The Daytime Evolution of the Atmospheric Boundary Layer and Convection over the Tibetan Plateau: Observations and Simulations

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Abstract

Based on field observations, theoretical analyses, and numerical simulations, this study investigates the structure and the evolution of the atmospheric boundary layer (ABL) and convection over the Tibetan Plateau during the dry season.

Both field observations and numerical simulations show that the convection over the plateau evolves from dry shallow convection in the morning to wet deep convection in the afternoon. The shallow convection is organized, and its major wavelength is controlled by mesoscale hills. The deep convection is not very regular. Both nonlinear scale interactions and latent heat release from convection may play significant roles in the development of the deep convection. However, the deep convection near mountains is related to an interactive process between mountain–valley circulations and rain evaporative cooling. The mountain–valley circulations in the afternoon can be either upslope or downslope.

The plateau ABL can extend to heights of almost 3 km above the ground surface, and is characterized by a well–mixed layer of potential temperature. The energy budget in the ABL indicates that the sensible heat is the dominant energy for sustaining the ABL growth, and radiations also play a significant role, but the rain evaporative cooling below the wet convection suppresses ABL development. The ABL evolution is strongly associated with the convective activities. The convection not only efficiently exchanges the quantities between the near–surface layer and the upper layer, but also enhances the air entrainment near the top of the ABL.

1. Introduction

The topography of the Tibetan Plateau comprises broad basins, low mountains and hills among several huge west–east running mountains. As the highest plateau in the world, Tibet has much thinner air and more intense solar radiation than lowland areas. It is well known that the plateau is a huge elevated heat source during the summer (Ye et al. 1979, Chapter 1). The heat and water vapor from the ground are directly transported to warm and moisten the middle troposphere layer over the plateau. It has been pointed out that the pla-
teau exerts profound thermal and orographic effects on atmospheric circulation patterns on all temporal and spatial scales (Gao et al. 1981). The onset of the Asian summer monsoon is an interactive process between the plateau-induced circulation and the circulation associated with the principal rainbelt migrating northward (Yanai et al. 1992).

The thermal and dynamic structures of the atmospheric boundary layer (ABL) are particularly important in the plateau meteorology. This layer has the most frequent synoptic-scale and mesoscale pressure systems of all regions in the same latitude belt and of similar elevation. Observational evidence indicates that many synoptic systems that cause the rainfall in the East China plain have their origin in the planetary boundary layer over the plateau and its surroundings (Gao et al. 1981; Tao and Ding 1981). Numerical experiments also show that the drought–flooding in the Yangtze-Huaihe River basin is sensitive to the thermal anomaly of the plateau surface (Xu et al. 2000). During the past thirty years, some characteristics of the plateau boundary layer have been revealed. This layer is dominated by a thermal low during the summer (Ye et al. 1979, Chapter 3). During the dry period preceding summer rains, this layer can extend to heights of 2–3 km, and is characterized by a nearly uniform profile of potential temperature, while humidity is not well mixed vertically (Luo and Yanai 1984; Yanai et al. 1992).

Intense convective activities are another striking feature of the plateau meteorology. According to Chen (Ye et al. 1979, Chapter 15) and Yuan (Ye et al. 1979, Chapter 7), there are more convective clouds and showers in Tibet than elsewhere in this latitude belt, and the density of mesoscale convective clouds is comparable to that over the tropical oceans during the summer. During the heavy rainfall in the Yangtze River area from June to July of 1998, “popcorn-like” convective clouds frequently occurred and developed over the middle and eastern parts of the plateau, and moved eastward during the flood period along the Yangtze River (Xu et al. 2002). The plateau ABL development is strongly associated with these convective activities in the boundary layer. The large-scale thermal low in the plateau boundary layer could be maintained by the convergence of the aggregated convective systems in this layer (Ye 1981). Yanai et al. (1992) speculated that dry convection might exist over the plateau during the dry season, and play a role in transporting heat from the hot surface to the upper layer.

In this study, we introduce the results of three recent experiments, i.e., the Tibetan Plateau Experiment of Atmospheric Sciences (TIPEX), the GEWEX (Global Energy and Water cycle Experiment) Asian Monsoon Experiment (GAME)/Tibet 1998, and the CEOP (Coordinated Enhanced Observing Period) Asia–Australia Monsoon Project (CAMP)/Tibet 2002. These projects reveal some of the features of the ABL and convection over the plateau. To understand the physical processes concerning these findings, a non-hydrostatic regional modeling system will be used to simulate the evolution of the convection and the ABL.

2. Field observations

The TIPEX was implemented by Chinese Scientists during the summer of 1998. Its objective was to detect the dynamic and thermal structures of the plateau boundary layer. The major findings were summarized in Zhou et al. (2000) and Xu et al. (2002). The GAME/Tibet and CAMP/Tibet were implemented by scientists from Japan, China, and Korea. Their overall goal was to investigate the water and energy cycles in the context of the Asian summer monsoon system. The latter two projects have an identical mesoscale experimental area (Lat. 30.5°–33°N, Lon. 91°–92.5°E), which is located in a valley 200–300 km wide, located between two huge mountain ranges running east-west: the Tanghla and the Nyainqen-tanghla Mountains, as shown in Fig. 1. In these experiments, some advanced instruments such as a Doppler sodar (Sound Detection And Ranging), a rainfall radar and two wind profilers were implemented for the first time in the plateau, in addition to routine observations using AWS (automated weather stations), SMTMS (soil moisture and temperature measuring systems), flux towers, radiosondes, and so on. The data obtained from these experiments form the basis of this study. This section briefly introduces the observed results on the convection and the boundary layer.
2.1 Convection

Figure 2 shows the evolution of vertical motion over the plateau detected by a Doppler sodar at the TIPEX Damxung site during the dry period. The vertical motion was measured every 15 seconds, but the small eddies were filtered out in the results presented in Fig. 2, using a 12-minute moving average. The vertical motion before 0830 LST shows weak turbulence, followed by thermal convection. The time scale of the vertical motion is in the range of 1.2–1.5 hr (Xu et al. 2002), and the maximum vertical motion was found to reach 1 m s\(^{-1}\) in the near-surface layer.

At the CAMP/Tibet Naqu site, the evolution of vertical motion on 22 August 2002, a monsoon-break day, was monitored by a wind profiler. The vertical motion was recorded every two minutes from the early morning to 1400 LST. Figure 3 shows the time–height section of the vertical motion after application of a 9-point moving average. Convection appeared from 0830 LST, and its height increased from 1 km at 0900 LST to 2 km at 1200 LST; the deeper convection started from 1300 LST, and extended up to a height of at least 3.5 km. The vertical motion of the convection was typically 1–2 m s\(^{-1}\). On the monsoon-break day, the sky conditions experienced a regular variation. During the morning, the sky was clear. Near noon, convective clouds appeared over the mountains, and then gradually covered the
valleys in the afternoon. The variation was similar to that observed during the dry season, as described by Zhou et al. (2000, p. 17). This variation suggests that the convection is usually dry in the morning, but wet in the afternoon. This suggestion is supported by a comparison between the convection level and the lifting condensation level (LCL). The LCL estimated from the observed surface parameters during this day increased from about 1 km at 0900 LST to 1.8 km at 1200 LST, and up to 2.1 km at 1500 LST. The LCL during the dry season is even higher than during this monsoon-break day. The estimated LCL during the dry season varies from 1.5 km at 0900 LST to 2.5 km at 1200 LST, and reaches 2.7 km at 1500 LST. It is clear that the convective cells during the morning (Fig. 3) have not or seldom have exceeded the LCL and thus they are usually dry, while the deeper cells during the afternoon have reached the LCL and thus they are wet.

In brief, the recent observations reveal that the plateau convection evolves from a shallower convection during the morning to a deeper convection by the afternoon. The convection is usually dry during the morning but becomes wet by the afternoon. The regular diurnal variation of atmospheric conditions suggests that the convective activities are mainly associated with local dynamical and thermal processes.

2.2 Boundary layer

In the plateau experiments, the radiosonde profiles of temperature and humidity are strongly affected by intensive convective activities, and thus it is difficult to interpret individual sounding profiles. Therefore, the composite temperature profile and water vapor profile are used to investigate the ABL evolution. Figure 4 shows the composite sounding profiles during 4–14 June 1998 (a dry period) for the GAME/Tibet Anduo site.

Figure 4 shows that the boundary layer is characterized by a well-mixed potential temperature profile and a non-uniform moisture profile. This phenomenon has also been observed in TIPEX and in earlier studies (e.g., Luo and Yanai 1984). Figure 4 also shows that the maximum ABL height is about 3 km AGL (above ground level). Gao (Ye et al. 1979, Chapter 8) estimated with the plateau quasi-steady thermal pressure system that the plateau ABL height was 2–3 km. In TIPEX, the height estimated by the Ekman spiral varies from between 1850 m and 2750 m. The ABL at the Anduo site is higher than at the TIPEX sites, because its elevation (4700 m ASL) is the highest among all of the radiosonde sites of the plateau projects. All of the findings show that the ABL over the plateau is much higher than over the lowlands, where the ABL height is usually no more than 1.5 km.
In addition, three heating periods in the ABL development could be roughly identified in Fig. 4a. During period I (sunrise ~ mid-morning), the near-surface layer becomes warm quickly, but the upper layer becomes slightly cool. The warming is due to the surface heating, while the cooling may be due to atmospheric long-wave emissions. The convective boundary layer extends to around 500 m AGL. During period II (mid-morning ~ mid-afternoon), not only is the ABL heated quickly, but also the ABL height extends rapidly. The height extends up to 2.0 km at 1200 LST and reaches about 3 km at 1500 LST. During period III (mid-afternoon ~ late afternoon), the height of the ABL becomes stable.

In short, the observations indicate that the evolution of the plateau ABL experiences three distinct periods and the ABL can extend up to near 3 km AGL during the afternoon. The ABL is characterized by a well-mixed layer of potential temperature and a non-uniform layer of specific humidity.

### 3. Theoretical analysis

To understand why the plateau ABL is much higher than the ABL over lowlands, in this section we evaluate the effect of surface elevation on the ABL height.

Based on a temperature profile in the early morning, it is easy to build up the relationship between the height of a mixed ABL and the energy required for its growth. We selected three morning sounding profiles to evaluate the elevation effect: a plateau sounding profile at the GAME/Tibet Anduo site (4700 m ASL), a lowland sounding profile at Kakoshima (4.2 m ASL), Japan, and an assumed profile with the plateau temperature profile and the lowland air density profile. The observed sensible heat fluxes at the Anduo site on a clear day (31 May 1998) are used to heat the three morning profiles. As a first approximation, the radiation, subsidence warming, and latent heating are ignored. Figure 5 shows the estimated variation in the ABL height during the daytime. This figure indicates that the maximum height of the estimated plateau ABL (3.4 km) can be as large as twice that of the lowland ABL (1.7 km) if the ABL growth is supported by the same amount of energy. In addition, the maximum ABL height for the assumed profile (2.2 km) is also larger than that for the lowland profile,

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**Fig. 4.** Temporal variations of (a) potential temperature and (b) specific humidity, averaged on 11 rain-free days (4–14 June 1998) at the GAME/Tibet Anduo site.

**Fig. 5.** The estimated temporal variation of the mixed layer when heated by observed surface heat fluxes at the GAME/Tibet Anduo site on 31 May 1998. The legends ‘Plateau’, ‘Lowland’, and ‘Assumed’ correspond to three temperature profiles (see details in the text).
indicating that the plateau atmospheric stratification is more favorable for ABL growth.

Therefore, the plateau ABL indeed has the potential to grow much higher than the ABL over lowlands, due to large sensible heat fluxes, low air density, as well as a result of the favorable atmospheric stratification over the plateau.

4. Numerical simulations

For the first time, the plateau field experiments have provided direct evidence of the evolution of convection over the plateau and revealed some of the characteristics of the ABL. However, the observations made are still too limited in number to clarify the spatial structure of convective cells, the mechanism of the ABL growth, the mountain–valley circulation, and so on. In this section, we have used a non-hydrostatic model to simulate the evolution of the convection and the boundary layer.

4.1 Model description

The Advanced Regional Prediction System (ARPS) (Xue et al. 2000; Xue et al. 2001) is used for simulations in this study. The ARPS is a non-hydrostatic model and has the capability of explicitly predicting convective clouds as well as larger-scale weather systems. The land surface scheme ISBA (interactions between soil, biosphere, and atmosphere) developed by Noilhan and Planton (1989) is used. This scheme is relatively simple, but it may over-predict the heat fluxes on bare soil surfaces and under-predict the ground temperature because it assumes an identical value for the thermal roughness length and the aerodynamic roughness length (Noilhan and Mahfouf 1996). In the plateau experimental area, the surface is almost entirely covered by bare soils during the dry season. With strong solar radiation during the day, the surface–air temperature difference can reach as high as 30 K. In this case, the thermal roughness length can be much smaller than the aerodynamic roughness length. Yang et al. (2002) improved the flux parameterization scheme, and incorporated it into the land surface scheme. The improved scheme is important because the surface heat flux is the most important energy source for the ABL development.

The ARPS provides several subgrid-scale (SGS) turbulence schemes, and we adopt the 1.5-order turbulent kinetic energy-based closure scheme after Deardorff (1980) and Moeng (1984). In this large-eddy simulation (LES) scheme, the eddy coefficients are proportional to the product of a length scale and a velocity scale:

\[ K_M = 0.1E^{0.5}l_h \]  
\[ K_M = 0.1E^{0.5}l_v \]  
\[ K_H = K_M \min[3, 1 + 2l_a/\Delta z] \]  
\[ K_H = K_M \min[3, 1 + 2l_a/\Delta z] \]

where \( K_M \) (\( \text{m}^2 \text{s}^{-1} \)) is the eddy mixing coefficient for momentum, and \( K_H \) (\( \text{m}^2 \text{s}^{-1} \)) is the eddy mixing coefficient for heat and other scalars. \( E \) (\( \text{m}^2 \text{s}^{-2} \)) is the SGS turbulent kinetic energy. The subscripts \( h \) and \( v \) denote the horizontal and vertical direction, respectively. \( l_h \) (m) is given by \((\Delta x \Delta y)^{0.5} \). For the unstable or neutral case, \( l_v = \Delta z \) (m), whereas for the stable case, \( l_v = \min(\Delta z, L_e) \) (m). Here, \( \Delta x, \Delta y \) and \( \Delta z \) (m) are the grid size in the three directions, \( L_e = 0.76E^{0.5}g^{0.5}T^{-0.5} \) (m), \( \theta \) (K) is the potential temperature, and \( g \) (m s\(^{-2} \)) is the gravitational acceleration.

In addition, the ARPS provides a radiative transfer scheme for calculating shortwave and longwave radiations absorbed by the atmosphere and the ground surface, so the radiative force can be evaluated in the ABL energy budget (Section 4.4).

4.2 Experiment design

In the plateau, the large-scale terrain variability is characterized by several 100–300 km wide valleys that extend in the west–east direction. Kuwagata et al. (2001) and Kurosaki et al. (2002) have indicated that valleys of this scale are important for water vapor transport and cloud generation over the plateau. In addition, westerly winds dominate over the plateau during the premonsoon season. Therefore, we carry out 2D simulations of the north–south cross-section of a broad valley between two bell-shaped mountains. The model valley is shown in Fig. 6. The top boundary of the model domain is flat and at 12.7 km ASL. To abate the contamination of lateral boundary conditions, a 50-km-wide absorbing layer (marked by ‘buffer’ in Fig. 6) is applied to both sides of the valley. We implement the periodic condition for the
lateral boundaries, a non-penetrative condition for the bottom boundary, and the zero-gradient condition to the top boundary. To explicitly solve the thermal-scale convection (~1 km), a horizontal resolution of 125 m is used. The vertical grid size stretches from 20 m near the surface up to 440 m at the top. The temporal step is 0.3 s for the integration of acoustically-active terms and 3 s for other terms. The influence of the Coriolis force is ignored in this instance.

The initialization for the soil variables and the surface parameters are based on observations and model calibrations. The soil temperatures are initialized with the mean values measured at seven soil temperature sites, and the initial soil wetness is calibrated as 0.3 m such that the land surface model can predict realistic heat fluxes for the bare soil surface. The aerodynamic roughness length was derived from the observations at various TIPLEX sites (Xu et al. 2002) and GAME/Tibet sites (Yang et al. 2003), and a value of 0.01 m was used in this study. The surface heat flux parameterized by Yang et al. (2002) is not very sensitive to the roughness length.

The atmospheric fields are initialized by the 0600 LST sounding profile averaged over the 1–7 June 1998 period at the Anduo site. The along-valley wind component, the potential temperature, and the specific humidity are assumed to be horizontally homogenous. In order to avoid initial conditions that are too smooth, a random perturbation of less than 1 K is added to the horizontally-homogeneous field of the potential temperature at each grid.

In this study, four cases are designed for investigating the effect of various conditions on the convection and the ABL. In all of the cases, the valley basin is 4700 m ASL (the elevation at the Anduo site), and the mountain ridges are 500 m higher. As shown in Table 1, the length scale of the valley is 150 km in Cases 1 and 2 but 100 km in Cases 3 and 4. In Case 1–Case 3, a number of 10 km wide hills are present in the valley and their heights vary randomly from hill to hill but are less than 100 m (Fig. 6). In addition, an ice microphysics package (Lin et al. 1983; Tao and Simpson 1989) is applied in the three cases for realistic simulations of the cloud processes over the plateau. Case 2 is similar to Case 1, but a 1 m s background cross-valley wind is used to examine its influence. Case 3 uses a narrower valley (100 km) for studying the influence of the valley scale. Case 4 is a dry run over a valley without mesoscale hills. It will be used in analyzing the role of mesoscale hills in organizing shallow convection as well.

**Fig. 6.** The idealized topography of the 2D simulation. L is the valley width, and ‘buffer’ is an absorbing layer.

<table>
<thead>
<tr>
<th>Condition</th>
<th>Initial cross-valley wind (m s⁻¹)</th>
<th>Valley scale (km)</th>
<th>Hills in the valley</th>
<th>Cloud microphysics</th>
</tr>
</thead>
<tbody>
<tr>
<td>Run</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Case 1</td>
<td>0</td>
<td>150</td>
<td>10-km-wide hills</td>
<td>On</td>
</tr>
<tr>
<td>Case 2</td>
<td>1</td>
<td>150</td>
<td>10-km-wide hills</td>
<td>On</td>
</tr>
<tr>
<td>Case 3</td>
<td>0</td>
<td>100</td>
<td>10-km-wide hills</td>
<td>On</td>
</tr>
<tr>
<td>Case 4</td>
<td>0</td>
<td>100</td>
<td>No hills</td>
<td>Off</td>
</tr>
</tbody>
</table>
as the interactions between the wet convection and mountain–valley circulations.

The simulations start from 0600 LST. The simulated instantaneous fields of wind and temperature at 5 minutes of simulation time are used to analyze the evolution of convection and the ABL in Sections 4.3 and 4.4.

4.3 Simulated convection evolution

All of the simulations show that the convection over the plateau valley evolves from shallow convection in the morning to deep convection in the afternoon, which is consistent with observations. To understand the evolution of the convection during the daytime, Figs. 7a

Fig. 7. Contours of simulated vertical velocity over the valley in Case 1. (a) 300 m AGL. Contour values: 0.5, 1.5, 2.5 m s⁻¹; (b) 3000 m AGL. Contour values: 1.2, ..., 8 m s⁻¹. The deep convective cells over the valley are marked with ‘A’, ‘B’, ‘C’. Cells near the mountains are marked with bold zigzagged lines.
and 7b show the temporal variations in the vertical velocity at 300 m and 3000 m AGL in Case 1, respectively. The vertical velocity at 3000 m AGL can be used to represent the wet deep convection, because the LCL in this season is less than 3000 m (see Section 2.1). The shallow convection has large vertical velocities at the lower level, and small vertical velocities at the upper level, but the deep convection would cause noticeable vertical velocities at both levels.

Figure 7a shows that a number of shallow convective cells develop in the valley during the mid-morning. Figure 8 shows that the shallow convection first appears at the top of a hill (Fig. 8a), and then propagates to the neighboring area (Fig. 8b). The convective cells gradually extend upward (Fig. 8c), move closer, merge together (Fig. 8d), and finally lead to a large cell over the top of the hill (Fig. 8e). In general, shallow convection in a stable boundary layer may deform the overtopping inversion layer and thus excite internal gravity waves, which in turn modulate the boundary layer convection (Clark et al. 1986; Balaji and Clark 1988). However, the hills serve as elevated heat sources, and can induce an upslope wind near the hills. This upslope wind may transport convective cells to the hills and lead to the large cells over the top of the hills. Therefore it is not surprising that the major wavelengths of the shallow convection are comparable to the horizontal scale of the hills. In the case without mesoscale hills (Case 4), the wavelengths of the shallow convection are much smaller (Fig. 9a). Therefore, the hills may play a significant role in controlling the wavelengths of convective cells.

The deep convection over the valley is not as organized as the shallow convection (Fig. 7b). Nevertheless, it is evident that the horizontal spacing between the major deep convection extends from about 10 km at 1300–1400 LST (marked with A) up to 20 km at 1500 LST (marked with B) and 30 km at 1700 LST (marked with C). In the early afternoon, the deep convection develops directly from the convective cells over the top of the hills, so it has a horizontal spacing close to the scale of the hills (10 km). Later, the spacing of deep convection favors a larger scale, implying distinct mechanisms responsible for the evolution of the deep convection. Etling and Brown (1992) reviewed previous studies and suggested two possibilities responsible for the increase in the horizontal spacing. One possibility is the nonlinear interaction of the small-scale convection. Theoretical investigations of Mourad and Brown (1990) have shown that the nonlinear scale interactions will lead to increasing wavelength with time, even in a non-stratified ABL. In the dry case (Case 4), the horizontal spacing between the convection indeed increases with time in the afternoon (Fig. 9a). This might support the idea that scale interactions play a role in increasing the spacing. The other possibility is that it results from the latent heat released from convection. In our simulations, the hori-

Fig. 8. Contours of vertical velocity near the valley center during the morning in Case 1. Contour interval: 0.5 m s\(^{-1}\), solid line: updraft, dash line: downdraft.
Horizontal spacing between deep convection in Case 1 (a wet case, Fig. 7b) is obviously larger than in Case 4 (a dry case, Fig. 9b), suggesting a relationship between the latent heat release from convection and the horizontal spacing. Early studies (Mapes 1993; Lin et al. 1998) suggest that the latent heating in the deep convection may radiate gravity waves, which modify the region around the convective clouds and initiate new convective cells. However, this mechanism is suggested for mesoscale convective systems and multicell storms, which are different from the simulated plateau convection. It is necessary in future studies to investigate how the increase of the horizontal spacing is realized through latent heat release from convection.

However, the convective activities over the mountain are related to an interactive process between the mountain–valley circulation (MVC) and the rain evaporative cooling. The upslope flow drives shallow convection moving...
from the valley to the ridge, and activates deep convection during the late morning. Afterwards, the deep convection releases precipitation, which then evaporates and cools the air in the boundary layer. During the dry season, the relative humidity is only 20–50% in the plateau ABL, so the cooling effect can make a strong difference in the temperature and pressure between the mountain boundary layer and its surroundings (Fig. 10a), and therefore a downslope flow starting from the ridge is induced after the occurrence of the precipitation over the mountains (Fig. 11a). The convergence between the downslope and the existent upslope triggers the deep convection propagating from the ridge to the valley. After the deep convection moves away from the ridge, the solar heating and air subsidence would warm the air over the ridge (Fig. 10b), and drive an upslope flow again (Fig. 11b). This initiates a propagation of the deep convection from the valley to the ridge. This propagation is indicated by the bold zigzagged lines in Fig. 7b. However, in the dry case (Case 4), the upslope flow continues for the whole of the afternoon and no deep convection propagates from the ridge to the valley (Fig. 9).

Therefore, the evaporative cooling exerts a great effect on the MVC, and the MVC in turn drives the propagation of the deep convection.

In the controlled experiments, Case 2 is initialized with a cross-valley wind. This background wind shifts the convective cells slightly northerly, but the coherent structure of the shallow convection is very similar to Case 1, which ignores the initial background wind (not shown). This is not surprising, because shallow convection depends mainly on the local thermal and dynamic conditions. The deep convection near the mountains also shows a similar feature, indicating that the interactions between the MVC and the evaporative cooling is a dominant mechanism responsible for the convective activities. However, Case 2 simulates more numerous and stronger deep convective cells over the valley than Case 1. Case 3 uses a narrower valley (100 km) than Case 1 (150 km), but the simulated convective structure in both cases is very similar in all aspects (not shown), suggesting that the topographic effect of the mountain is constrained to a limited range, and the 50-km wide absorbing layer in Fig. 6 is thus sufficient in the simulations.

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Fig. 10. Simulated potential temperature field near the mountain in Case 1. (a) at 1200 LST and (b) at 1400 LST. A 40-point (or 5 km) moving average has been applied.
4.4 Simulated ABL development

Figure 12 shows the simulated evolution of the horizontally averaged potential temperature over the central part of the valley (−25 km and 25 km) in Case 1. The observed ABL heights in Fig. 4 (500 m at 0900 LST, 2 km at 1200 LST, and near 3 km at 1500 LST) are well simulated. From the analysis of the energy budget in the boundary layer, it appears that the ABL evolution is affected by the sen-
sensible heat fluxes, radiative convergence, horizontal temperature advection, and latent heat released from convection/exhausted by rain evaporation. Figure 13 shows the simulated energy budget in the boundary layer in Case 1. The sensible heat fluxes are calculated by the land surface scheme. The local storage, radiative convergence (both longwave and shortwave), and horizontal temperature advection are integrated from the ground to 3 km above ground level. The latent heat term in the boundary layer cannot be calculated directly from the ARPS output, and thus it is estimated from other terms through the ABL energy budget. The calculation of the energy budget ignores an entrainment flux at the height of 3 km, because it is much smaller (see Fig. 14) than other terms. All of the terms are then horizontally averaged over the central part of the valley (from −25 km to 25 km). During the early morning (sunrise to 800 LST), the radiation, temperature advection, and latent heat are negligible, so the sensible heat fluxes from the ground are the major contributors to the heating of the ABL. From the mid-morning to mid-afternoon (900~1500 LST), the heat storage in the boundary layer is more than 200 Wm$^{-2}$, resulting in a rapid warming. The sensible heat fluxes may reach 300 Wm$^{-2}$, much larger than other energy sources. The radiation can reach near 100 Wm$^{-2}$ and thus plays a secondary role. From the mid-afternoon, the sensible heat fluxes and the radiation begin to decrease, while the evaporative cooling is enhanced rapidly. As a result, the net warming rate decreases quickly, and finally becomes negative at about 1700 LST when the ABL height reaches its maximum. Later, the ABL begins decaying. Therefore, the simulated energy budget could account for the observed three heating periods shown in Fig. 4.

To further understand the growth of the ABL height, it is necessary to analyze the processes responsible for the upward transport of the sensible heat. Figures 14a-c show the partition of the sensible heat transferred by local turbulent mixing and by nonlocal convection mixing. Before the onset of convection, the upward heat flux is mainly transported by turbulent diffusion (Fig. 14a). The turbulent diffusion is a local mixing and thus transfers heat upward gradually, so the air temperature near the ground increases very rapidly during this period, while the air temperature above changes little. After the onset of convection, the ABL height is affected mainly by convective activities. Although the turbulent mixing is still important near the surface, it is negligible above. The convection has a much larger spatial scale than the SGS turbulence, and the convective mixing is therefore the dominant mode for transporting heat and water vapor from the hot
surface to the upper boundary layer (Figs. 14b–c). Furthermore, the convection can enhance entrainment near the top of the ABL, which results in a downward heat transport at the upper ABL, and thus speeds up the development of the boundary layer. To further address the importance of the convective mixing, we run a 1D simulation with identical conditions as in the 2D cases. The potential temperature in the 1D simulation exhibits a deep superadiabatic profile (not shown) instead of a well-mixed one, which suggests that the local turbulence mixing cannot fulfill the ABL growth, and that the nonlocal convective mixing plays a crucial role.

Other 2D case studies show very similar characteristics of the ABL evolution as Case 1 (not shown). Case 2, which is initialized with a background cross-valley wind, has slightly lower temperatures in the late afternoon than the other cases. This case produces more active deep convection, which can result in more rain evaporative cooling in the ABL. In Case 4, the dry case, the ABL is slightly thicker and the temperature is slightly higher than other cases due to the lack of evaporative cooling in the ABL.

5. Summary and remarks

This study investigated the characteristics and the daytime evolution of the convection and the atmospheric boundary layer over the Tibetan Plateau during the dry season.

Observations show that the plateau convection evolves from dry shallow convection to wet deep convection during the daytime, and the evolution is reproduced in our simulations. The shallow convection may be organized, while the deep convection is not very regular. The shallow convective cells develop in the morning, and gradually merge together to become large cells. This upscaling process is usually related to the modulation of gravity waves in the stable boundary layer (Clark et al. 1986; Balaji and Clark 1988). However, mesoscale hills over the plateau may play a more important role in organizing the shallow convection and in controlling the major wavelength of the convection. The deep convection is generated over mountains from the late morning, and over valleys from the early afternoon. Simulations show that the deep convection tends to increase its horizontal spacing over time. The evolution of
deep convection over the valley may be related to both nonlinear scale interactions (Mourad and Brown 1990) and latent heat release from convection, but the convective activities over mountains are related to an interactive process between mountain–valley circulations and the rain evaporative cooling.

Observations show that the plateau ABL can extend up to near 3 km AGL, and is characterized by a well-mixed layer of the potential temperature. The high ABL of the plateau results from large sensible heat fluxes, low air density, as well as an atmospheric stratification favorable to ABL growth. Simulations successfully produce the ABL daytime evolution. The ABL evolution and the energy budget in the ABL suggest that there are three heating periods. During Period I (from sunrise to mid-morning), the ABL heating is constrained to the near-surface layer and is mainly supported by the sensible heat fluxes. During Period II (from mid-morning to mid-afternoon), the ABL is heated quickly and its height extends rapidly; the sensible heat is the major energy source for the ABL growth, and radiation plays a secondary role. During Period III (from mid-afternoon to late afternoon), the ABL develops slowly because the rain evaporative cooling in the boundary layer partially offsets the warming of the sensible heat and radiation. The development of the ABL is strongly related to the convective activities in the boundary layer. The convection not only efficiently mixes the quantities between the lower part and the upper part of the ABL, but also enhances the downward entrainment near the top of the ABL. Without the convective mixing, the ABL cannot extend up to 3 km AGL.

Because the plateau convective cells are much smaller than the model resolution in the general circulation models (GCMs) and large-scale circulation models, and the convective mixing dominates the flux transport from the near-surface layer to the upper boundary layer, it is necessary to parameterize its contribution to the atmosphere in these models, as addressed by Avisser and Chen (1993) and Pielke et al. (1993). Such a parameterization would contribute to the prediction of the plateau ABL height, and improve our understanding of the role of the plateau in the onset and variability of the Asian summer monsoon.

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