low 850 hPa vortex (west rain band) made significant contributions to the development and sustenance of mesoscale convection and heavy precipitation at 12 BJT.

3 Model configuration and experimental designs

The model used in this study is WRF version 3.9.1.1 (Skamarock et al., 2008). All domains used one-way nesting without feedback and same vertical discretization of 51 levels, including 11 layers within the boundary layer (below 1500 m). The model was driven by the NCEP FNL Operational Global Analysis Data (1° × 1° grid) as initial and boundary conditions for the outer domains. The 30-arc-second USGS terrain data and 21-categories, 2-m resolution MODIS land use/cover data were prescribed for static surface conditions. The model physics include the KF cumulus scheme for the outer grids, the Morrison microphysics (Morrison et al., 2009), the Noah Land Surface Model (Chen et al., 2004), and the RRTMG longwave and shortwave radiation scheme (Iacono et al., 2008).

Liang et al. (2019) put forward a pragmatic and effective approach in precipitation forecast to avoid the challenge in representing convection across the gray zone. The convective gray zone refers to the model grid spacing around 1-10 km, where
parameterized and resolved convective clouds could exist simultaneously. In the gray
zone, many widely-used cumulus parameterization assumptions become invalid and
difficult to realistically represent the convection (Yano et al., 2010). Liang et al. (2019)
found that double nesting simulations using the WRF model with large grid ratio (15:1
or 9:1) outperformed the traditional triple nesting with a middle 3- or 5- km grid in the
extreme rainfall forecast of Jiangsu province. Especially the outer grid using the Kain-
Fritsch scheme (Kain, 2014) to parameterize cumuli at 15 km with explicitly resolving
convection at 1 km produced the best forecast of hourly rainfall variation. Therefore,
we followed the same double nesting experiments using the KF cumulus scheme in the
outer coarse grids (15-, and 9- km) and explicitly resolving convection at the inner 1-
km grid. Our focus, however, is to examine the PBL effects on the extreme rainfall
prediction.

Three PBL schemes were used and the related diffusion equations and eddy
diffusivity coefficient computations are listed in Table 1. These schemes are all based
on the K-gradient transport theory, which determines the turbulent flux by multiplying
the eddy moment or heat diffusivity coefficient with the vertical gradient of grid-mean
variables. Their main differences are the closure assumptions used to define the eddy
diffusivity coefficient and the non-local effect for the energy exchange between model
layers. The YSU scheme is a first-order closure scheme which determines the K-profile
in the mixed layer by introducing the PBL height (eq. c) but depends on the mixing
length and Richardson number above the entrainment zone (eq. a,b). A counter gradient
transport term (eq. g) related to surface buoyancy flux is included for the nonlocal effect
especially for unstable boundary layers (Hong et al., 2006). As higher-order schemes, both MYNN (hereinafter refers to MYNN 2.5 in WRFv3.9.1.1) and MYJ (Janjic and Zavisa, 1994) are based on turbulent kinetic energy (TKE) closure to parameterize the eddy diffusivity but without considering the nonlocal effect in the unstable layer. They differ in the diagnostic equations for dimensionless stability function (S) and turbulent length scale (l). The MYJ (Janjic and Zavisa, 1994) scheme represents a non-singular implementation of the Mellor-Yamada level 2.5 turbulence closure model by adding limitation to the turbulent length scale. The MYNN scheme was considered to better represent the mixing in the convective boundary layer than MYJ because it considers the buoyancy effect on stability function and turbulent length scale (Srinivas et al. 2018). In WRFv3.9.1.1, the MYNN scheme includes several options to improve the coupling of the PBL scheme with radiation (icloud_bl=1) and microphysics (bl_mynn_cloudmix=1), and two options (bl_mynn_edmf=1, bl_mynn_mixlength=2) to use the cloud-specific and scale-aware mixing length following Ito et al. (2015). As suggested by the WRF physics documentation (Wang et al., 2017), the YSU scheme adopts the MM5 (Jimenez et al., 2012) surface layer scheme, and the MYJ and MYNN2.5 schemes both use the Monin-Obukhov (Monin and Obukhov, 1954) surface layer scheme.

Table 2 summarizes the configurations for two groups of double nesting simulations. The mesoscale grids (15 and 9 km) used KF parameterized convection, while the 1-km grid used fully explicit convection (EC). All experiments were initialized at 08 BJT on May 24, 2018, and integrated for 48 hours up to 08 BJT on May
244 26, 2018. For convenience, D1 and D2 are denoted as the outer and inner domains respectively. Nesting grid configurations are denoted directly by the grid spacing in km in sequential order. For example, 15-1 km denotes a double nesting configuration between D1 at 15 km and D2 at 1 km. The model results were bilinearly interpolated to the AWS stations to facilitate the comparison.

249 The evaluation methods include pattern correlations (COR, Barnston and Anthony, 1992), root mean square errors (RMSE, Barnston and Anthony, 1992), threat score (TS, Wilks, 2011), and bias score (BS, Wilks, 2011) for different precipitation intensity thresholds (i.e., 0.1, 10, 25, 50, 100 mm represent light, moderate, large, heavy, and extreme rain respectively). Larger COR and smaller RMSE indicate a better forecast of the spatial pattern of rainfall amount. The threat score (TS) is also called the critical success index and larger TS indicates higher predictive skills (perfect = 1) for the corresponding precipitation intensity. The bias score (Bias) denotes the frequency of rainfall forecasts compared with observations, and BS = 1 indicates an ideal prediction of the relative area size of corresponding precipitation intensity. The equations of COR, RMSE, TS, and BS are as follows:

\[
\text{COR}(\text{fore}, \text{obs}) = \frac{\sum (\text{fore}_i - \text{fore})(\text{obs}_i - \text{obs})}{\sqrt{\sum (\text{fore}_i - \text{fore})^2} \sqrt{\sum (\text{obs}_i - \text{obs})^2}}
\]

\[
\text{RMSE}(\text{fore}, \text{obs}) = \sqrt{\frac{1}{N} \sum_{i=1}^{N} (\text{fore}_i - \text{obs}_i)^2}
\]

\[
\text{TS} = \frac{a}{a + b + c}
\]

\[
\text{BS} = \frac{a + b}{a + c}
\]
where \( \text{fore} \) denotes the daily mean precipitation simulations, \( \text{obs} \) denotes the daily mean precipitation observations, and \( N \) denotes the number of observed stations. Four categories of hit (a), miss (b), false alarm (c) and correct non-rain forecast (d) are used to refer to the occurrence/non-occurrence of a rain event at each threshold, and is listed in Table 4. For example, both observed and predicted precipitation between 0.1 and 10 denote a hit for light rain.

4 Sensitivity of heavy rainfall simulations to PBL schemes

This section examines the sensitivity of heavy rainfall simulations to three PBL schemes in the mesoscale grids (15 and 9 km) and the high-resolution grids (1 km). The model evaluation will be focused on the spatial distribution of 24-hour accumulated precipitation and hourly variations of regional mean precipitation over the core rain band (24-hour accumulated precipitation > 100 mm).

4.1 24-hour accumulated precipitation

Figure 4 compares the 24-h accumulated precipitation from 00 BJT of 25 May to 23 BJT of 25 May using different grid spacing (15 km, 9 km, 1km) and PBL (YSU, MYJ, and MYNN) schemes. For the D1 15-km grid, all PBL schemes failed to capture the location of the observed rain band and the extreme rainfall amount was largely underestimated. As grid spacing was reduced to 9 km, these dry biases for all PBL schemes were reduced, but this still underestimated the west rain band and overestimated the coverage of the east rain band. The model differences between these three PBL schemes were not significant, implying that the extreme rainfall simulations
at coarser grids are not sensitive to the choice of PBL scheme. In the inner D2 1-km grid simulations, the model biases in the outer coarser grids were greatly reduced and the simulation of rainfall structure was refined, but the differences of using three PBL schemes became more significant. Specifically, using the YSU scheme in 1-km grids generally underestimated the east rain band nearby the Yangtze River Delta. The general pattern of the core rain band is better represented in the experiments using the MYJ and MYNN schemes, but the former produced a rain band that is deviated too far north.

Figure 5 further compares the statistics of TS, BS for different daily rainfall intensity thresholds, as well as COR and RMSE for all 5207 stations in the inner 1-km grid from all the double nesting simulations. In all the 1-km grid simulations, TS scores for light rain were generally higher than those for moderate-large rain. This is similar to the study of Liang et al. (2019), in which the WRF model using the KF cumulus parameterization scheme at 15- and 9-km grids produced higher TS scores for light-moderate rain but lower scores for all other categories in a monthly prediction over Jiangsu province. Thus, we speculate that this might be because the WRF model has limitations in representing physical (microphysics, PBL, etc.) and dynamic processes, leading to lower skills in predicting the convective cells than the general rainfall patterns. More importantly, three PBL schemes showed notable differences in predicting the intensity of extreme rainfall and the size of its relative area. For instance, in double nesting of 9-1 km simulations, the MYNN scheme had superiority in predicting the intensity and relative area size of extreme rainfall with the highest TS
and more realistic BS value; while the YSU scheme only performed better in predicting the intensity of light rain even performed worst in predicting the intensity and relative area size of extreme rainfall with the lowest TS and BS value. Using the MYJ scheme showed advantages in predicting the moderate rain with the highest TS and a more realistic BS (close to 1), but had medium predictive skills for the intensity and area size of extreme rainfall. Furthermore, double nesting of 9-1 km slightly performed better than the 15-1 km in predicting the entire precipitation distributions with systematically higher COR and lower RMSE.

Therefore, among all the double nesting simulations, using MYNN in the 1-km grid nested with 9-km grid produced the best predictive skill for the entire precipitation distribution, intensity, and size of the core rain band. This can be demonstrated by the largest COR for all the stations and TS as well as BS (0.68, 0.34, 0.93) for the extreme rainfall threshold compared to YSU (0.63, 0.12, 0.40) and MYJ (0.76, 0.24, 0.60).

### 4.2 Hourly rainfall variations

Figure 6 compares the hourly variations of observed and simulated rainfall averages over the two sub-regions (Fig. 1b) and the entire core rain band in the inner 1-km grid as well as the outer D1 coarser grids (15, and 9 km). In observations, the rainfall amount was gradually increasing from midnight 00 BJT and reached an early morning peak around 07 BJT. The primary rainfall peak occurred around 12 BJT and then weakened until 16 BJT, after that the rainfall increased again followed by a late-afternoon secondary peak at 18 BJT due to the intensification of nocturnal LLJ and the
mesoscale convective systems (Fig. 2). In the outer 15-km or 9-km grid (Fig. 6a, b), all three simulations systematically underestimated the primary and secondary rainfall peaks. The YSU scheme showed slight advantages over the MYNN and MYJ schemes in capturing the peak time as well as the intensity, while the MYJ scheme had the worst performance by producing a 1-hour earlier peak with the lowest intensity.

In the inner 1-km grid nested with the outer 15-km or 9-km grid (Fig. 6c, d), consistent advantages exist in the MYNN scheme over the other schemes in capturing the primary rainfall peak and intensity. Using the YSU scheme was capable of capturing the primary rainfall peak time, but with lower intensity than that of the MYNN scheme. However, using the MYJ scheme would still produce a 1-hour earlier rainfall peak with the lowest intensity in the double nesting of 15-1 km, which was improved in the double nesting of 9-1 km.

By analysing the hourly rainfall simulations over two sub-regions, we found that using the MYNN scheme in the double nesting of 9-1 km also well reproduced the primary rainfall peaks over both sub-regions (Fig. 6f, h). However, using the MYNN scheme in double nesting of 15-1 km would overestimate the peak intensity over the east rain band (Fig. 6e), also produce an earlier and weaker peak over the west rain band (Fig. 6g) compared to the observations. Furthermore, the largely underestimated rainfall peaks simulated by the MYJ scheme for both 15-1 km and 9-1 km were all mainly due to the dry biases over the west rain band. Therefore, we will focus on the double nesting of 9-1 km using the MYNN scheme, and to explore the possible causes for its superiority in simulating the primary rainfall peak intensity. The analyses mainly
include the PBL effects on the boundary layer structure, convective thermodynamic, and large-scale forcing variations.

5 Process understanding of PBL sensitivities

5.1 Simulations of boundary layer structure

Figure 7 compares the time-height sections of vertical eddy transport of equivalent potential temperature and water vapor mixing ratio. They are averaged over the core rain band from the double nesting of 9-1 km using three PBL schemes. The simulated vertical transport of heat and water vapor in the boundary layer all gradually became stronger from the early morning to noon, corresponding to the time when the primary rainfall peak occurred. The major difference exists with the vertical transport and mixing process at the top of the boundary layer between the nonlocal (YSU) and local PBL (MYNN & MYJ) schemes. The YSU scheme included the nonlocal effect with the counter gradient transport and produced the strongest vertical mixing below 2000 m than that of the local MYNN scheme before 12 BJT. The MYJ scheme showed intermediate vertical mixing between the YSU and MYNN schemes supported by Srinivas et al. (2018). This would cause differences in the vertical distribution of heat, water vapor from the boundary layer to the lower troposphere, and, thus, to affect the stability of the atmosphere as well as the development of convection and precipitation, which will be discussed as follows. It is noteworthy that there is high vertical equivalent potential temperature and moisture transports at 20 BJT in the experiment with the MYNN scheme, this may explain the high precipitation at 21 BJT shown in Fig. 6.
Figure 8 compares the hourly variation of vertical distributions of equivalent potential temperature, water vapor mixing ratio, horizontal wind speed, and vertical velocity averaged over the core rain band from the surface to the lower troposphere. The vertical profiles of these corresponding variables at 12 BJT in Fig. 9 are also combined to discuss the different effects of PBL schemes. For the thermal structures, three PBL schemes all produced an inversion layer near the surface and the depth was decreasing from the early morning till noon, but the local MYNN scheme produced the shallowest inversion layer at 12 BJT around 1000 m compared to the nonlocal YSU scheme (1200 m) or the local MYJ scheme (1300 m) (Fig. 9a). Meanwhile, they all produced a warm layer in the middle-upper boundary layer, resulting in an unstable layer above. For the water vapor distribution, three PBL schemes showed relatively little difference, but the water vapor was still more evenly distributed within the inversion layer (<1000 m) during the morning till noon in the nonlocal YSU scheme. The local MYJ and MYNN schemes both showed a more concentrated water vapor center in the bottom layer, but the MYJ scheme produced slightly higher water vapor mixing ratios from the top of the boundary layer to the middle troposphere at 12 BJT (Fig. 9b). For the wind structure, the local MYNN scheme produced stronger horizontal wind speed and upward motion than the nonlocal YSU scheme from the lower troposphere to the upper levels. The wind fields simulated by the MYJ scheme were strong at the lower troposphere, however, they rapidly reduced from the lower troposphere to upper levels (Fig. 9c, d). However, the nonlocal YSU scheme produced the lowest wind speed from the low-level troposphere (2500 m; 13.60 m/s) to upper
levels (6000 m; 17.42 m/s) due to its strongest vertical mixing at the top of the unstable boundary layer (Fig. 7a, d).

Therefore, the local MYNN scheme produced the shallowest and most humid inversion layer in the bottom, the least warm middle boundary layer, but stronger horizontal wind and upward motion from the top of the boundary layer to the upper levels. YSU produced a deeper inversion layer than MYNN in the bottom, relatively warm and humid middle boundary layer, and the weakest wind fields due to the strongest vertical mixing at the top of the boundary layer. The local MYJ scheme produced an inversion layer with the depth between the other two schemes, but the warmest middle boundary layer and the wind fields of MYJ at the boundary layer top were strong but quickly weakened above 5000 m. The MYJ scheme has slightly higher wind speeds in the BL as compared to others, but cautions should be taken here, considering the uncertainty of wind observation and computational errors. The following will focus on the effects of these different boundary layer structures on the development of convection.

5.2 Simulations of convective thermodynamic variations

Figure 10 compares the hourly variations of simulated convective available potential energy (CAPE) and convective inhibit energy (CIN) averaged over the core rain band from the double nesting of 9-1 km simulations with three PBL schemes. CAPE is defined as the accumulated buoyancy energy from the level of free convection to the equilibrium level, which represents the net kinetic energy that a rising air parcel
can gain from the environment and hence reflects the potential strength of convective systems and associated precipitation. CIN is defined as the accumulated negative buoyant energy from the air parcel starting position to the level of free convection, which represents the inhibit energy that the rising air parcel has to overcome to become free convection. Under stormy conditions, the PBL height could be defined by some alternative methods such as using a cloud base or lifting condensation level (LCL, Stull, 2011; Wisse and Vilá-Guerau de Arellano, 2004), and a higher LCL usually denotes a stronger vertical mixing. Table 3 also lists the simulated values of CAPE, CIN, and LCL as well as the horizontal wind speed and upward motions at the top of the boundary layer (around 1500 m), the low-level troposphere (around 2500 m), and the middle troposphere (around 6000 m) at 12 BJT for a quantitative comparison.

The nonlocal YSU scheme produced the lowest CAPE peak at around 08 BJT, which was two hours earlier than the primary rainfall peak. Although the YSU scheme produced comparable CIN with the MYNN scheme, it systematically had the highest LCL compared to the other two schemes. This implies that compared to the local MYNN scheme, the air parcel in the YSU scheme needs to be uplifted to a higher altitude to achieve condensation; nevertheless, the unstable energy in the upper boundary layer is insufficient to sustain the convection development.

However, the local MYJ scheme systematically produced the largest CAPE and CIN during the morning through the afternoon, indicating that the convection was strongly suppressed and the CAPE was slowly released. The LCL in the MYJ scheme was also higher than the MYNN scheme. The larger CIN, the slower release rate of
CAPE, and the higher LCL with the MYJ scheme all determine that the convection is more difficult to initiate and develop compared to using the MYNN scheme. The atmospheric instability, low-level moisture convergence, and vertical motion are the prerequisites for the development and sustenance of deep convection and mesoscale convective systems that often lead to heavy rainfall events (Srinivas et al. 2018). Therefore, in the following, we will discuss the PBL effects on the large-scale forcings, including the low-level moisture supply and the upward motions.

5.3 Simulations of large-scale forcing variations

Figure 11 compares the time-height sections of horizontal (a-c) and vertical fluxes (d-f) of water vapor mixing ratio and vertical velocity averaged over the core rain band from the surface to the troposphere in the double nesting of 9-1 km simulations with three PBL schemes. The water vapor transport simulated by these three PBL schemes mainly differed in their contributions to the primary rainfall peak during the noon. The local MYNN scheme systematically produced the strongest horizontal water vapor transport, as well as the upward motion from the top of the boundary layer to the middle troposphere during 10-12 BJT. This can also be identified from Table 3, in which the MYNN scheme showed slightly stronger wind speeds and vertical velocity at the boundary layer top and the middle troposphere at 12 BJT compared to YSU. Although the MYJ scheme produced larger wind speed and upward motion at 12 BJT at the top of the boundary layer, they weakened rapidly and produced less moisture transport to
the upper levels. Therefore, the MYJ scheme also predicted a lower intensity of primary rainfall peak compared to the MYNN scheme.

Figure 12 presents the longitude-height sections of moisture convergence and the zonal-vertical wind averaged over 31-33° N in the inner 1-km grid from the double nesting of 9-1 km simulations at 12 BJT and 14 BJT using three PBL schemes. Here we chose to demonstrate the moisture convergence at 14 BJT because the ECMWF reanalysis data was available. In ECMWF data, at 14 BJT, there was a strong low-level moisture convergence over the region of 120-122° E associated with a strong upward motion from the lower to upper troposphere, corresponding to the east rain band. Using MYNN produced stronger upward motion than other PBL schemes, which affected the amount of vertical moisture transport and produced the strongest low-level moisture convergence over this region. Weak low-level moisture convergence and upward motion in the lower troposphere were also shown in ECMWF data over the region of 117-118° E, corresponding to the relatively weak rainfall in the west rain band. However, using the MYNN scheme could only partly capture the upward motion and moisture convergence over this region but with underestimated intensity yet; the other two schemes could hardly capture the upward motion and the MYJ scheme even produced weak moisture divergence in the lower troposphere. This also explains why the experiments with the YSU and MYJ schemes have largely underestimated the extreme rainfall over the west rain band. Besides, at 12 BJT, MYNN also produced stronger low-level moisture convergence and upward motion in the lower troposphere over the region of 120-122° E and 117-118° E.
Therefore, the nonlocal YSU scheme with the strongest vertical mixing produced
the deepest inversion layer in the bottom with the highest LCL, the lowest CAPE, and
the weakest wind fields as well as low-level moisture transport. The MYJ scheme with
the intermediate vertical mixing produced larger CIN, slower release of CAPE, and
higher LCL compared to the MYNN scheme, which suppressed the convection
development. The wind fields in the MYJ scheme at the low-level troposphere was the
strongest but quickly weakened upwards, resulting in less moisture transport and
convergence in the lower troposphere, especially over the west rain band. However, the
MYNN scheme, with the weakest vertical mixing, produced the shallowest and most
humid inversion layer in the bottom with the lowest LCL, but stronger wind fields and
upward motions from the boundary layer top to upper levels. These all facilitate the
development of deep convection and moisture transport for intense precipitation.

6 Conclusions

This study conducted high-resolution predictions using the WRF model based on
an extreme event using double nesting with two different grid ratios (15:1 and 9:1). The
sensitivity of the precipitation forecast was examined using three PBL parameterization
schemes including the nonlocal first order YSU and local higher-order MYJ, MYNN
schemes. This extreme precipitation event occurred in Shanghai and its nearby area on
25 May 2018. It involved the interaction of multi-scale systems from the synoptic
circulation, mesoscale LLJ, and low-level shear line to local convective activities.
Various statistical measures were adopted to quantitatively evaluate the PBL effects on
daily rainfall distributions and the hourly variations of extreme rainfall. The impacts of
PBL schemes on the simulations of the boundary layer structure, convective
thermodynamic, and large-scale forcing variations were further analysed. The main
conclusions are as follows.

First, in the outer mesoscale grids (15 km and 9 km), all experiments failed to
capture the location of the observed rain band and the extreme rainfall amount was
largely underestimated. The model differences between these three PBL schemes were
not significant. In the inner 1-km grid simulations, the model biases in the outer coarser
grids were greatly reduced and rainfall structure simulation was refined. The east rain
band was generally underestimated with the YSU scheme while using both the MYJ
and MYNN scheme could capture the general pattern of core rain band, but the former
produced a rain band deviated further north. Among all the double nesting simulations,
using MYNN in 1-km grid with explicitly resolving convection nested with 9-km grid
with the Kain-Fritsch cumulus scheme, produced the best predictive skill for the entire
precipitation distribution and the intensity as well as the size of the core rain band.

Second, in the outer mesoscale grids (15km and 9km), all three simulations
systematically underestimated the primary and secondary rainfall peaks. In the inner 1-
km grid nested with the outer 15-km or 9-km grid, the MYNN scheme showed
consistent advantages over the other two PBL schemes in capturing the primary rainfall
peak and intensity. Using the MYNN scheme in double nesting of 15-1 km, however,
the peak intensity over the east rain band was overestimated, and an earlier, as well as
weaker peak over the west rain band, was produced. Nevertheless, the MYNN scheme
in the double nesting of 9-1 km had superiority over other 1-km simulations in predicting the primary rainfall peaks over both the west and east core rain bands.

Third, three PBL schemes differ in their vertical mixing processes, leading to different vertical profiles of temperature, water vapor, and wind field from the boundary layer to the lower troposphere, which affect the atmospheric stability and moisture transport for the development of convection and precipitation. Among these PBL schemes, for this extreme rainfall event, the local MYNN scheme showed the weakest vertical mixing and produced the shallowest as well as the most humid inversion layer in the bottom and the lowest LCL, which are conducive to the development of convection. At the same time, the MYNN scheme produced stronger horizontal wind and upward motion from the top of the boundary layer to the upper levels than the nonlocal YSU scheme, which facilitates the moisture transport and convergence in the lower troposphere for intense precipitation. The combined effects led to the strongest and most realistic rainfall peak in the MYNN scheme. However, the nonlocal YSU scheme with the strongest vertical mixing produced a deep inversion layer, the highest LCL and the lowest CAPE, and weaker wind fields, which tend to inhibit the convection development and moisture transport for heavy rainfall. This explains the largely underestimated peak of primary rainfall in the YSU scheme. On the other hand, The MYJ scheme with the intermediate vertical mixing produced larger CIN and higher LCL with a slower release of CAPE compared to the MYNN scheme, which also suppressed the convection development. Although the horizontal wind of the MYJ scheme was comparably strong with the MYNN scheme in the upper boundary layer,
it rapidly weakened in the lower troposphere, resulting in less moisture transport than the MYNN scheme. This especially led to the underestimated rainfall amount over the west rain band.

Although our study was based on one single day case over Shanghai and its adjacent region, the model performance of these three PBL schemes in this case was highly similar to those of previous high-resolution model studies such as Srinivas et al. (2018), but over different regions. On the other hand, the PBL sensitivities with the varying model resolution and convection treatment also provided a better understanding of the interaction of the boundary layer with the large-scale environment and the organization of mesoscale convective systems. Future work will conduct the high-resolution simulations over a longer time and under different types of climate regimes to verify the typical characteristics of different PBL schemes. Moreover, the degradation of 1-km simulations downscaled from the coarser grids (15 km and 9 km) in predicting the weak secondary rainfall peak requires to be further studied through improving the cumulus or microphysics representations.

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The ERA-Interim Global Atmospheric Reanalysis Data (0.125° × 0.125° grids) provided by European Centre for Medium-Range Weather Forecasts (ECMWF) was used to analyse the evolution of horizontal wind speed and can be found in https://apps.ecmwf.int/datasets/data/interim-full-daily/levtype=sfc/. The model simulations were conducted at the ECNU Multifunctional Platform for Innovation 001 facilities and data was deposited in ECNU public data server with IP 49.52.29.112. The views expressed are those of the authors and do not necessarily reflect those of the sponsoring agencies.
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Table 1. Description of diffusion equations and vertical diffusivity coefficient computation for three PBL parameterization schemes used in model experiments.

<table>
<thead>
<tr>
<th>PBL Schemes</th>
<th>YSU</th>
<th>MYNN</th>
<th>MYJ</th>
</tr>
</thead>
<tbody>
<tr>
<td>Reference</td>
<td>Hong et al., 2006</td>
<td>Mellor and Yamada, 1982</td>
<td>Janjic and Zavisa, 1994</td>
</tr>
<tr>
<td>Eddy Diffusivity Coefficient</td>
<td>( K_{\text{m,loc}} = l^2 f_{\text{m},c}(R_{\text{ig}}) \left( \frac{\partial U}{\partial z} \right) )</td>
<td>( K_{h,m} = l q S_{h,m} )</td>
<td>( K = Slq )</td>
</tr>
<tr>
<td></td>
<td></td>
<td>(d)</td>
<td>(e)</td>
</tr>
<tr>
<td>Rig</td>
<td>( \frac{g(\theta - \theta_s)(h - Z_{\text{mix}})}{\partial(\theta - \theta_s)(h - Z_{\text{mix}})^3} )</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>( K_m = kw_s z \left( 1 - \frac{z}{h} \right)^f )</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Diffusion Equation</td>
<td>( \frac{\partial C}{\partial t} = \frac{\partial}{\partial z} \left[ \frac{K_c}{\partial C} \left( \frac{\partial C}{\partial z} - \gamma_c \right) \right] + \frac{\partial}{\partial z} \left[ \frac{l}{\partial(\theta_s)} \right] + P_s + P_b )</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>(f)</td>
<td>(h)</td>
</tr>
<tr>
<td>Variable Description</td>
<td>( K_{\text{m,loc}} ) denotes the diffusivity coefficient used above the entrainment zone;</td>
<td>( S_{h,m} ) is a dimensionless stability item which is a function of Richardson number (Rig);</td>
<td>( S_{\text{is}} ) is a dimensionless stability item which is a function of Richardson number (Rig);</td>
</tr>
<tr>
<td></td>
<td>( K_m ) denotes the momentum eddy diffusivity in the mixed-layer;</td>
<td>( K_{h,m} ) denotes the heat, water vapor and momentum eddy diffusivity</td>
<td></td>
</tr>
<tr>
<td></td>
<td>( K_c ) denotes the eddy diffusivity;</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>( Z ) is the height from the surface and ( h ) is the height of the PBL;</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>( C ) denotes the prognostic variables for wind fields (u, v, w), water vapor (q) and potential temperature (( \theta ));</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>( \omega' c' ) denotes the flux for prognostic variables (u, v, q, ( \theta ));</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>( b ) is a constant of proportionality;</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>( \gamma_c ) is a local gradient correction item.</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Table 2. Two groups of model experiments using different configurations of grid nesting and PBL parameterization schemes as well as the convection treatments.

<table>
<thead>
<tr>
<th>Experiments groups</th>
<th>Outer domain [D1]: CUP&amp;PBL Grid cells</th>
<th>Inner domain [D2]: EC&amp;PBL Grid cells</th>
</tr>
</thead>
<tbody>
<tr>
<td>[1] 15-1 km</td>
<td>15 km KF YSU/MYJ/MYNN 407 x 297</td>
<td>1 km EC YSU/MYJ/MYNN 1156 x 1141</td>
</tr>
<tr>
<td>[2] 9-1 km</td>
<td>9 km KF YSU/MYJ/MYNN 677 x 494</td>
<td>1 km EC YSU/MYJ/MYNN 1153 x 1135</td>
</tr>
</tbody>
</table>
Table 3. Simulated values of CAPE, CIN and LCL as well as the horizontal wind speed, vertical velocity and water vapor mixing ratio at the top of boundary layer (around 1500 m), the low-level troposphere (around 2500 m) and the middle troposphere (around 6000 m) at 12 BJT.

<table>
<thead>
<tr>
<th>PBL</th>
<th>CAPE (J/kg)</th>
<th>CIN (J/kg)</th>
<th>LCL (m)</th>
<th>Height of inversion layer (m)</th>
<th>Vertical velocity (m/s)</th>
<th>Horizontal wind speed (m/s)</th>
<th>Water vapor mixing ratio (g/kg)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1500 2500 6000</td>
<td>1500 2500 6000</td>
<td>1500 2500 6000</td>
</tr>
<tr>
<td>YSU</td>
<td>216.93</td>
<td>11.04</td>
<td>1438</td>
<td>1200</td>
<td>0.01 0.10 0.34</td>
<td>13.60 17.42 13.80</td>
<td>10.90 4.70</td>
</tr>
<tr>
<td>MYJ</td>
<td>402.97</td>
<td>7.86</td>
<td>1411</td>
<td>1300</td>
<td>0.02 0.21 0.37</td>
<td>9.63 13.93 18.44</td>
<td>14.10 11.40 4.70</td>
</tr>
<tr>
<td>MYNN</td>
<td>236.6</td>
<td>11.08</td>
<td>1381</td>
<td>1000</td>
<td>0.03 0.17 0.43</td>
<td>9.39 13.94 18.9</td>
<td>13.40 10.80 4.90</td>
</tr>
</tbody>
</table>
Table 4. Four categories of the occurrence/non-occurrence for a rain event at each threshold.

<table>
<thead>
<tr>
<th>Rainfall Event</th>
<th>Forecast</th>
<th>Variable</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>yes</td>
<td>no</td>
</tr>
<tr>
<td>Observation</td>
<td>yes</td>
<td>a</td>
</tr>
<tr>
<td></td>
<td>no</td>
<td>b</td>
</tr>
<tr>
<td></td>
<td>a</td>
<td>c</td>
</tr>
<tr>
<td></td>
<td>b</td>
<td>d</td>
</tr>
<tr>
<td>a=hit</td>
<td>b=false alarm</td>
<td></td>
</tr>
<tr>
<td>c=miss</td>
<td>d=correct non-rain forecast</td>
<td></td>
</tr>
</tbody>
</table>
Figure 1. The WRF experiment domain configurations (a), geographic distributions of the observed 24-hour accumulated rainfall amounts on 25 May 2018 from the Automated Weather Stations (b), and the location of Shanghai (c).
Figure 2. Hourly variation of AWS observed precipitation on 25 May averaged over the core rain band (AVE 379) and averaged over the two typical regions in Fig. 1b: East region (East 227) and West region (West 152) (unit: mm/hour).
Figure 3. Geopotential heights (black solid contours, unit: gpm), wind field (black vectors), low-level jets (wind speed $\geq 12$ m s$^{-1}$, thick blue vectors) at 850hPa, and moisture convergence (red solid contours, unit: $10^{-7}$ g/(cm$^2$ hPa s)) at 850 hPa from the FNL data, 3-hourly AWS accumulated rainfall (color dots, unit: mm) at 08 BJT (a), 14 BJT (b), 20 BJT (c) 25 May and 02 BJT 26 (d) May, 2018.
Figure 4. Spatial distributions of 24-hour accumulated rainfall on 25 May 2018 simulated by WRF in D1 (15 km, 9 km) and D2 1km grids with three different PBL schemes (YSU, MYJ, and MYNN) (unit: mm).
Figure 5. Statistics of TS (a), BS (b), and COR/RMSE (c) for all 5207 stations with parameterized PBLs (YSU, MYJ and MYNN) for the inner D2 grids.
Figure 6. Hourly variations of observed (OBS) and simulated rainfall averages over the core rain band (a-d) and two sub-regions (East: (e-f), West: (g-h)) on 25 May 2018 by the WRF model using three different PBL schemes (unit: mm/hour).
Figure 7. Time-height cross-sections of vertical eddy transport of equivalent potential temperature (a-c, unit: K · s) and water vapor (d-f, unit: kg/(cm · hPa · s)) averaged over the core rain band on 25 May in the inner 1-km grids (9-1) simulations using YSU, MYJ and MYNN.
Figure 8. Time-height sections of the equivalent potential temperature (a-c, unit: K), water vapor mixing ratio (d-f, unit: g/kg), horizontal wind speed (g-i, unit: m/s) and vertical velocity (j-l, unit: m/s) averaged over the core rain band 1-km grids (9-1) using YSU, MYJ and MYNN.
Figure 9. Profiles of equivalent potential temperature (a, unit: K), water vapor mixing ratio (b, unit: g/kg), horizontal wind speed (c, unit: m/s) and vertical velocity (d, unit: m/s) from 100 to 6000 m over the core rain band at 12 BJT in the 1-km (9-1) grids using YSU, MYJ and MYNN.
Figure 10. Hourly variations of simulated convective available potential energy (CAPE, unit: J kg\(^{-1}\)) and convective inhibit energy (CIN, unit: J kg\(^{-1}\)) and hourly variations of lifting condensation level (LCL) (m, unit: m) averaged over the core rain band at 1-km (9-1) grids on 25 May using three PBL schemes.
Figure 11. Time-height sections of horizontal (a-c, unit: kg/(s·hPa·cm)), vertical water vapor mixing ratio flux (d-f, unit: kg/(s·hPa·cm)) and vertical velocity (black solid contours, unit: m/s) averaged over the core rain band on 25 May in the inner 1-km (9-1) grids using YSU, MYJ and MYNN.
Figure 12. Longitude-height sections of moisture convergence (shaded, unit: $10^{-7} \text{g/(cm}^2 \cdot \text{hPa} \cdot \text{s})$) and U-W wind (black vectors, unit: $U \times 100 \text{W m/s}$) averaged over 31-33°N on 25 May in the inner 1-km nested with 9-km grid using three PBL schemes.