Large Temperature Fluctuations due to Cold-Air Pool Displacement along the Lee Slope of a Desert Mountain

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Abstract

Large temperature fluctuations (LTFs), defined as a drop of the near-surface temperature of at least $3^\circ$C in less than 30 min followed by a recovery of at least half of the initial drop, were frequently observed during the Mountain Terrain Atmospheric Modeling and Observations (MATERHORN) program. Temperature time series at over 100 surface stations were examined in an automated fashion to identify and characterize LTFs. LTFs occur almost exclusively at night and at locations elevated 50–100 m above the basin floors, such as the east slope of the isolated Granite Mountain (GM). Temperature drops associated with LTFs were as large as $13^\circ$C and were typically greatest at heights of 4–10 m AGL. Observations and numerical simulations suggest that LTFs are the result of complex flow interactions of stably stratified flow with a mountain barrier and a leeside cold-air pool (CAP). An orographic wake forms over GM when stably stratified southwesterly nocturnal flow impinges on GM and is blocked at low levels. Warm crest-level air descends in the lee of the barrier, and the generation of baroclinic vorticity leads to periodic development of a vertically oriented vortex. Changes in the strength or location of the wake and vortex cause a displacement of the horizontal temperature gradient along the slope associated with the CAP edge, resulting in LTFs. This mechanism explains the low frequency of LTFs on the west slope of GM as well as the preference for LTFs to occur at higher elevations later at night, as the CAP depth increases.

1. Introduction

Fluctuations of near-surface air temperature are a common feature of the stable nocturnal boundary layer. In this study, we define a large temperature fluctuation (LTF) as a drop of at least $3^\circ$C in less than 30 min followed by a recovery of at least half the initial temperature drop within one hour of the temperature minimum. LTFs can indicate substantial modification of the horizontal and vertical structure of the stable boundary layer.

The potential mechanisms that could produce LTFs can be split into two primary groups. The first group involves significant horizontal advection of cold air and includes synoptic fronts (Shafer and Steenburgh 1987; Johnson et al. 2014), and slope and valley flow fronts (Doran and Horst 1981; Blumen et al. 1999). The second group involves the displacement of a cold-air pool (CAP; Whiteman et al. 2001) such that cold air ascends and descends the slopes confining the CAP, producing LTFs in the area that the CAP periodically envelops.

CAPs can form via in situ cooling, advection of cold air into basins and valleys by katabatic flows (Neff and King 1989; De Wekker and Whiteman 2006), or by warm-air advection above the basin (Wolyn and McKee 1989). The depth of a mature CAP can vary widely, from less than 100 m to nearly 1000 m (Clements et al. 2003; Lareau et al. 2013). While CAPs typically will deepen and strengthen through the nocturnal period, variability in CAP strength and depth can occur because of synoptic and mesoscale influences (Steinacker et al. 2007).

Numerous mechanisms for CAP displacement have been advanced in the literature. The strong stratification
observed in CAPs permits the development of wave motions such as internal gravity waves (Lareau and Horel 2015b) or solitary waves (Cheung and Little 1990) that can produce temperature and wind variations at the surface. These wave motions can be induced by flow collisions (Lin et al. 1993), convection (Stull 1976), or by the interaction of flow with terrain such as a mountain barrier (Brown et al. 2003; Epifanio 2003). Additionally, CAPs can be displaced by their interaction with slope or valley flows (Yoshino 1984; Mahrt et al. 2010; Burns and Chemel 2015).

When a stably stratified flow encounters a terrain barrier, four distinct three-dimensional flow regimes can develop (Epifanio 2003): 1) small-amplitude waves, 2) wave breaking, 3) upstream stagnation and flow splitting, and 4) lee vortices. Two parameters, $\epsilon$ and $\beta$, can be used to determine, for a given flow and barrier, which class of motions will be observed. The first parameter, $\epsilon$, is the nondimensional mountain height (Jackson et al. 2013):

$$\epsilon = \frac{N h_0}{U},$$

where $N$ is the Brunt–Väisälä frequency, a measure of the static stability of the flow impinging on the barrier, $U$ is the upstream wind speed normal to the barrier, and $h_0$ is the terrain height. The second parameter, $\beta$, expresses the horizontal aspect ratio of the barrier with large $\beta$ indicating mountains elongated in the cross-flow direction.

For $\epsilon \ll 1$, small-amplitude waves or wave breaking are the likely flow regimes that will develop. The flow is relatively insensitive to $\beta$ in these cases. However, the atmospheric stability and flow field during these regimes are not conducive to maintaining a strong CAP (and therefore a large temperature gradient) that would allow for LTFs to occur (Lee et al. 1989; Lareau and Horel 2015a).

Conditions with $\epsilon \gg 1$ and large values of $\beta$ typically result in blocked flow at low levels. Situations with values of $\epsilon \gg 1$ and smaller $\beta$ are likely to result in flow splitting or the formation of an orographic wake in the lee of the barrier. Orographic wakes are potentially conducive to LTF development. During an orographic wake, potentially warmer air descending in the lee of terrain and relatively cold, stable air flowing around the terrain can produce a quasi-stationary boundary that resembles a density current in the lee of the barrier (Epifanio and Rotunno 2005). If the position of this boundary fluctuates, LTFs may be produced at the surface.

Because of their small spatial scale and short temporal duration, LTFs present a challenge for numerical models. Models may not resolve the phenomena that contribute to LTFs or may have difficulty parameterizing those phenomena. Strong CAPs, an important component of LTFs, are particularly difficult to simulate and often display excessive vertical mixing (Holtslag et al. 2013; Neemann et al. 2015). Simulations of mountain waves, a potential LTF causal mechanism, have been shown to be highly sensitive to initial and boundary conditions and model parameterizations (Reinecke and Durran 2009a,b). No previous investigators have specifically attempted to replicate observed LTFs in complex terrain using numerical simulations.

The Mountain Terrain Atmospheric Modeling and Observations (MATERHORN) program has the primary goal of addressing deficiencies in our understanding of boundary layer processes in complex terrain and deficiencies in the representation of such processes in mesoscale models (Fernando et al. 2015). LTFs are the surface manifestation of a significant disturbance of the nocturnal stable boundary layer, and they were observed in many locations across the MATERHORN study domain in northern Utah. A thorough understanding of the causal mechanisms of LTFs will be beneficial for understanding boundary layer processes in complex terrain and insight gained will aid the model verification and improvement efforts of the MATERHORN Program and others.

To investigate LTFs, an automated algorithm is applied to systematically identify LTFs at over 100 automatic weather stations (AWSs). The events identified by the algorithm will be shown to occur primarily at night on slopes elevated above the main basin and with air temperature variations of up to 13°C in less than 30 min. A case study and composite (average of multiple events) analyses of LTFs, combined with high-resolution numerical simulations, will be used to conclude that the primary causal mechanism for the observed LTFs is the displacement of a CAP by an orographic wake and vertical axis vortex that forms in the lee of a topographic barrier when a persistent stably stratified nocturnal flows impinges on the barrier.

2. Field experiment and model simulation

The observational component of the MATERHORN Program was conducted at Dugway Proving Ground (DPG) in northwest Utah. DPG was chosen because of its complex terrain, accessibility, and an extensive network of preexisting meteorological sensors. Terrain at DPG is characterized by relatively tall, steep-sided mountain ranges interspersed with wide, flat valleys (see Fig. 1). The climate in the area is semiarid with a
mean annual precipitation of around 200 mm. The western part of DPG, labeled “Playa” in Fig. 1, consists of extremely flat (slopes as low as 0.0001%) and largely vegetation-free mud flats or playas. The basin in the eastern portion of DPG, labeled “East Basin,” is covered by sparse, low brush and grass with a mean slope of around 0.002%. Differences in soil thermal conductivity and other land surface contrasts between the two basins cause the east basin to develop a CAP that is stronger than its counterpart over the playa (Rife et al. 2002). Separating these two basins is Granite Mountain (GM), which rises up to 800 m above the basin floors. The upper half of GM has numerous slopes exceeding 100% while the lower slopes consist mainly of alluvial fans whose slopes are much gentler. One particular alluvial fan, on the southeast side of GM (see inset in Fig. 1), was the location of intensive observation during the MATERHORN field experiment. Data collected there have been used in multiple previous studies (Lehner et al. 2015; Grachev et al. 2016). This area will be referred to here as the east slope and will be the primary focus for analysis in this paper.

An extensive array of surface AWSs was installed across DPG during the MATERHORN field campaign. This includes 81 permanent and 40 portable stations that are shown as colored circles in Fig. 1. There were 104 stations that measured 2-m air temperature, 2-m wind speed and direction, atmospheric pressure, and 2-m relative humidity. The remaining 17 stations, installed on a transect across the southern end of GM, only measured temperature and relative humidity at approximately 1.5 m AGL. Five-minute averages of all AWS variables will be analyzed. Four AWSs, labeled L1–L4 in Fig. 1, will be a focal point for analysis of LTFs on the east slope.

In addition to the AWSs, several 32-m atmospheric flux measurement towers were deployed during the MATERHORN field campaign. In this study, we will be using data from one of these towers located at 1338 m ASL near the bottom of the east slope (labeled ES2 in Fig. 1). Air temperature, wind speed, wind direction, vertical velocity, turbulent kinetic energy, and sensible heat flux are available for four levels on ES2: 0.5, 4, 10, and 20 m AGL. All data were processed using the Utah Turbulence in Environmental Studies Processing and Analysis Code (Jensen et al. 2016). Data were quality controlled following Vickers and Mahrt (1997) and a two-sector planar fit, divided between upslope and downslope conditions.

FIG. 1. Map (UTM zone 12T) of the study area at DPG with the southern part of Granite Mountain detailed in the inset. The color-coded circles indicate the daily frequency of LTFs at various sites at DPG. Elevation contours are 80 m in the main plot and 30 m in the inset, with the 1315-m contour annotated in both. The transect in the inset marked L1–L4 is the primary slope transect in this study. The location of the flux tower ES2 and the Granite Ridge and West slope sites are annotated in the inset.
downslope winds, was applied (Wilczak et al. 2001). A 5-min averaging period was used to maintain high temporal resolution for the purpose of this study. The L2 tower is nearly collocated with the L2 AWS.

The flows in and above the topographic gap between the southern end of GM and the Dugway Range were observed with a sodar wind profiler. The sodar recorded 10-min average wind speed and direction at 10-m intervals up to 200 m AGL. The wind measured by the sodar at 150 m AGL in the gap is used here to characterize the flow impinging on the southern end of GM. Despite the location in the topographic gap, the gap rarely exhibits classical orographically forced gap winds in the manner of the Columbia River Gorge. Rather, the impact of the gap is generally relegated to minor channeling of mesoscale and synoptic-scale flows and contributing to the pattern of local thermally driven flows. Despite the limited height of the sodar measurements, these are the most representative observations available for the impinging flow because of the relatively uniform, flat terrain in the upwind fetch of the sodar. More detail on the instrumentation used here can be found in the supplemental materials of Fernando et al. (2015).

To supplement the observations, numerical simulations were conducted using the Weather Research and Forecasting (WRF) Model, version 3.4.1 (Skamarock et al. 2008). The model configuration included four one-way nested domains, 50 vertical levels (22 levels below 600 m), and a horizontal resolution of 500 m in the innermost domain, which is a 60-km square centered on the east slope of GM. At 500-m horizontal resolution, the steepest slopes on GM are approximately 20°C, precluding the need for alternative grid methods such as those described in Lundquist et al. (2012). Physical parameterizations include the Lin et al. (1983) microphysical scheme, the RRTM longwave radiation parameterization (Mlawer et al. 1997), the Dudhia shortwave radiation parameterization (Dudhia 1989), and the quasi-normal scale elimination (QNSE) planetary boundary layer (PBL) scheme (Sukoriansky et al. 2005). The model also included the improved land cover and Noah land surface model as described in Massey et al. (2014). The National Centers for Environmental Prediction Final Operational Model Global Tropospheric Analyses (NCEP 2000) were used for initial and boundary conditions. This configuration, in particular the PBL scheme, was identified by Dimitrova et al. (2016) to offer the best representation of flows in the gap and near the east slope. While multiple PBL schemes were tested, only the results from the QNSE will be shown. A WRF simulation of one case study night, 1–2 October 2012, will be used in this paper and consists of instantaneous values of modeled variables at 20-min intervals. Data intervals shorter than 20 min did not provide significantly more detail. The simulation was initialized at 1700 mountain standard time (MST) 30 September 2012 and was 48 h long. This provided a 28-h lead time before the first observed LTF on the night of 1–2 October 2012.

3. LTF identification algorithm

We developed an automated algorithm to identify LTFs that was applied to all 121 AWS datasets from 26 September 2012 to 29 October 2012. This time period corresponds to the MATERHORN fall intensive field campaign. LTFs were identified using two criteria. First, a temperature drop of at least 3°C must occur within 30 min (using data at a resolution of 5 min). This effectively isolates the sporadic LTF occurrences from the regular fluctuations of 1°–2°C amplitude that are observed at DPG. Second, at least half of the original temperature drop must be recovered within 1 h of the minimum temperature. It is relatively rare that temperatures recover to their initial value, as the LTFs are usually superimposed on a nocturnal cooling trend. This last requirement excludes synoptic frontal passages, which are relatively common in this area during the fall months. Each LTF identified by the algorithm at the four AWSs that comprise the L1–L4 transect was verified manually to ensure the quality of the algorithm.

More restrictive thresholds, such as a larger temperature drop criterion or a shorter time period criterion, lower the number of LTFs identified but do not fundamentally change the structure of the composite LTF. Less restrictive thresholds dramatically increase the number of LTFs identified at all stations. Additionally, the composite LTF becomes much less coherent when the temperature drop requirement is set below 2.5°C, as many small temperature fluctuations that exhibit substantially different characteristics are included in the sample. A 3°C drop within 30 min provides the optimum balance between having a large sample size while preserving the characteristics of the large fluctuations.

4. Temporal and spatial distribution of LTFs

The spatial distribution of LTF frequency is shown by color-coded circles in Fig. 1. The colors indicate the mean daily number count of LTFs at each AWS as identified by the automated algorithm. LTFs are rare at the bottom of both the east basin and the playa. LTFs are also very rare at mountaintop locations. LTFs are frequent on the slopes slightly above the valley floors, particularly near GM. The inset in the upper left of Fig. 1...
shows the southern end of GM in more detail. Locations near the bottom of the east slope of GM average up to three LTFs per night. This is the highest frequency observed at DPG. The frequency of LTFs decreases dramatically at an elevation only 50 m higher up on the slope and continues to decrease up to the ridgetop. Sites on the more sparsely instrumented west slope of GM exhibit much lower LTF frequencies than the east slope. Because of the high frequency of LTFs on the east slope of GM and the extensive observational network there, our analysis will focus on this area.

The mean daily count of LTFs on the east slope of GM as a function of the time of day is shown in Fig. 2. Overall, LTF frequency increases dramatically near sunset (between 1730 and 1820 MST), is elevated through the night, then decreases to near zero by 1000 MST. LTFs are extremely rare between 1000 and 1700 MST. The only LTFs observed between 1200 and 1400 MST are attributable to convective outflows observed at L1 on 25 October 2012. Manual analysis indicated that these were the only convective outflows identified as LTFs by the automated algorithm at L1–L4. LTF frequency at L1 (bottom of the slope) quickly reaches its peak at 2200 MST, after which it decreases through 1000 MST. At L2, at an elevation 35 m higher up, the highest LTF frequencies are observed. A broad plateau in frequency is observed here between 2200 and 0500 MST. At L3, at an elevation 23 m above L2, the overall frequency is again lower, although a steady increase is seen to a pronounced maximum at 0500 MST. At L4, at an elevation 74 m higher than L3, a similar pattern to L3 is shown, but a lower maximum is observed.

While the hourly frequencies of LTFs at L1–L4 differ, LTFs at the four levels tend to occur on the same days and under the same general atmospheric conditions. Figure 3a shows daily counts of LTFs at L1–L4 for each day of the 34-day study period. LTFs were observed at one or more levels on the slope on 28 of the 34 days. LTFs were observed at L4 on only 11 days, marking the lowest frequency of LTFs along the L1–L4 transect. LTFs tend to be observed at each level of the slope on the same nights. This is particularly true at L2 and L3 where there are only three days LTFs are observed at one but not the other. On nearly all days the largest number of LTFs is observed at L2 and the fewest at L4.

To investigate a possible connection between LTF occurrences and synoptic or mesoscale conditions, ERA-Interim reanalyses and additional observations at DPG were used. The ERA-Interim, a global reanalysis dataset operated by the European Centre for Medium-Range Weather Forecasts, is run with 60 vertical levels and a reduced Gaussian grid with 79-km spacing for surface and other gridpoint fields (Dee et al. 2011).
Using the ERA-Interim data, we could not identify a synoptic pattern during which LTFs were preferentially found. Instead, LTFs were found to correlate most strongly on the static stability of the atmospheric layer below the GM ridge and the speed and direction of the wind in the vicinity of GM as measured at and above the surface. To characterize the static stability near GM during LTFs we use the mean bulk nocturnal lapse rate of air temperature of two layers on the west side of GM (Fig. 3b). Data from the west side of GM are used as it is typically the upstream side during the nocturnal period. Bulk lapse rates are calculated for two layers using the Granite Ridge, West Slope, and Playa Station AWSs (locations shown in Fig. 1). The lower of the two layers is approximately 100 m thick and its lapse rate (shown in red in Fig. 3b) provides an estimate of the CAP strength west of GM. The upper layer (shown in orange in Fig. 3b) is 280 m thick and characterizes the static stability of the upper nocturnal boundary layer (UNBL) above the CAP and below the GM ridgeline. The highest numbers of LTFs, particularly at L2 and L3, are observed when the UNBL is isothermal or stable and the bulk lapse rate of the CAP layer exceeds 30°Ckm\(^{-1}\). Multivariable linear regression was performed with the predictand being the sum of the daily counts of LTFs at L1–L4 and the two lapse rates as the predictors. The results of the regression indicated that these two lapse rates can explain 86% of the day-to-day LTF variance at L1–L4. The bulk lapse rate determined from air temperature readings between Granite Ridge and Playa (shown in blue in Fig. 3b) will be later used to characterize the static stability upstream of GM.

Mean nocturnal wind speed and direction in the gap at 150 m AGL is shown in Fig. 3c for the 34-day study period. The mean wind speed tends to be less than 4 m s\(^{-1}\) and the wind direction tends to be southwesterly on nights with a large number of LTFs. In Fig. 3c and throughout this work we use the wind direction with respect to true north (0° = N, 90° = E, 180° = S, and 270° = W). An easterly mean nocturnal wind is observed on the night of 7–8 October despite this night displaying the highest number of LTFs. This is due to easterly winds during the first half of the night that switched to westerly after 0100 MST, producing a mean vector that is easterly, despite the LTFs occurring with westerly winds.

The lack of LTFs on three nights, 12–13, 16–17, and 23–24 October, is likely due to light precipitation and extensive cloud cover that limited the development of a nocturnal stable boundary layer. The latter two nights also exhibited relatively high winds. The other 3 nights without LTFs are among the nights with the weakest static stability during the 34-day period, despite the absence of precipitation and high winds.

Figure 4 illustrates the relationship between the bulk lapse rate (determined from air temperature readings at Granite Ridge and Playa Station) and the sodar-derived wind direction at 150 m AGL during individual LTF events. It shows that LTFs occur most often when the atmosphere is strongly statically stable and the flow impinging on GM is from the southwest. Over 80% of observations made during an LTF occur with southerly winds impinging on GM and lapse rates more stable than isothermal. LTF observations concurrent with northeasterly flow are often seen at nights when low-level flow reversals (up to 150 m AGL, not shown) in the gap are observed. These cases often coincide with particularly strong LTFs.

There is a noticeable reduction of northeasterly flow in the gap once the bulk lapse rate drops below 1°C km\(^{-1}\). The bias toward southwesterly flow in the gap under increasingly stable conditions (bulk lapse rate is highly negative) offers insight into the mechanism driving the flow in the gap. During the 34-day study period over a wide range of synoptic-scale weather conditions, westerly or southwesterly flow dominates over GM and through the gap from 2200 to 1300 MST (not shown). The onset of this flow typically occurs 4–5 h after sunset and the cessation of the flow lags sunrise by a similar amount of time. Local horizontal thermal contrasts due to the playa surface remaining much warmer at night than the east basin (Rife et al. 2002) would be expected to produce an easterly or north-easterly flow in the gap at night. The observed flow in the gap is directly opposed to this. The strong bias toward southwesterly flow under higher static stability may indicate that regional-scale boundary layer heterogeneity (due to topographic and land surface contrasts on a regional scale) may instead be responsible for the nocturnal flow observed in the gap.

5. Composite structure of east slope LTFs

After establishing that LTFs observed on the east slope occur under distinct mesoscale conditions, we composite data from L2 and the adjacent ES2 tower during detected LTFs to identify the key characteristics of the LTFs.

To make this composite, the time of minimum temperature of each LTF was designated as t = 0. Time series of air temperature, wind, pressure, vertical velocity and near-surface static stability for the 60 min preceding and the 40 min following each LTF were then averaged. For air temperature and static stability, each individual time series was normalized by its mean value over the 100-min period before being averaged across LTF events. Here, the near-surface static stability was
determined as a bulk lapse rate of air temperature calculated from two tower levels. While this follows the same methodology used to calculate the static stability of the layer up to the height of the GM crest, note that the vertical scale is substantially different. The individual pressure time series were also normalized, and detrended, prior to averaging. The composite wind speed and direction were derived from a vector average of the individual events. The vertical velocity time series represents a simple average. Manual subjective analysis of each LTF at ES2 was performed to establish that the composite pattern is representative of the individual cases. Since only 4 of 43 LTFs deviated substantially from the composite evolution, we conclude that the composite mean is representative of the evolution of LTFs in general. Figure 5 shows the composite mean evolution of air temperature, wind, pressure, vertical velocity, and static stability during an LTF at ES2. Interquartile ranges are included in Fig. 5a to visualize the variability of the temperature evolution of the LTFs included in the composite.

Prior to the initiation of the LTF, the flow field (Fig. 5b) and stability (Fig. 5c) at ES2 indicate the presence of a katabatic flow. The wind is out of the west (downslope), with a jet structure with a peak observed wind speed of 4 m. Relatively strong near-surface stratification of 1°C m⁻¹ is also observed (Fig. 5c). These flow characteristics are indicative of katabatic flows (Haiden and Whiteman 2005; Grachev et al. 2016).

At \( t = -20 \) min the temperature drop associated with the composite LTF begins. The temperature drops sharply at all levels of the tower (Fig. 5a), the wind speed at all tower levels drops below 0.5 m s⁻¹ (Fig. 5b), and the near-surface stratification (Fig. 5c) begins to weaken. The maximum rate of temperature change occurs at \( t = -15 \) min. At this time, the katabatic jet structure is no longer present and the wind direction has shifted clockwise to northerly or northeasterly at all levels. Vertical velocity at 20 m changes sign from negative (downslope) and positive (upslope) at \( t = -25 \) min coincident with the beginning of the temperature fall. It reaches a maximum positive value at \( t = -5 \) min.

The temperature minimum of the composite LTF is reached at \( t = 0 \). The temperature drops observed at the 4- and 10-m levels are 40% larger than at 0.5 or 20 m (Fig. 5a). This leads to the weakest stratification near the surface and strongest stratification in the 10–20-m layer during the composite LTF event at \( t = -5 \) (Fig. 5c). Directional wind shear and weak winds speeds on the tower (Fig. 5b) indicate that the katabatic flow has not redeveloped at this point.

The rise in air temperature ends by \( t = 35 \) min and the katabatic flow has reestablished as indicated by the near-surface jet structure, decreasing directional shear with
westerly winds, and increased near-surface stratification. In the composite LTF, the temperature only recovers 35% of the original drop, instead of the 50% prescribed by the algorithm. This is because the post-LTF temperature recovery often occurs as a sharp increase, after which the temperature continues its nocturnal decrease, rather than remaining elevated. As the exact timing of the temperature recovery varies, the magnitude seen in the composite LTF is smaller than in the individual cases.

The composite LTF pressure trace is shown in Fig. 5a. Prior to LTF initiation the pressure decreases. The pressure begins to increase shortly before the temperature drop begins and continues rising until the minimum temperature is reached. However, the amplitude of this fluctuation is less than 0.05 hPa. Individual cases display a similar, slowly varying, pressure evolution that does not indicate the presence of solitary waves or transient gravity waves like those observed in Cheung and Little (1990). Potential causes of a pressure signal of this magnitude include hydrostatic effects of the cooling air, a pressure signal associated with an orographic wake, or a combination of the two.

Turbulence kinetic energy (TKE) and sensible heat flux (SHF) show no consistent pattern in the LTF composites from ES2 and are not shown in Fig. 5. Large spikes in turbulent kinetic energy and sensible heat flux often occur in individual LTF cases with the onset or cessation of a LTF, but their timing relative to the occurrence of the LTF minimum temperature are not consistent enough for them to be presented in the composites.

6. Case study of 1–2 October 2012

To investigate the causal mechanisms of LTFs, observations from the ES2 tower (Fig. 6) and numerous AWSs (Fig. 7) will be compared with WRF simulations in a case study for the night of 1–2 October 2012. This particular night was the second intensive observation period of the MATERHORN fall field campaign. Five LTFs were observed at L2. The synoptic environment, inferred from ERA-Interim reanalysis data, was relatively quiescent as a high-amplitude ridge was centered just to the west of Utah with 700-hPa wind speeds less than 6 m s$^{-1}$ throughout the period. The observational data for this night will be described first, followed by a comparison with WRF simulations. Observations from ES2 and L2 will be compared with WRF simulations of 2-m data at the nearest WRF grid point (150 and 130 m southeast of ES2 and L2, respectively and 16 m lower in elevation). The WRF simulation reproduced one LTF event, but at a later time. For our analysis of the LTF event, we therefore shifted the simulations by 80 min, to synchronize the modeled and observed events. All times shown in the figures and discussed in the text are those corresponding to the observed dataset. With small-scale, discrete phenomena such as LTFs, simulations with temporal or spatial errors can still be quite valuable, such as with simulations of severe convection (Lack et al. 2010).

Figure 6 shows a time series of numerous observed variables at L2 and ES2 along with the WRF simulated values. Local sunset occurs at roughly 1815 MST, after which a katabatic flow develops on the east slope (Fig. 6c). This flow is marked at ES2 by the onset of weak westerly winds with a jet structure and an observed peak wind speed at the 4-m level. At 2100 MST the flow in the gap becomes southwesterly (sodar observations, not shown). At this time the katabatic flow signature on the slope disappears and the wind speed increases at all levels of the tower.

Examination of AWSs early in the evening of 1–2 October indicate a growing stable nocturnal boundary layer in the east basin (not shown). Because of the confinement by terrain, this stable boundary layer forms a 150-m-deep CAP that includes a distinct 30-m-deep surface-based inversion.

At 2140 MST, a CAP has formed in the vicinity of GM as indicated by lapse rates calculated between surface stations on the valley floor and on top of GM (not shown). With the CAP in place and southwesterly flow over GM, the first LTF occurs. A 7°C air temperature drop is observed at the 4- and 10-m levels of the tower during this LTF. Less pronounced temperature drops are observed at 0.5 and 20 m. As a result, near-surface static stability decreases dramatically, consistent with the composite LTF structure.

The near-surface temperature and wind fields from surface observations in the vicinity of the east slope site are shown in Fig. 7 for periods before and after the second observed LTF, together with the results of the WRF simulation (shifted by 80 min). At 2300 MST, 50 min prior to the initiation of the second LTF, L1 is still inundated by the CAP (Fig. 7a). The flow through the gap at 150 m AGL is relatively strong (7 m s$^{-1}$) from the southwest (not shown) and L2 displays weak southwesterly winds. At 2345 MST (Fig. 7b), the flow at L1 is northerly. The temperature drop associated with the second LTF at ES2 begins at 2350 MST. At 0015 MST (Fig. 7c) the minimum temperature during the second LTF has nearly been reached. Northerly or northwesterly flow predominates at L2 and at areas lower on the slope. Only a moderate temperature drop is observed at L3.
After another brief warm-up, a third LTF begins at 0025 MST (see time series in Fig. 6). After this LTF the CAP fully retreats from all but the lowest portions of the east slope (Fig. 7d). The cold air is replaced by relatively warm southwesterly flow which initiates a warming trend that lasts for nearly 90 min at L2 and ES2 (Fig. 6a). Additionally, the near-surface stratification strengthens and a katabatic flow signature begins to reappear (Fig. 6b). The fourth LTF begins at 0210 MST as the CAP envelops ES2 again. After brief warming, a fifth marginal LTF occurs at 0305 MST. After this LTF, the CAP retreats from ES2. While small temperature fluctuations continue during the rest of the night, none meet the requirements for a LTF.

The signals in the TKE time series (Fig. 6d) vary widely among the LTFs. While large increases in TKE are observed at the onset of the first and fourth LTFs, other TKE spikes are observed that seem unrelated to LTF evolution. This agrees with a composite analysis of TKE during LTFs that did not reveal any consistent TKE patterns associated with LTFs (not shown). Vertical velocity (Fig. 6e) tends to change sign with the
onset of each LTF, particularly at the 20-m level of ES2. Similar to TKE, SHF (Fig. 6f) shows no consistent pattern during LTF events.

**WRF simulation of 1–2 October 2012**

Only one LTF occurred in the WRF simulation, corresponding to a combination of the second and third LTFs in the observations. We will not perform a systematic verification of the skill of WRF; rather, we present a single case to aid in diagnosing the causal mechanism of this LTF and, by extension, LTFs in general.

Figure 8 compares a time series of observations at L2 with data from the WRF grid point closest to L2. At L2, several air temperature fluctuations are superimposed upon the general temperature trend that can be represented by a smoothed time series. Averaging the air temperature observations to the same 20-min time interval for which WRF data are available produces much better agreement, although this indicates that there are variations in the CAP edge that are not resolved by WRF at 20-min sampling. The biggest wind direction discrepancies between the observations and the WRF simulation are seen between 2300 and 0030 MST. Wind speed discrepancies are largest immediately before and after the LTF simulated by WRF.

Figure 7a demonstrates that the pre-LTF conditions are well simulated by WRF with southwesterly flow...
through the gap, westerly flow over the east slope and northeasterly flow over the east basin. However, the lowest portions of both the west slope and east slope are warmer than the measured air temperatures in WRF. At 2345 MST (Fig. 7b), immediately prior to the initiation of the LTF, a vertically oriented vortex is visible in both the observations and WRF, with the CAP moving up the slope in the WRF simulation.

At 0015 MST (Fig. 7c) the CAP has enveloped L1 as the vertically oriented vortex advects cold air across the slope. Agreement between the observations and WRF is reasonable, although strong northeasterly winds are observed on the lower slope where WRF has weak northerly or northeasterly winds. Post-LTF conditions (Fig. 7d) display westerly winds and higher air temperatures on the east slope. Also, the low-level flow in the gap has reversed. This reversal is likely due to the propagation of the vertically oriented vortex into the gap (not shown). WRF resolves this reversal, although the observed flow soon returns to southwesterly, while WRF maintains the low-level northeasterly flow in the gap.

Based on the evidence presented in Fig. 7, the LTF is linked to an orographic wake and the associated vertically oriented vortex that forms in the lee of GM. WRF cross sections across GM at the latitude of ES2 are shown in Figs. 9 and 10 and help in investigating the orographic wake further. These cross sections represent the time just before the CAP reaches ES2. Relatively warm, strong flow can be seen descending the east side of GM (Fig. 9) with a positive potential temperature anomaly there. The flow on the west side of GM is blocked, with a flow reversal producing an easterly component of the flow in the lowest 100 m. Easterly flow is also present in the lowest levels of the CAP on the east slope, with relatively undisturbed southwesterly flow just above it. The spatial variation in the height of the CAP seen in Fig. 9 indicates that the fluctuations of the CAP edge along the slope are not only a result of a deepening or lowering CAP top, but that they are also affected by local erosion or depression (impinging wake) and advection (with the vertically oriented vortex) of the CAP edge.

The nondimensional mountain height $\epsilon$ can be used to test whether the conditions upstream of GM at this time would be conducive to blocked flow and the formation of an orographic wake (Epifanio 2003). Parameter $\epsilon$ is calculated here using a bulk method for the determination of $N$. At the initiation of this LTF, $\epsilon = 3$ because of a wind speed at 150 m AGL in the gap of 4.2 m s$^{-1}$ and a Granite Ridge to Playa Station temperature difference of 9$^\circ$C. Since GM has a $\beta$ value of approximately 3, this value of $\epsilon$ supports the occurrence of blocked flow upstream of GM and a lee wake.

7. Discussion

The analysis of the frequency of LTFs on the east slope of GM indicated that LTFs occurred most frequently when a stably stratified westerly or southwesterly nocturnal flow interacted with GM. These conditions resemble those covered by the idealized simulations of Epifanio and Rotunno (2005) where a flow with nonuniform stratification interacted with a uniform terrain barrier with $\beta = 3$. Nonuniform stratification implies that multiple layers with varying stratification comprise the flow. In the Epifanio and Rotunno (2005) case, a bulk calculation across the full flow depth yields $\epsilon = 4$. A nonuniform stratification is a better approximation of the conditions observed upstream of GM than uniform stratification below the ridgetop height, and may indicate that the nonuniform stratification observed at DPG is important for developing the orographic wake and vertically oriented vortex. In the simulation by Epifanio and Rotunno (2005), the low-level
flow is blocked on the upwind side of the barrier. Warming occurs on the lee side as air with a higher potential temperature descends the lee side of the mountain, replacing cold air near the leeside surface. This results in a strong potential temperature gradient at the base of the mountain that resembles a density current of cold air that is continuously propagating upward against the flow. Furthermore, Epifanio and Rotunno (2005) demonstrate that the lowering of the isentropes on the lee side relative to adjacent areas undisturbed by the barrier leads to positive buoyancy and baroclinic vorticity generation that induces the formation of counterrotating vortices on both ends of the barrier.

Our 1–2 October 2012 case study of LTFs on the east slope of GM, based on observations and a high-resolution WRF Model simulation, revealed evidence for an orographic wake and an associated vertically oriented vortex in the vicinity of the east slope that interacted with the CAP that had developed in the east basin. The vertical temperature gradient associated with the CAP, projected onto the slope, results in an along-slope temperature gradient. This temperature gradient can be magnified by the leeside warming due to the cross-barrier flow. Small changes in the speed and stratification of the flow approaching GM are expected to result in variations in the strength and position of the orographic wake and the vertically oriented vortex. The displacement of the CAP due to interactions with the evolving orographic wake and vertically oriented vortex leads to changes in the position of the strong temperature gradient along the slope, which results in LTF events observed at the surface stations along the slope. The signature in the evolution of the near-surface variables observed at the ES2 tower during the case study matches the signature of the composite of all LTF events. For example, this mechanism explains why the winds in the composite LTF at ES2 are consistently northerly or even northwesterly during the temperature drop. Typically, a deepening of the east basin CAP is observed during the night, and the associated along-slope temperature gradient is seen at higher elevations on the east slope. This explains why LTFs, produced by interactions between the CAP and the orographic wake and vertically oriented vortex, occur later in the night at higher elevations of the east slope. This means that CAP edge fluctuations that produce LTFs are not triggered by a variation of the basinwide depth of the CAP but rather by local processes along the CAP edge near GM. These local fluctuations will however, occur around a higher mean CAP depth as the night progresses.

To summarize, we have identified a mechanism based on the complex interactions of stably stratified southwesterly to westerly flow with GM that leads to an orographic wake and a vertically oriented vortex that displaces a CAP along the lee slope. This mechanism explains the formation, evolution, mean characteristics, and frequency of occurrence of LTFs on the east slope of GM.

We think that the identified mechanism is responsible for the majority of LTFs observed on the east slope of GM, despite the fact that the WRF simulation only reproduced a single LTF in our case study. The reason for this is the highly nonlinear response of the strength and position of the wake and vortex to small changes in the stability and directions of the flow interacting with GM. We would not necessarily expect WRF to simulate the exact number of LTF for this reason, although the horizontal and vertical resolution of the model and limitations of the PBL scheme may contribute to this shortfall.

Both the WRF simulations and observations under conditions with high LTF frequency show a clear resemblance to the idealized model results by Epifanio and Rotunno (2005). The WRF cross section shown in Fig. 9 clearly displays the upstream blocking, lee warming, and the density current at the base of the terrain, which are key features identified by Epifanio and Rotunno (2005). Convergent flows near the surface near ES2 in Fig. 10 show a vortex that advects the east basin CAP up the slope in the shape of a density current.

Observed mean values of $\epsilon$ and $\beta$ during LTF events indicate that conditions are conducive to developing an orographic wake. The average wind speed observed by the sodar at the 150-m level during LTFs is 3 m s$^{-1}$, and the average temperature at Granite Ridge is 6°C higher than at Playa. Assuming a mean barrier height of 500 m for GM, $\epsilon \approx 3$ under these conditions, meaning blocked flow is expected on the upstream side of GM, with potential for the development of an orographic wake. While assessment of $\epsilon$ in real-world situations can be difficult, the bulk stability method used here provides the best characterization of the incoming flow for situations where low-level blocking is expected (Reinecke and Durran 2008).

The presence of a CAP upstream of GM is likely necessary to produce the flow blocking that allows the orographic wake and vertically oriented vortex to form. Calculated values of $\epsilon$ in the UNBL alone, below GM but above the height of the CAP west of GM, shown in Fig. 3b, are rarely sufficient to generate the orographic wake. This indicates that the added stratification due to the CAP is important for initiating the flow blocking that caused the orographic wake to form. However, the CAP on the eastern (lee) side of GM remains relatively quiescent until it is displaced by the dynamically induced
flows, and seems to play no active role in the initiation of the complex flow interactions culminating in LTFs.

Dynamic terrain interactions have previously been advanced by Serafin et al. (2016) as a possible mechanism for temperature perturbations in the vicinity of GM. They identified high-frequency perturbations in the wind, air temperature, and pressure fields in WRF simulations and used them as a proxy for identifying areas of flow separation at DPG. Our results confirm that dynamic terrain interactions are the cause of temperature perturbations, at least on the east slope of GM. Serafin et al. (2016) also speculate that this is in fact the cause for the LTFs (referred to as “collisions”) discussed in Lehner et al. (2015), a conclusion that is also confirmed by this research.

Moderate sensitivity of the orographic wake to WRF model configuration was found when comparing two separate planetary boundary layer schemes. Simulations utilizing the Yonsei University planetary boundary layer scheme (Hong et al. 2006) produced an orographic wake and vertically oriented vortex at approximately the same time as the simulation utilizing the QNSE. However, the vertically oriented vortex formed at greater distance from GM and did not produce an LTF at the location of ES2. Additionally, the modeled east basin CAP was weaker than the one observed. We expect that since the development of LTF events on the east slope of GM are dependent on not just the presence but also the positioning of the orographic wake, vertically oriented vortices, and the CAP on both the west and east sides of GM, modeled LTFs are likely sensitive to model configuration. Accurate representation of the stratification of the CAP, stratification of UNBL, and of the boundary layer wind field are all critical to reproducing the orographic wake, the vertically oriented vortex, the CAP displacement, and the resulting LTFs.

Much of the analysis in this study has focused on LTFs observed at the L2 level on the east slope. CAP displacements by orographic wakes and vortices are likely the primary causal mechanism of LTFs at L1, L3, and L4 as well. This is due to the fact that LTFs are observed on many of the same nights at each AWS and that the composite LTF structure at those stations is similar to that produced by LTFs at L2 (not shown). The clear differences in hourly LTF frequency at L1–L4 are likely not attributable to a different mechanism, but to the deepening of the CAP through the night in the east basin. Observations at L2 exhibit more LTFs than the other AWSs likely because it is at an elevation corresponding to the mean depth of the CAP. The height of L1 corresponds to the CAP top only in the evening hours, leading to the evening peak in LTFs there (Fig. 2). The top of the CAP is almost always below the height of L4, and only very vigorous CAP displacements reach that level. If another mechanism were responsible for LTFs observed anywhere on the slope, it would likely be in the early
evening at L1. From 1800 to 1900 MST, the boundary layer may not be sufficiently stable to produce an orographic wake and vortex, or the southwesterly flow has not yet developed. LTFs at this time at L1 are likely due to the displacement of a shallow CAP by mechanisms other than the orographic wake and vortex mechanism.

Other areas of DPG not discussed here see high frequencies of LTFs (see Fig. 1). One location is at the northern tip of the Dugway Range. Based on evidence from the WRF simulation (not shown), this area is likely impacted by wakes forming in the lee of the Dugway Range in a similar fashion as those impacting the east slope of GM. A large number of LTFs are also observed at one AWS in the southern end of the east basin at a similar elevation to L3. It is possible that an orographic wake affects this location because of southerly flow impacting the mountain to its south, disturbing the CAP. A third AWS, on the northern end of the east basin, also exhibits a large number of LTFs. The location of this AWS is very close to the elevation of L3, indicating that CAP displacement may be the cause of LTFs there. However, the mechanism for CAP displacement there remains unknown.

8. Conclusions

This study analyzed meteorological data collected from more than 100 surface stations during the MATERHORN experiment that was conducted at Dugway Proving Ground in northwestern Utah in autumn of 2012. An automated algorithm was used to identify large temperature fluctuations, defined as a temperature drop of at least 3°C within 30 min followed by a partial temperature recovery, in a systematic fashion. Analysis of the spatial and temporal characteristics of LTFs showed that they are most common from 2200 to 0800 MST on the slopes of GM, with the frequency of LTFs on the east slope much higher than on the western slopes of GM.

LTFs occur most often under stably stratified westerly or southwesterly flow impinging on GM. This flow is blocked at low levels and an orographic wake with associated vertically oriented vortices is formed in the lee of GM. This wake and the vortices interact with the nocturnal CAP on the leeward side and intensifies the preexisting horizontal air temperature gradient along the slope that results from the vertical air temperature gradient of the CAP projected on the east slope of GM. Changes in strength or location of the orographic wake and the vertically oriented vortex lead to a displacement of the CAP and cause the location of this temperature gradient to fluctuate along the slope, leading to LTFs. This mechanism is likely to impact the entire east slope of GM.

The composite LTF compiled for the location with the highest frequency of LTF shows that the temperature drop is largest at the midlevels of a 20-m tower. This leads to a decrease in near-surface static stability during an LTF. Winds are typically westerly before the LTF and show the signature of a katabatic flow, northerly during the temperature drop, and then return to westerly after the temperature recovers. LTFs display amplitudes of 3°–13°C and occur on time scales of 30–90 min.
Numerical simulations of phenomena such as orographic wakes are frequently conducted using idealized or semi-idealized numerical simulations. This study offers unique observational evidence for the flows identified in numerical simulations. The uniquely dense observational network on the east slope of DPG made such a comparison possible, and further investigation of mountain-wave-type flows could be pursued using the available dataset.

The presented results offer MATERHORN investigators a much-improved understanding of flow interactions with GM that may be useful for future model verification and development. A logical extension of this study would be to conduct a thorough model verification and sensitivity study using the MATERHORN observational dataset. Sensitivity to boundary layer scheme, advective scheme, or other model parameters could be tested. The ability of WRF to simulate LTFs under different large-scale conditions could be assessed, as well as an analysis of the variability of the orographic wake from night to night.

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