



1 **Summertime Turbulence Characteristics across China'**
2 **Planetary Boundary Layer: Roles of Wind Shear, Static**
3 **Stability, Cloud and Topography-Land-Cover Conditions**

4
5 Tian Xian¹, Yuping Sun^{2,6*}, Tianmeng Chen^{2,6}, Hui Xu², Xiaoran Guo², Zhen Zhang²,
6 Zhongwang Wu¹, Huaquan Hu¹, Juan Chen³, Liping Zeng⁴, Yifei Wang², Shuairu
7 Jiang^{2,5}, Pengzhan Yao^{2,5}, Ning Li², Honglong Yang⁶, Jianping Guo^{2,6*}

8
9 ¹National Key Laboratory of Space Target Awareness, School of Space Information,
10 Space Engineering University, Beijing 101416, China

11 ²State Key Laboratory of Severe Weather Meteorological Science and Technology &
12 Specialized Meteorological Advanced Support Technology Research Center, Chinese
13 Academy of Meteorological Sciences, Beijing, 100081, China

14 ³AVIC Leihua Electronic Technology Research Institute, Wuxi 214063, China

15 ⁴Guizhou New Meteorological Technology Co., Ltd, Guiyang 550001, China

16 ⁵Department of Atmospheric and Oceanic Sciences & Institute of Atmospheric
17 Sciences, Fudan University, Shanghai 200438, China

18 ⁶CMA Field Scientific Experiment Base for Low-Altitude Economy Meteorological
19 Support of Unmanned Aviation in Guangdong-Hong Kong-Macao Greater Bay Area,
20 Shenzhen 518108, China

21

22

23

24 Correspondence to: Dr./Prof. Jianping Guo (Email: jpguocams@gmail.com)

25 and Yuping Sun (Email: yupingsun99@hotmail.com)



Abstract

The spatio-temporal pattern of turbulence dissipation rate (ϵ) and its evolution remain elusive inside and above the planetary boundary layer (PBL). Using summertime 2022 data from a nationwide network of 132 radar wind profilers (RWPs), we contrast ϵ within the inner PBL ($0.3 z_i \leq z \leq z_i$) and above it ($z_i < z \leq 3 z_i$). Results reveal pronounced spatial heterogeneity: ϵ is strongest over Northwest and North China within the 30–40°N belt and weakest over the Tibetan Plateau and Southwest China, and it is systematically weaker aloft than within the PBL; ϵ also peaks near 1300 local time in both layers and persists day-to-day for roughly 9 days. We examine four modulating factors—topography and land cover, cloud conditions, static stability, and vertical wind shear (VWS). Of these, terrain complexity and land-cover type partly shape the observed heterogeneity, with stronger near-PBL turbulence over rough arid regions and weaker over water-dominated basins. Cloud cover enhances ϵ throughout the vertical column, with a stronger amplification above the PBL ($\Delta \log_{10} \epsilon = 0.16 \text{ m}^2 \text{ s}^{-3}$) than within it ($0.1 \text{ m}^2 \text{ s}^{-3}$), mainly through intensified VWS. The dominant controls also differ with height: within the PBL, static stability and VWS jointly govern turbulence, whereas above the PBL turbulence is driven overwhelmingly by shear-generated disturbances. Spectral analysis further reveals distinct vertical stratification of turbulent energy density that strengthens under stronger VWS. These findings demonstrate the value of national RWP networks for resolving turbulent dynamics across China and provide observational constraints for PBL turbulence parameterization.



47 **Short Summary**

48 Turbulence in the planetary boundary layer (PBL) affects cloud and precipitation, yet
49 little is known about it across China. With summertime 2022 data from 132 radar wind
50 profilers, we map the turbulence dissipation rate (ϵ) within and above the PBL. Stability
51 and wind shear jointly govern turbulence within the PBL, whereas shear dominates
52 above it; clouds enhance turbulence aloft, and topography and land cover shape its
53 regional pattern, offering new constraints for weather and climate models.



54 **1 Introduction**

55 The planetary boundary layer (PBL), the lowest part of the atmosphere, plays a
56 critical role in the exchange of heat, moisture, momentum, and pollutants between the
57 surface and the free atmosphere above (Cimini et al., 2020). Turbulence within the PBL
58 is a fundamental mechanism driving these exchanges, facilitating the transfer of
59 moisture and energy to the upper atmosphere, sustaining atmospheric motion, and
60 influencing weather and climate systems (Li et al., 2023). These processes have
61 profound implications, particularly for the aviation industry, where turbulence directly
62 impacts flight safety and operational efficiency (Kim et al., 2011; Fritts et al., 2012;
63 Sharman et al., 2012).

64 However, direct observational constraints remain inherently challenging due to
65 turbulence's small spatial scales, high spatiotemporal variability, and transient nature.
66 These challenges are compounded by the complex interplay of turbulence's multi-scale
67 structure, rapid temporal evolution, and inherently chaotic dynamics, rendering it
68 particularly elusive in both theoretical modeling and observational detection (Frisch,
69 1995; Sreenivasan & Antonia, 1997; Heilman et al., 2023). As a result, our
70 understanding of turbulence remains incomplete, particularly in terms of its spatial and
71 temporal variability.

72 To address these challenges, significant progress has been made through
73 observational studies. Early research relied on aircraft observations, which provided
74 turbulence quantities throughout the entire PBL but were limited to specific moments
75 in time (LeMone, 1976; Hildebrand & Bernice, 1984; Cho et al., 2003; Shi et al., 2014).
76 Tower-based measurements partially overcame these limitations by enabling
77 continuous monitoring at various heights, allowing for the analysis of surface and lower
78 atmospheric turbulence under diverse meteorological conditions (Feigenwinter et al.,
79 1999; Poulos et al., 2002; Sun et al., 2016; Zhang et al., 2023). However, these
80 observations are primarily confined to the surface layer, the inertial sublayer, and lower



81 mixed layer, leaving turbulence processes at higher altitudes under-sampled and poorly
82 understood.

83 Great advancements in remote sensing technologies have significantly expanded
84 our ability to study turbulence. Doppler lidar, for example, has demonstrated the
85 capability to detect low-frequency clear-sky turbulence features in the PBL with high
86 spatial and temporal resolution (Harvey et al., 2023; Rios et al., 2023; Xian et al., 2025).
87 In contrast, radar wind profilers (RWP) have proven particularly effective in capturing
88 turbulence structures even under cloudy or even overcast conditions (Meng et al., 2023).
89 Recent studies have used RWP turbulence profile measurements to characterize the fine
90 structure of the PBL in specific regions of China, yielding promising and reasonable
91 results (Solanki et al., 2022; Meng et al., 2023; Yang et al., 2023). Nevertheless,
92 previous work was mostly confined to limited sites or short-term campaigns, lacking
93 nationwide observational constraints on how turbulence varies geographically,
94 diurnally, and vertically, as well as the relative importance of terrain, land cover, cloud
95 conditions and atmospheric forcing. This knowledge gap motivates the present work
96 using the national RWP network dataset.

97 Given aforementioned research gaps, this study leverages high-resolution
98 measurements from the national RWP network to investigate summertime planetary
99 boundary-layer turbulence across China. Two core scientific objectives are addressed
100 in this work: (1) Characterize the nationwide spatiotemporal variability of summertime
101 turbulent dissipation rate within and above the PBL; (2) Quantify the respective roles
102 of topographic-land-cover conditions, cloud properties, and thermodynamic-dynamic
103 forcing in modulating turbulence at different altitudes, and exploit spectral features of
104 vertical-wind perturbations to unravel the associated energy-distribution and
105 stratification mechanisms. The rest of this paper is structured as follows: Section 2
106 describes the data and methods used in this study; Section 3 presents a detailed analysis
107 of nationwide spatiotemporal variability of turbulence, explores multiple modulating
108 factors, including surface conditions, cloud effects, and meteorological forcing, and a



109 spectral analysis of turbulence as well; and Section 4 summarizes the key findings and
110 their broad implications.

111

112 **2. Data and methods**

113 **2.1 RWP measurements**

114 The dataset used to estimate the turbulent dissipation rate (ϵ) consists of RWP
115 measurements by the China Meteorological Administration (CMA, Guo et al 2021).
116 These measurements offer a vertical resolution of 120 meters and a temporal resolution
117 of 6 minutes, providing detailed information on atmospheric wind fields, radial
118 velocities, Doppler spectrum widths, and signal-to-noise ratios (SNR). The data span
119 from June 1, 2022, to August 31, 2022, a period when surface heating and turbulent
120 mixing are strongest and the modulation of turbulence by underlying surface conditions
121 is most evident.

122 The spatial distribution of the RWP network in China is illustrated in Fig. 1. This
123 network comprises 132 RWP stations, covering a broad geographical area. Additionally,
124 to mitigate the influence of near-surface clutter (Guo et al., 2023), this study only uses
125 RWP measurements taken at altitudes above 0.3 km and excludes data affected by
126 precipitation.

127 **2.2 Meteorological data**

128 This study integrates three types of meteorological data, including ground-based
129 in-situ observations, FY-4A satellite cloud products, and ERA5 reanalysis data, to
130 match its core objectives of characterizing summertime ϵ within and above the PBL.
131 These datasets are specifically selected to quantify the individual and combined
132 modulations of VWS, static stability, and cloud conditions on turbulent dynamics.

133 Hourly near-surface precipitation and sea level pressure observations were
134 collected from conventional meteorological stations collocated with the 132 RWP sites.
135 All ground measurements were strictly quality-controlled by the China Meteorological



Administration (CMA) through extreme value verification, spatiotemporal consistency checks, and abnormal data removal. To quantify cloud modulation on PBL turbulence, this study employs 15-minute cloud fraction (CF) and cloud top temperature (CTT) products from the FY-4A geostationary satellite (Zhang et al., 2019). The two parameters effectively represent cloud coverage and developmental intensity, enabling an assessment of cloud-induced turbulence variations within and above the PBL. Given the mismatched temporal resolution between satellite cloud products and hourly RWP data, cubic spline temporal interpolation was used to resample the original 15-minute cloud datasets to hourly intervals. This temporal unification guarantees synchronous variations between cloud evolution and turbulent activity, supporting quantitative evaluation of cloud amplification effects on turbulence dissipation rate.

Hourly ERA5 reanalysis data (Hersbach et al., 2023) were utilized to characterize vertical atmospheric dynamic and thermodynamic structures for turbulence mechanism analysis. The high-resolution vertical temperature and horizontal wind profiles from ERA5 are critical for calculating static stability and VWS, the two key drivers of turbulence variability. Unified hourly preprocessing of ground, satellite, and reanalysis data establishes a consistent dataset framework, facilitating layered comparison of turbulence control mechanisms and spectral analysis of vertical wind perturbations across different atmospheric layers.

2.3 Estimations of ϵ , PBL height, vertical wind shear and gradient Richardson number

Here, the PBL height (PBLH, z_i) is calculated using the improved threshold method (ITM) proposed by Liu et al., (2019). This method begins by normalizing the SNR profiles obtained from the RWP dataset to account for variations across different time periods and measurement conditions. Next, based on the diurnal variations in atmospheric turbulence, distinct thresholds for the normalized SNR are established for daytime and nighttime: 0.75 for daytime and 0.90 for nighttime. Starting from the top of the normalized SNR profile, the algorithm searched downwards the ground,



164 identifying the first height at which the SNR value exceeds the corresponding threshold.
165 This height is then designated as z_i .

166 Vertical wind shear (VWS) is calculated based on RWP measurements of wind
167 components at different heights (Markowski et al., 2006), which is shown as follows:

168
$$VWS = \left[\left(\frac{u_2 - u_1}{z_2 - z_1} \right)^2 + \left(\frac{v_2 - v_1}{z_2 - z_1} \right)^2 \right]^{1/2} \quad (1)$$

169 where u_1 and u_2 represent the values of low- and high-level zonal wind speed
170 components, respectively, while v_1 and v_2 represent the values of low- and high-level
171 meridional wind speed components, respectively. The formula calculates the change in
172 wind speed vector (including wind speed and direction) between two different heights.

173 Following previous studies (i.e., Stull, 1988; Vogelezang et al., 1996), the gradient
174 Richardson number (Ri) can be derived from the following formula:

175
$$Ri = \frac{g}{\theta_v} \frac{\partial \theta_v / \partial z}{\left(\frac{\partial u}{\partial z} \right)^2 + \left(\frac{\partial v}{\partial z} \right)^2} \quad (2)$$

176 where θ_v is the virtual potential temperature, u and v is the zonal and meridional
177 wind speed components, and z is the height from ERA5.

178 **2.4 Estimation of slopes in the energy-containing range and transition region**

179 The energy-containing range was defined using three criteria, with its upper
180 bound idx_{ee} taken as the maximum of: (i) the bin where the cumulative spectral energy
181 first reached 55% of the total energy, based on trapezoidal integration; (ii) the first bin
182 beyond the spectral peak where the local slope dropped below -0.5 ; and (iii) five bins
183 above the spectral peak as a minimum-span constraint. Local slopes were estimated
184 from a moving-window linear regression of $\log_{10}(S)$ on $\log_{10}(f)$ using a ± 3 -bin
185 window. The slope in the energy-containing range was then obtained from a linear fit
186 to the $\log_{10}(S)$ – $\log_{10}(f)$ relationship between the spectral peak and idx_{ee} .

187 The transition region was defined as the interval from the bin immediately
188 following the end of the energy-containing range, i.e. $idx_{ee} + 1$, to the highest-
189 resolution bin corresponding to a 6 min period (see Fig. 2 for details).



190 **3 Results and discussion**

191 **3.1 Spatiotemporal variability of turbulent**

192 Summer turbulence dissipation rates exhibit pronounced spatial heterogeneity
 193 across China (Fig. 1). Overall, turbulence is stronger in Northwest China and North
 194 China, with mean $\log_{10}\epsilon$ values of -1.64 and -2.29 , respectively. It is weaker over the
 195 Tibetan Plateau and Southwest China, with mean values of -3.78 and -3.39 . These
 196 results indicate a clear regional gradient in summer turbulent dissipation rate. From a
 197 spatial perspective, turbulence is relatively stronger in the mid-latitude belt ($30^{\circ}\text{--}40^{\circ}\text{N}$),
 198 whereas it is weaker at low latitudes ($<25^{\circ}\text{N}$) and high latitudes ($>40^{\circ}\text{N}$).

199 In terms of vertical structure, regions with stronger turbulence generally
 200 correspond to larger vertical gradients (Fig. 1c), and the larger the amplitude, the more
 201 negative the gradient (Fig. 3a), indicating a more pronounced difference between
 202 turbulence within and above the PBL in high-turbulence regions. The national mean
 203 vertical gradient is -29.2% , suggesting that turbulence aloft is generally weaker than
 204 that within the PBL. The spatial correlation coefficient between turbulence within and
 205 above the PBL is high ($r = 0.950$), indicating strong consistency in their spatial
 206 distribution.

207 In addition to spatial heterogeneity, summertime turbulence shows a marked
 208 diurnal cycle (Fig. 3b). Both turbulence within and above the PBL reach their maxima
 209 at 1300 LT (local time), with peak values of -2.10 and -2.72 , respectively. The vertical
 210 gradient also exhibits a clear diurnal variation: stratification is strongest at 1200 LT ($-$
 211 31.57%) and weakest at 2300 LT (-24.17%), indicating a transition from strong
 212 stratification toward enhanced mixing over the course of the day (Fig. 3c).

213 Further analysis from the perspectives of temporal continuity and spatial scale
 214 shows that the lagged cross-correlation between turbulence within and above the PBL
 215 peaks at lag = 0, demonstrating synchronous variability (Fig. 4a). The lag-1-day
 216 autocorrelation of daily mean turbulence is $r = 0.85$, and the time scale at which the



217 correlation decays to $1/e$ is approximately 9 days (Fig. 4b), indicating strong day-to-
218 day persistence. Spatial autocorrelation analysis further shows that turbulence
219 correlation decreases with distance (Fig. 4c), with a spatial decorrelation length of
220 several degrees of latitude and longitude, implying that turbulence maintains a degree
221 of spatial continuity at the mesoscale.

222 **3.2 Influences of topography and land-cover on turbulence**

223 To further investigate the relationships between topography, land cover, and
224 turbulence, we compared afternoon values with daily mean values by
225 computing $\Delta \log_{10} \epsilon$ to characterize the change in turbulence during the afternoon (1200–
226 1800 LT) relative to the daily mean. To exclude the influence of synoptic weather
227 systems, days affected by large-scale weather systems were first removed based on 3-
228 hour pressure changes, and only rain-free days were retained, thereby forming a quasi-
229 stationary sample set. Figure 5 shows the spatial distribution of $\Delta \log_{10} \epsilon$ within and
230 above the PBL. The results indicate that turbulence generally intensifies in the
231 afternoon, but the magnitude of the increase varies spatially. $\Delta \log_{10} \epsilon$ is higher in
232 Northwest China and lower in the Yangtze River basin, with other climate zones falling
233 in between.

234 This difference may be related to variations in topographic conditions and surface
235 characteristics across regions. Northwest China has relatively pronounced surface relief
236 and more rough underlying surfaces, which may enhance mechanical turbulence
237 development in the near-surface layer. By contrast, the Yangtze River basin and parts
238 of eastern China have more heterogeneous underlying surfaces, and grassland, water
239 bodies, and croplands differ in thermal properties and surface roughness, which may
240 jointly modulate the magnitude of afternoon turbulence enhancement. Increased surface
241 roughness favors momentum exchange and turbulence generation, whereas thermal
242 contrasts affect sensible heat transport and boundary-layer mixing intensity. Both may
243 therefore contribute to the spatial distribution of $\Delta \log_{10} \epsilon$.



244 Furthermore, there is a significant positive correlation between turbulence and
245 terrain complexity, and this relationship exists both within and above the PBL, but is
246 stronger within the PBL (Fig. 5c-d). This indicates a robust link between topographic
247 relief and turbulence enhancement, which is more readily manifested within the
248 boundary layer.

249 In addition to topography, turbulence is modulated by surface type, with distinct
250 behavior arising across different land-cover classes (Fig. 6). Above the
251 PBL, $\Delta \log_{10} \epsilon$ remains generally weak across all land-cover types. Within the
252 PBL, $\Delta \log_{10} \epsilon$ is enhanced for all land-cover types and the differences become more
253 pronounced. This suggests that the modulation of turbulence by land cover is stronger
254 within the PBL, and that surface conditions exert a more direct influence on near-
255 surface turbulence development.

256 Overall, both topography and land cover affect the spatial distribution of summer
257 turbulence, with this influence being more pronounced within the PBL. Higher surface
258 roughness, complex topography, and stronger heterogeneity of underlying surfaces in
259 Northwest China may jointly promote afternoon turbulence enhancement, whereas the
260 weaker relief and more homogeneous surface conditions over the Yangtze River basin
261 and some plain areas may limit the magnitude of this enhancement.

262 **3.3 Cloud-driven modulation of turbulence**

263 To quantify cloud impacts on ϵ within and above the PBL, we analyze ϵ under
264 three cloud regimes: 1) Clear-sky (daily CF <20%) vs. cloudy-sky (CF>80%); and 2)
265 CTT change rates ($dCTT/dt > 1^\circ\text{C h}^{-1}$ vs. $< 0.5^\circ\text{C h}^{-1}$). Figure 7 shows the diurnal
266 variation of the vertical profile of ϵ under clear- and cloudy-skies. Overall, ϵ exhibits a
267 consistent diurnal pattern, with ϵ peaking near 1400 Local Standard Time
268 (LST), coinciding with maximum surface solar heating. By comparison, ϵ reaches the
269 minimum at about 0400 LST, when surface cooling stabilizes the atmosphere. This
270 tends to form a nocturnal inversion layer that suppresses vertical mixing, leading to
271 minimal ϵ values. Vertical profiles of ϵ further reveal a monotonic decrease with



272 altitude across all conditions, which suggests that the land surface plays a significant
 273 role in the dissipation process in the vertical direction. This is also consistent with well-
 274 established notion of turbulence decaying quickly above the PBL under normal
 275 conditions (Stull, 1988).

276 Further comparison revealed that ε under cloudy sky was slightly stronger overall
 277 than that under clear sky, with an average ε of $-2.78 \text{ m}^2\text{s}^{-3}$ compared to $-2.94 \text{ m}^2\text{s}^{-3}$
 278 under clear sky (Fig. 7). This indicates that increased cloud cover is conducive to
 279 maintaining a higher level of turbulence activity, especially in summer when high
 280 clouds and enhanced surface convection often occur simultaneously (Yang et al., 2020).
 281 Further analysis of ε at different heights (Fig. 8) revealed that cloud cover produces
 282 stronger enhancement of ε above the PBL ($\Delta\log_{10}\varepsilon = 0.16 \text{ m}^2 \text{ s}^{-3}$) than inside the PBL
 283 ($\Delta\log_{10}\varepsilon = 0.1 \text{ m}^2 \text{ s}^{-3}$). This vertical contrast arises mainly from intensified vertical wind
 284 shear (Wang et al., 2024).

285 Given the complexity of cloud modulation of turbulence, we further conduct
 286 synergic analysis of cloud-driven turbulence variability. The results show that a rapid
 287 CTT declines ($d\text{CTT}/dt > 1^\circ\text{C h}^{-1}$), indicative of cloud growth, amplify ε below the
 288 PBL by 17% (Fig. 9). This likely stems from radiative cooling at cloud tops
 289 destabilizing the lower atmosphere, whereas stable stratification above the PBL mutes
 290 this effect.

291 These results underscore that cloud impacts on ε are height- and mechanism-
 292 dependent, necessitating disentanglement of microphysical (e.g., shading) and dynamic
 293 (e.g., shear) drivers. In Section 3.3, these linkages will be explicitly explored through
 294 meteorological factor analysis.

295 **3.4 Thermodynamic and dynamic controls on turbulence across vertical layers**

296 ε within and above the PBL is governed by both thermodynamic and dynamic
 297 processes. Near-surface thermal instability favors the development of buoyancy-driven
 298 convection, while VWS generates mechanical mixing (Kim et al., 2011; Zhang et al.,
 299 2022; Chechin et al., 2023). To disentangle these drivers, we analyze the variability of



300 ϵ using Ri as a proxy for atmospheric stability (Fig. 10a) and VWS profiles (Fig. 10b–
301 c) as dynamic indicators.

302 In the upper and lower terciles of Ri , the vertically resolved ϵ profiles indicate
303 that near-surface buoyancy mainly enhances turbulence within the PBL (Fig. 10a).
304 Under unstable stratification, buoyancy promotes momentum and heat exchange (Wang
305 et al., 2022), thereby strengthening convective updrafts and turbulence generation (Fig.
306 11a-b). Above the PBL, this thermodynamic influence weakens, consistent with the
307 reduced coupling to surface fluxes.

308 As far as the dynamic forcing concerns, VWS exerts height-dependent effects (Fig.
309 10b-c). Low-level shear ($z < z_i$) moderately enhances ϵ throughout the PBL, likely via
310 mechanical mixing from surface drag. By comparison, upper-level shear ($z_i - 3z_i$) tends
311 to drive ϵ to increase above the PBL, as shear-induced Kelvin-Helmholtz instabilities
312 overcome stable stratification.

313 Composite analysis confirms that both Ri (thermodynamic) and wind shear
314 (dynamic) contributes positively to ϵ at all heights (Fig. 12). However, their dominance
315 shifts vertically. Within the PBL, turbulence is primarily thermally driven (regression
316 coefficient = 0.012 between ϵ and Ri). In contrast, above the PBL, dynamical forcing
317 prevails, with regression coefficient = 0.0095 between ϵ and upper-level shear.

318 These results align with conceptual models where surface heating initiates
319 turbulence, while shear sustains it aloft (Stull 1988; Zhang et al., 2022). The weaker
320 low-level shear impact suggests overlapping mechanical and convective mixing near
321 the surface.

322 **3.5 Power spectral characteristics of vertical wind disturbances**

323 Spectral analysis of wind perturbations provides critical insights into turbulent
324 dynamic processes, particularly within the PBL. While frequency and wavenumber
325 spectral methods have been extensively adopted in turbulence investigations (Larsén et
326 al., 2016; Rios et al., 2023), the power spectrum of vertical wind speed (ω) is governed
327 by intricate synergies among atmospheric stratification, vertical wind shear, and surface



328 turbulent fluxes. To disentangle these intertwined modulating effects, we conduct high-
 329 frequency profile spectral analysis, which enables continuous temporal observations
 330 and direct quantification of turbulent mixing dynamics across the PBL (Wood et al.,
 331 2010; Sun et al., 2012).

332 Spectral analysis of vertical wind perturbations reveals distinguishable spectral
 333 regimes below and above the PBL height under diverse stability and shear conditions
 334 (Fig. 13). Across all examined cases, the ω power spectrum exhibits a characteristic
 335 camel-hump-shaped distribution with pronounced peaks at intermediate frequencies.
 336 The dominant peak frequency (f) ranges from 0.00056 to 0.0008, occupying
 337 approximately 39.6–57.1 % of the full frequency spectrum.

338 No matter it is under stable conditions (Fig. 13b) or strong wind shear (Fig. 13d,
 339 f), turbulence shifts toward lower frequencies (longer wavelengths) with increasing
 340 altitude, indicating dominance of small-scale eddies within the PBL. This transition
 341 coincides with enhanced vertical stratification, which decouples the lower and upper
 342 PBL. During the stable atmospheric period, the existence of the inversion layer
 343 effectively suppresses vertical mixing and heat exchange between the upper and lower
 344 atmospheric layers. Compared to the convective boundary layer, turbulent activity
 345 within the stable boundary layer is significantly reduced, which implies a corresponding
 346 decrease in momentum and material exchange between the layers, further exacerbating
 347 the decoupling phenomenon between them. Moreover, the enhancement of wind shear
 348 weakens momentum exchange between the layers and strengthens the structure of the
 349 stable layer, making it difficult for effective vertical mixing to occur between
 350 atmospheric layers at different heights, ultimately leading to decoupling.

351 In shear-dominated regimes (e.g., $z_i < z < 3z_i$), decoupling intensifies (Fig. 13d,
 352 f), consistent with the shear-instability-driven turbulence identified in Fig. 11. This
 353 aligns with Sun et al. (2012), who attributed such behavior to shear-induced overturning
 354 eddies generating local thermal instabilities. However, our results indicate that the



355 effect of high-level wind shear on decoupling is greater than that of low-level wind
356 shear.

357 To further elucidate the spectral morphology and energy cascade characteristics,
358 the slopes in the energy-containing and transition ranges are examined. The spectral
359 exhibit a clear scale-dependent response of boundary-layer turbulence to stratification
360 and shear (Fig. 14). In the energy-containing range, the slopes remain relatively weak,
361 ranging from approximately -0.846 to -0.705, indicating that the dynamics are still
362 governed primarily by large-scale organized motions. By contrast, the transition range
363 exhibits a pronounced steepening, with slope values ranging from approximately -1.665
364 to -1.483, approaching the classical Kolmogorov inertial-subrange scaling of $-5/3$,
365 which indicates a clearly enhanced energy cascade in this interval. Across the different
366 Ri regimes, increasing stratification tends to suppress this transfer, leading to a flatter
367 spectral structure and indicating weakened turbulence mixing and reduced cascade
368 efficiency. Shear exerts the opposite tendency: it enhances turbulent production and
369 facilitates the redistribution of energy toward smaller scales, thereby driving the
370 transition range closer to the classical inertial-subrange behavior. These results suggest
371 that the observed spectral evolution is controlled by the competition between stabilizing
372 buoyancy effects and shear-induced turbulence generation. Notably, the transition
373 range appears to be the most sensitive part of the spectrum to this competition, whereas
374 the energy-containing range responds more moderately, reflecting the persistent
375 dominance of large-scale motions in that scale interval.

376 Overall, despite the pronounced meteorological variability, the RWP spectra
377 consistently capture the transition toward the Kolmogorov $-5/3$ scaling across heights
378 and conditions. This indicates that, even though the 6-min temporal resolution is not
379 sufficient to fully resolve the inertial-subrange frequencies, the RWP still provides a
380 robust depiction of the larger-scale signatures of turbulent energy transfer relevant to
381 PBL mixing. Compared with the 1-s resolution recommended by Rios et al. (2023) for
382 coastal sites, these results highlight the advantage of the RWP in reliably characterizing



383 turbulence structure under varying atmospheric conditions, while also showing that the
384 temporal sampling requirement may depend on terrain, site characteristics, and
385 instrument response.

386 4. Summary and concluding remarks

387 This study investigates the nationwide spatiotemporal variability, as well as
388 multiple modulating drivers of summertime boundary-layer turbulence dissipation rate
389 (ϵ) across China, based on high-resolution summertime measurements from a national
390 network of 132 radar wind profiler (RWP) stations.

391 Turbulent dissipation rate exhibits pronounced regional and diurnal variability.
392 Turbulence peaks in the early-afternoon across most sites over China, consistent with
393 the maximum surface solar heating, and ϵ generally decreases with increasing height.
394 The spatial pattern of ϵ shows strong heterogeneity: the largest afternoon-mean $\Delta \log_{10} \epsilon$
395 is found over Northwest China ($0.18 \text{ m}^2 \text{ s}^{-3}$), whereas the Yangtze River basin shows
396 the weakest signal ($0.04 \text{ m}^2 \text{ s}^{-3}$).

397 Our analyses highlight important influences of topography and land-cover on
398 turbulence. Greater topographic complexity is positively correlated with enhanced
399 turbulence, and this correlation is stronger within the PBL. Large differences in
400 afternoon turbulence amplification are also observed across land-cover types: desert
401 and grassland surfaces favour stronger turbulence enhancement, while water bodies
402 suppress turbulent activity. Such surface-related effects decay with height but remain
403 detectable even above the PBL.

404 Cloud fraction exerts an additional modulating effect on ϵ . Turbulence is stronger
405 under cloudy-sky compared with clear-sky conditions, and this cloud-related
406 enhancement is more prominent above the PBL, which is largely associated with
407 amplified vertical wind shear accompanying cloud development.

408 Thermodynamic and dynamic forcing exert vertically differentiated controls on ϵ .
409 Within the PBL, turbulence is jointly regulated by buoyancy-driven thermal instability
410 and mechanical vertical-wind-shear. Above the PBL, turbulence is predominantly



411 sustained by shear-generated disturbances, which can induce atmospheric decoupling
412 between lower and upper layers under strong-shear conditions. Such vertical
413 decoupling is further supported by spectral analysis of vertical-wind perturbations.
414 Spectral slopes in the energy-containing and transition ranges reveal competing effects
415 between static stability (suppressing energy cascade) and wind shear (promoting energy
416 transfer toward smaller eddy scales). The core modulating factors and their influential
417 pathways are conceptually summarized in Fig.15.

418 Despite these observational advances, several knowledge gaps remain regarding
419 surface-roughness effects, fine-scale topographic forcing, and aerosol-cloud-turbulence
420 interactions. Future work should incorporate high-resolution terrain datasets to better
421 quantify mechanically-generated turbulence, and large-eddy-simulation (LES)
422 experiments are required to disentangle topographic versus thermodynamic
423 contributions.

424 These nationwide observational findings provide useful observational constraints
425 for improving boundary-layer-turbulence parameterization schemes in weather and
426 climate models, especially for regions subject to frequent extreme-weather impacts
427 under ongoing global change.

428

429 **Data Availability**

430 The turbulence retrievals from three RWPs in China are accessible from
431 <https://doi.org/10.5281/zenodo.15334907> (Accessed on 04-05-2025), land cover types
432 were obtained from the MODIS Land Cover Type Yearly Climate Modeling Grid
433 product (MCD12C1; [https://ldsweb.modaps.eosdis.nasa.gov/missions-and-](https://ldsweb.modaps.eosdis.nasa.gov/missions-and-measurements/products/MCD12C1)
434 [measurements/products/MCD12C1](https://ldsweb.modaps.eosdis.nasa.gov/missions-and-measurements/products/MCD12C1)), and the ERA5 reanalysis data used here
435 (Hersbach et al., 2023) can be acquired via Copernicus Climate Change Service (C3S)
436 CDS <https://doi.org/10.24381/cds.bd0915c6> (Accessed on 21-01-2025).



437 **Author contributions**

438 JG and YS designed the research framework and conceptualized this study; TX and
 439 JG conducted the experiment and drafted the initial manuscript; YS, TC, XG, HY and
 440 ZZ helped the data collection from the meteorological tower and carried out data quality
 441 control; TX, TC, HX participated in the results analysis and discussions. All authors
 442 contributed to the revision of the manuscript.

443 **Competing interests**

444 The contact author has declared that none of the authors has any competing
 445 interests.

446 **Acknowledgments**

447 This work was under the auspice of the National Natural Science Foundation of
 448 China under grants 42325501, 42505002 and 42561160141, Henan Provincial Major
 449 Industrial Key Technology Tackling Project under grant 261000320600, the
 450 Department of Science and Technology of Guizhou province under grant
 451 KXJZ[2024]033, and the Chinese Academy of Meteorological Sciences under grants
 452 2024Z003, 2026Z001 and 2026Z004.

453 **References**

- 454 Chechin, D. G., Lüpkes, C., Hartmann, J., Ehrlich, A., and Wendisch, M.: Turbulent
 455 structure of the Arctic boundary layer in early summer driven by stability, wind
 456 shear and cloud-top radiative cooling: ACLOUD airborne observations, *Atmos.*
 457 *Chem. Phys.*, 23, 4685–4707, <https://doi.org/10.5194/acp-23-4685-2023>, 2023.
- 458 Cho, J. Y. N., Newell, R. E., Anderson, B. E., Barrick, J. D. W., & Thornhill, K. L.
 459 (2003), Characterizations of tropospheric turbulence and stability layers from
 460 aircraft observations. *Journal of Geophysical Research: Atmospheres*, **108**(D20):
 461 8784. <https://doi.org/10.1029/2002JD002820>



- 462 Cimini, D., Haeffelin, M., Kotthaus, S., Löhnert, U., Martinet, P., O'connor, E., Walden,
463 C., Coen, M.C., & Preissler, J. (2020), Towards the profiling of the atmospheric
464 boundary layer at European scale—Introducing the COST Action PROBE.
465 *Bulletin of Atmospheric Science and Technology*, **1**: 23–42
- 466 Feigenwinter, C., Vogt, R., & Parlow, E. (1999), Vertical Structure of Selected
467 Turbulence Characteristics above an Urban Canopy. *Theoretical and Applied*
468 *Climatology*, **62**(1-2): 51–63. [https://doi.org/ 10.1007/s007040050074](https://doi.org/10.1007/s007040050074)
- 469 Frisch, U. (1995), Turbulence: The legacy of AN Kolmogorov, Cambridge University.
470 Press, Cambridge.
- 471 Fu, P., Zhong, S., Huang, J., & Song, G. (2012), An observational study of aerosol and
472 turbulence properties during dust storms in northwest China, *Journal of*
473 *Geophysical Research: Atmospheres*, **117**, D09202.
474 <https://doi.org/10.1029/2011JD016696>
- 475 Guo, J. P., Chen, X. Y., Su, T. N., Liu, L., Zheng, Y. T., Chen, D. D., Li, J., Xu, H., Lv,
476 Y. M., He, B. F., Li, Y., Hu, X. M., Ding, A. J., & Zhai, P. M. (2020). The
477 Climatology of Lower Tropospheric Temperature Inversions in China from
478 Radiosonde Measurements: Roles of Black Carbon, Local Meteorology, and
479 Large-Scale Subsidence. *Journal of Climate*, **33**(21), 9327–9350,
480 <https://doi.org/10.1175/JCLI-D-19-0278.1>
- 481 Guo, J., Liu, B., Gong, W., Shi, L., Zhang, Y., Ma, Y., Zhang, J., Chen, T., Bai, K.,
482 Stoffelen, A., de Leeuw, G., & Xu, X. (2021). Technical note: First comparison of
483 wind observations from ESA's satellite mission Aeolus and ground-based radar
484 wind profiler network of China. *Atmospheric Chemistry and Physics*, **21**, 2945–
485 2958, <https://doi.org/10.5194/acp-21-2945-2021>.
- 486 Guo, X., Guo, J., Zhang, D., & Yun, Y. (2023), Vertical divergence profiles as detected
487 by two wind-profiler mesonets over East China: Implications for nowcasting
488 convective storms. *Quarterly Journal of the Royal Meteorological Society*, **149**:
489 1629–1649. <https://doi.org/10.1002/qj.4474>



- 490 Harvey, N.J., Hogan, R.J., & Dacre, H. F. (2013), A method to diagnose boundary-
491 layer type using Doppler lidar. *Quarterly Journal of the Royal Meteorological*
492 *Society*, **139**: 1681–1693
- 493 Heilman, W. E. (2023), Atmospheric turbulence and wildland fires: a
494 review. *International Journal of Wildland Fire*, **32**: 476–495.
- 495 Hersbach, H., Bell, B., Berrisford, P., Biavati, G., Horányi, A., Muñoz Sabater, J.,
496 Nicolas, J., Peubey, C., Radu, R., Rozum, I., Schepers, D., Simmons, A., Soci, C.,
497 Dee, D., & Thépaut, J.-N. (2023), ERA5 monthly averaged data on single levels
498 from 1940 to present. Copernicus Climate Change Service (C3S) Climate Data
499 Store (CDS). <https://doi.org/10.24381/cds.fl7050d7>
- 500 Hildebrand, P. H., & Ackerman, B. (1984), Urban Effects on the Convective Boundary
501 Layer. *Journal of The Atmospheric Sciences*, **41**(1): 76–91.
502 [https://doi.org/10.1175/1520-0469\(1984\)041<0076:UEOTCB>2.0.CO;2](https://doi.org/10.1175/1520-0469(1984)041<0076:UEOTCB>2.0.CO;2)
- 503 Kim, J., & Chun, H. (2011), Statistics and Possible Sources of Aviation Turbulence
504 over South Korea. *Journal of Applied Meteorology and Climatology*, **50**: 311–324.
505 <https://doi.org/10.1175/2010JAMC2492.1>
- 506 Larsén, X. G., Larsen, S. E., & Petersen,
507 E. L. (2016), Full-scale spectrum of boundary-layer winds. *Boundary-Layer*
Meteorology, **159**(2): 349–371. <https://doi.org/10.1007/s10546-016-0129-x>
- 508 LeMone, M. (1976), Modulation of turbulence energy by longitudinal rolls in an
509 unstable planetary boundary layer. *Journal of The Atmospheric Sciences*, **33**:
510 1308–1320
- 511 Li, X., & Pu, Z. X. (2023), Dynamic Mechanisms Associated with the Structure and
512 Evolution of Roll Vortices and Coherent Turbulence in the Hurricane Boundary
513 Layer: A Large Eddy Simulation During the Landfall of Hurricane Harvey.
514 *Boundary-Layer Meteorology*, **186**(3): 615–636
- 515 Liu, B. M., Ma, Y. Y., Guo, J. P., Gong, W., Zhang, Y., Mao, F. Y., Li, J., Guo, X. R.,
516 & Shi, Y. F. (2019), Boundary layer heights as derived from ground-based radar



- 517 wind profiler in Beijing. *IEEE Transactions on Geoscience and Remote Sensing*,
 518 **57**: 8095–8104. <https://doi.org/10.1109/tgrs.2019.2918301>
- 519 Markowski, P., & Richardson, Y. (2006), On the classification of vertical wind shear
 520 as directional shear versus speed shear. *Weather Forecast*, **21**: 242–247
- 521 Meng, D. L., Guo, J. P., Guo, X. R., Wang, Y. J., Li, N., Sun, Y. P., Zhang, Z., Tang,
 522 N., Li, H. R., Zhang, F., Tong, B., Xu, H., & Chen, T. M. (2024), Elucidating the
 523 boundary layer turbulence dissipation rate using high-resolution measurements
 524 from a radar wind profiler network over the Tibetan Plateau. *Atmospheric*
 525 *Chemistry and Physics*, **24**: 8703–8720. [https://doi.org/10.5194/acp-24-8703-](https://doi.org/10.5194/acp-24-8703-2024)
 526 2024
- 527 Poulos, G. S., Blumen, W., Fritts, D. C., Lundquist, J. K., Sun, J., Burns, S. P., Nappo,
 528 C., Banta, R., Newsom, R., Cuxart, J., Terradellas, E., Balsley, B., & Jensen, M.
 529 (2002), CASES-99—a comprehensive investigation of the stable nocturnal
 530 boundary layer. *Bulletin of The American Meteorological Society*, **83**: 555–581
- 531 Rios, G., & Ramamurthy, P. (2023), Turbulence in the Mixed Layer Over an Urban
 532 Area: A New York City Case Study, *Boundary-Layer Meteorology*, **188**: 419–440,
 533 2023
- 534 Sharman, R. D., Trier, S. B., Lane, T. P., & Doyle, J. D. (2012), Sources and dynamics
 535 of turbulence in the upper troposphere and lower stratosphere: A
 536 review. *Geophysical Research Letters*, **39**(L12803). <https://doi.org/10.1029/2012GL051996>
- 537 Shi, X. K., Liu, J. W., Li, Y. D., Huang, B., & Tan, Y. Q. (2015), A diagnostic method
 538 for aircraft turbulence based on high-resolution numerical weather prediction
 539 products. *Natural Hazards*, **77**: 867–81
- 540 Solanki, R., Guo, J. P., Lv, Y. M., Zhang, J., Wu, J. Y., Tong, B., & Li, J. (2022),
 541 Elucidating the atmospheric boundary layer turbulence by combining UHF radar
 542 wind profiler and radiosonde measurements over urban area of Beijing. *Urban*
 543 *Climate*, **43**(101151). <https://doi.org/10.1016/j.uclim.2022.101151>



- 545 Sreenivasan, K. R. & Antonia, R. A. (1997), The phenomenology of small scale
546 turbulence. *Annual Review of Fluid Mechanics*, **29**: 435-472.
- 547 Sun, J. L., Mahrt, L., Banta, R. M., & Pichugina, Y. L. (2012), Turbulence Regimes
548 and Turbulence Intermittency in the Stable Boundary Layer during CASES-99.
549 *Journal of the Atmospheric Sciences*, **69**(1): 338-351.
550 <https://doi.org/10.1175/JAS-D-11-082.1>
- 551 Sun, J. L., Lenschow, D. H., LeMone, M. A., & Mahrt, L.: The Role of Large-Coherent-
552 Eddy Transport in the Atmospheric Surface Layer Based on CASES-99
553 Observations. *Boundary-Layer Meteorology*, **160**(1): 83-111.
554 <https://doi.org/10.1007/s10546-016-0134-0>
- 555 Stull, R. B. (1988), Mean Boundary Layer Characteristics, in: An Introduction to
556 boundary layer meteorology, edited by: Stull, R. B., Springer Netherlands.
557 Dordrecht, 1–27. https://doi.org/10.1007/978-94-009-3027-8_1
- 558 Vogelezang, D. H. P., & Holtslag, A. A. M. (1996), Evaluation and model impacts of
559 alternative boundary-layer height formulations. *Boundary-Layer Meteorology*,
560 **81**(3-4): 245-269. <https://doi.org/10.1007/BF02430331>
- 561 Wang, M. Z. & Zhang, J. T. (2022), The relationship among summer atmospheric
562 boundary layer height over the Taklimakan Desert, its land surface parameters and
563 Eurasian circulation. *Atmospheric Science Letters*, **23**(e1122).
564 <https://doi.org/10.1002/asl.1122>
- 565 Wang, Y. J., Zeng, X. B., Xu, X. D., Galarneau Jr, T. J., Li, F., & Zhao, Y. (2024),
566 Power Spectra and Diurnal Variation of Low-Level Horizontal Winds Observed
567 by a Wind Profiler Radar Network Over China. *Geophysical Research Letters*,
568 **51**(3). <https://doi.org/10.1029/2023GL105652>
- 569 Wood, C. R., Lacser, A., Barlow, J. F., Padhra, A., Belcher, S. E., Nemitz, E., Helfter,
570 C., Famulari, D., & Grimmond, C. S. B. (2010), Turbulent flow at 190 m height
571 above London during 2006–2008: a climatology and the applicability of similarity



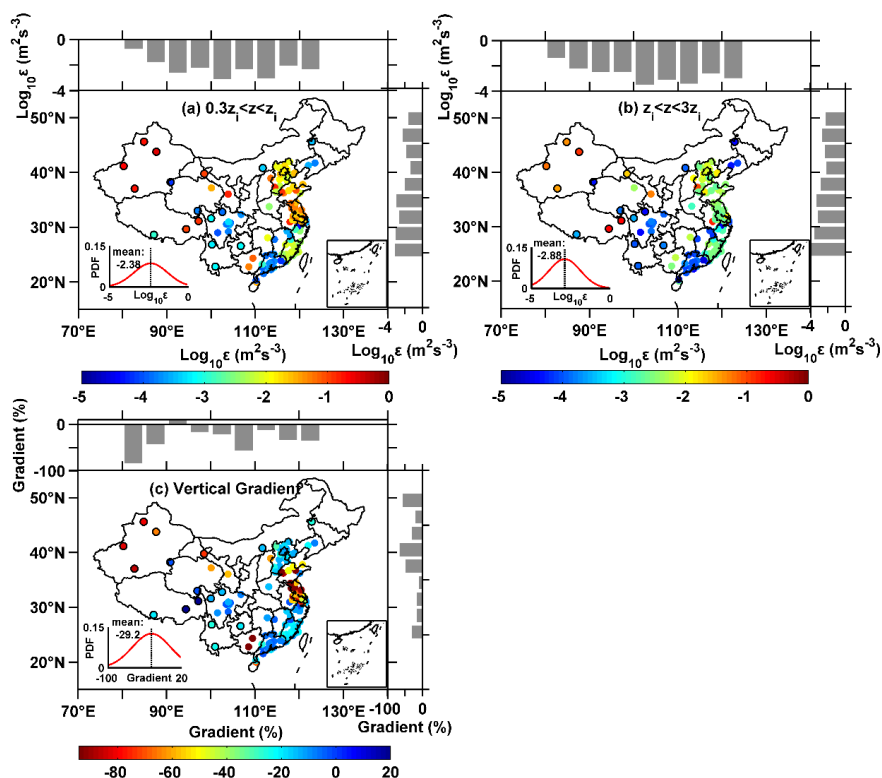
- theory. *Boundary-Layer Meteorology*, **137**(1): 77–
 96. <https://doi.org/10.1007/s10546-010-9516-x>
- Xian, J., Qiu, Z., Luo, H., Hu, Y., Lin, X., Lu, C., Yang, Y., Yang, H., & Zhang, N.
 (2025), Turbulent energy budget analysis based on coherent wind lidar
 observations. *Atmospheric Chemistry and Physics*, **25**, 441–457,
<https://doi.org/10.5194/acp-25-441-2025>.
- Yang, Y. K., Zhao, C. F., & Fan, H. (2020), Spatiotemporal distributions of cloud
 properties over China based on Himawari-8 advanced Himawari imager data.
Atmospheric Research, **240**. <https://doi.org/10.1016/j.atmosres.2020.104927>
- Yang, R. F., Guo, J. P., Deng, W. L., Li, N., Fan, J. H., Meng, D. L., Liu, Z., Sun, Y.
 P., Zhang, G. L., & Liu, L. H. (2023), Investigation of Turbulent Dissipation Rate
 Profiles from Two Radar Wind Profilers at Plateau and Plain Stations in the North
 China Plain. *Remote Sensing*, **15**(16): <https://doi.org/10.3390/rs15164103>
- Zhang, L., Zhang, H. S., Li, Q. H., Wei, W., Cai, X. H., Song, Y., Mamtimin, A., Wang,
 M. Z., Yang, F., Wang, Y., & Zhou, C. L. (2022), Turbulent mechanisms for the
 deep convective boundary layer in the Taklimakan desert. *Geophysical Research
 Letters*, **49**(15). <https://doi.org/10.1029/2022gl099447>
- Zhang, P., L. Zhu, S. Tang, L. Gao, L. Chen, W. Zheng, X. Han, J. Chen, & Shao, J.
 (2019), General comparison of FY-4A/AGRI with other GEO/LEO instruments
 and its potential and challenges in non-meteorological applications. *Frontier in
 Earth Science*, **6** (2019), 224, <https://doi.org/10.3389/feart.2018.00224>.
- Zhang, Q., & Wang, S. (2009), A Study of the Atmospheric Boundary Layer Structure
 During a Clear Day in the Arid Region of Northwest China. *Acta Meteorologica
 Sinica*, **23**(3): 327-337
- Zhang, Z., L. Liu, Y. Shi, K. Zhang, K. Liu, S. Wang, & F. Hu (2023). Analysis of
 urban turbulence intensity observed by Beijing 325-m tower and comparison with
 the IEC turbulence model for small wind turbines. *Journal of Wind Engineering
 and Industrial Aerodynamics*, 241: 105511.



600 Zhou, L. T., & Huang, R. (2014), Regional differences in surface sensible and latent
601 heat fluxes in China. *Theoretical and Applied Climatology*, **116**(3-4): 625-637.
602 <https://doi.org/10.1007/s00704-013-0975-0>
603



604 Figures

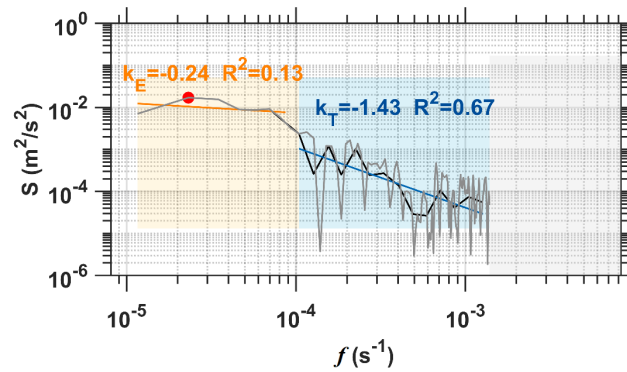


605

606 **Figure 1.** Spatial distribution of the $\log_{10}\epsilon$ (a-b) and (c) vertical gradient of $\log_{10}\epsilon$
 607 as calculated from the RWP measurements across China in summer over China,
 608 derived from radar wind profiler (RWP) observations for heights between 0.3
 609 and 3 times the boundary layer height ($0.3z_i-3z_i$). The horizontal and vertical
 610 bars represent the longitude and latitude variations of $\log_{10}\epsilon$, respectively. The
 611 insets in each panel present the corresponding probability density function (PDF)
 612 of $\log_{10}\epsilon$.
 613



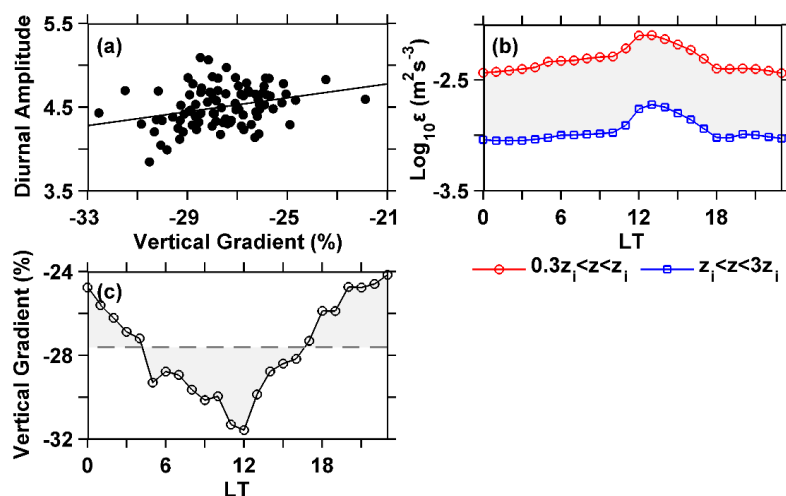
614



615

616 **Figure 2.** Two-regime partition of the turbulent energy spectrum of ω from the
 617 RWP measurements across China. The gray line shows the raw power spectrum,
 618 and the black line shows the log-binned smoothed spectrum (25 bins). The red
 619 circle marks the spectral peak. The yellow and blue shaded areas denote the
 620 energy-containing range and transition region, respectively; the orange and blue
 621 solid lines indicate their linear fits in log space, with k_E and k_T representing the
 622 slopes and R^2 the goodness of fit. The gray shaded area marks the classical
 623 Kolmogorov inertial subrange (10^2 – 10^2 Hz), which is inaccessible here due to
 624 sampling-frequency limitations.

625

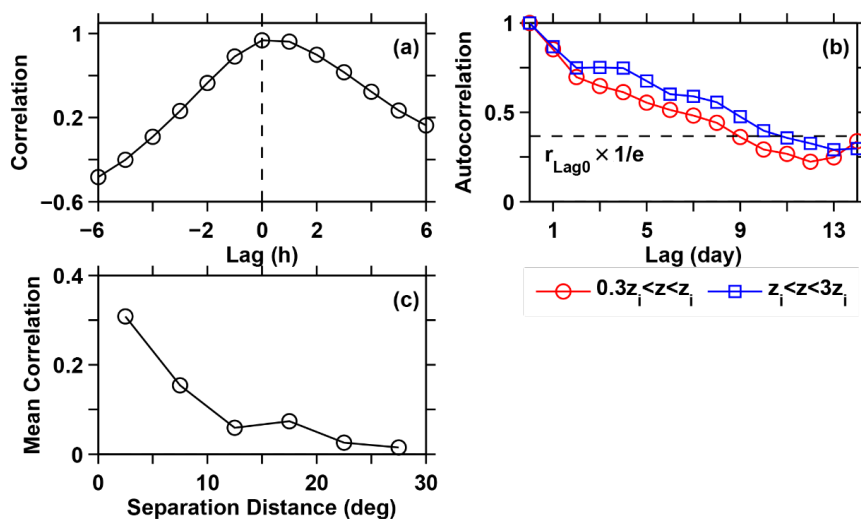


626
 627 **Figure 3.** (a) Scatter plot showing the relationship between vertical gradients and
 628 corresponding amplitudes of wind disturbances; (b) diurnal variations of
 629 turbulence dissipation rate within and above the PBL; (c) diurnal variations in
 630 vertical gradients.

631
 632



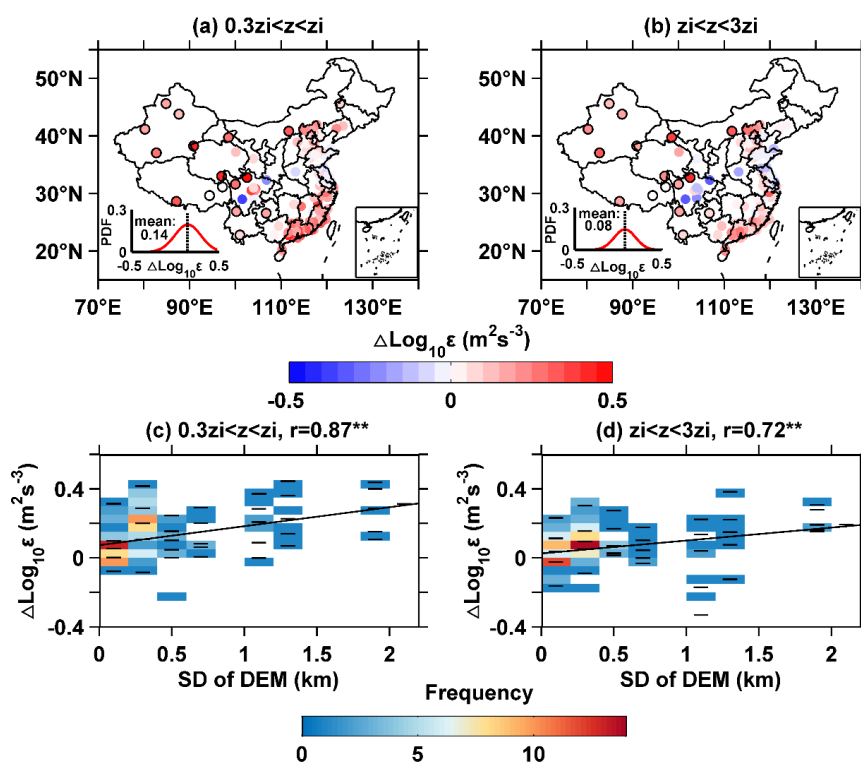
633



634

635 **Figure 4.** (a) Lagged cross-correlation between the diurnal turbulence time series
 636 within and above the PBL, with the dashed vertical line marking the zero-lag peak; (b)
 637 autocorrelation of the nationally averaged daily mean turbulence within ($0.3 z_i < z < z_i$,
 638 red circles) and above ($z_i < z < 3 z_i$, blue squares) the PBL; and (c) mean inter-site
 639 correlation as a function of station separation distance. The dashed horizontal line in (b)
 640 marks the $1/e$ level relative to the lag-0 autocorrelation ($1/e$), and the lag at which the
 641 autocorrelation decays to this level (~ 9 days) gives the e-folding time scale.

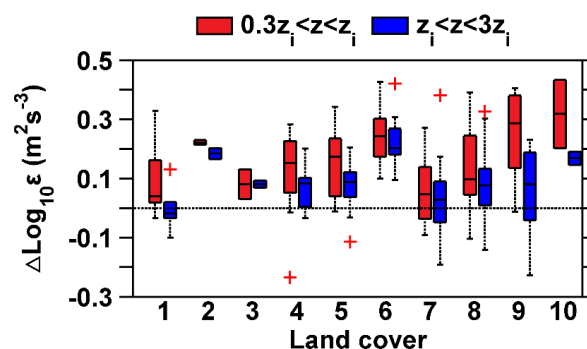
642



643
 644 **Figure 5.** Spatial distribution of mean $\Delta \log_{10} \epsilon$ (afternoon minus all-day) from the
 645 RWP measurements across China within (a) and above (b) the PBL. The insets in each
 646 panel present the corresponding PDF of $\Delta \log_{10} \epsilon$. (c-d) Density scatter plots illustrating
 647 the variations in $\Delta \log_{10} \epsilon$ and DEM standard deviation, with all reported Pearson
 648 correlation coefficients statistically significant at the 1% confidence level.
 649



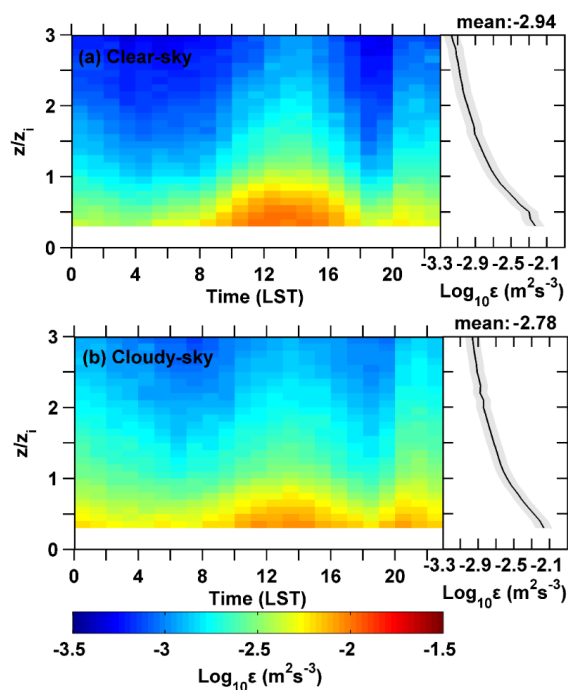
650



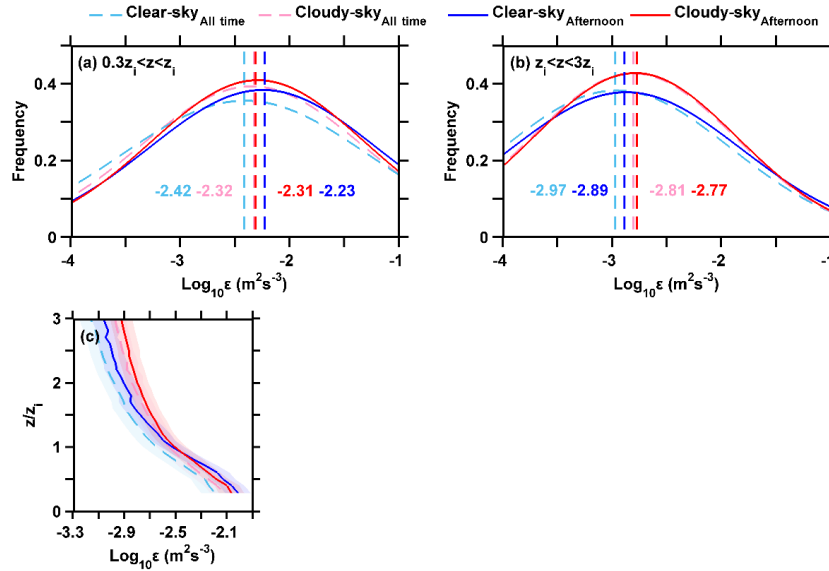
1: Water Bodies 2: Evergreen Broadleaf Forests 3: Deciduous Broadleaf
 Forests 4: Woody Savannas 5: Savannas 6: Grasslands 7: Croplands
 8: Urban and Built-up Lands 9: Cropland/Natural Vegetation Mosaics
 10: Barren

651

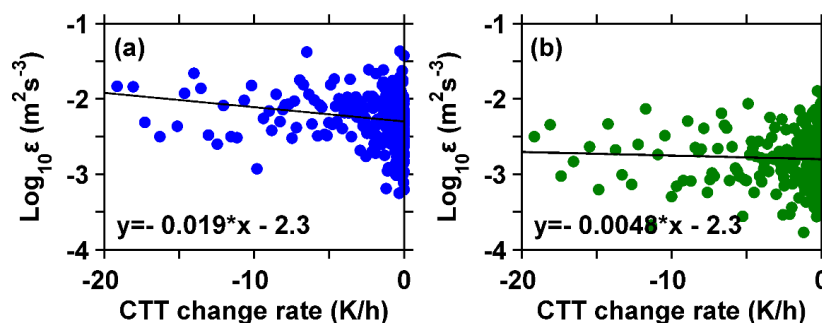
652 **Figure 6.** Box-and-whisker plots of $\Delta \text{Log}_{10} \epsilon$ (afternoon minus daily-mean ϵ , in units of
 653 $\text{m}^2 \text{s}^{-3}$) as a function of land-cover type. Red (blue) boxes denote turbulence within
 654 (above) the PBL, i.e., $0.3 z_i < z < z_i$ ($z_i < z < 3 z_i$). The categories are: 1, water bodies;
 655 2, evergreen broadleaf forests; 3, deciduous broadleaf forests; 4, woody savannas; 5,
 656 savannas; 6, grasslands; 7, croplands; 8, urban and built-up lands; 9, cropland/natural
 657 vegetation mosaics; and 10, barren. The dotted horizontal line marks $\Delta \text{Log}_{10} \epsilon = 0$, and
 658 red crosses indicate outliers.
 659



660
 661 **Figure 7.** Diurnal evolution of $\log_{10} \epsilon$ (color shading; $\text{m}^2 \text{s}^{-3}$) from the RWP
 662 measurements across China as a function of normalized height z/z_i and local solar time
 663 under (a) clear-sky (daily cloud fraction $< 20\%$) and (b) cloudy-sky (cloud fraction $>$
 664 80%) conditions. The side panels show the corresponding mean vertical profiles, with
 665 overall means of -2.94 and $-2.78 \text{ m}^2 \text{s}^{-3}$ for clear- and cloudy-sky conditions,
 666 respectively.
 667



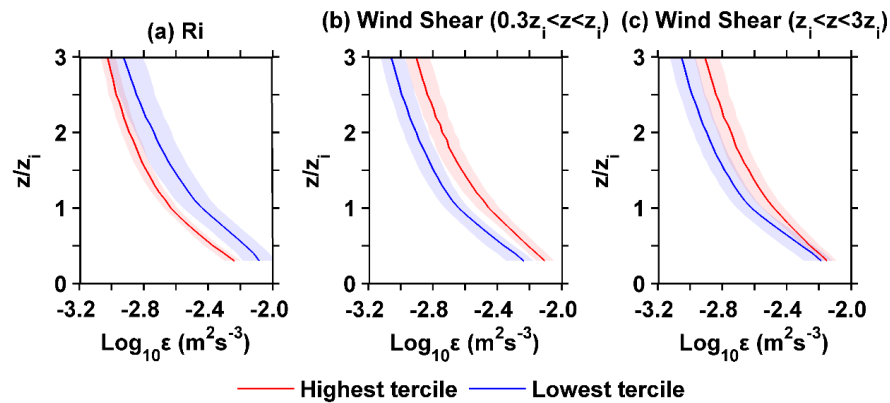
668
 669 **Figure 8.** (a-b) Probability density functions of $\log_{10} \epsilon$ from the RWP
 670 measurements across China for (a) $0.3 z_i < z < z_i$ and (b) $z_i < z < 3 z_i$, for the full day
 671 (dashed lines) and afternoon (solid lines) under clear-sky (blue) and cloudy-sky (red)
 672 conditions. The vertical dashed lines mark the corresponding distribution means, which
 673 are -2.42, -2.32, -2.31, and -2.23 in (a) and -2.97, -2.89, -2.81, and -2.77 in (b),
 674 respectively. Panel (c) shows the corresponding mean vertical profiles of $\log_{10} \epsilon$ ($\text{m}^2 \text{s}^{-3}$).
 675
 676



677
 678 **Figure 9.** Scatterplot of summer afternoon $\log_{10}\epsilon$ (m^2s^{-3}) from the RWP measurements
 679 across China against cloud-top temperature (CTT) change rate (K h^{-1}) (a) within and (b)
 680 above the PBL. Solid lines denote linear least-squares fits, with slopes of -0.019 and
 681 -0.0048 per K h^{-1} for (a) and (b), respectively.
 682



683

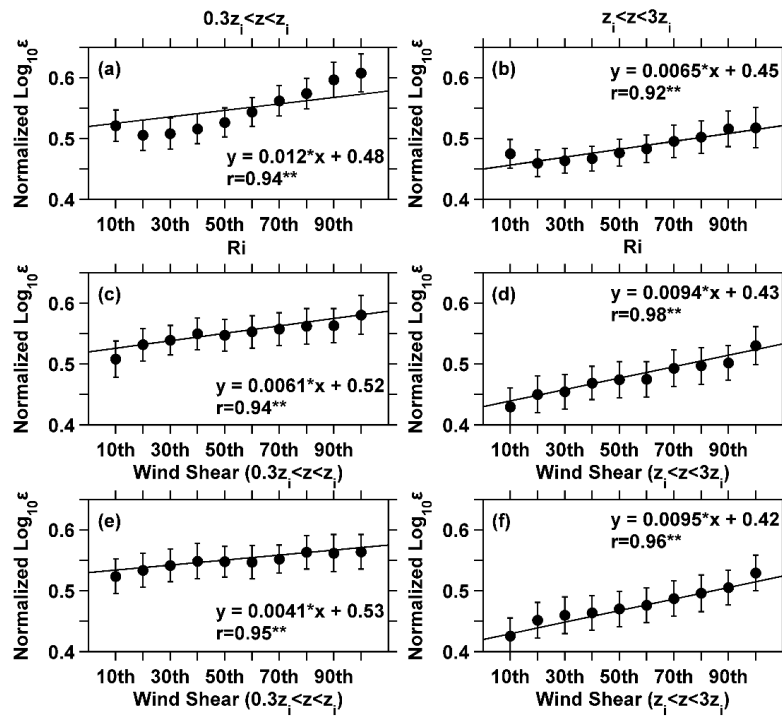


684

685 **Figure 10.** Vertical profiles of summertime afternoon $\log_{10} \epsilon$ ($\text{m}^2 \text{s}^{-3}$) from the RWP
 686 measurements across China, composited for the lowest (blue) and highest (red) terciles
 687 of (a) the gradient Richardson number (Ri), (b) vertical wind shear within the PBL
 688 ($0.3z_i - z_i$), and (c) vertical wind shear above the PBL ($z_i - 3z_i$). Shaded envelopes denote
 689 the corresponding confidence intervals.

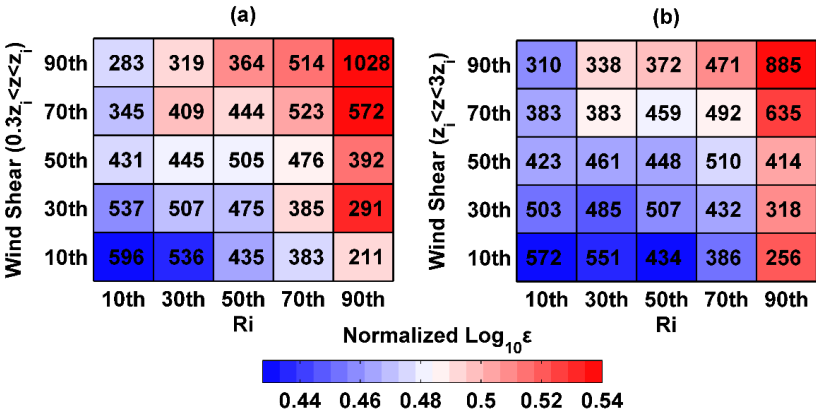


690



691

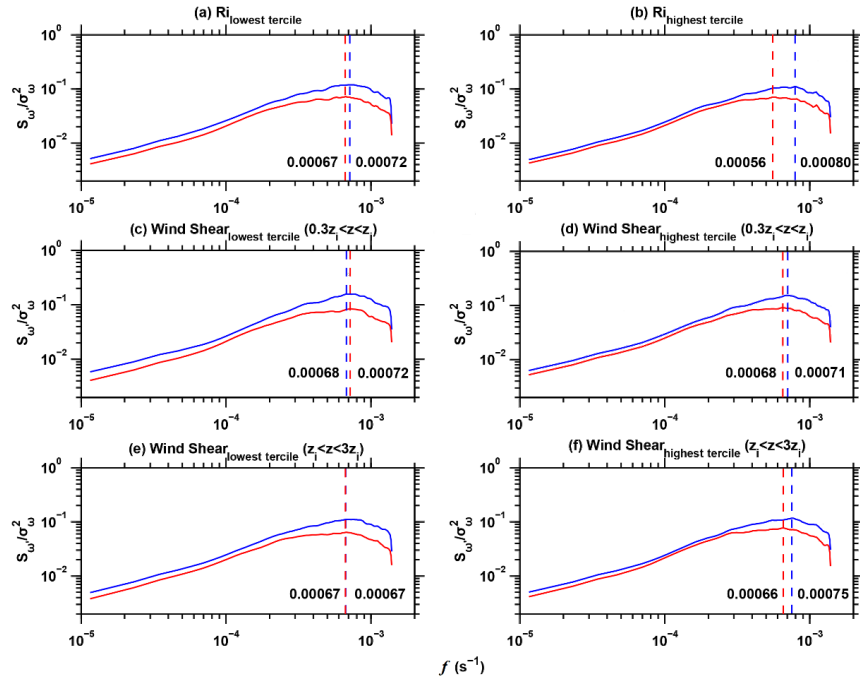
692 **Figure 11.** Relationship among standardized $\log_{10}\epsilon$, Ri (a, b), wind shear within
 693 the PBL (c, d), and wind shear above the PBL (e, f) during summer afternoon at
 694 different percentiles, with Ri sorted in descending order for consistency. The left
 695 column represents conditions within the PBL ($0.3z_i - z_i$), and the right column
 696 represents conditions above the PBL ($z_i - 3z_i$). The equation describes the linear
 697 correlation between $\log_{10}\epsilon$ and the respective variables. The Pearson correlation
 698 coefficient (R) and linear regression line are shown, with ** indicating
 699 significance at the 99% confidence level.



700
701 **Figure 12.** Joint dependence of $\log_{10}\epsilon$ during summer afternoon derived from
702 the RWP measurements across China on (a) wind shear within the PBL ($0.3z_i$ –
703 z_i), wind shear above the PBL (z_i – $3z_i$), and T_s – T_a within the PBL, and (b) above
704 the PBL. The numbers in each cell denote the total number of samples used.
705
706



707

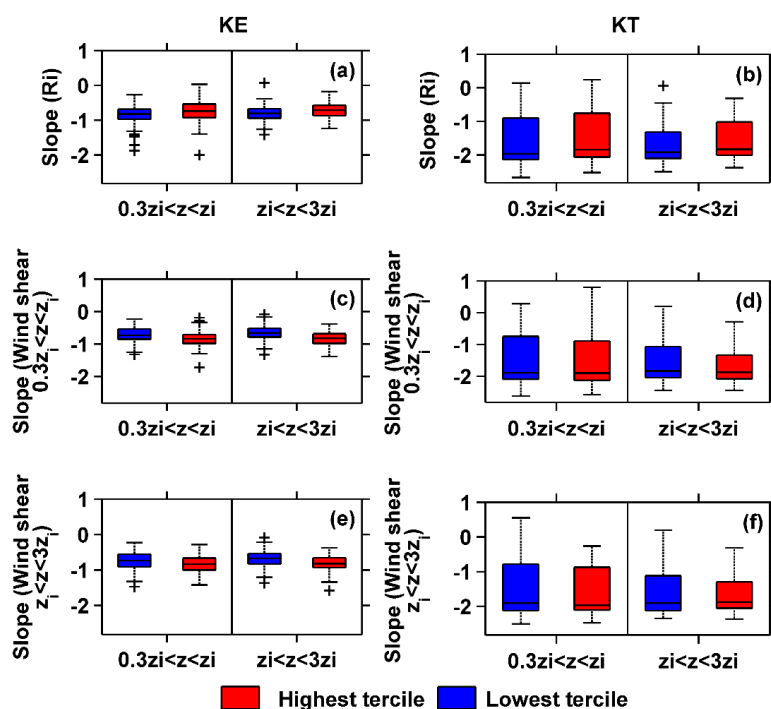


708

709 **Figure 13.** Power spectra of the vertical velocity (S_{ω}), normalized by the w
 710 variance (σ_{ω}^2), as a function of the frequency (f) under different Ri (a-b), wind
 711 shear (c-d) within the PBL ($0.3z_i < z < z_i$), and (e-f) wind shear above the PBL ($z_i -$
 712 $3z_i$) conditions. The left column represents the lowest tercile of each
 713 meteorological factor, and the right column represents the highest tercile. The
 714 blue line represents conditions within the PBL ($0.3z_i - z_i$), the red line represents
 715 conditions above the PBL ($z_i - 3z_i$).

716

717



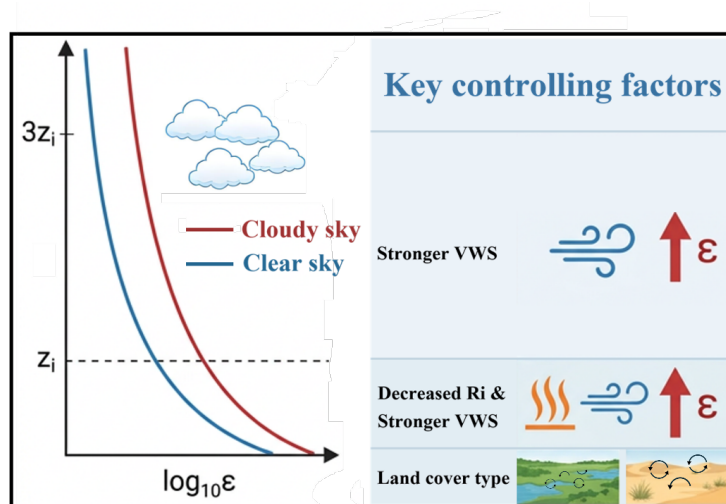
718

719 **Figure 14.** Box plots of slope ranges for the energetic (KE) and transition (KT)
 720 zones within and above the PBL under different atmospheric stability and wind
 721 shear conditions.

722



723



724

725 **Figure 15.** Schematic diagram of turbulence characteristics and controlling
 726 factors within and above the PBL based on nationwide RWP network
 727 observations.