An investigation into the vertical structures of low-altitude atmosphere over the Central Taklimakan Desert in summer

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Abstract
In this study, 1-month continuous radiosonde observational data were applied to present the low-altitude vertical structures and their evolutions over the Central Taklimakan Desert (CTD). The primary focus is to highlight the vertical structures near the ground with the high-resolution (10 m in height and 6 hr in time) radiosonde data. One of the unique features evident in our results is an obvious diurnal transition in lower layers near the ground due to strong surface heating or cooling. Unlike a traditional vertical structure in the boundary layer observed over a nondesert surface, both superadiabatic and inversion layers are distinct during the day. More specifically, the superadiabatic layer is obvious in the daytime because of strong solar radiation over the desert, and the superadiabatic can reach up to 0.2 km. In contrast, an apparent inversion layer forms in the nighttime due to the surface cooling. It is found that the surface forcing mainly dominates the structures in the boundary layer. At last, the vertical structures from the observations are compared with those from the ERA-Interim and MERRA2 reanalysis data sets. The results indicate that both reanalysis products can provide similar vertical profile patterns and diurnal variations. However, the diurnal transitions of temperature and wind profiles over the CTD are underestimated severely by both reanalysis data. Besides, the reanalysis data sets completely miss the superadiabatic near the ground in the daytime.

KEYWORDS
radiosonde observation, superadiabatic and inversion, the Central Taklimakan Desert, vertical structure of the atmosphere

1 | INTRODUCTION
The Taklimakan Desert (TD) is the second-largest desert in the world, which plays a significant role in weather and climate in the mid-latitude region of the Northern Hemisphere, even the whole world. The TD covers over 337,000 km² with an average elevation of about 1.1 km, which is known as the Tarim Basin surrounded by the Tianshan Mountains and the Kunlun Mountains on the north and south, respectively. As a result, the TD is the main dust source region of East Asia. Besides, the TD is a typical extreme arid area with rare rainfall. Overall, the TD is highly susceptible to the weather and climate over the arid areas of northwestern China and downwind regions.
The Central Taklimakan Desert (CTD) is fully covered by shifting sand, expanding several hundred kilometers with the same land-use category (Sun and Liu, 2006). To date, there are still key knowledge gaps in our understanding of the atmospheric structure and its interaction with the pure shifting sand ground over the CTD due to little observation available. Wang et al. (2019) proposed that the peak value of the atmospheric boundary layer (ABL) in the CTD can reach up to 5.0 km in July. Various factors have effects on ABL, such as convection system (Wang et al., 2016), sensible heating flux (Xu et al., 2018a). Although much progress has been made in numerical models in the past decade, the atmospheric structure over the CTD was not well described by the numerical models at present (Xu et al., 2018b). Overall, lack of adequate observations at appropriate spatial and temporal scales might be responsible for our limited ability to understand the atmospheric thermal structure in pure desert environments.

Over the last 100 years, many observational and numerical experiments have been conducted to investigate the boundary layer (Kristovich et al., 2018). However, much attention has been paid to the boundary layer over urban areas, especially for air pollution. To the best of our knowledge, relatively little work has been performed to estimate the atmosphere vertical structures over the CTD because of sparse observation over this area. As a result, the vertical structure over the CTD was rarely reported in the previous studies. Here, we attempt to investigate the vertical structures over the CTD using 1-month continuous L-band radiosonde observations in July 2016. One of the advantages of the L-band radiosonde observation is high vertical and temporal resolution. L-band radiosonde observations have been extensively tested and used in atmospheric structures studies, particularly in the lower boundary layer (e.g., Guo et al., 2016; Jiang et al., 2017; Han et al., 2019; Hu et al., 2019). In the present study, our main objectives are to analyze the vertical thermal structures over the moving-sand area of the CTD and to quantify the diurnal variability of the desert thermal profiles.

The remainder of this paper is organized as follows. Section 2 gives an overview of the observational data sets. Section 3 provides detailed results from the observations. Vertical structures from the observations are compared with those from the ERA-Interim and MERRA2 reanalysis data in Section 4. Finally, conclusions and discussions are provided in Section 5.

2 | DATA AND METHOD

2.1 | Data sets

The L-band radiosonde data were obtained at the Tazhong site (located at 83.38°E, 39.02°N) with four times observations per day at 0000, 0600, 1200, and 1800 local standard time (LST, UTC + 6) in July 2016. The L-band radiosonde has been widely used in operation by the China Meteorological Administration (CMA) since 2010. One of the advantages of the L-band radiosonde data is high resolution in vertical, with a record each second as a sounding balloon ascent. In other words, the vertical resolution is ~5–8 m. Besides, the four-time daily observations can be used to detect the diurnal transitions of atmospheric vertical structures. To make all the observations on the same level above the ground, the observational data were linearly interpolated into vertical levels with resolutions of 10 m. In this study, composite profiles were produced based on the total data sets of 124 soundings in July 2016. Besides, net radiation at 10 m and air temperature at 2 m altitude above the ground, and soil temperature at 0 cm depth were used to represent the evolution of the surface forcings.

To date, there are several reanalysis products available over the world with various temporal and spatial resolutions. However, only a few of them provide model-level products. To make the reanalysis products comparable with the high vertical resolution of the observations, the European Center for Medium-Range Weather Forecasts (ECMWF) Reanalysis-Interim (ERA-Interim) and Modern-Era Retrospective analysis for Research and Applications version 2 (MERRA2) model level products were selected to compare with the observations. The ERA-Interim data were obtained from the ECMWF with 60 model levels up to 0.1 hPa in vertical (Dee et al., 2011). One of the features of the ERA-Interim is that the 4-dimensional data assimilation (4D-Var) method was utilized for the upper-air atmospheric state, and a set of parameters was updated for the majority of satellites. As for the new generation reanalysis data, the MERRA2 data, managed by the NASA/GSFC Global Modeling and Assimilation Office, provide 72 model levels with the model top of 0.01 hPa (Gelaro et al., 2017). Much improvement was made in the MERRA2 by taking some significant steps toward the Global Modeling and Assimilation Office’s targets. Besides, space-based observations of aerosols were taken into account in the MERRA2 data.

2.2 | Method

The atmospheric lapse rate (LR) is defined as a change of temperature with a change of altitude, which can be calculated according to vertical temperature profiles from the radiosonde observations. The lapse rate can be expressed as
lapse rate \([^\circ C (100m)^{-1}]\) = \(\frac{dT}{dz} = \frac{T_{k+1} - T_k}{Z_{k+1} - Z_k} \times 100\)

(1)

Here, \(T\) is temperature, \(Z\) is geopotential height in meters, and \(k\) is the vertical level index. Multiplying by 100 is to convert the unit into \(^\circ C (100 m)^{-1}\). Note that using all observed levels caused obvious noise in the lapse rate. Therefore, the observed lapse rates were obtained from two levels with (or near) 100 m apart. As for the model-level reanalysis products, lapse rates were calculated from two adjacent levels in vertical. It should be noted that the lowest level of the observation is 10 m above the ground, and those of the ERA-Interim and MERRA2 products are 10 and 16 m, respectively. Therefore, the lapse rate close to the ground is not included at present. The dry adiabatic lapse rate is defined by a lapse rate of 0.98 \(^\circ C (100 m)^{-1}\). When the air temperature dropped at a rate larger (less) than the dry adiabatic lapse rate, superbadiabatic (subadiabatic) occurred. In contrast, a temperature inversion is where air temperature increasing with height.

3 | RESULTS

3.1 | Vertical properties

Figure 1 shows the averaged profiles of potential temperature and horizontal wind. It is obvious that there is a typical transition of the potential temperature below 4 km, and it appears a little difference at the upper levels. In the evening hours (0000 and 0600 LST) before sunrise, the soil temperature is less than that of the atmosphere (Figure 2e), and thus the cooling of the surface results in an inversion near the ground. After sunrise, the inversion decays gradually with the enhancement of the solar radiation, and a superadiabatic layer is obvious near the surface at 1,200 and 1,800 LST. Generally speaking, the ground radiative heating and cooling play a crucial role in the diurnal transition of the atmosphere vertical structure—from inversion to sub-adiabatic, adiabatic, and superadiabatic states. As for the wind profile, there is an obvious wind shear from the surface to the upper levels. More specifically, the wind direction changes continuously with height due to the vertical variation of horizontal wind due to surface roughness and shear forces predominantly in the boundary layer/Ekman layer. Figure 1b shows strong synoptic westerly jets (11–14 m\(s^{-1}\)) at 6.0 km which strengthen during the evening or late evening hours whereas the surface winds are more or less moderate (2–5 m\(s^{-1}\)) northeasterly winds. The statistical results show that 0–6 km bulk wind shear increases from roughly 3.6 to 28.9 m\(s^{-1}\), with an overall average of 12.4 m\(s^{-1}\). The level at which the surface winds change from northeasterly to westerly winds can be recognized as the transition layer below which boundary layer can be noticed at this desert location. The wind profiles show that the lowest extent of transition is about 1.65 km in the early morning at 0600 LST and the highest extent is about 3.5 km between 1,200 and 1,800 LST, subsequently it gradually descends to the lower regions in the night.

The transition of the thermal near the ground is illustrated in Figure 2. Before sunrise, there is an obvious inversion layer up to 200 m. As the day progresses, the inversion decays gradually, and a sub-adiabatic or superadiabatic layer occurs and persists up to sunset. The transition, from stable to convective and back to stable conditions, is consistent with the diurnal variation of solar radiation. Near the ground, the lowest temperature occurs at 0600 LST with the mean value of 23 \(^\circ C\), and the temperature reaches its peak value of 36 \(^\circ C\) at 1,800 LST. It should be noted that the standard deviations of temperature show a little variation at different times and levels,
indicating that the observation shows close variability patterns at all levels.

It is found that the transition in lower levels is dominated directly by the surface forcings in terms of strong heating or cooling (Figure 2e,f). The evolution of the surface forcings can be approximately represented by the daily variation of 10 m net radiation above the ground (Figure 2e). After sunrise at near 0700 LST, the soil temperature becomes warmer than the atmosphere since the soil warming is much faster than the air above it. Consequently, the superadiabatic layer results from strong surface heating because the soil temperature is much higher than that of the air above it during the daytime (Figure 2f). In contrast, an inversion forms near the ground owing to the surface cooling because the soil temperature is lower than that of the air during the nighttime (Figure 2e). The diurnal variation of the strong wind shear is also consistent with the change of the surface forcings. At 1,200 LST, the boundary layer is lifted because of the gradual enhancement of the surface heating (Figure 2f). Yuan et al. (2019) proposed that dust particles were able to be lifted to high altitude levels up to 8 km due to the enhanced sensible heating in summer, and the strong vertical wind shear has greatly influenced on horizontal dispersion (Walcek, 2002).

**FIGURE 2** (a–d) Transition of the average temperature at different hours, and the bars represent the standard deviation of each profile. (e) Indicates net radiation at 10 m altitude above the ground. (f) Presents temporal evolutions of air temperature at 2 m (T2m) altitude above the ground and soil temperature at 0 cm soil depth (TS0).
3.2 Vertical lapse rates

Profiles of lapse rates in terms of per 100 m are presented (Figure 3). One can see that superadiabatic is obvious below 0.2 km in the daytime (1,200 and 1,800 LST), and inversion is apparent during the nighttime (0000 and 0600 LST). During the daytime, dry adiabatic occurs from 0.2 to 0.6 km, and the lapse rate is close to dry adiabatic up to 2.4 km. In the upper levels, the lapse rate approaches to wet adiabatic gradually. Given the daily variation of the net radiation (Figure 2e), we can conclude that surface heating is responsible for the superadiabatic temperature profile in the daytime, and the inversion results from surface cooling in the nighttime.

The superadiabatic layer near the surface level has an average lapse rate of \(-3.11^\circ\text{C} (100 \text{ m})^{-1}\) at 0.1 km at 1,200 LST, with the largest value of \(-4.86^\circ\text{C} (100 \text{ m})^{-1}\). As for the inversion in the nighttime, the mean lapse rate is \(2.85^\circ\text{C} (100 \text{ m})^{-1}\) at 0600 LST, with the largest lapse rate of \(8.48^\circ\text{C} (100 \text{ m})^{-1}\). Comparatively speaking, the inversion at 0600 LST is much stronger than that at 0000 LST. The inversion is connected to the surface radiative cooling in the nighttime, and thus the strongest inversion occurs before sunrise near 0600 LST in the morning. However, there is a small difference in the superadiabatic layer between 1,200 and 1,800 LST. The difference of the superadiabatic may be explained by the strongest net radiation occurring at 1,200 LST.

To date, there have been many studies about the vertical atmosphere structure in different regions over the world. To the best of our knowledge, however, relatively little attention has been paid to the atmosphere vertical structures over the CTD. In this study, vertical atmosphere structures and their evolutions over the CTD were presented. It is found that the superadiabatic and inversion in lower layers over the CTD are more obvious, comparing with those over the Sahara desert (e.g., Messager et al., 2010; Garcia-Carreras et al., 2015; Ryder et al., 2015). The most likely reason for this is that the high vertical resolution radiosonde data were used in this study. Another possible reason is that the vertical structure atmosphere over the Sahara desert is turned by the West Africa monsoon (Skonieczny et al., 2019), while the Asian monsoon has little influence on the vertical atmosphere structure over the CTD because of the orographic blocking of the Qinghai-Tibet Plateau. Superadiabatic and inversion in the CTD is much stronger than those observed in the nondesert conditions of urban (Day et al., 2010; Melecio-Vázquez et al., 2018), flat terrain (Angevine et al., 2001), Basin area (Feng et al., 2020), inland Indochina Peninsula (Ogino et al., 2010), Glacier (Van Den Broeke, 1997), and mountains (Das et al., 2021).

4 COMPARED WITH REANALYSIS DATA

Compared with the observation, both the ERA-Interim and MERRA2 show similar vertical distribution patterns of the potential temperature and wind over the CTD. Overall, both ERA-Interim and MERRA2 can reproduce the wind shear layer. Especially, the ERA-Interim shows very close vertical structures of PT to the observation. Comparatively speaking, the ERA-Interim performs much better than the MERRA2. The MERRA2 reanalysis products have less potential temperature diurnal variation at lower levels, especially near the ground, compared with that of the observations (Figure 1a).

As for the wind profiles, both the ERA-Interim and MERRA2 reanalysis products show the ability to generate atmospheric vertical profile patterns. However, the diurnal variations are seriously underestimated over the CTD. More specifically, both ERA-Interim and MERRA2 products almost present constant heights of the wind shear layer (Figure 4a,b) near 2.7 and 2.9 km, respectively, while there is an obvious diurnal variation from the observation (Figure 1b). As has been mentioned above, the height of the maximum wind shear layer shows an obvious diurnal variation from 2.2 km at 0000 LST to 3.2 km at 1,200 LST, which corresponds to the PBL height.

Figure 5 shows the diurnal variations of the vertical thermal structures below 0.6 km from the ERA-Interim and MERRA2. Generally speaking, both reanalysis
products appear similar diurnal transitions as that of the observations. There is an inversion layer at 0000 and 0600 LST (Figure 5a,e), and the inversion disappears gradually with the enhancement of surface heating. Then a dry adiabatic layer occurs near the ground and lasts until sunset. Although the transition, from stable to...
convective and back to stable condition, agrees well with the observation, there are still large gaps between the reanalysis products and the observation. The daily temperature range near the ground is not as great as the observation. The ERA-Interim and MERRA2 have daily temperature ranges of 10.5 (from 25.5 to 36°C) and 8 (from 25 to 33°C), respectively, which is much less than that of observation with the values of 13 (from 23 to 36°C). Besides, the ERA-Interim cannot reproduce the minimum temperature as low as the observation at 0600 LST, while the MERRA2 misses both the minimum temperature and maximum temperatures.

Lapse rate profiles from the ERA-interim and MERRA2 reanalysis data are illustrated in Figure 6. In general, both reanalysis products are unable to capture the observed superadiabatic over the CTD near the ground in the daytime. One can see that most of the lapse rates are located near the dry adiabatic line. The superadiabatic layer is barely visible at all near the surface, even no superadiabatic layer. On the contrary, the ERA-interim has a larger lapse rate than observation from 0.2 to 2.3 km, showing as superadiabatic. As for the inversion, the ERA-interim shows a similar inversion pattern to the observations, which is better than that of the MERRA2.

One of the main possible factors is that the underlying surface of the desert and its interactions between atmosphere and land were not well described in both reanalysis systems, especially in the land models and/or planetary boundary layer schemes. Besides, low horizontal resolutions of the global models cannot resolve all the boundary layer structures under the pure desert region where there exist obvious diurnal variations in temperature and turbulence in the lower atmosphere (Wei et al., 2019). Additionally, convective parameterization used in the low-resolution global model may prevent the superadiabatic layer from developing. Moreover, little available observation assimilated in both ERA-Interim and MERRA2 systems has a significant effect on the quality of the reanalysis products. Apart from the above factors, the low model vertical resolution is another contribution to the poor representation and underestimation of super adiabatic layers in these two data sets. Usually, physical processes in most global models are vertically one-dimensional models, and thus the performance of the physical processes is significantly influenced by the model vertical resolution.

5 CONCLUDING REMARKS AND DISCUSSION

Desert atmospheric vertical structures, especially in the lower layers near the ground, form as a direct response to the surface forcing. In this study, we investigate the thermal vertical structures over the CTD. It is found that local forcing mainly dominates the structure of the lower layers near the ground, and the transition of profile structures in the boundary layer is consistent with the diurnal variation of solar radiation (surface heating and cooling).

One of the unique features evident in our results is the interaction between strong heating and cooling near the ground, leading to an obvious diurnal transition in lower layers. Unlike a traditional boundary layer observed over a nondesert surface, the superadiabatic layer is obvious because of strong solar radiation over the desert in the daytime, and the superadiabatic reaches up to 0.2 km due to little water vapor in the atmosphere. Besides, an obvious inversion layer can be found near the ground in the nighttime, which results from the surface cooling.

Compared with the observations, the ERA-interim and MERRA2 reanalysis data capture the main atmosphere profile patterns over the CTD. However, the diurnal variations of temperature and wind profiles are underestimated severely. Besides, the superadiabatic near the ground in the daytime are completely missed. It seems that the air temperature in the reanalysis system cannot increase (decrease) rapidly when surface heating ascends(descends) gradually. This suggests that there are

Figure 6 Same as Figure 3 but from (a) ERA-interim and (b) MERRA2 reanalysis data.
some deficiencies in land surface and planetary boundary layer (PBL) parameterization schemes with much attention to the moving-sand ground, dust, and other factors in the future. Additionally, Yin et al. (2015) proposed that the ERA-Interim products underestimate cloud fraction significantly over the TD, which might have some effects on the diurnal variation of temperature.

Although we present the main thermal characteristics of the desert environment here, additional work is required to understand the detailed turbulent properties within the desert boundary layer. It should be noted that the results from 1-month data might not be enough to represent the whole desert environment, and a long enough period to investigate the seasonal and intra-seasonal variability of the reported phenomena in the future. Besides, it also should be pointed out that the comparison results with ERA and MERRA only indicate the performance of the analysis data in July 2016. Owing to the limitation of observation, large-eddy simulations (LES) may be used to investigate the atmospheric vertical structures in the CTD since LES can provide some realistic representation of turbulence internal boundary layers.

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