Characteristics of the Summer Atmospheric Structure over the Bayinbuluke Basin in the Tianshan Mountains

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Abstract

Using data from the Bayinbuluk National Basic Meteorological Station in summer during 2014–2024, tethered airship observations and rocket radiosonde data from August 2014, and concurrent ERA5 reanalysis data, this study analyzed the characteristics of the summer atmospheric structure in the region and evaluated the applicability of ERA5 reanalysis data in alpine mountainous areas. The results show that: (1) The mean air temperature in the Tianshan Bayinbuluk Basin is 11.14℃, and easterly winds are the dominant near-surface wind direction in summer. Strong radiation inversions and low specific humidity are prominent features of the low-level atmosphere in this region. During the daytime, air temperature decreases with height, and the specific humidity in the near-surface layer, 5~6g·kg-1, is higher than that in the upper layer, 3~4g·kg-1. At night, near-surface wind speed fluctuates markedly, while during the daytime the wind speed at 200~800 m is approximately vertically uniform; a strong wind belt with wind speeds greater than 30m·s-1 exists in the middle and upper layers. (2) The atmospheric boundary layer height exhibits a pronounced diurnal variation. Under stable stratification at night, the boundary layer height determined by the potential temperature gradient method (mean 274.0~356m) is significantly higher than that determined by the wind profile method (mean 164.9~166.3 m), whereas the two tend to be consistent under convective mixing. The stratification in the Tianshan Bayinbuluk Basin is dominated by stable conditions, and stability increases with height. (3) ERA5 reanalysis data show high correlations with observed temperature and humidity data in the free atmosphere, with correlation coefficients of 0.99~1.00; however, biases exist in the simulation of near-surface inversion processes, with a temperature correlation coefficient of 0.67 at 02:00 and a specific humidity correlation coefficient of 0.39 at 14:00. This study deepens the understanding of boundary-layer physical processes in alpine mountainous areas and provides a theoretical basis for improving the physical parameterization schemes of numerical weather prediction and climate models in the Tianshan Bayinbuluk Basin.

Full Text

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Characteristics of the Summer Atmospheric Structure over the Bayinbuluke Basin in the Tianshan Mountains

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Abstract: Using summer data from the Bayinbuluke National Basic Meteorological Station for 2014–2024, tethered-airship sounding and rocket-sounding data from August 2014, and concurrent ERA5 reanalysis data, this study analyzes the characteristics of the summer atmospheric structure in the region and evaluates the applicability of ERA5 reanalysis data in high-cold mountainous areas. The results show that: (1) The mean air temperature in the Bayinbuluke Basin of the Tianshan Mountains is 11.14 °C, and easterly winds are the dominant near-surface wind direction in summer. Strong radiative inversions and low specific humidity are prominent features of the lower atmosphere in this region. During the daytime, air temperature decreases with height; the specific humidity in the near-surface layer, 5–6 $g \cdot kg^{-1}$, is higher than that in the upper layers, 3–4 $g \cdot kg^{-1}$. At night, near-surface wind speed fluctuates markedly; during the day, wind speed at 200–800 m is nearly vertically uniform, and a strong-wind belt with wind speeds greater than 30 $m \cdot s^{-1}$ exists in the middle and upper layers. (2) The atmospheric boundary-layer height exhibits pronounced diurnal variation. Under stable stratification at night, the boundary-layer height determined by the potential-temperature-gradient method, averaging 274–356 m, is significantly higher than that determined by the wind-profile method, averaging 164.9–166.3 m; under convective mixing, the two tend to converge. The stratification of the Bayinbuluke Basin in the Tianshan Mountains is dominated by stable types, and stability increases with height. (3) ERA5 reanalysis data are highly correlated with observed temperature and humidity data in the free atmosphere, with correlation coefficients of 0.99–1.00; however, deviations exist in its simulation of the near-surface inversion process, with a temperature correlation coefficient of 0.67 at 02:00 and a specific-humidity correlation coefficient of 0.39 at 14:00. This study deepens understanding of boundary-layer physical processes in high-cold mountainous areas and provides a theoretical basis for improving the physical parameterization schemes of numerical weather prediction and climate models for the Bayinbuluke Basin in the Tianshan Mountains.

Keywords: Bayinbuluke; atmospheric structure; atmospheric boundary-layer height; stratification stability; tethered-airship sounding; rocket sounding

The atmospheric boundary layer (ABL) is located at the bottom of the troposphere and is directly influenced by the surface. It is an important region for the exchange of matter, energy, and water vapor between the land surface and the free atmosphere. The intensity of vertical turbulent exchange of heat, water vapor, and momentum in this layer is far greater than that in the upper troposphere[1]. Characteristics of the atmospheric boundary layer generally include atmospheric temperature and humidity profiles, atmospheric boundary-layer height, turbulent characteristic quantities, and heat and water-vapor budgets. As one of the most critical parts of the atmosphere, the atmospheric boundary layer is closely related to human activities, and the vertical structure and variations of its meteorological elements have long been important research topics in atmospheric physics, atmospheric environment, and atmospheric chemistry[2].

In recent years, research on the structure of the atmospheric boundary layer under complex terrain conditions has become a focus of boundary-layer meteorology[3]. As an important geographical unit in the middle section of the Tianshan Mountains, Bayinbuluke has a distinctive high-cold mountain climate. Existing studies have mainly relied on long-term observational data from surface meteorological stations and hydrological stations, revealing, at the macroscopic scale, the interdecadal trends, abrupt-change characteristics, and statistical patterns of temperature and precipitation in the Bayinbuluke Basin[4–6]. Against the background of macroscopic climate, studies by Zhang Yichi et al.[7], Diao Peng et al.[8], and Li Xiaoqi et al.[9] further explored the effects of climate change on runoff changes in the Kaidu River Basin, the evolution of the surface-water environment of high-cold wetlands, the response of flood-season runoff to climate fluctuations and human regulation, and the characteristics and impacts of changes in atmospheric saturation vapor-pressure deficit. In terms of specialized monitoring, Yao Junqiang et al.[10] assessed the comprehensive impacts of climate change on water resources and grassland ecosystems in high-cold basins through dynamic monitoring. In addition, in cloud-precipitation physics and boundary-layer detection, studies have begun to use specialized equipment for observations. Ming Hu et al.[11] effectively retrieved raindrop spectra for a stratiform-cloud precipitation process in the central Tianshan Mountains using the velocity power spectrum of the vertical beam of a wind-profiler radar; Li Yanwei et al.[12] used a disdrometer to analyze raindrop-spectrum characteristics and distribution patterns in the Tianshan Mountains; Yang Jinming et al.[13] studied changes in the dielectric constants of dry snow and wind-blown snow in the Bayinbuluke Basin.

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Although substantial work has been carried out on the climatic characteristics around the Bayinbuluk Basin, previous studies have mostly focused on fixed-point surface observations or analyses of the physical characteristics of individual precipitation processes. Relatively little research has addressed atmospheric vertical structure, boundary-layer height, and stratification stability; in particular, studies of the vertical structure of meteorological elements based on tethered aerostat sounding, rocket sounding, and other observational methods remain limited. This has, to some extent, constrained a deeper understanding of boundary-layer physical processes in this region and the development of more refined numerical prediction.

This study uses near-surface meteorological data from the Bayinbuluk National Basic Meteorological Station during the summers of 2014–2024, first-hand data from tethered aerostat sounding and rocket sounding experiments conducted in the Bayinbuluk Basin in August 2014 under the National Science and Technology Support Program project "Research, Development, and Application of Key Technologies for Artificial Rain and Snow Enhancement in the Tianshan Mountains," and concurrent ERA5 reanalysis data to conduct an in-depth analysis of the characteristics of the summer atmospheric structure in this region. The aim is to deepen understanding of land–atmosphere interactions and boundary-layer physical processes in alpine mountainous areas, and to provide an important theoretical basis for improving the accuracy of physical parameterization schemes in numerical weather prediction and climate models for alpine mountainous regions.

1 Data and Methods

1.1 Overview of the Study Area

The Bayinbuluk Basin is located in the central hinterland of the Tianshan Mountains in Xinjiang. It is about 270 km long from east to west and 136 km wide from north to south, with an area of approximately 2.3×10^4 km^2 (Fig. 1). The basin floor is flat, with an elevation of 2400–2600 m. Except for the Gongnaisi Valley in the northwest, which is affected by moist westerly valley airflow, the basin is essentially enclosed on the other three sides, forming a distinctive alpine mountain climate. The underlying surface is alpine meadow; the annual mean temperature is $-4.6\,^{\circ}$C, the extreme minimum temperature is below $-48\,^{\circ}$C, annual precipitation is 276 mm, the number of snow-cover days reaches 139.3 d, and the maximum frozen-soil depth is 750 cm[^14].

1.2 Data Processing and Quality Control

The data were obtained from near-surface observations during June–August in the summers of 2014–2024 at the Bayinbuluk National Basic Meteorological

Fig. 1 Schematic diagram of the study area and tethered balloon sounding

Figure 1: Fig. 1 Schematic diagram of the study area and tethered balloon sounding

Station (station number 51542), including meteorological elements such as air pressure, air temperature, relative humidity, wind speed, and wind direction. The station has an elevation of 2458.9 m and is located in the hinterland of the Bayinbuluk–Yulduz Basin in the middle section of the Tianshan Mountains (43.02°N, 84.09°E). It is equipped with a standardized ground automatic weather station (DZZ) and conducts continuous all-weather monitoring in strict accordance with the *Specifications for Surface Meteorological Observations*. The data underwent strict quality control by the National Meteorological Information Center; abnormal values and systematic errors were removed, and the data completeness exceeded 99%. The dataset is therefore reliable and can meet the requirements of this study for analyzing the near-surface climatic background of the alpine basin.

Tethered aerostat sounding data collected in the Bayinbuluk Basin from 5 to 15 August 2014 were used. Eight soundings were conducted per day, with a temporal resolution of 3 h; there were 54 valid flights in total. The sounding times included 02:00, 05:00, 08:00, 11:00, 14:00, 17:00, 20:00, and 23:00, with the numbers of valid soundings at these times being 10, 4, 10, 5, 10, 3, 7, and 5, respectively. To ensure data quality and reliability, obvious errors or data not conforming to statistical characteristics in the raw data were processed. Observations from the ascent stage of the tethered aerostat were used. If any one of air temperature, air pressure, relative humidity, or wind speed was missing or abnormal in a single sounding, that sounding was discarded. When repeated measurements occurred at the same altitude, the mean of all valid data at that altitude was used[^15]. Physically unreasonable outliers were removed, with the criteria defined as air temperature $> 60\,^{\circ}\mathrm{C}$, instantaneous wind speed $> 50\ \mathrm{m \cdot s^{-1}}$, wind direction $> 360^{\circ}$, and humidity $> 100\%$.

In addition, rocket sounding data at 18:08 on 9 August, 17:08 on 13 August, and 03:08, 05:08, and 10:08 on 19 August (Beijing time) were incorporated, with a temporal resolution of 1 s. Because the observational method is relatively costly and the amount of data actually obtained was limited, only data from the above sounding periods were selected for analysis. For missing values in the data, cubic interpolation was used to

Figure 1. Schematic diagram of the study area and tethered balloon sounding.

Visible figure annotations: (a) schematic diagram of the study area; Bayinbuluk meteorological station; meteorological station; elevation/m; 5241; 1051; scale 0–70 km. (b) tethered aerostat sounding photograph.

processing. When the number of consecutive missing data points for a meteorological element does not exceed 5 (i.e., the duration of continuous missing values does not exceed 5 s), cubic interpolation is used for completion; if the number of consecutive missing data points exceeds 5, or if any element contains a serious anomalous value, the entire record is deleted [16].

ERA5 is the fifth-generation atmospheric reanalysis dataset released by the European Centre for Medium-Range Weather Forecasts and launched by the Copernicus Climate Change Service funded by the European Union. The dataset has a horizontal resolution of $0.25^{\circ} \times 0.25^{\circ}$, a temporal resolution of 1 h, and contains 37 pressure levels from 1 to 1000 hPa in the vertical direction. In this study, the world time of ERA5 was converted to Beijing time, the ERA5 grid values were interpolated to the station locations using bilinear interpolation, and the altitude was uniformly converted to height above ground level (all heights in this paper refer to height above ground level).

1.3 Research Methods

According to the needs of data analysis, the measured data were converted by formula into specific humidity q ($\mathrm{g \cdot kg^{-1}}$) and potential temperature θ (K).

$$r_s = \frac{\varepsilon e_s(T)}{P} \tag{1}$$

$$e_s = 6.1078 \times 10^{\left(\frac{7.5t}{t+237.3}\right)},\ t < 0 \tag{2}$$

$$e_s = 6.112 e^{\left(\frac{17.67t}{t+243.5}\right)},\ t \geq 0 \tag{3}$$

$$e = \frac{RH}{100} \times e_s \tag{4}$$

$$q = \frac{0.622 \times e}{P - 0.378 \times e} \tag{5}$$

$$\theta = T \left(\frac{P_o}{P}\right)^{0.286} \tag{6}$$

where r_s is the saturation mixing ratio ($\mathrm{g \cdot kg^{-1}}$); ε is a dimensionless physical constant, $\varepsilon = M_v/M_d \approx 0.622$; M_v is the molar mass of water vapor ($\mathrm{g \cdot mol^{-1}}$); M_d is the molar mass of dry air ($\mathrm{g \cdot mol^{-1}}$); e_s is the saturation water vapor pressure (hPa); e is the water vapor pressure (hPa); t is the air temperature in degrees Celsius (°C); RH is relative humidity (%); P is air pressure (hPa); P_o is the standard pressure (hPa), taken as 1000 hPa; T is the air temperature in Kelvin (K), $T = t + 273.15$.

1.4 Method for Determining Atmospheric Boundary-Layer Height

In view of the complex terrain and local climatic characteristics of the Bayanbulak alpine basin, this study adopts a comprehensive identification method combining the potential-temperature gradient method and the wind-profile method to determine the atmospheric boundary-layer height, so as to reduce the uncertainty of a single criterion. The potential-temperature gradient method determines the boundary-layer height by analyzing the vertical distribution characteristics of the potential-temperature profile [17]. Within the mixed layer, air temperature and water vapor are uniformly distributed, and the vertical gradient of potential temperature is relatively small; above the mixed-layer height, the potential-temperature gradient is relatively large. Therefore, the height of the maximum potential-temperature gradient is determined as the boundary-layer height [18]. The wind-profile method makes judgments based on differences in the dynamical characteristics at the top of the boundary layer. In the stable boundary layer, nocturnal near-surface radiative cooling forms an inversion, and stable stratification suppresses vertical turbulence, obstructing momentum exchange between the upper airflow and the surface friction layer. Under inertial oscillations, low-level jets are often induced near the top of the inversion layer. In this paper, the height at which the maximum low-level wind speed occurs is defined as the nighttime boundary-layer height. In the convective boundary layer, during daytime the surface is heated, and momentum mixing in the deep mixed layer is uniform. The top of the boundary layer is located in the entrainment layer, which is the interface between the turbulent layer and the free-atmosphere momentum exchange; wind speed and wind direction often change sharply there. Therefore, the height of the maximum wind speed or maximum wind shear is identified as the daytime convective boundary-layer height [19].

1.5 Methods for Determining Atmospheric Stratification Stability

1.5.1 Gradient Richardson Number Method This method is a dimensionless parameter introduced by Richardson in 1920 on the basis of the energy budget equation, and is used to quantify atmospheric turbulence intensity and the stability characteristics of atmospheric stratification [20]. In essence, it reflects the ratio of the suppressing effect of buoyancy on turbulence to the mechanical generation effect of wind shear on turbulence [21]. The formula is as follows:

$$R_i = \frac{g}{\theta_v} \frac{(\partial \theta_v / \partial z)}{(\partial u / \partial z)^2 + (\partial v / \partial z)^2} \tag{7}$$

$$\theta_v = \theta \times (1 + 0.61q) \tag{8}$$

where θ_v is virtual potential temperature; g is gravitational acceleration (usually taken as 9.8 $\mathrm{m \cdot s^{-2}}$); and z is height above sea level. In actual calculations, the gradient terms are approximated by finite differences $\Delta\theta_v/\Delta z$, $\Delta u/\Delta z$, and $\Delta v/\Delta z$, where the increments of θ, u, and v over the vertical distance Δz are taken from the vertical profiles obtained by actual observations or numerical simulations. In this paper, atmospheric boundary-layer states are classified into three types according to the R_i value [22]: unstable atmospheric stratification ($R_i < 0$), near-neutral atmospheric stratification ($0 \leq R_i < 0.25$), and stable atmospheric stratification ($R_i \geq 0.25$).

1.5.2 Virtual Potential-Temperature Gradient Method The virtual potential-temperature gradient method is a method based on the atmospheric hydrostatic equation that determines atmospheric stratification stability by analyzing the rate of change of virtual potential temperature with height [23]. To eliminate the interference of instrumental measurement errors (± 0.1°C) with stratification determination, the judgment threshold is set to $\delta = 0.001\ \mathrm{K \cdot m^{-1}}$. Accordingly, based on the vertical gradient of virtual potential temperature ($\partial\theta_v/\partial z$), atmospheric stratification is divided into the following three categories:

$$\text{Stratification state} = \begin{cases} \text{unstable}, & \dfrac{\partial\theta_v}{\partial z} < -0.001 \\ \text{near-neutral}, & -0.001 \leq \left|\dfrac{\partial\theta_v}{\partial z}\right| < 0.001 \\ \text{stable}, & \dfrac{\partial\theta_v}{\partial z} \geq 0.001 \end{cases} \tag{9}$$

2 Results and Analysis

2.1 Summer Atmospheric Boundary-Layer Characteristics

2.1.1 Characteristics of Near-Surface Atmospheric Elements As can be seen from Fig. 2a, during 2014-2024, the summer near-surface mean air temperature in the Bayanbulak Basin was 11.14°C, varying overall within the range of 9.75-12.50°C. 2023–

(a) Air temperature (b) Specific humidity

(b) Wind speed (d) Wind direction

Figure 2. Interannual variations of near-surface meteorological elements in the Bayanbulak Basin during summer from 2014 to 2024

Fig. 2 Interannual variations of surface meteorological elements in the Bayanbulak Basin during summer from 2014 to 2024

The mean temperature in 2024 was higher than in the other years; the summer mean temperature reached its maximum value of 12.50 °C in 2024. The mean temperature in 2020 was relatively low, with the summer mean temperature falling below 10 °C (9.75 °C). The extreme maximum temperature was 29 °C in 2015. The summer extreme temperature range in every year exceeded 25 °C; in 2018 the extreme temperature range reached 34.20 °C, with a maximum temperature of 23.90 °C and a minimum temperature of −10.30 °C. As shown in Fig. 2b, the summer mean specific humidity was 6.32–7.48 $g \cdot kg^{-1}$. The mean specific humidity was highest in 2024, at 7.48 $g \cdot kg^{-1}$, and lowest in 2014, at 6.32 $g \cdot kg^{-1}$. The maximum specific humidity, 12.84 $g \cdot kg^{-1}$, occurred in 2015, while the minimum specific humidity, 0.95 $g \cdot kg^{-1}$, occurred in 2018; the difference in specific humidity reached 10.13 $g \cdot kg^{-1}$.

As shown in Figs. 2c and 2d, near-surface wind speed exhibited pronounced fluctuations on the interannual scale. During the summers from 2014 to 2024, the mean wind speed ranged from 2.82 to 3.45 $m \cdot s^{-1}$. Among these years, the mean wind speed was highest in 2024, at 3.45 $m \cdot s^{-1}$, while the extreme maximum wind speed of 16.60 $m \cdot s^{-1}$ occurred in 2015. The mean wind speeds in 2021 and 2023 were relatively low, at 2.96 $m \cdot s^{-1}$ and 2.82 $m \cdot s^{-1}$, respectively, and the corresponding maximum wind speeds also decreased to 13.10 $m \cdot s^{-1}$ and 12.40 $m \cdot s^{-1}$. In summer, near-surface easterly winds were the prevailing wind direction in this region, accounting for as much as 21.71%; southwesterly winds ranked second, with an occurrence frequency of 17.08%, together accounting for 38.79%. In addition, the occurrence frequencies of northerly, northeasterly, northwesterly, and westerly winds ranged from 11.40% to 13.30%. Southeasterly and southerly winds occurred very infrequently, at only 5.52% and 5.88%, respectively.

2.1.2 Characteristics of Atmospheric Elements Based on the Tethered Aerostat

Figure 3a shows that the temperature profile has pronounced diurnal variation characteristics. During the daytime, temperature exhibits an insolation-type pattern in which it decreases with height; at night, it exhibits a radiative-type pattern in which temperature increases with height, that is, an inversion. In the daytime, the land surface absorbs solar radiation and warms, and heat is transferred to the near-surface atmosphere through turbulent sensible heat flux, causing the near-surface air temperature to rise rapidly and forming a temperature distribution that decreases upward from the surface. At night, radiative cooling of the surface occurs and sensible heat flux is transported downward, causing the near-surface atmosphere to cool; the magnitude of cooling weakens with increasing height. As a result, an inversion layer develops upward from the surface, and the temperature gradient is positive. At 02:00, the near-surface air temperature was relatively low, at 7.4 °C, and temperature increased with height, showing typical nocturnal inversion characteristics. At this time, the top height of the inversion layer was 225 m, and the inversion intensity was 1.55 °C. By 08:00, as incoming solar radiation at the ground strengthened, the inversion layer began to dissipate from the bottom, while the residual inversion layer rose to 350 m and the inversion intensity weakened to 1.23 °C. From 14:00 to 20:00, the surface air temperature increased to about 16.0 °C, the vertical distribution of temperature tended to become uniform, and temperature as a whole decreased with height, consistent with the characteristics of the dry adiabatic lapse rate. The formation of inversions in this region is mainly influenced jointly by topography, arid climate, and local radiative cooling. At night, cold air produced by strong radiative cooling on slopes sinks along the mountainsides and accumulates at the bottom of the basin; topographic barriers inhibit the diffusion of cold air, while the upper atmosphere maintains a relatively high temperature, thereby forming a stable inversion structure. Figure 3b

Fig. 3 Tethered balloon sounding profiles of meteorological elements

Panels: (a) Air temperature; (b) specific humidity; (c) wind speed; (d) wind direction. Legend: 02:00, 08:00, 14:00, 20:00.

This indicates that specific humidity decreases with increasing height. From 02:00 to 08:00, the near-surface specific humidity is $5.0 \sim 6.0\ \mathrm{g \cdot kg^{-1}}$, gradually decreasing upward to $3.0 \sim 4.0\ \mathrm{g \cdot kg^{-1}}$. Weak moisture-inversion layers exist within 35 m and 74 m, respectively, where specific humidity reaches maximum values of $5.6\ \mathrm{g \cdot kg^{-1}}$ and $5.3\ \mathrm{g \cdot kg^{-1}}$. At 14:00, daytime convection strengthens, vertical mixing of water vapor intensifies, and the vertical distribution of specific humidity tends to become uniform. At 20:00, near-surface specific humidity reaches the daily maximum of $5.9\ \mathrm{g \cdot kg^{-1}}$; an inflection point appears at 870 m, where specific humidity increases slightly to $4.7\ \mathrm{g \cdot kg^{-1}}$.

The variation in wind speed depends on surface roughness and atmospheric stability [23]. Figure 3c shows that near-surface wind speed at night is generally low. Within $0 \sim 200$ m, wind speed fluctuates greatly under the influence of surface friction and basically conforms to the logarithmic law; above 200 m, wind speed is relatively steady. Within the surface-based inversion layer, wind speed exhibits a maximum. As shown in Fig. 3d, within the effective detection height, the prevailing wind direction at night is westerly. During 08:00–14:00, wind speed in the convective boundary layer is relatively uniform; within the height range of $200 \sim 800$ m, the vertical variation in wind speed is slight, and the wind-speed profile is approximately vertically homogeneous. In general, when the atmosphere is unstable, vertical turbulent transport of momentum [24] makes wind speed vary little with height; the prevailing wind directions during this period are southwesterly and southerly. At 20:00, wind speed shows multi-peak fluctuations, with disordered inflection points at various height levels. Low-level wind-speed fluctuations are evident, the maximum wind speed reaches $4.3\ \mathrm{m \cdot s^{-1}}$, and the wind direction shifts to southwesterly and westerly. Combining Fig. 3a and Fig. 3c, it can be seen that the vertical distribution of boundary-layer wind speed is closely related to the temperature stratification. The inversion layer formed at night suppresses turbulent mixing and the downward transport of upper-level momentum, causing wind speed below the inversion layer to be generally low. Thermally driven circulations caused by the diurnal temperature difference also affect the daily variation of wind speed: daytime surface heating enhances turbulent mixing and may strengthen low-level wind speed; at night, surface cooling increases the near-surface temperature gradient and pressure gradient, thereby strengthening the wind.

2.1.3 Characteristics of variations in atmospheric boundary-layer height

The summer atmospheric boundary-layer height over the Bayinbuluk Basin, determined using the potential-temperature gradient method and the wind-profile method, exhibits a diurnal variation characterized by "lower at night and higher in the daytime." Moreover, the consistency between the two methods shows obvious stage-dependent differences as the atmospheric stratification state changes. As shown in Figs. 4a and 4b, at night and in the early morning (02:00 and 08:00), under the influence of surface radiative cooling, the atmospheric stratification is stable. The thermal boundary-layer heights determined by the potential-temperature gradient method are mainly distributed within 183 $\sim$ 526 m (with mean values of 274 m and 356 m, respectively), whereas the dynamic boundary-layer heights determined by the wind-profile method are significantly lower, mainly concentrated within 122 $\sim$ 284 m (with mean values of 166.3 m and 164.9 m, respectively). A significant systematic negative bias exists between the two (with mean biases of -107.7 m and -191.1 m, respectively). The potential-temperature gradient method identifies the height corresponding to the top of the radiative inversion layer, where thermal diffusion is restricted, whereas the wind-profile method characterizes the upper limit of dynamic mixing under the combined effects of surface friction and stratification suppression. As shown in Figs. 4c and 4d, with enhanced solar radiation in the afternoon and evening (14:00 and 20:00), surface heating drives the development of a deep convective mixed layer, and the atmospheric boundary-layer height rises rapidly. The heights determined by the potential-temperature gradient method jump to 420 $\sim$ 951 m (with mean values of 660.6 m and 636.0 m, respectively), and those determined by the wind-profile method also rise synchronously to 336 $\sim$ 944 m (with mean values of 610.6 m and 640.8 m, respectively), reflecting an intense turbulent mixing process. During this period, the calculated results of the two methods are highly consistent: the mean bias at 14:00 decreases to -50.6 m, and the mean bias at 20:00 is only 4.8 m. This indicates that within the convective mixed layer, strong turbulent activity promotes sufficient vertical mixing of heat and momentum, causing the heights of the thermal and dynamic boundaries to tend toward coincidence. This confirms that the summer daytime and evening boundary-layer structure in this region has a high degree of thermo-dynamic coupling.

2.1.4 Characteristics of atmospheric stratification stability

A comparison between the gradient Richardson number method and the virtual-potential-temperature gradient method shows (Fig. 5) that the Bayinbuluk Basin is dominated by stable atmospheric stratification, with proportions of 69.43% and 66.23%, respectively.

Figure text

Legend: blue–potential-temperature gradient method; red–Richardson number method

(a) 02:00

(b) 08:00

(c) 14:00

(d) 20:00

Vertical-axis label for panels (a)-(d): Boundary-layer height/m

Legend for panels (e)-(h): orange–$\Delta H > 0$ (Richardson-number method biased high); green–$\Delta H < 0$ (potential-temperature method biased high)

(e) 02:00 Mean bias: -107.7 m

(f) 08:00 Mean bias: -191.1 m

(g) 14:00 Mean bias: -50.6 m

(h) 20:00 Mean bias: 4.8 m

Vertical-axis label for panels (e)-(h): Height-difference value/m
Horizontal-axis label: Time/month–day

Fig. 4 Characteristics of atmospheric boundary-layer height variation from August 5 to 15, 2014

The main differences between the two methods lie in the fact that the proportion of unstable stratification identified by the R_i method (28.30%) is higher than that identified by the gradient method (11.89%), whereas the proportion of near-neutral stratification identified by the gradient method (21.89%) is far greater than that identified by the R_i method (2.26%). This indicates that the R_i method is relatively sensitive to shear turbulence and tends to classify some transitional states as unstable, whereas the gradient method, which considers only thermal factors, can identify near-neutral stratification in greater detail. In terms of vertical distribution, stable stratification accounts for a significant proportion in all height layers. The difference is that the unstable stratification identified by the R_i method is distributed relatively uniformly across all height layers, whereas the virtual

(a) Stratification stability distribution by the R_i method
Vertical axis: Height/m; horizontal axis: Time. Color scale: Richardson number –stable, neutral, unstable.

(b) Stratification stability distribution by the virtual-potential-temperature-gradient method

Vertical axis: Height/m; horizontal axis: Time. Color scale: Stability index—stable, neutral, unstable.

Legend: unstable; near-neutral; stable.

(c) Vertical distribution of stratification stability by the R_i method
Vertical axis: Height layer/m; horizontal axis: Proportion/%.

(d) Vertical distribution of stratification stability by the virtual-potential-temperature-gradient method
Vertical axis: Height layer/m; horizontal axis: Proportion/%.

Fig. 5. Characteristics of atmospheric stratification stability

The potential-temperature-gradient method shows pronounced stratified features: unstable stratification is confined to the near-surface layer of 0–100 m. As height increases, thermal turbulence is suppressed, and the stratification gradually shifts toward near-neutral conditions. The gradient method can effectively capture thermally driven stratification dominated by surface heating in the near-surface layer, whereas the R_i method can sensitively identify dynamically unstable states in the upper layers induced by wind shear associated with local circulation. Unstable atmospheric stratification mainly occurs from 11:00 to 20:00. From 23:00 to 11:00 the next day, under the influence of surface radiative cooling, stable stratification is overwhelmingly dominant, effectively suppressing the vertical exchange of turbulence.

2.2 Characteristics of upper-air meteorological elements from rocket sounding

Based on the rocket-sounding data, Fig. 6a shows that air temperature at each time exhibits the typical feature of decreasing with height. Near-surface air temperature ranges from 6.0 to 15.6 °C, with the highest temperature occurring at 18:08 on 9 August. With increasing height, the temperature decreases to −33.0 °C, −31.1 °C, −32.1 °C, −41.1 °C, and −38.1 °C at 6692 m, 6636 m, 7125 m, 7707 m, and 7824 m, respectively. Fig. 6b shows that near-surface specific humidity reaches its maximum value of 6.6 $\mathrm{g \cdot kg^{-1}}$ at 03:08 and its minimum value of 3.4 $\mathrm{g \cdot kg^{-1}}$ at 18:08, approaching zero near an altitude of 8000 m. Figs. 6c and 6d show that wind speed generally increases with height. Wind speed is relatively low in the lower layer, ranging from 2.0 to 10.4 $\mathrm{m \cdot s^{-1}}$, and increases markedly in the middle and upper layers. At 03:08 and 05:08, wind speed increases to about 20 $\mathrm{m \cdot s^{-1}}$; at 10:08 on 19 August, it exceeds 30 $\mathrm{m \cdot s^{-1}}$, indicating a strong-wind-layer structure. In the lower and middle layers, wind direction is dominated by westerly components, with obvious directional shear with height. During the night of 19 August, wind direction is relatively stable: it is southeasterly in the lower layer and turns clockwise to southerly with height.

2.3 Comparison between ERA5 reanalysis data and observational data

To evaluate the applicability of ERA5 reanalysis data in the Bayanbulak Basin, ERA5 reanalysis data from the same period were selected and compared with tethered-airship observations and rocket-sounding data.

As shown in Fig. 7, the air temperature from ERA5 and from rocket sounding exhibits extremely high consistency, with the correlation coefficient R close to 1.0. Differences between ERA5 reanalysis data and tethered-airship temperature observations are affected by the near-surface inversion, and the agreement is better during non-inversion periods. The consistency is best at 14:00, with $R = 0.92$ and RMSE = 1.17 °C; at 02:00, 08:00, and 20:00, however, R decreases to 0.61–0.67, and the RMSE values are relatively high, being 2.36 °C, 2.68 °C, and 2.56 °C, respectively. This indicates that ERA5 has difficulty accurately capturing the fine-scale thermal structure of the near-surface layer under strong nocturnal radiative inversions in the Bayanbulak Basin. The specific humidity from ERA5 and from rocket sounding still maintains an extremely high correlation, with $R >$

Figure 6 (panels): (a) Temperature; (b) Specific humidity; (c) Wind speed; (d) Wind direction.
Legend times: 08-09 18:08; 08-13 17:08; 08-19 03:08; 08-19 05:08; 08-19 10:08.

Fig. 6 Rocket sounding profiles of meteorological elements

0.99, verifying its reliability in simulating water vapor throughout the layer. However, comparison with the specific humidity from tethered-balloon observations shows that, except at 08:00, the degree of dispersion increases at the other times, especially at 14:00, when the correlation weakens significantly and R is only 0.39, while the RMSE is relatively high, reaching 1.59 $g \cdot kg^{-1}$. At 02:00 and 20:00 the correlations are moderate, with R values of 0.74 and 0.67, and RMSE values of 1.49 $g \cdot kg^{-1}$ and 0.93 $g \cdot kg^{-1}$, respectively. These diurnal differences arise from the diurnal cycle of thermodynamic and dynamic processes in the summer boundary layer over the Bayinbuluk Basin, as well as from the resolution and parameterization limitations of the ERA5 reanalysis data. In the afternoon, the convective boundary layer develops deeply, and strong turbulent mixing makes the temperature field tend to be uniform; ERA5 can better capture the large-scale thermal field, so the correlation is high. During the nighttime inversion and transition periods, rapid nonequilibrium processes exceed the analytical capability of ERA5, resulting in reduced correlations. The diurnal variation of specific humidity reflects the scale dependence of water-vapor transport. At 08:00, turbulence is weak, the water-vapor distribution is dominated by large-scale processes, and ERA5 performs best in its representation; at 14:00, turbulence is intense and water-vapor heterogeneity cannot be resolved, leading to the lowest correlation.

3 Discussion

Based on integrated observations in the Bayinbuluk Basin, this study reveals the unique characteristics of the summer boundary-layer structure over an underlying surface of alpine meadow. The results indicate that the prominent nocturnal inversion, the specific wind–humidity vertical structure, and the relatively low boundary-layer height in this region are mainly influenced by its unique topography and underlying-surface conditions. Unlike typical arid regions where sensible heat flux dominates and readily promotes the development of a deep convective boundary layer [25–28], as well as regions with special topographic conditions such as the Tibetan Plateau [29–32], the underlying surface of the Bayinbuluk Basin is alpine meadow, with relatively low surface albedo, high soil moisture, and large vegetation coverage. A large portion of the net radiation absorbed by the surface is consumed as latent heat flux, which limits the sensible heat flux that drives near-surface turbulent motion. Meanwhile, the relatively large heat capacity and high thermal conductivity of meadow soil allow energy to be conducted into upper soil layers, further weakening near-surface atmospheric thermal uplift. This energy-partitioning pattern dominated by latent heat constrains the development of the convective boundary-layer height, and is the thermal mechanism causing the boundary-layer height in this region to remain far lower than that in desert regions. In addition, the coupling between topography and stratification has a significant influence on the diurnal variation process. At night, under the influence of basin topography, cold air accumulates near the bottom, and, together with intense surface radiative cooling, forms a strong and structurally stable radiation inversion. This stable stratification effectively suppresses vertical turbulent exchange of momentum, causing the near-surface layer to exhibit low wind-speed characteristics; during the day, turbulent mixing strengthens, and only then does the vertical distribution of wind speed tend toward uniformity.

In terms of water-vapor distribution, the vertical structure of specific humidity is closely related to the boundary-layer dynamics. The overall specific humidity in the Bayinbuluk Basin is relatively low and its diurnal variation amplitude is small, which contrasts sharply with the Tibetan Plateau and the Taklimakan Desert. Some regions of the Tibetan Plateau are affected by moisture transport from the Indian monsoon, and water vapor fluctuates strongly during the wet season [33–34]; the Taklimakan Desert, by contrast, is greatly affected by local evaporation driven by extremely high sensible heat. Because of its enclosed terrain, high altitude, weak external water-vapor transport, and limited local evaporation, the Bayinbuluk Basin has overall low specific humidity. This low-humidity environment enhances nocturnal radiative cooling and promotes the formation of a strong inversion layer, while the extremely stable stratification at night in turn markedly weakens vertical turbulent exchange of water vapor.

In evaluating the applicability of ERA5 reanalysis data, it was found that ERA5 can reasonably reproduce large-scale trends in temperature and humidity variations, showing relatively high correlations especially for air temperature and specific humidity in the rocket-sounding data. However, it has clear deficiencies in capturing the strong near-surface inversion. Although ERA5 reanalysis data have relatively high accuracy in simulating atmospheric temperature and specific humidity, considering the complex terrain and

Legend: data points; regression line; 1:1 reference line.

- **(a) 02:00**
  $y = 0.75x - 0.07$, RMSE $= 2.36$, $R = 0.67$
  x-axis: tethered-sounding air temperature/°C; y-axis: ERA5 air temperature/°C.
- **(b) 08:00**
  $y = 0.87x - 1.60$, RMSE $= 2.68$, $R = 0.63$
  x-axis: tethered-sounding air temperature/°C; y-axis: ERA5 air temperature/°C.
- **(c) 14:00**
  $y = 0.91x + 0.25$, RMSE $= 1.17$, $R = 0.92$
  x-axis: tethered-sounding air temperature/°C; y-axis: ERA5 air temperature/°C.
- **(d) 20:00**
  $y = 0.68x + 1.31$, RMSE $= 2.56$, $R = 0.61$
  x-axis: tethered-sounding air temperature/°C; y-axis: ERA5 air temperature/°C.
- **(e) 02:00**
  $y = 0.95x + 1.51$, RMSE $= 1.49$, $R = 0.74$
  x-axis: tethered-sounding specific humidity/$(\mathrm{g} \cdot \mathrm{kg}^{-1})$; y-axis: ERA5 specific humidity/$(\mathrm{g} \cdot \mathrm{kg}^{-1})$.
- **(f) 08:00**
  $y = 0.78x + 1.62$, RMSE $= 1.11$, $R = 0.92$
  x-axis: tethered-sounding specific humidity/$(\mathrm{g} \cdot \mathrm{kg}^{-1})$; y-axis: ERA5 specific humidity/$(\mathrm{g} \cdot \mathrm{kg}^{-1})$.
- **(g) 14:00**
  $y = 0.29x + 4.22$, RMSE $= 1.59$, $R = 0.39$
  x-axis: tethered-sounding specific humidity/$(\mathrm{g} \cdot \mathrm{kg}^{-1})$; y-axis: ERA5 specific humidity/$(\mathrm{g} \cdot \mathrm{kg}^{-1})$.
- **(h) 20:00**
  $y = 0.62x + 2.18$, RMSE $= 0.93$, $R = 0.67$
  x-axis: tethered-sounding specific humidity/$(\mathrm{g} \cdot \mathrm{kg}^{-1})$; y-axis: ERA5 specific humidity/$(\mathrm{g} \cdot \mathrm{kg}^{-1})$.
- **(i) 03:00**
  $y = 1.15x - 0.83$, RMSE $= 3.32$, $R = 1.00$
  x-axis: rocketsonde air temperature/°C; y-axis: ERA5 air temperature/°C.

- **(j) 05:00**
  $y = 0.98x - 1.49$, RMSE $= 1.47$, $R = 1.00$
  x-axis: rocketsonde air temperature/°C; y-axis: ERA5 air temperature/°C.
- **(k) 10:00**
  $y = 1.14x - 1.48$, RMSE $= 4.23$, $R = 1.00$
  x-axis: rocketsonde air temperature/°C; y-axis: ERA5 air temperature/°C.
- **(l) 03:00**
  $y = 1.51x + 0.67$, RMSE $= 1.62$, $R = 0.99$
  x-axis: rocketsonde specific humidity/$(\mathrm{g} \cdot \mathrm{kg}^{-1})$; y-axis: ERA5 specific humidity/$(\mathrm{g} \cdot \mathrm{kg}^{-1})$.
- **(m) 05:00**
  $y = 2.29x + 0.26$, RMSE $= 2.03$, $R = 1.00$
  x-axis: rocketsonde specific humidity/$(\mathrm{g} \cdot \mathrm{kg}^{-1})$; y-axis: ERA5 specific humidity/$(\mathrm{g} \cdot \mathrm{kg}^{-1})$.
- **(n) 10:00**
  $y = 2.21x + 0.04$, RMSE $= 1.69$, $R = 0.99$
  x-axis: rocketsonde specific humidity/$(\mathrm{g} \cdot \mathrm{kg}^{-1})$; y-axis: ERA5 specific humidity/$(\mathrm{g} \cdot \mathrm{kg}^{-1})$.

Fig. 7 Comparison and analysis between ERA5 reanalysis data and observations

meteorological conditions; relying solely on ERA5 reanalysis data may fail to provide sufficient accuracy, especially in detailed simulations of the lower atmosphere. Therefore, for applications involving fine-scale low-level structures, it is recommended that ERA5 reanalysis data be corrected by combining them with direct observational methods such as tethered-aerostat measurements.

The limitations of this study are mainly reflected in the spatiotemporal coverage of the data. Because of the complex environmental conditions in alpine mountainous areas, direct observations are relatively difficult, and it was not possible to comprehensively cover different seasons and extreme weather processes. In the future, more systematic integrated observational data should be obtained and combined with high-resolution numerical simulations to deepen understanding of the fine evolution of atmospheric boundary-layer height, layer stability, and turbulent exchange processes in the Bayinbuluk Basin, and to develop boundary-layer parameterization schemes suitable for alpine mountainous areas, thereby improving the application accuracy of numerical weather prediction and climate models in similar alpine mountainous regions.

4 Conclusions

This study used summer near-surface data from the Bayinbuluk National Basic Meteorological Station, tethered-aerostat observations, and rocketsonde data, together with ERA5 reanalysis data, to analyze the characteristics of the summer atmospheric structure in this region. The main conclusions are as follows:

(1) From 2014 to 2024, the mean summer air temperature in the Bayinbuluk Basin was 11.14 °C. Near-surface easterly winds were the dominant summer wind direction, accounting for 21.71%. In summer, the lower atmosphere exhibited distinct nocturnal strong radiative inversion and low specific humidity. Near-surface specific humidity was only $5 \sim 6\ \mathrm{g \cdot kg^{-1}}$ and decreased with increasing height. At night, near-surface wind speed was relatively low and fluctuated strongly; during the daytime, within the height range of $200 \sim 800$ m, the vertical distribution of wind speed was approximately uniform. Westerly winds prevailed at night, while during the daytime they shifted to southwesterly and southerly winds. In the upper part of the convective layer, there was a strong-wind zone with wind speed greater than $30\ \mathrm{m \cdot s^{-1}}$.

(2) The diurnal variation in boundary-layer height was significant. Under nocturnal stable stratification, the thermal boundary-layer height (mean $274 \sim 356$ m) was significantly higher than the dynamic boundary-layer height (mean $164.9 \sim 166.3$ m); the two tended to be consistent only under strong daytime convective mixing. The regional atmospheric stratification was dominated by stable conditions, and unstable stratification was confined to the $0 \sim 100$ m near-surface layer from noon to evening. With increasing height, the stratification rapidly evolved toward near-neutral or stable conditions, and stratification stability increased with height.

(3) The ERA5 reanalysis data showed good ability to reproduce the temperature and humidity fields of the free atmosphere over the Bayinbuluk Basin, with correlation coefficients greater than 0.99 for observed air temperature and specific humidity. However, the dataset showed biases in simulating near-surface nocturnal inversion and afternoon water-vapor variations: at 02:00, the air-temperature correlation coefficient decreased to 0.67, and at 14:00, the specific-humidity correlation coefficient was only 0.39.

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Summer Atmospheric Structure Characteristics over the Bayanbulak Basin in the Tianshan Mountains

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Abstract: Based on the summer data from the Bayanbulak National Basic Meteorological Station from 2014 to 2024, tethered balloon and rocket sounding data in August 2014, and concurrent ERA5 reanalysis data, this study analyzed the characteristics of the summer atmospheric structure in this region and evaluated the applicability of ERA5 reanalysis data in alpine mountainous areas. The results show that: (1) The average summer temperature in the Tianshan Bayanbulak Basin was 11.14 °C, with easterly winds dominant near the surface. Strong radiation inversion and low specific humidity were prominent features of the low-altitude atmosphere. During the daytime, the temperature decreased with increasing altitude, and the specific humidity in the near-surface layer (5–6 $\mathrm{g \cdot kg^{-1}}$) was higher than that in the upper layer (3–4 $\mathrm{g \cdot kg^{-1}}$). At night, the wind speed in the near-surface layer exhibited large fluctuations. During the daytime, the wind speed between 200 and 800 m was approximately vertical, and a strong wind belt with wind speeds greater than 30 $\mathrm{m \cdot s^{-1}}$ occurred in the middle and upper layers. (2) The atmospheric boundary layer height showed significant diurnal variation. Under stable stratification at night, the boundary layer height determined by the potential temperature gradient method (averaging 274-356 m) was significantly higher than that determined by the wind profile method (averaging 164.9-166.3 m); however, the two methods yielded consistent results under convective mixing. The atmospheric stratification in the Tianshan Bayanbulak Basin was predominantly stable, with stability increasing with altitude. (3) The ERA5 reanalysis data showed a high correlation with the measured temperature and humidity data in the free atmosphere, with correlation coefficients ranging from 0.99 to 1.00. However, notable deviations occurred in simulating the near-surface temperature inversion process, with the temperature correlation coefficient at 02:00 dropping to 0.67 and the specific humidity correlation coefficient at 14:00 dropping to 0.39. This study enhances understanding of boundary layer physical processes in alpine mountainous areas.

Keywords: Bayanbulak; atmospheric structure; atmospheric boundary layer height; stratification stability; tethered balloon sounding; rocket sounding

Note: Figure translations are in progress. See original paper for figures.