Simulation of the Effect of Small-Scale Mountains on Weather Conditions During the May 2021 Ultramarathon in Gansu Province, China

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Abstract Hypothermia killed 21 participants of the 100 km cross-country Ultramarathon Mountain race in Baiyin, Gansu Province in northwestern China on 22 May 2021, when a large-scale cold front passed by the race site. The hypothermia was caused by low temperatures, high winds, and hail-like precipitation, as reported by survivors. However, there is no meteorological station near the race site, leading to a lack of data on the exact weather conditions during the race. Considering the complex topography of the race site, the weather conditions must be re-built using high-resolution model simulations, so that the tragedy can be investigated.

In this study, the weather conditions in this mountainous area were simulated using the Weather Research and Forecasting model with a 333 m horizontal grid size. The results show that interaction between the mountain wind and orographic precipitation led to a 6.7°C decrease in temperature and a 12 m s⁻¹ increase in wind speed. Precipitation increased owing to orographic accent and convergence, and fell as graupel. High-speed downslope winds were the result of large-amplitude trapped lee waves. The combination of low temperatures, high winds, and spillover graupel on the lee resulted in the blizzard-like weather during the race, thus the hypothermia amongst the runners. The apparent temperature was estimated to be as low as −10°C. Our study shows that mesoscale numerical models with a grid size of hundreds of meters should be used to improve weather forecasts in mountainous regions and prevent future tragedies for special events.

1. Introduction

A mountain ultramarathon has been held in recent years at the Yellow River Stone Forest tourist site in Baiyin, Gansu Province, in northwestern China. The 2021 race route had nine checkpoints (CPs) and was on a west-east-oriented mountain, which is approximately 5.6 km wide, 15.1 km long, and 0.7 km above the closed northern and southern plains (Figure 1). The mountain is the eastern extension of the Qilian Mountains, and is separated from another larger mountain to the east by the Yellow River. On 22 May 2021, the race route experienced unexpected weather, with high-speed winds, a sharp decrease in temperature, and possible graupel precipitation. According to the official survey report from the government of Gansu Province (official report; http://www.gansu.gov.cn/gsszf/c100002/c100010/202106/1643566.shtml), this severe weather resulted in the death of 21 runners due to hypothermia.

The tragedy mainly occurred between the CP2 at the foot of the mountain and CP3 near the mountain crest, between 11:00 and 14:00 Beijing Time (BJT). A brief review of the event was given in the official report and by Zhang et al. (2021). Here it is summarized as: (a) the 2 m temperature decreased by 5–7°C between 08:00 and 13:00 BJT; (b) precipitation began at 10:30 BJT, increased between 11:00 and 12:00 BJT, and decreased after 14:00 BJT; (c) graupel fell at high altitudes on the mountain; and (d) the average wind speed between CP2 and CP3 was approximately 10.8–17.1 m s⁻¹, with gusts of 17.2–24.4 m s⁻¹, indicating a fresh gale (17.2–20.5 m s⁻¹) or strong gale (20.5–24.4 m s⁻¹) event. The official investigation was based on the field damage survey, survivors’ accounts, and empirical linear analysis using surface observations from the surrounding automatic weather stations (AWSs) and the altitude of the CPs. However, plausible empirical estimations of the temperature, wind speed, and precipitation are difficult to linearly estimate because of the nonlinear effects of the mountains. As a result, the official report relied strongly on the field survey and survivor accounts. Alternatively, high resolution simulations can sufficiently resolve the effects of mountains, and are therefore a useful tool to review the event, as well as to help understand the mechanisms of mountain flows on the occurrence of high-speed winds and graupel.
Small- to meso-scale mountains can largely impact multiple atmospheric processes through thermally and

dynamically driven mountain flows, which may further interact with microphysics to influence precipitation.

There have been many in-depth studies on dynamically driven mountain flows, such as large-amplitude gravity

waves (Lilly, 1978; Smith et al., 2007) and associated severe downslope windstorms (Brinkmann, 1974), as well

as orographic precipitation triggered or enhanced by mountains (e.g., Chu & Lin, 2000; Kikuchi & Wang, 2008;

Smith et al., 2003). Among others, the combination of solid precipitation and high-speed winds results in extreme

weather (blizzards), which is challenging to forecast and can result in devastating ecological losses and deaths

(Coleman & Schwartz, 2017; Meyers & Steenburgh, 2013; Mo et al., 2019). These weather conditions are usually

common on very high mountains and sometimes may result in deaths, such as on Mount Everest (Moore &

Semple, 2006, 2011) and Denali (McIntosh et al., 2008). The solid precipitation and high-speed winds can also

occur on low-altitude mountains in unexpected seasons, which can have even more severe impacts, such as the

blizzard-like weather during the Spring 2021 ultramarathon. However, there have only been a few studies on the

Figure 1. Topography around the Yellow River Stone Forest tourist site from 3′′ SRTM digital elevation data. The black rectangle approximately shows the width and length of the mountain, CP1–CP9 are the race checkpoints, and the red line represents the race route.

Small- to meso-scale mountains can largely impact multiple atmospheric processes through thermally and
interaction mechanisms between the solid precipitation and downslope flows. Poulos et al. (2002) and Meyers et al. (2003) determined that devastating downslope winds (DSW) can be caused by large-amplitude lee waves, as the cross-barrier flow in the lower troposphere was strengthened due to reflected mountain waves by a critical upper layer over the mountain. In other studies, it was found that strengthened low-level wind may advect hydrometeors from the windward to leeward of mountains, resulting in the “spillover” phenomenon (Mo et al., 2019; Siler & Durran, 2016). Therefore, it is important to determine if similar processes occurred during the 2021 ultramarathon, and how downslope winds interacted with precipitation.

The large-scale process associated with the 2021 ultramarathon tragedy has been identified as a cold front passing event, which can be predicted by the operational numerical weather prediction (NWP) models (Zhang et al., 2021). However, the meso- and micro-scale processes remain unclear. For the large-scale pattern, a 500 hPa trough and upper-level jet were located over the west side of CP3 in the morning before the tragedy (Figures 2a and 2b). This indicates that updrafts and precipitation was a possibility due to the secondary circulation caused by the jet, and the positive vorticity advection of the trough (Cressman, 1981; Whitney, 1977). Furthermore, between 02:00 and 08:00 BJT, the low-level wind direction turned from westerly to northerly over CP3, and the low-level temperature decreased by 6°C at the 700 hPa level (see position of the 0°C contour in Figures 2c...
and 2d). This indicates that a cold front passed by CP3 just before the tragedy. As a result, the location of the ultramarathon tragedy was dominantly affected by cold air from northerly winds at 700 hPa (Figure 2d), while the upper layer experienced southwesterly wind at 500 hPa (Figure 2b). The wind shear between the two layers may have reflected the upward propagating mountain waves and induced high wind speeds on the lee. However, it is difficult to investigate how the mountain affected the detailed mechanisms during the ultramarathon tragedy. This is because the mountain cannot be explicitly resolved by the reanalysis and the operational regional NWP model with horizontal grid sizes of 3–9 km, as horizontal topographies covering less than 4 grid size are generally filtered out (Wang et al., 2008). Therefore, higher resolution simulations that sufficiently resolve the mountain topography could be useful for investigating the effects of the mountain on the windstorms and precipitation.

In this study, simulations at a grid spacing of 333 m were conducted, with the aim of understanding the impacts of the small-scale mountain on the combination of downslope wind and precipitation, for investigating the weather conditions during the 2021 ultramarathon. This will hopefully help improve weather forecasts and NWP services for events in mountainous regions. The remainder of this paper is organized as follows. The model configuration and data used are presented in Section 2. Results and mechanisms are provided in Sections 3 and 4, respectively, and our conclusions and discussions are given in Section 5.

2. WRF Model Setup and Data

The Weather Research and Forecasting (WRF) version 4.1 is used to simulate weather conditions in three nested domains (D01–D03) with grid-spacings of 3, 1 and 0.333 km, respectively (Figure 3a). There were 86 vertical layers, with 21 layers in the lowest 2 km above the ground level (AGL). A previous study successfully used such layers to simulate a gale event on the lee of a small-scale mountain (Xue et al., 2020). The time steps were 3, 1, and 0.33 s for the three domains, and the model top was set at 10 hPa. The dynamics options used were the default settings (see the WRF user's guide at https://www2.mmm.ucar.edu/wrf/users/docs/user_guide_v4/contents.html), except for the imposition of a damping layer in the upper 5 km of the model to prevent model contamination by unphysical wave reflections from the model lid (Klemp et al., 2008). The rapid radiative transfer model (RRTMG; Iacono et al., 2008) was used for the short-wave and long-wave radiations, respectively. The convection scheme was turned off and the Thompson aerosol-aware scheme (Thompson et al., 2008) was used for clarifying the microphysical effects. The Noah land model (Ek et al., 2003) was used for the surface processes along with the Shin-Hong scale-aware planetary boundary layer (PBL) scheme (Shin & Hong, 2015), to account for the “gray zone” simulation of the PBL. A turbulent orographic form drag parameterization scheme (Beljaars et al., 2004) was used in D01, which has been previously implemented in the WRF model and can significantly improve the simulations of surface wind and precipitation over mountainous regions in South China and on the Tibetan Plateau (Xue et al., 2021; Zhou et al., 2018). The finest geographical data set in the WRF is of a resolution of 30′ (approximately 1 km in mid-latitude regions), which is not sufficient to represent a mountain with a horizontal scale of approximately 5 km. Therefore, we adopted the 3′ Shuttle Radar Topography Mission digital elevation data (SRTM; Farr et al., 2007) for D03 (Figure 3b). To ensure the stability of the numerical integration, a high-pass filter (Beljaars et al., 2004) was applied to the SRTM data to exclude scales below 600 m. The model configuration was defined as the control (CTL) run. Although the main features of the terrain are displayed on D01, with a 3 km grid size, much finer details could be resolved in D03, with a 333-m grid size, which refers to the original 3′ SRTM digital elevation data (Figure 3c). The terrain in the National Centers for Environment Prediction (NCEP) Final Analysis products, with a resolution of 0.25° (NCEP-FNL025), is nearly flat in the D03 (Figure 3c). A parallel simulation with the same configuration, except for the flat terrain in D03, was carried out for comparison (FLT run). The terrain height in the FLT run was set to be transiting smoothly from its original value to its domain mean in a band of 10 grids at the four boundaries. The simulated atmosphere in the FLT run was considered to be the background atmospheric reference that was not contaminated by the local small-scale mountains.

The WRF model was initiated at 20:00 BJT (12:00 UTC) on 21 May 2021, and integrated for 24 hr, with the first 6 hr regarded as the spin-up time. NCEP-FNL025 was used as the initial state and the lateral boundary conditions. Observations from 1 national and 25 local AWSs around the mountain (Figure 3b) were used to verify the simulations. All observations were checked for quality and are provided as the hourly data from surface meteorological stations in China (HDSMSC) by the China Meteorological Data Service Centre (CMDC) of the
China Meteorological Administration (CMA). It is important to note that 11 and 16 AWSs recorded the wind and relative humidity, respectively, while all 25 stations recorded the temperature and precipitation.

3. Results

3.1. Verification for the Fine-Scale Simulation

Figure 4 shows the time series of 10 m wind speed \( V_{10} \), 2 m temperature \( T_2 \), 2 m relative humidity \( RH_2 \), and precipitation at each AWS. The spatial mean and standard deviations (STDs) represent the temporal trend and spatial variation, respectively. The observations show that \( V_{10} \) increased before 11:00 BJT, after which it decreased (Figure 4a). A similar \( V_{10} \) trend was observed in the CTL and FLT simulations, though the peak was delayed by approximately 1 hr compared to the observations (Figure 4a). The observed and simulated \( V_{10} \) at some locations was closed to 20 m s\(^{-1}\), indicating that a gale occurred around these stations in D03 (Figure 3a). However, it is important to note that the recorded \( V_{10} \) were the 2 min means, therefore, shorter instantaneous gusts could have been more severe. Both simulations overestimated the mean \( V_{10} \), while the CTL run was closer to the observations (see the mean value in Figure 4a and bias in Table 1). The overestimated wind speed may be partly explained by the short time step (0.33 s) used in the inner-most domain, which made the wind more “instantaneous” than the observations. The STD of the \( V_{10} \) was approximately 4 m s\(^{-1}\) for the observations and CTL run, while it was underestimated in the FLT run. This suggests there was a reasonable spatial variation in the CTL run (Figure 4a). Figure 4b shows the time series of the \( T_2 \) (Note that the altitude correction had been applied to the simulated \( T_2 \) with a climatological lapse rate of \(-6.5^\circ C\) km\(^{-1}\)), wherein a sharp cooling period can be observed.
between 10:00 and 12:00 BJT in both the observations and simulations. The $T_2$ STD in the CTL run were again closer to the observed results than those in the FLT run (Figure 4b). Figure 4c shows the time series of $RH_2$ indicating that there was a sharp increase during the morning. Although both simulations underestimated the $RH_2$, the periods in which $RH_2$ increased, decreased and peaked were generally well reproduced by the model. Figure 4d shows the time series of hourly precipitation, which started at 10:00 BJT, peaked at approximately 13:00 BJT, and stopped after 15:00 BJT. The CTL run accurately reproduced the precipitation process and the magnitude of precipitation, with the STDs being very close to the STDs of the observation. The precipitation started during the sharp cooling period showed by $T_2$, indicating that the cooling effect from the melting and evaporation of precipitation may account for a large part of surface cooling. As the cooling effect of the melting and evaporation of precipitation has been investigated in great depth (Market et al., 2006; Thériault et al., 2006; Wexler et al., 1954), this was not measured quantitatively in this study.

Table 1 shows the cross-station statistical metrics (bias, absolute mean error (ABE), root mean square error (RMSE), and correlation) for $T_2$, $RH_2$, $V_{10}$, surface pressure, and precipitation from the CTL and FLT runs. Most of these metrics were closer to the observations in the CTL run than in the FLT run. In the CTL run, the mean
Table 1
Mean Statistical Metrics [Mean Bias (MB), Mean Absolute Error (MAE), Root Mean Square Error (RMSE), and Correlation (Corr; %)] for \( T_2 \) (°C), Relative Humidity (RH) (%), \( V_{10} \) (m s\(^{-1}\)), Surface Pressure (PS; hPa) and Precipitation (Precip; mm h\(^{-1}\)) From Both Control (CTL) and FLT Runs Compared to the Hourly Observations Between 02:00 BJT on 21 and 20:00 BJT on 22 May 2021

| Variable     | MB  | MAE | RMSE | Corr |
|--------------|-----|-----|------|------|
|              | CTL | FLT | CTL  | FLT  |
| \( T_2 \)°C | 0.86| −0.32| 2.21 | 3.01 |
| \( RH \)%   | −15.39| −15.03| 19.28| 19.65 |
| \( V_{10} \)m s\(^{-1}\) | 2.01| 2.13| 3.16| 3.40 |
| PS/hPa       | 0.65| −1.68| 2.12| 10.35 |
| Precip/mm h\(^{-1}\) | −0.00| −0.03| 0.06| 0.08 |

Note: The bold number indicate that the differences of the quantities between CTL and FLT are significant at 95% confidence level.

absolute errors of \( T_2 \), \( V_{10} \), and precipitation were approximately 2°C, 3 m s\(^{-1}\), and 0.1 mm, respectively. These values were comparable to those from multiple operational regional NWP models with horizontal grid sizes of 3–9 km during the same day over Gansu Province, although precipitation was much better simulated (private communication with the NWP verification team of CMA). The STDs of the relevant variables were well simulated and similar to those observed at the surrounding AWSs. This illustrates the advantage of high-resolution simulations that can explicitly resolve more details of the heterogeneity induced by topography. As the CTL run more closely simulated the key metrics, it is considered to be more credible than the FLT run for determining the relevant cold-air outbreaking processes and spatial variation during the 2021 ultramarathon tragedy.

3.2. Blizzard-Like Weather on the Mountain Lee

Figure 5 shows the three-dimensional distribution of \( V_{10} \) over the strong gale threshold, \( T_2 \) below 3°C, and graupel precipitation on the mountain between 11:00 and 14:00 BJT from D03 of the CTL run. High-speed winds and graupel clearly occurred on the mountain lee, with lower temperatures on the mountain top. The wind speed appears to have been higher at lower altitude on the lee, but the graupel precipitation occurred at relatively high altitudes. This is surprisingly consistent with the accounts in the official report, which states that some runners were exposed to the graupel precipitation after their clothes were torn off by the high-speed winds. Given that the temperature was not low enough to be considered deadly, the key cause of hypothermia was likely the combination of graupel and high-speed winds on the lee.

The influence of the mountain on the ultramarathon tragedy can be well illustrated by a comparison of the relevant quantities from the CTL run, to those from the FLT run during the period of the tragedy. The domain means in the FLT are shown in Figure 6, as these quantities were nearly horizontally homogeneous over D03. However, the relevant quantities for the CTL run along the track between CP2 and CP3 (36 grid boxes) are shown. It is clear from the figure that numerous locations along the track experienced a fresh gale, while some areas experienced a strong gale. However, the mean wind speed in the D03 of the FLT run was much lower, which decreased during the event (Figure 6a). Therefore, even if there were higher speed gusts, the wind on flat terrain is unlikely to have been a fresh gale event, given a mean absolute error (MAE) of approximately 3 m s\(^{-1}\) (Table 1). The D03 mean \( T_2 \) was between 9 and 11°C in the FLT run, which was significantly higher than the temperature of 2°C observed at many locations on the track in the CTL run (Figure 6b). Given the correction of the \( T_2 \) due to the elevation difference between CP3 and terrain height in the FLT run, with a climatological lapse rate of −6.5°C km\(^{-1}\) or a local lapse rate of −7.8°C km\(^{-1}\) that calculated from the mean temperature gradient in the FLT run, the lowest temperature can be constrained to 5°C (Figure 6b). Therefore, the linearly-corrected \( T_2 \) cannot fully account for the temperature difference between the two runs. It is understood that temperature is affected by many processes, including the mean wind advection, precipitation evaporation (melting) cooling and downdraft warming. If the maximum difference in \( T_2 \) at CP3 between the CTL run and the FLT run corrected linearly is considered to be the result from the total nonlinear effects, the nonlinear effects could account for a temperature difference of up to 3.2°C, in addition to the 3.5°C difference from cooling due to the linear lapse. Furthermore, the graupel shown
on the track in the CTL run was absent in the FLT run (Figure 6c). The cooling effect of the graupel may be a key reason for the hypothermia experienced by the runners.

There have been several metrics accounting for the combined cooling of exposed skin from low temperatures, high winds and humidity. Two of them are the wind chill equivalent temperature (WCT) and facial frostbite time (FFT; Moore & Semple, 2011). The former is defined as the temperature that would result in the same steady-state facial heat loss as occurs at a given temperature and wind speed (see the caption of Figure 6), and the latter is defined as the time it takes facial flesh to freeze (Moore & Semple, 2011). The two metrics have been successfully applied on the windy and cold conditions on Mount Everest to explain the tragedy in May 1996 (Moore & Semple, 2006, 2011). However, FFT is only applicable in a restrict condition which requires the temperature lower than −5.0°C. Therefore, only WCT is provided in Figure 6d, in which the lowest WCT can be −6.0°C in the CTL run against about 0.0°C in the FLT run with an altitude correction. Another metric is apparent temperature (AT), accounting for the temperature, humidity, and wind speed (Steadmann, 1984), which is shown in Figure 6f. In the FLT run, the minimum AT and AT calculated from the altitude-corrected $T_2$ at CP3, were approximately 0 and −4°C, respectively. The lowest AT on the track between CP2 and CP3 in the CTL run was approximately −10°C (Figure 6d). Given that the formula of AT (see the caption of Figure 6) does not account for a wet body due to the precipitation, the AT of −10°C for a runner without sufficient clothes would have been deadly. The above results indicate that local topographies can nonlinearly amplify temperature changes, high-speed winds, and precipitation, and may induce deadly weather conditions during a normal cold air outbreak in spring. Notably, the nonlinear effect of the mountain can account for a 12 ms$^{-1}$ higher wind speed, 3.2°C decrease in temperature, and the occurrence of solid precipitation on the mountain lee.

Figure 5. Three-dimensional map of high wind speeds (>21 m s$^{-1}$; green arrows), low temperature (<3°C; blue rendered), and graupel precipitation (white dots) on the mountain between 11:00 and 14:00 BJT. The red sphere, cone, and line are CP2, CP3, and the ultramarathon route, respectively.
Figure 6. Temporal series of (a) $V_{10}$ (m s$^{-1}$), (b) $T_2$ (°C), (c) graupel precipitation (mm (6 min)$^{-1}$), (d) wind chill temperature (WCT; °C), and (e) apparent temperature (AT; °C). The brown dots show quantities on the 36 grid boxes along the track between CP2 and CP3 in the control (CTL) run. The thick blue lines show the D03 mean in the FLT run. The thin gray and black lines in (b) shows the linearly corrected $T_2$ with the climatological lapse of $-6.5$°C km$^{-1}$ and the local lapse of $-7.8$°C km$^{-1}$, respectively, at CP3. The thin gray and black lines in (d) and (e) show the corresponding quantities calculated from the corrected $T_2$. $WCT = 13.12 + 0.621T_2 - 13.96V_{10}^{0.16} + 0.4867T_2V_{10}^{0.16}$ and $AT = -2.7 + 1.04T_2 + 0.2e - 0.65V_{10}$, where $e$ is the vapor pressure in hPa.
4. Mechanisms

4.1. Background Atmosphere

Figure 7a shows the equivalent temperature ($\theta_e$) from D01 at 700 hPa level at 11:00 BJT. A cold front can be identified as a large southwest-northeast trending $\theta_e$ gradient band between 320 and 325 K, behind of which the northerly and northwesterly winds were dominant. Therefore, the 325-K $\theta_e$ was assumed to be the front interface and is highlighted in Figure 7b. The interface was close to the ground in the south, but rose to a height of 5 km in the north, forming a typical 3D cold-front structure, that is, the cold northerly surface wind and southwesterly upper-level wind (Figure 7b).

The D03-mean height-time section of meridional wind ($v$) showed a clear two-layer troposphere, with southerly wind bounded above the northerly wind, and an interface at approximately 4 km above sea level (ASL; Figure 8a). Basically, the lower layer became more and more deeper during the period (Figures 8a and 8b), indicating the cold-front passing by the domain. During the night before the onset of precipitation, the wind speed increased in both layers, resulting in a large wind shear and potential Kelvin-Helmholtz instability at the interface between the two layers. The low-level temperature continually decreased with time and stabilized the atmosphere (Figure 8b), which may have been a result of the precipitation cooling effect. An increase in the water vapor mixing ratio ($q_v$) after 10:00 BJT (Figure 8c) may have been the result of the integrated water vapor flux (IWVF) convergence during the morning (Figure 8d). This would have resulted in an increase in precipitable water (PW) in the atmosphere (Figure 8d).

The height-latitude cross sections of $v$, water vapor flux (WVF), and reflectivity at 12:00 BJT are shown in Figure 9. The upper southerly wind layer was slightly lifted by the lower northerly wind layer from the south to north of the domain (Figure 9a). A WVF convergence and Kelvin-Helmholtz instability (Richardson number less than 0.25) were observed near the interface (Figure 9b), which were higher than the mean lifting condensation level (LCL) and mean 0°C level (Figure 9c). Therefore, shallow convection was triggered and ice clouds were formed, as indicated by the high reflectivity at latitudes between 37.1 and 37.3°N (Figure 9c). As the 0°C level was more than 1 km AGL, there should have been enough time for the ice in the cloud base to melt before it reached the ground surface (Gao et al., 2018). The Hovmöller diagram of precipitation shows that the precipitation propagated from the north to south of D03 within approximately 1 hr, and was nearly homogeneously distributed from the north to south (Figure 9d). These results indicate that the large-scale background atmosphere had provided favorable conditions for the northerly cold-front wind and light precipitation.

4.2. Downslope Windstorm and Lee

It is well understood that lee waves can form as upward propagating gravity waves are reflected back toward the surface when a critical layer is present over a mountain (e.g., Clark and Peltier, 1984; Eliassen & Palm, 1960). Thus, downslope windstorms can be generated by large-amplitude lee waves (Brinkmann, 1974; Jiang & Doyle, 2004; Mobbs et al., 2005; Xue et al., 2020). Figure 10a shows the height-latitude cross section of $v$ across CP3 at 12:00 BJT, wherein an apparent shooting flow with a speed over 20 m s$^{-1}$ can be seen on the lee of the mountain. $\theta_e$ contours show a trapped lee wave (TLW)-like feature downstream below 4 km ASL that likely occurred because of the reflection from the interface between the lower northerly flow and upper southerly flow. The widely used nondimensional mountain height ($H_m = Nh_m/U$) was used to determine the wave regime, where $N$ is the mean buoyancy frequency, $h_m$ is the mountain peak height and set as 700 m (see Figure 10), and $U$ is the mean incident wind speed perpendicular to the mountain (Doyle & Reynolds, 2008; Durran, 1986; Sauer et al., 2016; Vosper, 2004; Xue & Giorgetta, 2021). When $H_m$ is much smaller than 1, the flow is linearly perturbed by the mountain and the linear TLWs are formed (Pearce & White, 1967; Scorer, 1949). However, when $H_m$ is significantly greater than 1, low-level blocking occurs (Reinecke & Durran, 2008). When $H_m$ is close to 1, gravity waves can be amplified nonlinearly into large-amplitude TLWs or low-level breaking waves, and downslope windstorms occur (Doyle & Reynolds, 2008; Vosper, 2004). As the quantities were approximately horizontally homogeneous over D03 in the FLT run, the domain mean could be used as the background to obtain $H_m$. However, $H_m$ is generally calculated in theoretical and idealized simulations based on the inviscid assumption and with uniform upstream $N$ and $U$ (Reinecke & Durran, 2008). It is obvious that the theoretical definition cannot be directly applied in the realistic atmosphere, in which the viscosity in the lower PBL is nonnegligible and both wind and stability vary with height. Based on a boundary theoretical study proposed by Belcher and
Figure 7. (a) Equivalent potential temperature ($\theta_e$) at 700 hPa. (b) Three-dimensional 325-K $\theta_e$ surface (gray surface), with $\vec{V}$ (brown arrows) and $\vec{V}_{10}$ (black arrows) at 11:00 BJT from D01 projected onto the surface. The blue lines with triangles in (a) and (b) are the cold front at 700 hPa and at the ground surface, respectively.
Figure 8. Height-time cross sections of (a) $v$ (ms$^{-1}$); (b) $\theta_e$ (K); (c) $q_v$ (g kg$^{-1}$); and (d) time series of precipitable water (kg m$^{-2}$) and integrated water vapor flux (IWVF) convergence (m s$^{-1}$ g kg$^{-1}$) averaged over D03 in the FLT (blue) and control (CTL) (brown) runs. IWVF convergence was calculated as $-\int_{z_0}^{z_{top}} \nabla \cdot F_w dz = -\nabla \cdot \int_{z_0}^{z_{top}} F_w d z = -\nabla \cdot \left[ \text{Int} \left( F_w \right) \right]$, where $F_w = u_q \vec{i} + v_q \vec{j}$ is the water vapor flux (WVF) vector and $z_{top}$ is the model top (Mo et al., 2019).
Wood (1996), the incident wind to calculate $H_N$ should perhaps be those in the middle or upper part of the atmospheric boundary layer (Teixeira et al., 2004). Therefore, an average method proposed by Reinecke and Durran (2008) was employed to calculate $U$ and $N$,

$$
\overline{M_0} = \frac{1}{z_2 - z_1} \int_{z_1}^{z_2} M_0 \, dz,
$$

where $M_0$ is the domain-mean $v$ or $N$, and $z_1 = z_i/2$ and $z_2 = \min(z_i, h_0)$, where $z_i$ is the PBL depth. Figure 11a shows the time evolution of $\overline{U_0}$ and $\overline{N_0}$, in which $\overline{N_0}$ increased and $\overline{U_0}$ maintained its value from 10:00 to 12:00 BJT, resulting in an increase in $H_N$ (Figure 11b). After 12:00 BJT, $H_N$ decreased due to the decrease in $\overline{N_0}$ given the decrease in $\overline{U_0}$ (Figure 11). The increase in $\overline{N_0}$ may have been largely related to the cooling effect of the precipitation that started at 10:00 BJT (Figures 4 and 9d). TLWs were observed in the downstream direction near UHM (Figure 10a), which was expected as $H_N$ was close to 0.5 (Figure 11b near 12:00 BJT). However, the nonlinear effect should not be neglected, as $H_N$ was not much smaller than 1 and the waves experienced nonlinear amplification. Note that the diabatic cooling effect of precipitation could also induce a down-valley/slope flow that may amplify the downslope wind (Steiner et al., 2003).

Figure 9. Height-latitude cross sections at 12:00 for (a) $v$ (ms$^{-1}$; shaded area), $\theta_e$ (contours; in 5 K), and wind vector (arrows), with fivefold amplification of vertical speed $w$, (b) WVF (s$^{-1}$ g kg$^{-1}$), (c) reflectivity (DBZ), and (d) time-latitude cross section of 6 min accumulated precipitation (mm) for D03 in the FLT run. The cyan points in (b) indicate where the Richardson number is smaller than 0.25. The black dashed and red dotted lines in (c) show the 0°C level and mean lifting condensation level (LCL), respectively. The long red arrows in (d) show the propagation direction of the precipitation. Quantities are averaged zonally. The black triangle indicates the latitude of CP3.

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4.3. Precipitation and Hydrometeor Drift

The mountain appears to have affected the orographic ascent and convergence of the upslope atmosphere, resulting in a deeper lower northerly-wind layer and uplifted upper southerly-wind layer in the upstream direction (Figure 10a). Convergence over the upslope led to a much greater WVF convergence over the mountains in the CTL run compared with that in the FLT run (Figures 9 and 10b). The terrain-affected ascent and WVF convergence in the CTL run may have resulted in heavier and longer-lasting precipitation over the mountains (Figures 9 and 10d). The domain-mean accumulated precipitation during this event increased from 1.1 mm in the FLT run to 2.2 mm in the CTL run. Given the similar domain-mean IWVF convergence and PW in CTL and FLT (Figure 8d), terrain-induced ascent may have been the dominant factor that doubled the precipitation in the CTL run. The STD of the vertical speed in the lowest 5 km AGL was also twice as high in the CTL (0.069 m s⁻¹) than that in FLT (0.031 m s⁻¹), between 10:00 and 15:00 BJT. Thus, the local dynamical effect of the topography may have been the main factor that amplified the precipitation efficiency (total precipitation divided by atmospheric total PW in Ye et al., 2014).

Another apparent influence of the mountain is that the 0°C level was lowered over the mountain in the CTL run compared to that in the FLT run (Figures 9c and 10c), which may be partly due to the accumulated cold air from orographic blocking, and partly due to the cooling from the melting and evaporation of the precipitation. The combination of the lowered 0°C level and elevated surface led to graupel precipitation on the mountain (Figure 5). Assuming that the atmosphere at \( z_1 = \frac{1}{2} z_i \) (~350 m) flowed over the mountain \( (h_m \sim 700 \text{ m}) \) and experienced a saturated(dry)-adiabatic ascent (descent), the warming should have been \( (\Gamma_d - \Gamma_s)(h_m - z_1) \sim 1.8^\circ\text{C} \), where \( \Gamma_s \) (\( \Gamma_d \)) is the saturated(dry)-adiabatic lapse with value of 9.76 (4.5) °C.

Figure 10. Same as Figure 9 but for the section across CP3 in the control (CTL) run. Note that the color bar scale in (b) is an order of magnitude larger than the scale in Figure 9b, and the thick black lines show the terrain height across the section.
km$^{-1}$. Given the warming of the Foehn effect, the combined cooling and warming effects of the mountain were to decrease the surface temperature on the lee (Figure 36b).

The last effect of the mountain is the “spillover” phenomenon due to the hydrometeor drift effect, which refers to a situation when the hydrometeors forming over the upwind side are advected downwind to fall as precipitation in the rain-shadow region in the leeward side (Browning et al., 1974; Mo et al., 2019; Siler & Durran, 2016). Notably, the hydrometeor drift effect associated with a severe winter storm was partially blamed for a massive avalanche in western Canada on 29 January 2016 that killed five snowmobilers (Mo et al., 2019). The hydrometeor drift effect is clearly the reason for the tilt in the reflectivity from the south to north in Figure 10c. It can also explain why the precipitation area was mainly located on the mountain top and lee in CTL (Figure 10d) rather than on the upslope in FLT (Figure 9d). The spillover phenomenon may be largely related to the mountain geometry. As the mountain (UHM) is narrow, the atmosphere experiences a quick upslope ascent and downward motion (Figure 10a). However, this phenomenon was much less significant over the other larger mountain just east to UHM (not shown).

To further investigate the hydrometeor drift effect, Figures 11a and 11b show the mean integrated graupel flux (FG) vector between 11:00 and 13:00 BJT in FLT and CTL, respectively. A clear northerly advection of graupel was shown over and around the mountain in CTL, which was absent in FLT. This result suggests that the graupel was not only absent on the surface (Figure 6c) but also absent in the whole layer over the mountain in FLT.
Nevertheless, in the CTL run, graupel was formed and fell on the surface, and the northerly flow was strengthened on the lee. Therefore, to investigate the relative contribution from the strengthened wind and the presence of the graupel due to the drift effect, a diagnosis of the difference of FG between CTL and FLT is employed as

\[
F_{G,FLT} - F_{G,CTL} = q_{g,FLT} V_{FLT} - q_{g,CTL} V_{CTL}
= q_{g,CTL} (V_{FLT} - V_{CTL}) + V_{CTL} (q_{g,FLT} - q_{g,CTL}) + (q_{g,FLT} - q_{g,CTL}) (V_{FLT} - V_{CTL}).
\] (2)

The term on the left side of Equation 1 and the three terms on the right side are denoted as term I, II, III and IV, respectively. -I and IV represent the total difference of \(F_g\) and the cross product of wind (V) difference and graupel \((q_g)\) difference from FLT to CTL, respectively. -II and -III represent the contribution from the V difference and \(q_g\) difference, respectively, from FLT to CTL. These terms are calculated in the lowest 12 model levels (about 800 m AGL) to exclude the southerly wind in the first trough of the trapped lee wave (Figure 10a) and averaged vertically. The time-mean values of the four terms between 11:00 and 13:00 BJT are presented in Figures 12e–12f. Clearly, the presence of the graupel dominated the increased \(F_g\) from FLT to CTL (Figures 12a and 12b), which seemed to be self-evident as the change in graupel from FLT to CTL was zero-to-one. However, we focus more on the distribution of term -III though the magnitude was relatively small. Negative and positive drift effects were clearly shown in the upstream and downstream, respectively, of the mountain (Figure 12e), which indicates that the downslope wind acted to drift the graupel over the lee during the tragedy. Term IV displayed a similar distribution as term -III (Figure 12f). The hydrometeor drift effect caused the graupel to be spilled over the narrow mountain and combined into the downslope wind, resulting in the blizzard-liking weather on the lee.

5. Conclusions and Discussions

This study investigated the extreme weather during the 2021 ultramarathon on a small-scale mountain in northwestern China, during which 21 participants died of hypothermia. This was done using two parallel WRF simulations with a grid resolution of 333 m, with and without the local mountains. The accuracy of both simulations was determined through comparisons with the observations of 26 AWSs around the mountains. This comparison showed that both simulations accurately replicated the processes of cold-air outbreak with a sharp temperature drop, high northerly winds, and precipitation propagation from the north to the south of the inner-most domain. However, the results of the CTL run, which incorporated high-resolution realistic terrain, had better accuracy for surface spatial variations of temperature, wind, and precipitation than the FLT run. The nonlinear effect of the small-scale mountain accounted for temperature decreases of up to 3.2°C, in addition to the 3.5°C decrease due to the linear variation caused by the elevated terrain. The wind speed on the lee between CP2 and CP3 was 12 m s⁻¹ higher than that in other areas because of the downslope-wind effect of the mountain, which resulted in a strong gale event (>20.5 ms⁻¹). Solid precipitation (graupel) also occurred on the lee, which would have been absent if the terrain was flat. This indicates that there was blizzard-like weather during the ultramarathon, even though it was spring.

A two-layer troposphere formed as the cold front passed by the focused domain, with a cold northerly wind in the lower layer and a southerly wind in the upper layer. As the interface of the two layers was higher than the domain mean LCL, convection may have been triggered by the Kelvin-Helmholtz instability as the wind shear increased overnight. Precipitation occurred because of the IWVF convergence and an increase in PW. The precipitation occurred as graupel owing to the elevation of the mountain surface. In the absence of mountains, liquid precipitation would have fallen on the surface owing to shallower convection as well as sufficient time for the ice to melt as the 0°C level was more than 1 km AGL.

The mechanism that caused the blizzard-like weather on the lee of the mountain (UHM) is summarized in Figure 13. When the cold front moved from the north to the mountain, the upstream lower layer stabilized because of the cooling effect of melting and evaporating precipitation. The increased low-level stability amplified the nonlinear effect on the trapped lee wave, resulting in downslope high winds on the lee. At the same time, the presence of the mountain induced stronger windward convergence and uplift, leading to solid condensation at higher elevation with cooler air (deeper convection). The 0°C level also lowered because of the influence of the elevated surface and strengthened cooling due to the increased precipitation melting and evaporation. The solid condensates were drifted downstream to fall as graupel on the mountain lee. The blizzard-like weather was the combination of strong downslope winds and spillover of graupel.
Figure 12. The integrated graupel flux vectors ($\int_{0}^{z_{0}} \vec{F}_G \, dz$) in (a) FLT and (b) control (CTL), where $\vec{F}_G = u q_g \hat{z} + v q_g \hat{y}$ is the graupel flux vector and $q_g$ is the graupel mixing ratio. (c)–(f) represents term -I, -II, -III and VI in Equation 1. All the quantities were calculated in the lowest 12 model levels (about 800 m AGL) and averaged vertically and between 11:00 and 13:00 BJT. The integrated graupel flux vectors are displayed in (c)–(f) as well for reference. The cyan dots show the ridge of UHM.
The ultramarathon case considered in this study shows how small-scale mountains influence the near-surface wind speed and its distribution around the mountain, the precipitation area (downslope), precipitation type (liquid or solid), and the cooling effect caused by the melting and evaporation of precipitation. Since these effects are very sensitive to mountain geometry, the horizontal model resolution needs to be increased for better representing mountains in forecasting simulations. However, the resolution required to sufficiently resolve the topography is contingent on how rapidly the terrain-height variance reduces with horizontal scale. A study of the terrain-height variance over Colorado, USA, showed an upper bound for a grid size of 0.1 km for mesoscale numerical models (Young & Pielke, 1983), which required significant computational resources given a small domain of 50 × 50 km; a 36-hr simulation required approximately 1 week of integration with 1,024 computational cores (Xue et al., 2020). Even if the numerical cost of such simulations may vary a lot between models, works on model dynamics and code optimization would be beneficial for an operational use of such high-resolution model simulations. Otherwise, sub-grid orographic parameterization is required to represent the missing gravity-wave drag, low-level blocking, and turbulent drag (Beljaars et al., 2004; Lott & Miller, 1997). However, most existing parameterization schemes have been developed for large-scale models with resolutions above 10 km, and orographic parameterization schemes for kilometer-scale models are not available to date. Moreover, we often suffer from a lack of observations to evaluate high resolution models, and some processes are not yet well understood. Dedicated observational campaigns such as the TEAMx (Multi-scale transport and exchange processes in the atmosphere over mountains – program and experiment; http://www.teamx-programme.org/), planned over the Alps in the coming years, are essential to make additional progress. The integrated meteorological observations at Dali National Climate Observatory, in Hengduan Mountains, China, may also contribute to the understanding of high-resolution simulations on complex terrain (Xu & Li, 2020). Otherwise, high-resolution simulations that compromise between resolving the main features of the topography and computational resources may still be the most feasible method of providing weather forecast services over mountainous regions for some public events.

Figure 13. Sketch illustration of the formation of the blizzard-like weather on the lee of UHM. Solid precipitation from the deeper convection caused by the orographic ascending fell on the lee side due to the hydrometeor drift effect, and downslope winds (DSW) formed as the trapped lee waves (TLW) were amplified due to the upstream low-level cooling, resulting in the blizzard-like weather on the lee.
Data Availability Statement
The WRF model is an open-source modeling system developed in the University Corporation for Atmospheric Research (UCAR) and National Center for Atmospheric Research (NCAR) (UCAR & NCAR, 2022) and is available at Release%20WRF%20Version%204.1%20wrf-model%20WRF%20GitHub. The NCEP final analyses used as the background atmosphere in our study are available at CISL%20RDA%20NCEP%20GDAS/FNL%2020%25%20Degree%20Global%20Tropospheric%20Analyses%20and%20Forecast%20Grids%20(ucar.edu). The hourly data from surface meteorological stations in China (HDSMSC) by the China Meteorological Data Service Centre (CMDC, 2022) of the China Meteorological Administration (CMA), and can be found at CMDC%20(cma.cn). The model simulations that are essential for this study are available at (Xue, 2022).

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