Common Marginal Cold Pools

L. MAHRT AND ROBERT HEALD
NorthWest Research Associates, Redmond, Washington

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ABSTRACT

This study examines marginal cold pools forming in a shallow valley. “Marginal” refers to cold pools that are generally weak and intermittent. Nineteen stations were instrumented with sonic anemometers and accurate slow-response temperature measurements. The vertical structure of the cold pool is determined from a 20-m tower on the valley floor that includes eight levels of sonic anemometers and eight additional levels of slow-response temperature. On the basis of the data analysis, the traditional concept of a cold pool must be generalized to include cold-pool intermittency, complex variation of temperature related to some three-dimensional effects of terrain, and a diffuse cold-pool top. Different types of cold pools are classified in terms of the stratification and gradient of potential temperature along the slope. The strength of the cold pool is related to a forcing temperature scale that is proportional to the net radiative cooling divided by the wind speed above the valley. The scatter is large partly because of nonstationarity of the marginal cold pool in this shallow valley.

1. Introduction

With the onset of radiational cooling at the surface, the coldest air may form initially at the bottom of the slope in the basin or valley (Bodine et al. 2009) or the flow may initially descend to the basin or valley floor (Sheridan et al. 2014). Such drainage flow on the sideslope may descend the slope as a microfront (Papadopoulos and Helmis 1999; Mahrt et al. 2010; Fernando et al. 2013) or may form simultaneously along the entire slope. Cuxart et al. (2007) found both drainage flow over the cold pool and thin, cold skin flow that penetrated under the pool. Different scenarios are related to valley geometry and orientation because the initial cooling occurs on shaded slopes (Magono et al. 1982; Papadopoulos and Helmis 1999). Whiteman et al. (2004) concluded that cold-pool strength increases with the “sky-view factor” because sidewall radiation in deep narrow valleys (smaller sky-view factor) reduces the surface cooling.

The initial downslope flow may be associated with isentropic surfaces that are parallel to the slope without formation of a cold pool at the bottom (sketched in the bottom-left panel of Fig. 1). Most theories of drainage flow are based on this regime in which the variation of the potential temperature along the slope can be neglected.

After this initial development of cold-air drainage, a cold pool develops within closed basins, depressions, or valleys with little downvalley slope (Heywood 1933; Yoshino 1975), as sketched in the top panel of Fig. 1. The cold-air drainage on the sideslopes is not as cold as the air at the valley bottom and therefore flows over the top of the cold pool. The sideslopes are steeper than the downvalley slope, leading to stronger acceleration, shear-driven mixing, and downward heat transport (Helmis and Papadopoulos 1996). Drainage flows are also nonstationary because of both internal shear instability and externally forced variations such as propagating transient modes (Mahrt 2014). In addition, sideslope flows are more vulnerable to external disturbances (Horst and Doran 1986). Sheltering by the sideslopes reduces downward mixing of momentum and warmer air down to the valley floor and might be the main catalyst for cold-pool formation at the valley bottom (Sheridan et al. 2014; Vosper et al. 2014). In addition, downvalley cold-air drainage limits the cold-air accumulation in the valley and ultimately limits the gradient of potential temperature from the valley floor up the slopes. This situation is analogous to the negative feedback of buoyancy-driven upslope circulations.
that limit the decrease of potential temperature up the slope (Mahrt 2006).

In this sense, cold-pool formation is likely when the downvalley slope is much weaker than the sideslopes. From another point of view, the potential temperature above the valley floor increases vertically more rapidly than the increase of the potential temperature with height along the slope, as depicted in the top panel of Fig. 1. With superpositioning of multiple slopes on different scales, drainage flows with different temperatures reach different equilibrium levels above the coldest air in the valley (Monti et al. 2002).

For deep depressions, closed basins, and valleys with little downvalley slope, a mature cold pool may form (Fig. 1, bottom right) where the isentropic surfaces are approximately level and the height dependence of temperature on the slope is similar to that above the center of the basin (Whiteman 2000). Sheridan et al. (2014) found that strong cold pools in enclosed basins are generally well protected from wind events above the cold pool, although they can be weakened or eliminated with onset of cloudy conditions. The mature cold pool may be perturbed by disturbances trapped within the bottom of the cold pool (Mori and Kobayashi 1996), downward intrusions of warmer air from outside the cold pool (Yao and Zhong 2009; Adler et al. 2011), and flow of colder air from outside the basin into the cold pool (Whiteman et al. 2010; Haiden et al. 2011). Nonetheless, such well-defined classical cold pools are relatively robust with clear skies and weak large-scale flow. Although steeper sideslopes provide more sheltering, they also lead to back radiation into the cold pool (Whiteman et al. 2004).

Cold pools are often complex as a result of a number of features that are not included in the discussions above. Neff and King (1989) and Coulter et al. (1991) found that cold-air drainage in tributary valleys flows over the top of the colder air in the main valley. The generality of this behavior is not known. Drainage flow may develop within the cold pool separate from the
drainage flow above the cold pool (Mori and Kobayashi 1996; Vosper and Brown 2008; Bodine et al. 2009). Cold-pool formation may be encouraged when the valley outflow is retarded by opposing ambient flow (LeMone et al. 2003).

Even with well-defined valleys, three-dimensional terrain structure occurs simultaneously on different scales, partly as a result of small-scale undulations and gullies on the sideslopes. Small tributary valleys empty into larger valleys, which empty into still-larger valleys. As a result, surface air in a given valley may serve as a cold pool with respect to warmer air at higher levels but may also flow over even colder air in larger valleys farther down the slope. Exceptions occur where the outflow in tributary valleys is colder than air in larger valleys farther downslope, as sometimes occurs in the current study. Complications also occur as a result of bodies of water on the valley floor (Kossmann et al. 2002), merging gullies, and cross-valley asymmetry (Clements et al. 1989; Barr and Orgill 1989). Cold micropools may occur within larger colder pools. Much of the earth’s surface is characterized by relatively disorganized topography such that the local microclimatological conditions play a major role in addition to topography on larger scales (Acevedo and Fitzjarrald 2001; Medeiros and Fitzjarrald 2014). Katurji et al. (2013) provide an example of dynamical interaction between disturbances induced by local terrain features on different scales.

**Marginal cold pools** are defined as those that are intermittently eliminated multiple times during the night, partly as a result of background nonstationarity of even weak wind fields. Marginal cold pools can also be eliminated by intermittent cloud cover and transient development of unfavorable pressure gradients. The weakness of marginal cold pools can be related to shallow topography and loss of cold air as a result of downvalley drainage. We return to the definition of marginal cold pools in section 3 and in the conclusions. Elimination or disturbance of the cold pool by nonstationary wind (Stone and Hoard 1989; Mori and Kobayashi 1996; Gryning et al. 1985) includes local generation of oscillations within the cold pool (Chemel et al. 2009). Marginal cold pools correspond to the top panel in Fig. 1, but the top panel also includes developing mature cold pools.

Although shallow valleys are common, their nocturnal flow patterns have not been rigorously studied. Marginal cold pools can temporarily restrict dispersion of atmospheric contaminants in a way that is not currently predictable. Marginal cold pools are sometimes associated with rapid cooling, whereas their frequent demise arrests such cooling and sometimes leads to temporary warming. Forecasting minimum surface temperature becomes problematic. Increased understanding of marginal cold pools should improve forecasting of dispersion, frost damage, and fog formation.

This study examines the complex structure of the marginal cold pool on the basis of a network of 19 stations with sonic anemometers and a central tower within the shallow valley. Although conditions during the field program were often windy, numerous marginal cold pools formed, with various strengths and spatial structures. Although marginal cold pools cannot be completely described by just a few parameters, this exploratory study will seek a simple set of quantities as guidance for representation of the strength of the marginal cold pool and its forcing.

### 2. Data and averaging

The Shallow Cold Pool Experiment (SCP) was conducted over semiarid grasslands in northeastern Colorado in October and November of 2012 (Mahrt et al. 2014) and is described in detail online (https://www.eol.ucar.edu/field_projects/scp). The domain and tower locations are shown in Fig. 2, and a photograph of the main valley is shown in Fig. 3. The main valley is relatively small, roughly 12 m deep and 270 m across. The width of the valley bottom averages about 5 m with an average slope of 2%, increasing to about 3% at the upper end of the valley. This study primarily analyzes data from the National Center for Atmospheric Research hygrothermometers (1 Hz) deployed at the 0.5- and 2-m levels at 19 stations and at 0.5, 1.5, 2.5, 3, 4, 6, 8, and 15 m on the main tower. The temperatures from the sonic anemometers extend to 20 m and are used for evaluation of the vertical profiles. The sonic-anemometer temperatures were calibrated using profiles from the more reliable hygrothermometers. Periods of near-neutral stratification for which the vertical temperature difference is difficult to estimate are excluded when the vertical difference of potential temperature between 0.5 and 20 m is $\Delta \theta < 0.1$ K. The potential temperature is defined with respect to the height of the valley floor at the main tower where the potential temperature is then the actual temperature. The nocturnal net radiation was computed from the longwave components measured by a Kipp & Zonen B.V. model CG4 pyrgeometer at 1.5 m, located 17 m to the southeast (147°) of the main tower.

We currently include nocturnal data from 1700 LST in the evening to 0700 of the following morning. Various quantities are averaged over 5-min periods. Such short-term averaging leads to large scatter within the nonstationary environment of the shallow marginal cold pool. Use of larger averaging windows incorporates more nonstationarity within the averaging window that
obscures any systematic relationship between variables. For some analyses, we composite 5-min averages for different intervals (bins) of an independent variable, as described further in section 5a.

Four different subsets of data will be analyzed: 1) a 30-min case-study transition period leading to formation of a cold pool, 2) the subsequent 2-h period of a quasi-equilibrium cold pool and downvalley drainage, 3) all of the nocturnal 5-min periods with weak regional flow and significant net radiative cooling at the surface within the 2-month field program, and 4) all of the 5-min periods for the 2-month field program.
3. Basic formulations

a. Temperature scales

Although the concept of a cold pool is an idealization, we proceed to define diagnostic quantities for representing the thermodynamic structure of the marginal cold pool. Cold pools can be defined in terms of significantly larger potential temperature on the sideslopes relative to the potential temperature at the valley floor. As in Sheridan et al. (2014), we define the cold-pool intensity (CPI) as the surface potential temperature at a station on the slope above the valley floor minus the surface potential temperature at the valley floor such that

\[ \text{CPI} = \theta_A - \theta_{cp}. \]  

Here, \( \theta_A \) is based on the 0.5-m potential temperature at one or more of the A stations and \( \theta_{cp} \) is the 0.5-m potential temperature at the bottom of the cold pool, nominally taken as the 0.5-m temperature on the main tower (Fig. 2). Because CPI depends on the choice of stations, the concept of CPI is somewhat qualitative. CPI depends not only on its height above the valley but also on its horizontal position with respect to subvalley features. As a result, we avoid specifying a threshold value for qualification as a cold pool, although some quantitative evidence emerges in section 5. Air from the upland stations will ride over colder air in the valley if CPI is sufficiently large to compensate for any cooling of the parcel en route to the valley floor. Positive CPI does not preclude air from the slopes contributing to the cold-pool air, because the air may cool as it flows down the slope and colder skin drainage flow may develop within the lowest 1 m above the ground (Manins 1992; Mahrt et al. 2001; Nadeau et al. 2013).

The buoyancy generation of downslope drainage flow is proportional to the deficit of the surface potential temperature relative to the area-averaged potential temperature at the same absolute level, defined as

\[ \theta^* = \theta_A - \theta(Z_{\text{ref}}), \]

where \( Z_{\text{ref}} \) is defined as the absolute height of the 0.5-m level at a station above the bottom of the cold pool (Fig. 4) and \( \theta(Z_{\text{ref}}) \) is the “free” air potential temperature above the valley. The buoyancy deficit of the cold-air drainage is \( -\theta^* \). This calculation will assume that an area-averaged potential temperature at height \( Z_{\text{ref}} \) can be approximated by the profile of potential temperature on the main tower. The validity of this approximation is not known.

A stratification temperature scale is defined as

\[ \Delta_z \theta = \theta(Z_{\text{ref}}) - \theta_{cp}, \]

where again \( Z_{cp} \) is taken as the 0.5-m level on the main tower (Fig. 4). For brevity, we refer to the vertical difference of potential temperature as the stratification even though we have not divided by the vertical distance between the two levels.

As depicted in Fig. 4, these three temperatures [Eqs. (1)–(3)] are related by the relationship

\[ \Delta_z \theta = -\theta^* + \text{CPI}. \]

For example, cold-air drainage is generated by the buoyancy deficit of air on the slope \( -\theta^* \) if the CPI is less than the vertical potential temperature difference \( \Delta \theta \). Generation of cold-air drainage vanishes when CPI approaches \( \Delta \theta \) (isentropic surfaces are horizontal). The latter case is the mature cold pool at the bottom-right side of Fig. 1.

Three nondimensional ratios can be constructed from the three temperature scales: \( -\theta^*/\text{CPI}, -\theta^*/\Delta_z \theta, \) and \( \text{CPI}/\Delta_z \theta \), or their inverses. The first two are useful for describing the forcing of downslope drainage flow. The third ratio,

\[ R_{cp} = \frac{\text{CPI}}{\Delta_z \theta}. \]
defines the relative cold-pool strength and will be examined in this study. When $R_{cp}$ is less than unity, buoyancy generates downslope cold-air drainage, which can be alternatively shown by dividing Eq. (4) by $\Delta_c \theta$. This ratio is most meaningful if the CPI is defined in terms of the same levels as the stratification $\Delta_c \theta$. Three regimes can be posed in terms of the $R_{cp}$:

1) $R_{cp} \approx 0$, corresponding to the early evening (Fig. 1, bottom left) for those cases in which buoyancy forces generate cold-air drainage that reaches the valley floor (the isentropic surfaces are parallel to the slope, and there is no cold pool),

2) $0 < R_{cp} < O(1)$ (Fig. 1, top panel), corresponding to a cold pool where the buoyancy deficit generates downslope flow that does not reach the valley floor (the case includes, but is not limited to, marginal cold pools), and

3) $R_{cp} \rightarrow 1$, corresponding to a mature cold pool where buoyancy forces on the slope are negligible and the isentropic surfaces are horizontal (Fig. 1, bottom right).

Thus, $R_{cp}$ is inversely related to the slope of the isentropic surfaces, which is maximum (equal to the terrain slope) for vanishing $R_{cp}$.

**b. Forcing**

Relating the above three temperature scales [Eqs. (2)–(4)] or their ratios to the stability must be done cautiously to avoid self-correlation. For example, choosing the Richardson number as the stability parameter leads to self-correlation through the occurrence of stratification in both $R_{cp}$ and the Richardson number. The Richardson number is not an external variable in that it already contains information on the structure of the cold pool. Self-correlation can be avoided by choosing more-external variables. For guidance, we start with a simplified surface energy balance in which the net radiative cooling is balanced by the downward heat flux:

$$R_{net} = -\rho C_p C_h V \Delta \theta,$$

where $\rho$ is the density, $C_p$ the specific heat capacity, $C_h$ is the transfer coefficient for the surface heat flux, and $V$ is the wind speed. This hypothetical balance suggests a scaling temperature for $\Delta \theta$ that is proportional to $R_{net}/V$. For example, we define a cold-pool-forcing temperature scale as

$$CF = -C \frac{R_{net}}{V}.$$  \(7\)

The effect of the radiative forcing on the cold pool is reduced by the airflow above the cold pool and attendant downward mixing of warmer air. Using the surface energy budget, $C$ could be defined as $1/(\rho C_p C_h)$. Here we avoid numerical commitment to the oversimplified surface energy budget, however, and use the cold-pool forcing $CF$ only as a scaling variable that becomes small with windy, cloudy conditions and becomes larger with clear-sky, weak-wind conditions. In fact, $C$ could be highly variable and not even well posed when advection of temperature or vertical divergence of the radiative flux becomes important. These conditions are most likely with the strong stratification of the mature cold pool that severely restricts downward heat flux. For the study of the marginal cold pool, $C$ is chosen as $0.1 \text{ K m s}^{-1} \text{kg}^{-1}$ to make $CF$ be the same order of magnitude as the other temperature-difference scales.

For deeper basins, Yao and Zhong (2009) found that the cold pool was most intense with strong radiational cooling and weak surface flow outside the basin, whereas here we use the airflow above the valley; $CF$ should also include representation of the valley geometry. In addition, vegetation can reduce the cold-pool strength (Kiefer and Zhong 2013) and warm-air advection above the cold pool can increase the cold-pool strength (Lu and Zhong 2013). It is not clear how to formulate dependence on valley dimensions, however, and the network represents only one site.

To provide perspective on the limitations of the current analysis, we note that in addition to $CF$, the marginal cold pool is influenced by the details of the topographical geometry (closed basins, valley depth, sideslope, downvalley slope, and so forth). The marginal cold pool is modified by the direction of the flow above the valley with respect to the valley axis and by non-stationarity of the wind field on time scales of minutes or more. With more-windy conditions, the stratification above the valley may partly determine whether the cold pool forms (Sheridan et al. 2014).

**4. Structure of the marginal cold pool**

**a. Cold-pool formation**

We begin with a brief examination of the formation of a marginal cold pool. The CPI will be defined as the averaged potential temperature of the higher stations (A1, A4, and A14) minus the averaged potential temperature of the valley stations (A8, A11, and the main tower; see Fig. 2). We first select a case study period during 2000–2030 LST on the evening of 17 November 2012. At 2000 LST, the cloud cover is eliminated and the regional flow at 20 m decreases from a little more than $3 \text{ m s}^{-1}$ to a little less than $2 \text{ m s}^{-1}$. A cold pool forms during the period from 2000 to 2030 LST. The entire network begins to cool, but cooling is more rapid at
valley stations (red and black lines in Fig. 5a) than at the upland station (blue line). During the transition period, the CF (Fig. 5b, blue), the CPI (black), the 1–20-m potential temperature difference on the main tower ($\Delta\theta_u$; red), and the relative cold-pool strength ($R_{cp}$; green) all increase sharply. These indices reach quasi-equilibrium values during the 2-h period from 2030 to 2230 LST, subject only to submesoscale variations on time scales of minutes or tens of minutes.

The forcing CF (Fig. 5b, blue) increases from a value of less than 1 K to values that are typically between 3 and 4 K during the transition period. Because the CPI (Fig. 5b, black) increases faster than $\Delta\theta_u$ (red), $R_{cp}$ (green line) increases rapidly during the transition period. Strength $R_{cp}$ shows little trend between 2030 and 2230 LST, with an average value of about 0.55 (Fig. 5b). The value of $R_{cp} = 0.55$ during the quasi-equilibrium period is about one-half of the strength of the mature cold pool (approximately unity). Because $R_{cp}$ does not increase systematically with time after 2030 LST, it is not evolving toward a mature cold pool. The flow is instead maintaining quasi-equilibrium in which the loss of cold air as a result of cold-air drainage is compensated by local cooling associated with surface radiational cooling. This loss of cold air does not occur in enclosed basins, allowing easier formation of mature cold pools.

During the partial demise of the cold pool between 2230 and 2300 LST, CF (blue) decreases from approximately 4 K to a value of 2 K because of an increase in the regional flow. As a result, $R_{cp}$ decreases rapidly to about 0.2. The reduced CPI and reduced downvalley buoyancy force lead to temporary elimination of the cold-air drainage down the valley (not shown).

b. Spatial structure

The values of CPI and $R_{cp}$ depend on the choice of stations, because the top of the cold pool is not well defined and the temperature of the upland stations varies according to their microtopography superimposed upon the main valley structure. In fact, one upland station might indicate a significant cold pool in the valley while another cooler upland station might not.

Idealized regimes for a given station are defined in Fig. 6 on the basis of the relative height $Z$ of the station and its potential temperature relative to the potential temperature profile of the “free” air above the valley floor (red curve). With this type of plot, Sheridan et al. (2014) found conditions to be favorable for sideslope drainage flow in the early evening, followed by evolution toward horizontal isentropic surfaces (mature cold pool) in a valley that is more sheltered with weaker downvalley slope than the SCP site that is used here. Examples for the
Fig. 6. Three flow regimes that are based on the relative height and potential temperature for a given station, where \( Z \) is the height of the station above the valley floor. The mean temperature profile at the main tower in the valley is schematically indicated by the red curve. The vertical dashed line indicates the surface potential temperature measured at 0.5 m on the main tower at the valley bottom. These two lines define three regimes. For stations with potential temperature that is less than the vertical dashed line, the surface potential temperature is colder than that at the valley bottom and no cold pool is established, at least in terms of those stations in this regime. This situation corresponds to a negative cold-pool intensity (CPI < 0) and significant buoyancy generation of downslope flow (\( \theta^* < 0 \)) most like the regime in the bottom-left corner of Fig. 1. The middle regime between the vertical dashed line and the red curve corresponds to buoyancy generation of downslope flow (\( \theta^* < 0 \)) but also formation of the cold pool (CPI > 0). This regime corresponds to the top panel in Fig. 1. The regime in which the station elevation and potential temperature coincide with the red potential temperature profile above the valley floor corresponds to a vanishing buoyancy force (\( \theta^* = 0 \)), which is the mature cold pool (Fig. 1, bottom-right corner). Potential temperatures that are greater than the vertical profile of potential temperature (red curve) correspond to an upslope-directed buoyancy force that sometimes occurs after sunrise.

SCP data are given for the transition period (Fig. 7a) and the quasi-equilibrium period (Fig. 7b) for all of the 5-min periods in the entire field program for which CF is greater than 3.5 K (Fig. 8). This threshold value is chosen on the basis of the analysis in section 5 and Fig. 9. The value of CF for the case study period just exceeds this threshold value.

For upland locations with potential temperature to the left of the vertical solid line in Fig. 7, the buoyancy is negative because the potential temperature is colder than that at the same height on the tower such that \( \theta^* < 0 \). Generation of downslope flow is expected. Since the potential temperature of this air is less than that at the floor of the valley (CPI < 0), adiabatically descending cold-air drainage is expected to reach the valley floor without encountering a cold pool. During the transition period, this regime occurs with respect to the stations in the tributary gullies (Fig. 7a, red symbols) in which case the airflow from these locations is expected to reach the valley floor, as depicted in the bottom-left corner of Fig. 1.

At the same time, air on the valley sideslopes during the transition period (Fig. 7a, black symbols) is generally warmer than at the valley bottom and is thus expected to flow over the top of the air on the valley floor. With respect to the air on the sideslopes (but not in the tributary gullies; red), the valley is a cold pool corresponding to the regime between the vertical dashed line and the mean vertical profile in Fig. 6. That is, air is initially accelerated down the slope but may override colder air in the valley. The 1-m sonic-anemometer measurements and the 0.5-m measurements from the instruments manufactured by the Handar Company (now Vaisala, Inc.) (see Fig. 2) indicate that drainage flow is common in the tributary gullies but is less common on the valley sideslopes where it is often eliminated by the regional flow.

The average over the entire field program for CF > 3.5 K (Fig. 8) shows behavior that is between that of the transition period and that of the quasi-equilibrium period in that the cold pool is well defined with respect to the sideslope stations (black symbols) and the two highest upland stations (cyan), as occurs in the transition period, but not with respect to the air in the upvalley and tributary gullies (red), which have a potential temperature close to the air in the valley. This intermediate behavior between the transition and quasi-equilibrium periods is probably partly due to the fact that many of the marginal cold pools are quickly eliminated before reaching the quasi-equilibrium state. In addition, the example in Fig. 7 must be viewed with the reservation that every cold pool is different. For example, in the case study above, northeast regional flow extended downward and eliminated cold air in the northern tributary gully but not in the southern gully.

None of the marginal cold pools in this field program come close to the mature cold pool where the potential temperature of the stations above the valley floor would approach the ambient vertical profile of potential temperature. Whiteman et al. (2004) found that the temperature profile that is based on sidewall temperatures within a closed basin was only slightly cooler than at the same height on the vertical profile away from the sideslope corresponding to a mature cold pool, or close to the...
idealized red curve in Fig. 6. The flow within and outside the shallow valley is more complex and less well defined in comparison with that of the mature cold pool. In addition, the 0.5-m temperature at each station above the valley floor is also influenced by its microtopography.

The air at each location on the slope, in the tributary gullies, or above these gullies will seek a different equilibrium level above the valley floor. The vertical distance between the station potential temperature in Fig. 7 and the same potential temperature in the valley profile describes the amount of sinking that would be required for an idealized air parcel to reach its equilibrium level in the valley when approximately conserving potential temperature. If the airflow above the cold pool roughly conserves potential temperature (limited mixing) and is in fact directed toward the valley, airflow from different upslope regions will descend to different levels above the valley floor. The contribution of different source regions to different levels above the valley as well as vertical diffusion would contribute to the lack of a definite top to the cold pool, as revealed by the smooth observed vertical profile of potential temperature above the valley floor.

A number of processes influence the horizontal distribution of the surface potential temperature on scales smaller than the main valley topography. For example, station A4 (lower cyan point) is a degree warmer than A1 (upper cyan point) because it is on a more significant slope when compared with the relatively flat terrain around A1. Stations A4 and A9 (highest black times signs) may also be warmed by lee mixing in cases of northerly flow over the upland miniplateau. In summary, Fig. 8 indicates that, in gentle terrain, attempts to empirically model the change of surface temperature with height are complex (Mahrt 2006).

5. Relation of cold-pool structure to forcing

a. Dependence on forcing

The CPI is now examined in terms of its dependence on CF (Fig. 9a). All of the nocturnal 5-min periods are collected into intervals (bins) of CF, regardless of synoptic situation and wind direction with respect to the topography. Because of these and other influences not represented by CF and because of large nonstationarity for weak wind conditions, the interval averaging is very different from an ensemble average and the scatter is large, as discussed further below. The large number of
points visually masks the systematic trend identified in terms of the interval averages (red dots in Fig. 9a).

This trend indicates three regimes:

1) CPI roughly vanishes as CF vanishes. The stratification $\Delta_z \theta$ does not vanish as CF vanishes, as discussed below.

2) CPI increases with increasing CF up to $\sim 6$ K, depending on choice of stations. The scatter is enormous, partly because of nonstationarity and failure of the cold pool to reach equilibrium with the forcing.

3) On average, CPI does not increase further for CF $> 6$ K. Such large values of CF correspond to small values of the wind speed based on the 5-min averages. As a result, submesoscale wind shear on time scales of less than 5 min probably generates most of the turbulence. Further increase of CPI is also constrained by the loss of cold air as a result of downvalley drainage.

Points below the composited values (Fig. 9a, line with red circles) tend to be periods in which the cold pool is just forming or intensifying as a result of increasing forcing where CPI lags CF. Conversely, points above this line tend to correspond to decay of the cold pool because of decreasing forcing. The numerical values of CF correspond to the specific topography of the SCP site and the choice of stations, both in terms of height above the valley floor and horizontal position with respect to subvalley structure.

b. Stratification

The stratification $\Delta_z \theta$ shows a similar dependence on the forcing CF except that it retains a significant positive value as CF vanishes (Fig. 9b). Some of these cases of near-zero CPI but nonzero weak $\Delta_z \theta$ correspond to the early stage of cooling (Fig. 1, bottom left), but most of the cases correspond to windy, weakly stratified flow that appears to follow the terrain. As a consequence, $R_{cp}$, on average, is near zero for vanishing CF.

The stratification $\Delta_z \theta$ increases with increasing CF and saturates to an asymptotic value at a threshold value of CF $\approx 2$ K (Fig. 9b). This threshold value is smaller than that for CPI (Fig. 9a). As a result, $R_{cp}$ continues to increase with increasing CF until about 6 K. The stratification $\Delta_z \theta$ reaches its asymptotic value at a lower threshold of CF because downvalley cold-air drainage becomes common for CF that is greater than about 2 K (Fig. 10). The cold-air drainage limits additional cooling in the valley and thus restricts further increase of the stratification.

c. Large scatter

The standard error of CPI within different intervals of CF in Fig. 9 is generally less than 0.1 K for CF $< 5$. These small values are due to the large number of points, ranging from several hundred to more than 1000. The significance of individual interval averages remains somewhat uncertain, however, because the individual points are not independent and their distribution deviates from Gaussian, which is required for formal interpretation.
of the standard error. The computed standard error increases with increasing CF and becomes significant for CF > 5. Because CPI becomes approximately independent of CF for CF > 5, the interval width can be increased to reduce the standard error for each interval of CF.

The scatter and standard deviation are large for the entire range of CF (Fig. 9), however, partly as a result of influences that are not represented by CF and partly because of nonstationary wavelike motions and other more-complex transient structures. In particular, many of the cold pools forced by large CF never reach the quasi-equilibrium stage before CF decreases as a result of increases in the regional flow or cloud cover. The scatter decreases with increasing averaging time. The threshold values of CF become more obscure with increasing averaging time, however, because a larger averaging window more likely includes subperiods of CF that is both above and below the threshold value.

CPI is more closely related to the Richardson number than to CF because the Richardson number contains information on the vertical structure of the temperature within the cold pool. In a similar way, the scatter is reduced by using the wind speed within the cold pool instead of the wind speed at 20 m. The wind speed along the valley floor is part of the cold-pool structure, and the downvalley wind component is both a result of the cold pool and acts to limit the cold-pool strength. The CF provides a more external predictor of the cold pool. Scatter also appears to be caused by the dependence on direction of the regional flow with respect to the valley axis and associated wind direction shear.

6. Conclusions

Cold pools created by shallow topography are relatively weak and intermittent but still lead to colder temperatures, weaker turbulence, and restricted dispersion relative to flatter surfaces. Such shallow topography pervades much of the earth’s land surface. The current study examined marginal cold pools within a shallow valley that is about 12 m deep with a downvalley slope of 2%. Marginal cold pools are frequently eliminated during the night because of transient increases of wind speed above the valley to a couple of meters per second that are associated with normal background submesoscale motions. Marginal cold pools are vulnerable because they form in shallow topography and generally lose cold air as a result of weak downvalley drainage. Because of significant nonstationarity, well-defined relationships emerge only after compositing extensive data.

Although marginal cold pools are far too complicated to be described by a few parameters, some aspects of marginal cold pools can be summarized in terms of three temperature scales (section 3). The stratification \( \Delta_\theta \) is computed as the vertical difference of potential temperature between the valley surface and some height above the valley, here the 20-m level on the main tower. Parameter \( \theta^* \) is the buoyancy of the air on the slope with respect to the free air at the same height above the valley floor. The cold-pool intensity is defined as the average of the potential temperature over one or more surface stations above the valley minus the potential temperature of the air on the valley floor.

The relative cold-pool strength \( R_{cp} (=\text{CPI}/\Delta_\theta) \) is the ratio of the cold-pool strength to the stratification and is used to define different cold-pool regimes:

1) \( R_{cp} \) that is small relative to unity, with either positive or negative values, corresponds to isentropic surfaces that are approximately parallel to the sloped surface (Fig. 1, bottom left) where buoyancy forces generate cold-air drainage that descends to the valley floor (this regime can develop in the early evening before a cold pool develops),

2) \( R_{cp} \) that is greater than some small value but is significantly less than unity (Fig. 1, top) corresponds to a cold pool forming where buoyancy forces generate downslope flow that does not reach the valley floor but instead flows over the cold pool (for the current field program, \( R_{cp} \) averages about 0.3 and often falls within this regime), and

3) \( R_{cp} \) close to unity corresponds to a mature cold-pool flow developing where the isentropic surfaces are approximately horizontal, buoyancy forces on the slope are negligible, and flow in the valley or basin is very weak (such conditions, not found in the shallow
of the SCP domain, are more likely with enclosed basins and depressions).

The exact numerical values of CPI and $R_{cp}$ depend on the choice of stations used to represent the surface air temperature above the cold pool. The marginal cold pool does not have a definite top, and the air temperature of the upland stations varies significantly, partly as a result of the microtopography on horizontal scales of meters to tens of meters. This complexity was evident in section 4.

The cold-pool forcing is proportional to the net radiative cooling divided by the regional wind speed above the cold pool [Eq. (7)]. Significant regional wind speed leads to mixing and reduction or elimination of the cold pool. Several regimes can be roughly defined:

1) For very small forcing (small CF), CPI essentially vanishes but the stratification retains nonzero values, albeit small. This situation corresponds to $R_{cp}$ being close to zero with isentropic surfaces parallel to the slope. Although this regime includes some cases of initial formation of drainage flow in the early evening, it also includes windy cases in which the weakly stratified flow follows the terrain; this case is not pursued in this study.

2) CPI increases systematically with increasing CF until CF reaches a threshold value of $\sim6$ K, depending on choice of stations.

3) On average, CPI does not increase further with increase of CF beyond $\sim6$ K, partly as a result of loss of cold air through downvalley drainage. The strength of the downvalley drainage flow and the transient variations of the flow on submesoscales (minutes or tens of minutes) become important velocity scales, and the regional/synoptic wind speed is no longer the main velocity scale.

Marginal cold pools correspond to significant variability of CF and CPI. The CF presumably depends on the specific topography of SCP and the choice of stations, both in terms of height above the valley floor and horizontal position with respect to subvalley structure. The large scatter in the dependence of the cold pool on CF is partly due to the fact that the cold pool adjusts to changes in the net radiation and the wind aloft on a time scale that is longer than the 5-min averaging period. Partially formed cold pools do not reach the quasi-equilibrium stage when CF exceeds the threshold value for only very short periods (generally less than 30 min). Variations of CF are caused by variations of cloud cover and always-present variation of the wind field on small time scales. Use of a larger averaging time incorporates more nonstationarity within the averaging window and obscures the relationship between CPI and CF. Here, 5-min values were subsequently averaged over intervals of an independent variable such as CF.

CPI is more closely related to the bulk Richardson number (not shown) than to CF because the gradient of potential temperature along the slope is constrained by the vertical gradient of potential temperature that is also the numerator of the Richardson number. Here, the intention is to statistically relate the cold-pool strength to more-external variables such as net radiation and wind speed above the cold pool instead of the surface wind. The current study focused on a limited number of forcing and response variables, and the bin averaging is incomplete. The influence of the direction of the regional flow and smaller-scale three-dimensional terrain features within the domain must be examined. More explicit examination of the influence of transient submesoscale shear is complex and initially requires more-extensive case-study analyses. Such analyses would incorporate the Handar data, pressure measurements, and fiber-optic data whose locations are shown in Fig. 2.

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