Surface Turbulent Fluxes during Persistent Cold-Air Pool Events in the Salt Lake Valley, Utah. Part I: Observations

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ABSTRACT

The land surface is coupled to the atmospheric boundary layer through surface turbulent fluxes. Persistent cold-air pools (PCAPs) form in topographic depressions where cold, dense air fills the valley basin and in the presence of air pollution is accompanied by poor air quality. For the first time, the surface turbulence dataset from seven monitors during the Persistent Cold-Air Pool Study conducted in Salt Lake Valley, Utah (December 2010–February 2011), are analyzed. We found that the surface sensible (H) and latent (LE) heat fluxes were lower during strong PCAP events compared with non-PCAPs. The higher ratio of heat flux to net radiation (H/R_n and LE/R_n) for strong PCAPs compared with weak PCAPs is suspected to be related to the presence of boundary layer clouds, which could enhance the turbulent mixing through cloud top–down mixing. The daily average ground heat flux (G) was a similar order of magnitude to H and LE during wintertime. The highest surface turbulent fluxes and energy balance closure occurred in the stability range of $-0.05 < \xi \leq -0.02$, or under slightly unstable conditions, near the neutral stability range. The median surface exchange coefficient (C_h), a crucial parameter to determine surface turbulent fluxes in land surface models, was slightly higher at the bare land site (BL) than the short vegetation sites (PH and CR) in wintertime, suggesting the importance of dynamic land-use information in numerical models.

1. Introduction

Cold-air pools (CAPs) occur when cold air is topographically trapped within a valley or basin and with warmer air aloft. CAPs are associated with limited mixing and weak surface winds. Thus, persistent CAPs (PCAPs), which last for more than one day, can lead to poor air quality due to an accumulation of pollutant concentrations. PCAPs have been observed in the Salt Lake Valley, Utah, in the western United States (Lareau et al. 2013). They have also been documented in similar valleys in many other areas, such as the United Kingdom (Sheridan et al. 2014; Vosper et al. 2014), Germany (Zängl 2005), and Spain (Pagès et al. 2017). CAP formation is generally related to regional-scale anticyclones and high pressure subsidence. Besides the large-scale forcing, the boundary layer evolution during CAPs also depends on heat and humidity transferred from the ground through surface turbulent fluxes (Tokinaga et al. 2006).

In a broad view, land–atmosphere exchange has associations with weather phenomena of various scales

through surface heat flux variations. For example, in the synoptic scale, the anomalies of air–sea surface turbulent fluxes were related to the propagating cyclones in the Gulf Stream area (Zolina and Gulev 2003). The sensitivity of frontal development to land–vegetation processes, related to evapotranspiration, was demonstrated through numerical sensitivity experiments in the southern United States (Holt et al. 2006). In the meso-scale, the magnitude of land surface sensible heat flux impacts the horizontal temperature gradients and drives sea and lake breezes (Crosman and Horel 2010). In the microscale, a linear relationship between the atmospheric boundary layer (ABL) height and surface turbulent heat flux was found based on shipboard radiosonde surveys in east Japan (Tokinaga et al. 2006). The storm scale is impacted by the spatial variation of surface sensible and latent heat fluxes, where changes of land surface properties altered the deep cumulus convection during summer in the tropics (Pielke 2001). Thus, it is of interest, due to their impact on multiple scales, to analyze the characteristics of the surface turbulent fluxes during PCAPs.

Realistic simulations of surface turbulent fluxes benefit the prediction of weather and climate. In numerical

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models, surface turbulent heat fluxes are calculated based on Monin–Obukhov (M–O) similarity theory (Monin and Obukhov 1954) that depends on atmosphere stability. However, it has been reported that the land surface models sometimes cannot capture the magnitudes of the surface turbulent fluxes (Dirmeyer et al. 2018; Lee et al. 2011; Massey et al. 2017), especially during wintertime (Karsisto et al. 2016) and over complex terrain (Kalverla et al. 2016). Therefore, more efforts are needed to understand how the atmospheric stability impacts the surface turbulent fluxes based on observational datasets.

In addition to the surface atmospheric stability impacts, the surface exchange coefficient (C_h) is calculated based on land-use types and is one crucial parameter to calculate fluxes in land surface models. Its magnitude indicates the coupling strength between the land and atmosphere. C_h over tall canopy (short canopy) were reported to be underestimated (overestimated) in the Noah land surface model based on 12 AmeriFlux sites (Chen and Zhang 2009). It is important to investigate the C_h over multiple land-use types in different regions to assess model deficiencies. C_h can be calculated from the eddy covariance (EC) flux data using bulk transfer theory.

The Persistent Cold-Air Pool Study (PCAPS) (Lareau et al. 2013) was conducted in the Salt Lake Valley (SLV), Utah, from December 2010 to February 2011. Notice here the capital “S” to abbreviate the name of the field study, while the lowercase “s” used above defines the term “persistent cold-air pools” (PCAPs), this paper will use these two conventions throughout. PCAPS provides us with a unique surface turbulence dataset for almost 2 months over multiple land-use types to study the land–atmosphere exchange during CAPs. This paper is, to our knowledge, the first time data analysis using the surface turbulence dataset collected from PCAPS has been published. The main objectives are to understand the behavior of surface turbulent fluxes during PCAPs in wintertime and investigate the surface atmospheric stability impact on surface turbulent fluxes. The data from PCAPS used in this paper and methods for data analysis are described in the following section. In the results section, we characterize the surface turbulent and radiation fluxes during non-PCAPs and PCAPs and analyze the land-use impacts on surface energy fluxes focusing on surface albedo and surface exchange coefficients. A discussion section follows with a multisite analysis of the variability of the surface energy balance terms due to atmospheric stability. This paper provides valuable results that can be further used to evaluate the applicability of M–O similarity theory utilized in numerical

FIG. 1. Topographic map of the turbulence measurement locations (circles) of the PCAPS field campaign, the KSLC airport (triangle), and the Hawthorne (HW) site (asterisk) for air quality measurement in the Salt Lake Valley, Utah.

models in estimating surface turbulent fluxes under different stability ranges in wintertime over complex terrain.

2. Data and methods

a. Observation datasets from PCAPS

During PCAPS, the surface turbulent fluxes and radiation components, as well as 2-m temperature, 2-m humidity, and 10-m wind speed, were measured at seven sites in the Salt Lake Valley (Fig. 1) from 1 December 2010 to 7 February 2011 (Lareau et al. 2013). The 2-m temperature and humidity were measured by Sensirion humidity and temperature (SHT) hygrometers. The 10-m wind speed was measured with R.M. Young (RMY) prop-vanes. In total, 10 intensive observation periods (IOPs) were conducted based on when a PCAP event was expected using a forecasted potential temperature deficit threshold of 8 K from the valley bottom to the valley ridge. During the PCAPS experiment, the duration of the IOPs were fine-tuned based on real-time, hourly observations. Readers are referred to Lareau et al. (2013) for a detailed description of the CAP formation mechanisms of each IOP. The SLV is surrounded by the Oquirrh Mountains to the west, the Wasatch Mountains to the east, and the Traverse Mountains to the south. The Great Salt Lake is located at the

TABLE 1. Details of the seven turbulence monitoring sites used during the PCAPS field campaign. The altitude of the seven sites is 1320 m (ASL).

Code	Land use	Site description	Latitude (°N)	Longitude (°W)	Sonic anemometer height (m)	Soil moisture ($\text{m}^3 \text{m}^{-3}$)
BL	Barren land	Within a playa	40.78	112.10	3	0.30
DH	Developed, high intensity	Urban concrete surface	40.74	111.91	10	0.20
DM	Developed, medium intensity	Within residential area	40.70	111.84	10	0.31
DL1	Developed, low intensity	Within residential area	40.68	111.99	10	0.27
DL2	Developed, low intensity	Within residential area	40.61	111.80	10	0.14
PH	Pasture/hay	In rural area	40.59	112.06	3	0.23
CR	Cultivated crops	In rural area	40.51	111.97	10	0.17

northwestern end of the valley. Salt Lake City, with a population of 0.19 million estimated in 2011, is located in this valley, where the overall SLV population is over 1 million people. The valley is dominated by a semiarid climate. It has an annual average temperature of 10.56°C and precipitation of 1.27 mm, and snow depth of 6.35 mm day^{-1} (2010–11 mean), measured from the Salt Lake City International Airport (KSLC).

To investigate the surface heterogeneity and different land-use types in the valley, monitoring locations for surface turbulence and radiation during PCAPS consisted of one site over barren land (BL), four developed urban sites with different intensities [developed, high intensity (DH); developed, medium intensity (DM); developed, low intensity (DL)], one site over pasture and hay (PH), and one site with cultivated crops (CR) (Table 1) (Foster 2015). Compared to the four developed urban sites (DH, DM, DL1, and DL2), the BL, PH, and cultivated crop (CR) sites are classified as undeveloped sites due to their low population density. CR still had some residual vegetation on the surface during wintertime. The site description and average soil moisture at each site are also presented in Table 1. These flux stations, Integrated Surface Flux Systems (ISFSs), were set up by the National Center for Atmospheric Research (NCAR) Earth Observing Laboratory (EOL) In-Situ Sensing Facility (UCAR/NCAR–Earth Observing Laboratory 1990). The ISFS is designed to study the land–atmosphere exchange processes and has been deployed in multiple field projects (Lehner et al. 2016; Medeiros and Fitzjarrald 2014; Rhodes and Lundquist 2013). The ISFS data were also quality controlled and managed by the NCAR EOL.

All four components of radiation were measured at the PCAPS observation sites. Pyranometers and pyrgeometers measure near-surface solar irradiance and infrared radiation, respectively. Kipp and Zonen (K&Z) CM21 pyranometers and CG4 pyrgeometers were used at the DM, DL1, PH, and CR sites. Epply PSP pyranometers and PIR pyrgeometers were

deployed on the BL, DH, and DL2 sites. Comparisons of these different instruments were conducted by Kohnsiek et al. (2007) and no significant differences were found based on the radiation measurements in the Energy Balance Experiment (EBEX-2000). Thus, we did not consider the error associated with instrument differences here.

To obtain the sensible and latent heat fluxes, two instruments were used. Three components of wind speed and sonic temperature were measured by Campbell Scientific, Inc. (CSI) CSAT3 sonic anemometer at 10 Hz and a CSI krypton hygrometer was used to measure the water vapor fluctuations, at the same frequency. Surface turbulent fluxes during PCAPS were calculated based on the eddy covariance (EC) method (Arya 2001; Sun et al. 2003). Under steady-state conditions over a specific period of time, any variable can be decomposed into mean and perturbation components, that is, Reynolds decomposition. The typical averaging time for Reynolds decomposition is 30 min (Charuchittipan et al. 2014) and this was used to process the ISFS data. Using the EC method, H (LE) is calculated based on the product of vertical velocity fluctuations and potential temperature (specific humidity) fluctuations. As is typically required for sonic anemometer data, several correction steps were done during data processing of the ISFS data by the NCAR EOL. The correction of the sonic temperature was done considering the effect of moisture on the speed of sound (Schotanus et al. 1983). The krypton hygrometer measurements were corrected for the absorption by oxygen (Van Dijk et al. 2003). The Webb–Pearman–Leuning (WPL) corrections were made to the vertical fluxes of water vapor density (Webb et al. 1980). Corrections for the spatial separation of the krypton hygrometer and sonic anemometer, which was meant to reduce flow distortion, were made by considering the associated flux loss (Horst and Lenschow 2009). The sonic wind coordinates were rotated to correct the tilt of the sonic anemometer (Wilczak et al. 2001). ISFS

data processing also used a despiking algorithm from [Hojstrup \(1993\)](#) to ensure data quality.

Soil heat flux was measured by the CSI Radiation and Energy Balance Systems (REBS) HFT3 heat flux plate placed at a depth of 5 cm in the soil. In addition to the soil heat flux, the soil heat storage was also considered in the calculation of ground heat flux based on soil temperatures measured at four layers (0.6, 1.9, 3.1, and 4.4 cm). The heat capacity of soil, which was used to determine the soil heat storage, was measured by laboratory tests on soil samples gathered from the sites. Canopy heat storage is another term of the surface energy budget but is often negligible ([Onceley et al. 2007](#)) and was not quantified during PCAPS.

b. Methods

The bulk stability was determined using the valley heat deficit, which represents the heat required to warm the air in a column per unit area to the potential temperature at a specific height ([Whiteman et al. 2014](#)). Similar to [Whiteman et al. \(2014\)](#) and [Holmes et al. \(2015\)](#), we chose the integration height to be 2200 m (MSL), which is the ridge height of the Oquirrh Mountains. The valley heat deficit was calculated from twice-daily (0000 and 1200 UTC) radiosondes launched near KSLC airport using the equation below:

$$H22 = c_p \int_{\text{sfc}}^{2200} \rho(z) [\theta_{2200\text{m}} - \theta(z)] dz, \quad (1)$$

where c_p is the specific heat of air ($\text{J kg}^{-1} \text{K}^{-1}$), ρ is air density (kg m^{-3}), θ is the potential temperature (K), and z is the altitude (m). Valley heat deficit has been used to quantify the CAP strength and reported to have a nearly linear relationship with $\text{PM}_{2.5}$ ([Silcox et al. 2012](#); [Whiteman et al. 2014](#)). The daily average $\text{PM}_{2.5}$ data used in our study come from the Hawthorne site ([Fig. 1](#)) in the Air Quality System (AQS) operated by the U.S. Environmental Protection Agency (EPA).

The surface energy balance represents a fundamental approach for understanding the surface–atmosphere interactions. The energy balance ratio (EBR), the ratio of surface turbulent heat fluxes to available energy, was calculated to analyze the surface energy balance closure:

$$\text{EBR} = (H + \text{LE})/(\text{Rn} - G), \quad (2)$$

where H (W m^{-2}) and LE (W m^{-2}) are the surface sensible and latent heat flux, respectively. Rn (W m^{-2}) represents net radiation, and G (W m^{-2}) is the ground heat flux.

Besides the variations of surface energy partitioning impacted by soil moisture, different land–atmosphere coupling strengths are expected over different land-use types. Surface exchange coefficients can be calculated based on the bulk transfer relationship using the equations below ([Chen and Zhang 2009](#)):

$$H = \rho c_p C_h U_a (T_s - T_a), \quad \text{and} \quad (3)$$

$$\text{LE} = \rho L_e C_e U_a (q_s - q_a), \quad (4)$$

where c_p is the specific heat capacity of air ($\text{J kg}^{-1} \text{K}^{-1}$). The environment variables U (m s^{-1}), T (K), and q (kg kg^{-1}) represent the wind speed, temperature, and specific humidity, respectively. Subscripts s and a denote the surface and air values at 2 m, respectively, and L_e is the latent heat of vaporization (J kg^{-1}). The surface skin temperature (T_s) is estimated based on the upward longwave radiation, which is proportional to the fourth power of T_s . The surface emissivity used to calculate T_s is based on land-use type at each measurement site based on the 2011 National Land Cover Database. Because the surface emissivity is an approximation based on satellite remote sensing data, it may lead to uncertainties in calculating T_s . The dimensionless parameters, C_h and C_e , are the surface transfer coefficients, which quantify the land–atmosphere coupling strength. No measurements are available to determine the humidity gradient thus C_e and we focus on C_h in this study.

The energy balance terms variations with atmospheric surface stability were explored. Surface turbulent fluxes are generally parameterized in surface-layer schemes based on M-O similarity theory ([Monin and Obukhov 1954](#)), which describes the mean wind speed and mean temperature profiles as functions of a dimensionless length parameter (ξ) in the surface layer. This M-O stability parameter (ξ) was used here to determine atmospheric stability. In this paper, we use previously determined stability classes ([Franssen et al. 2010](#)) to stratify the results based on stability class and discuss the results in [section 4](#). The stability classes include very stable ($\xi > 1.0$), slightly stable ($0.025 < \xi \leq 1.0$), neutral ($-0.02 < \xi \leq 0.025$), slightly unstable ($-1.0 < \xi \leq -0.02$), and very unstable ($\xi \leq -1.0$). ξ is formulated as

$$\xi = \frac{z}{L} = \frac{zkgH}{-\rho c_p u_*^3 T}, \quad (5)$$

where z is the sensor height (m), L is the M-O length (m), k is the von Kármán constant (0.4), g is the gravitational acceleration (m s^{-2}), u_* is the friction velocity (m s^{-1}), and T is the air temperature (K).

FIG. 2. Twice-daily valley heat deficit (H22) at the Salt Lake City Airport and daily $\text{PM}_{2.5}$ concentration at site Hawthorne ($40^{\circ}44'\text{N}$, $111^{\circ}52'\text{W}$) during PCAPS. IOPs, corresponding to persistent cold-air pool events, are marked as shaded areas. Strong (weak) PCAPs that last for more (less) than 4 days are in dark (light) pink. IOP numbers are marked on the top.

3. Results

a. Valley heat deficit

The IOPs, corresponding to 10 PCAP events, are indicated by shaded areas in Fig. 2. The average H22 during the whole observation period was 3.30 MJ m^{-2} . The H22 during PCAPs (average 3.89 MJ m^{-2}) was more than 2 times higher than non-PCAPs (average 1.64 MJ m^{-2}). Using the bulk Richardson number method with a critical value of 0.25 (Sun et al. 2017), PCAPs had a lower average boundary layer height (average 113.10 m) compared with non-PCAPs (average 369.78 m). Higher $\text{PM}_{2.5}$ concentrations were observed during PCAPs (average $24.89 \mu\text{g m}^{-3}$) compared with non-PCAPs (average $6.93 \mu\text{g m}^{-3}$) due to the limited vertical mixing and lower boundary layer heights (i.e., decreased mixing volume) associated with the high atmospheric stability (H22). The Pearson correlation coefficient between $\text{PM}_{2.5}$ and CAP strength (represented by H22) was 0.71 in our case, suggesting the potential inimical impact of PCAP events on human health.

Based on the method given in Lareau et al. (2013), the CAPs that lasted for less (more) than four days were classified as weak PCAPs (strong PCAPs). There were four strong PCAPs and six weak PCAPs identified in the SLV during the 2010/11 winter. Strong and weak PCAPs had an average H22 of 5.08 and 2.70 MJ m^{-2} , respectively. This indicates that the classification based on CAP episode length is also physically related to the valley heat deficit, which is a commonly used variable to measure atmosphere stability in cold-air pool studies (Lu and Zhong 2014; Pagès et al. 2017; Whiteman et al. 2001).

b. Meteorological variables

The average wintertime daily mean 2-m temperature (T) over the seven sites during the more than two-month observation period was -0.54°C (Fig. 3a). Strong PCAPs (-2.35°C) had a lower average T compared with weak PCAPs (1.58°C), which was mostly related to one cold period that included IOP5. The developed sites (DH, DM, DL1, and DL2) had a higher average T than the undeveloped sites (BL, PH, and CR) for the whole observation period. The higher near-surface temperature in the developed areas could be attributed to anthropogenic heat sources that can be associated with higher surface turbulent fluxes.

The overall mean 2-m mixing ratio (r) was 3.36 g kg^{-1} (Fig. 3a). Variation of the daily mean r coincided with variation of T , which is probably related to surface evaporation. The strong PCAPs had the lowest average mixing ratio (3.23 g kg^{-1}). Due to the higher temperature in developed areas, the average r in developed sites (3.39 g kg^{-1}) was higher than undeveloped sites (3.28 g kg^{-1}).

The daily mean 10-m wind speed (WS) ranged from 1 to 7 m s^{-1} (Fig. 3b). The mean WS during PCAPs (1.99 m s^{-1}) was weaker compared with non-PCAPs (3.08 m s^{-1}). It is well known that the large-scale forcing can enhance WS. For example, a surface low pressure controlled SLV during 30 to 31 January and lead to an increase in WS, as well as a drop in T and r (Fig. 3a). In addition, strong winds, such as a gap flow or downslope windstorm, can also be dynamically induced due to topographic impacts. The SLV on 28 December was under a region of relatively high surface pressure compared with the surrounding areas. The strong pressure

FIG. 3. Time variations of the daily mean meteorological parameters: (a) 2-m temperature (T) and water vapor mixing ratio (r); (b) 10-m wind speed (WS) averaged over the seven ISFS sites, and twice-daily observed snow depth at the Salt Lake City Airport.

gradient in the SLV directed southerly flow into the valley and generated strong winds due to the mountainside funneling effect. When the flow descended at the lee sides of mountains, daily average temperature increased slightly on 28 December due to the adiabatic heating of air, suggesting the outbreak of foehn winds.

Snow is another factor that can impact the CAP evolution because the snow surface contributes to CAP persistence due to the increased surface albedo and decreased surface temperature (Whiteman et al. 2014). The low surface sensible heat flux that results from reduced net radiation because of the albedo increase cannot break the boundary layer stability from the ground (i.e., surface heating is weak). This is true during PCAPS since three out of the four strong PCAPS were snow covered (Fig. 3b). Note that the snow measurement data came from the KSLC airport. Spatial differences of snow coverage among the seven measurement sites existed based on the satellite images from the National Operation Hydrologic Remote Sensing Center (<https://www.nohrsc.noaa.gov/nsa/>). However, this spatial difference should not impact the overall mesoscale process that impact CAP formation. Additionally, the surface turbulent fluxes will be impacted by the spatial heterogeneity of the snow coverage, but this can be accounted for using the surface albedo (Table 2). However, the snow-covered period from

15 to 22 December was not identified as a PCAP period. Accompanied by the presence of a stationary front, the high wind speed, both at the surface (maximum 4.79 m s^{-1} , average 3.09 m s^{-1}) and at the valley ridge height (maximum 20.06 m s^{-1} , average 8.70 m s^{-1}), are suspected to preclude the inversion by the mechanism of turbulent erosion. Additionally, the strong winds at the valley top can yield air advection and contribute to the formation of Kelvin–Helmholtz waves, leading to the redistribution of heat and momentum (Lareau and Horel 2015). The less stratified atmosphere during this period corresponded to a lower H22 compared to other snow-covered periods (Fig. 2).

c. Radiative fluxes

1) DAILY VARIATIONS DURING THE WINTERTIME

The daily mean variations of the four-component radiative fluxes, as well as the daily minimum and maximum net radiation (R_n), during the whole PCAPS field campaign period are presented in Fig. 4. The mean downward shortwave radiation (DSR) over this 2010/11 winter was 80 W m^{-2} . DSR is determined by the solar zenith angle and atmospheric transmissivity, which is inversely proportional to the amount of clouds. The higher mean midday ($\pm 3 \text{ h}$ around the solar noon) cloud amount observed at the KSLC airport during non-PCAPS (6.2/10)

TABLE 2. Mean midday (± 3 h around the solar noon) values of downward shortwave radiation (DSR), upward shortwave radiation (USR), downward longwave radiation (DLR), upward longwave radiation (ULR), net radiation (Rn), and albedo (α) at the seven observation sites. Standard deviation ($\pm$) is shown in parentheses.

Site	DSR (W m^{-2})	USR (W m^{-2})	DLR (W m^{-2})	ULR (W m^{-2})	Rn (W m^{-2})	α
BL	263 (± 120)	114 (± 96)	268 (± 41)	316 (± 24)	108 (± 94)	0.43 (± 0.35)
DH	255 (± 119)	98 (± 96)	270 (± 41)	322 (± 23)	112 (± 90)	0.38 (± 0.36)
DM	254 (± 125)	119 (± 87)	273 (± 41)	323 (± 26)	91 (± 82)	0.47 (± 0.29)
DL1	277 (± 126)	129 (± 95)	270 (± 41)	323 (± 29)	97 (± 79)	0.47 (± 0.28)
DL2	251 (± 136)	124 (± 85)	275 (± 40)	317 (± 25)	94 (± 88)	0.49 (± 0.29)
PH	294 (± 118)	145 (± 88)	262 (± 41)	315 (± 25)	103 (± 78)	0.49 (± 0.25)
CR	291 (± 128)	145 (± 101)	268 (± 42)	323 (± 27)	97 (± 87)	0.50 (± 0.30)
Average	269 (± 124)	125 (± 92)	269 (± 41)	320 (± 25)	100 (± 86)	0.46 (± 0.31)

translated to its lower daily average DSR (average 66 W m^{-2}) compared with PCAP events (4.6/10, average 86 W m^{-2}).

Both the DSR and surface albedo impact the upward shortwave radiation (USR). The average surface albedo (0.61) and daily mean USR (average 53 W m^{-2}) were highest during strong PCAPs, which was related to its long-term snow cover and high DSR. The snow-cover period (15–22 December) during non-PCAPs contributed to its higher USR (average 33 W m^{-2}) compared with weak PCAPs (average 20 W m^{-2}).

Mean daily longwave radiative fluxes had larger magnitudes than the shortwave (Fig. 4). The average downward longwave radiation (DLR) during this winter was 271 W m^{-2} . DLR depends on cloud cover, the temperature of the atmosphere and water vapor concentration (Sicart et al. 2006). Daily variations of DLR and DSR showed distinctive patterns where they clearly mirrored each other (Fig. 4) due to the cloud impacts. Non-PCAPs had the highest average DLR (273 W m^{-2}), likely associated with its higher amount of clouds.

The average upward longwave radiation (ULR) during the whole observation period was 306 W m^{-2} . ULR was dominant over DLR, which led to a daily net longwave cooling of 34 W m^{-2} on average. ULR is proportional to the product of surface emissivity and the fourth power of surface skin temperature. Thus, the ULR variations can be interpreted as the surface skin temperature variations, which can be approximated by the near-surface T . Weak PCAPs had the highest ULR (average 316 W m^{-2}) due to its highest near-surface T .

Net radiation (Rn) is calculated by the sum of the four radiation components discussed above. The overall net shortwave radiative heating and net longwave radiative cooling generated an Rn of 13 W m^{-2} on average. Due to the low average Rn values in winter season, the daily maximum and minimum Rn are shown in Fig. 4. The average daily maximum (minimum) Rn was 157 W m^{-2} (-51 W m^{-2}), respectively. The daily maximum Rn was low during periods with snow coverage with an average value of 68 W m^{-2} . Rn provides available energy to drive H and

FIG. 4. Time variations of the daily mean four component radiation fluxes, including downward shortwave radiation (DSR), upward shortwave radiation (USR), downward longwave radiation (DLR), and upward longwave radiation (ULR) and daily maximum and minimum net radiation (Rn) averaged over the seven ISFS sites.

FIG. 5. Average diurnal variations of (a) downward shortwave radiation (DSR), (b) upward shortwave radiation (USR), (c) downward longwave radiation (DLR), (d) upward longwave radiation (ULR), and (e) net radiation (Rn) under non-PCAP, weak PCAP, and strong PCAP events over the seven ISFS sites. Error bars indicate the standard deviations.

LE that govern the heat and humidity exchange between the land and atmosphere.

2) DIURNAL VARIATIONS DURING NON-PCAP, WEAK PCAP, AND STRONG PCAP

The average diurnal variations of radiation components under non-PCAPs, weak PCAPs, and strong PCAPs are presented in Fig. 5. Peak shortwave radiative fluxes occurred around noontime, as well as the largest shortwave radiation differences in the three

scenarios. The mean midday value of wintertime DSR was 273 W m^{-2} . The lower mean midday DSR of non-PCAPs (228 W m^{-2}) compared with PCAPs (295 W m^{-2}) was related to its higher average cloud amount as discussed in the last subsection. Although the midday average cloud amount during strong PCAPs (5.1/10) was higher than weak PCAPs (4.2/10), the mean midday DSR difference between weak and strong PCAPs was relatively small (2 W m^{-2}). It is possible that the clouds were thinner during strong

FIG. 6. Time variations of the daily maximum and minimum surface turbulent fluxes (H and LE), ground heat fluxes (G), and daily mean friction velocity (u_*) averaged over the seven ISFS sites.

PCAPs, yielding more transmittance of solar radiation.

Similar to the daily means of DSR, strong PCAPs had the highest midday USR, with an average value of 181 W m^{-2} . Weak PCAPs generated a lower USR (midday average 64 W m^{-2}) compared with non-PCAPs (midday average 109 W m^{-2}). The different midday USR variations in the three cases compared with DSR (i.e., weak PCAP had a higher midday DSR than non-PCAP) suggest the importance of the surface albedo impact.

There were no obvious diurnal variations of DLR observed in all the three cases, indicating less dependence on temperature in our case. The diurnal variations of ULR exist, where peak values also occurred around noontime, but were much less significant compared with the shortwave radiative fluxes. Unlike the large midday differences in DSR and USR from the three cases, the differences in ULR from the three cases remained relatively constant in the daytime and nighttime. The weak PCAP had the highest peak ULR with an average midday value of 339 W m^{-2} .

Rn was dominated by the net shortwave radiation at midday with an average value of 112 W m^{-2} and dominated by the net longwave radiation at midnight with an average value of -25 W m^{-2} . The weak PCAPs had the highest Rn at midday with a net shortwave radiation of 224 W m^{-2} and a net longwave radiative cooling of -73 W m^{-2} . At midnight, Rn was largely impacted by the surface temperature and strong PCAPs had relatively weak radiative cooling due to their low surface T .

3) SPATIAL VARIATIONS DURING THE WINTERTIME

The land-use impacts on radiation components are addressed here. Similar to Loridan and Grimmond

(2012), the midday radiation fluxes were used to characterize the energy transfer from the surface to the atmosphere in Table 2. DSR was the dominant contributor to midday Rn. Developed sites received less shortwave radiation compared with undeveloped sites, probably because of the absorption and scattering of shortwave radiation from the higher aerosol loadings associated with developed areas and the higher building densities (Bornstein 1968).

The average midday albedo (α) for all seven sites during the observation period was 0.46. Developed sites had a lower average α than undeveloped sites. The BL site had a much lower albedo than most of the developed sites (except for DH), which might be attributed to its land cover and high soil moisture content (Table 1). The daily average DLR over all sites was 269 W m^{-2} . The higher surface temperature at the developed sites led to their overall higher ULR compared to the undeveloped sites. All the measurement sites had more longwave radiation emitted than received at the surface.

d. Turbulent fluxes

1) DAILY VARIATIONS DURING THE WINTERTIME

Same as Rn, the daily maximum and minimum variations of surface turbulent fluxes, including sensible heat flux (H), latent heat flux (LE), and ground heat flux (G), as well as the daily mean friction velocity (u_*), during this wintertime are presented in Fig. 6. The average H (4.46 W m^{-2}) during the whole observation period were lower than LE (10.84 W m^{-2}), but $|H|$ was larger than $|LE|$. This indicates that the energy exchange associated with heat transfer at the surface was more active than the energy related to evaporation during wintertime. The daily maximum H (average 48.99 W m^{-2}) and LE

FIG. 7. As in Fig. 5, but for (a) sensible heat flux (H), (b) latent heat flux (LE), and (c) friction velocity (u_*).

(average 39.16 W m^{-2}) had a similar magnitude most of the time, while the magnitude of minimum H (average -31.27 W m^{-2}) was larger than LE (average -3.90 W m^{-2}).

G is an important term in the surface energy budget. However, compared with H and LE, less focus has been given to the measurements and studies of G . Neglecting the G term can lead to the degradation of the surface energy balance closure. The daily maximum G (average 16.13 W m^{-2}) was smaller than H and LE. The daily minimum G (average -14.04 W m^{-2}) was of the same magnitude of H and larger than LE. Overall, the average G during the wintertime was 6.33 W m^{-2} and was comparable to the magnitudes of H and LE. This indicates the importance of including G in the surface energy budget during wintertime when the surface turbulent fluxes are small. The daily maximum and minimum G were nearly the same (both close to zero) during periods with snow coverage. This is expected since the snow layer can act as an insulator (Zhang 2005) and leads to a lower G .

The daily variations of u_* during the wintertime fluctuated around 0.2 m s^{-1} . The mean friction velocity (u_*) was 0.21 m s^{-1} . Non-PCAPs had the highest daily

mean u_* (0.28 m s^{-1}) and strong PCAPs had the lowest (0.15 m s^{-1}), indicating dampened surface mechanical mixing during CAPs.

2) DIURNAL VARIATIONS DURING NON-PCAP, WEAK-PCAP, AND STRONG PCAP

The diurnal variations of surface turbulent fluxes (Fig. 7) show that H dominated over LE during daytime but had the opposite sign during nighttime, which made its daily mean value lower than LE. The midday ($\pm 3 \text{ h}$ around the solar noon) and midnight (12 h after midday, $\pm 3 \text{ h}$ average) average values of H were 38.29 and -8.01 W m^{-2} , respectively. Negative H suggests that heat was transferred from the atmosphere to the ground due to surface radiative cooling. LE remained positive all day, with an average value of 23.57 and 5.21 W m^{-2} at midday ($\pm 3 \text{ h}$ around solar noon) and midnight (12 h after midday, $\pm 3 \text{ h}$ average), respectively. The positive LE indicates that on average the surface lost heat due to evaporation and/or sublimation all day. At nighttime when the incoming energy was low, the magnitudes of H and LE were highest during non-PCAPs and lowest in strong PCAPs. However, when R_n began to increase after sunrise, the incoming energy and atmospheric

TABLE 3. Mean midday (± 3 h around solar noon) values of surface heat fluxes (H and LE), friction velocity (u_*), net radiation (R_n), and flux ratios (H/R_n and LE/R_n). Standard deviation ($\pm$) is shown in parentheses.

Terms	Non-PCAPs	Weak PCAPs	Strong PCAPs
H (W m^{-2})	38.11 (± 54.26)	51.30 (± 44.08)	25.46 (± 29.99)
LE (W m^{-2})	26.26 (± 23.96)	31.54 (± 16.03)	12.92 (± 14.76)
u_* (m s^{-1})	0.34 (± 0.23)	0.24 (± 0.14)	0.19 (± 0.12)
R_n (W m^{-2})	98 (± 89)	166 (± 85)	72 (± 86)
H/R_n	0.40	0.32	0.39
LE/R_n	0.32	0.20	0.21

stability together impact the surface turbulent flux magnitudes by providing more available surface energy and enhancing (unstable)/suppressing (stable) surface turbulence, respectively.

The midday H and LE were suppressed in strong PCAPs compared with non-PCAPs (Fig. 7). Weak PCAPs did not show a smaller midday H and LE compared with non-PCAPs. The higher H and LE during weak PCAPs are attributed to its higher R_n compared with non-PCAPs. Statistics of the surface flux ratios (Table 3) shows that even though weak PCAPs had higher H than non-PCAPs, the ratio H/R_n for weak PCAPs was lower than non-PCAPs. This means the partitioning of R_n to H was subdued under weak PCAPs. The distribution of surface energy fluxes exhibited an approximate normal distribution (versus lognormal distribution). Thus, the statistics shown in Table 3 can appropriately represent the characteristics during different PCAP scenarios. A possible explanation of the higher H/R_n ratio of strong PCAPs (0.39), together with its higher cloud coverage, compared with weak PCAPs (0.32) is the boundary layer cloud impacts. Boundary layer clouds could modify the thermodynamic structure of the boundary layer through cloud top-down convection and thus enhance the surface turbulence intermittently. For example, in IOP 9, widespread and persistent stratocumulus clouds were present over SLV in the later phase of the CAP (after 28 January) (Crosman and Horel 2017; Lareau et al. 2013). The midday average R_n decreased from 268.43 W m^{-2} on 27 January without cloud presence to 206.74 W m^{-2} on 29 January with cloud presence, while the midday H increased from 60.64 to 75.69 W m^{-2} . Therefore, the ratio of H/R_n increased from 0.23 to 0.37.

The average midday u_* (0.26 m s^{-1}) was 43.96% higher than at midnight (0.18 m s^{-1}). The diurnal variation of u_* was less pronounced during strong PCAPs during daytime. For all CAP cases, the diurnal variation of u_* differences was not as significant as the surface heat fluxes, that is, the midday and midnight u_* differences

TABLE 4. Mean midday (± 3 h around the solar noon) values of flux ratios (H/R_n and LE/R_n), energy balance closure (EBR), Bowen ratio (β), and friction velocity (u_*) with standard deviation in parentheses at the seven observation sites.

Site code	H/R_n	LE/R_n	EBR	β	u_* (m s^{-1})
BL	0.30	0.24	0.46	1.26	0.19 (± 0.12)
DH	0.38	0.18	0.49	2.13	0.28 (± 0.13)
DM	0.42	0.23	0.62	1.82	0.27 (± 0.17)
DL1	0.38	0.23	0.61	1.69	0.34 (± 0.15)
DL2	0.26	0.16	0.43	1.61	0.22 (± 0.14)
PH	0.21	0.23	0.45	0.93	0.20 (± 0.13)
CR	0.41	0.29	0.69	1.42	0.35 (± 0.15)
Average	0.34	0.22	0.54	1.55	0.26 (± 0.15)

tended to be nearly constant. The average daily maximum u_* was only 0.20 m s^{-1} during strong PCAPs.

3) SPATIAL VARIATIONS DURING THE WINTERTIME

The net radiation partitioning into sensible and latent heat flux depends on land-use type, of which soil moisture is one important impactor (Yuan et al. 2017). R_n was partitioned mostly to sensible heat flux (0.34 on average) during the midday (Table 4). One exception was the site PH, where the contribution of evaporation process (LE/R_n) was slightly higher than the sensible heat flux (H/R_n) due to its high soil volumetric water content ($0.8 \text{ m}^3 \text{ m}^{-3}$). Urbanization impacts the energy partitioning between sensible and latent heat fluxes due to less vegetation and extra heat generated by human activities. The average Bowen ratio ($\beta = H/LE$) of the four sites in developed areas (1.81) was higher than the three sites in the undeveloped areas (1.20). Specifically, β was lowest (0.93) at the PH site where the land use is classified as pasture and hay.

The EBR, which indicates the surface energy balance closure, also has spatial variations. The highest EBR was observed at the CR site (Table 4), which had less heterogeneous surfaces compared with developed sites. It is hypothesized that surface heterogeneity could induce secondary circulation and generate eddies at a larger time scale than those measured by the EC method (Foken et al. 2006; Maronga and Raasch 2013). EBR can also be impacted by atmospheric stability and will be discussed in section 4.

C_h , indicating the coupling strength between the land and atmosphere, also depends on land-use type. Considering the most active land-atmosphere exchange occurs during daytime, the midday values (± 3 h around solar noon) of C_h were investigated (Chen and Zhang 2009). The box and whisker plots of C_h (Fig. 8) show that the means are higher than the medians. The average median value of midday C_h over the seven observation

FIG. 8. Midday values (± 3 h around solar noon) of C_h derived from PCAPS observations over seven sites with different land-use types. The y axis is in log scale. The square dots and solid lines in the boxes indicate the mean and median values, respectively. The lower and upper limits of the bars represent the 25th and 75th percentiles. The whiskers represent the 10th and 90th percentiles.

sites in the SLV was 2.11×10^{-3} . The mean value of C_h derived from these winter observations was the same order of magnitude but smaller compared to AmeriFlux network based values for short vegetation surfaces during spring (around 4.0×10^{-3}) (Chen and Zhang 2009). This is expected since vegetation activities during growing season would yield a rougher surface and thus a higher C_h . For example, smooth water surfaces have a lower C_h (1.1×10^{-3}) (Xiao et al. 2013) compared with our case. The median C_h for the four developed sites (2.60×10^{-3}) was nearly 2 times the value for the three undeveloped sites (1.47×10^{-3}). The higher C_h is attributed the higher roughness lengths at the developed sites. The bare land site (BL) had a slightly higher C_h than short vegetation sites (PH and CR). The C_h difference between bare land and short vegetation during a wintertime nongrowing season in our case is less pronounced than growing season (Chen and Zhang 2009).

4. Discussion

Our results show that strong PCAP events had lower magnitudes of surface turbulent fluxes compared with non-PCAPs. This suggests that the surface turbulent fluxes, normally in higher measurement frequency (half hourly) compared with atmospheric soundings (twice daily), might be useful to quantify the CAP strength. High-temporal-resolution surface turbulent fluxes would also be valuable to investigate the atmospheric physics changes throughout the day during CAPs compared with radiosonde observations. Understanding the CAP

turbulent fluxes would require more comprehensive investigations based on long-term observations. Meanwhile, numerical models need to be able to capture the subdued surface turbulence for accurate representation of the strong PCAP evolution. One interesting thing we found is that the average midday surface turbulent fluxes were even higher during weak-PCAP compared with non-PCAP. The weak (strong) PCAP had an average H_{22} of 2.70 MJ m^{-2} (5.08 MJ m^{-2}), which is lower (higher) than the threshold of 4.04 MJ m^{-2} that corresponds to daily $\text{PM}_{2.5}$ values greater than $17.5 \mu\text{g m}^{-3}$ (i.e., half of the $\text{PM}_{2.5}$ standard) based on a climatology study by Whiteman et al. (2014). This suggests that CAP events that last for more than 4 days (strong PCAP) should have more focus for air quality considerations.

The boundary layer cloud impacts that reduce R_n but enhance surface turbulence due to cloud top-down mixing has also been suggested by Holmes et al. (2015). Large-eddy simulation of the cold-air pool event of IOP 9 by Crosman and Horel (2017) also simulated increased surface sensible heat flux with the presence of low clouds. However, they briefly explained it as the “insulating” impact of clouds and did not compare their results with the ISFS surface turbulent flux data. This cloudy top-down turbulent mixing implies that dry CAPs might lead to more serious air pollution conditions compared with cloudy CAPs. Indeed, VanReken et al. (2017) found decreased primary pollutant concentrations with the onset of boundary layer clouds in the Yakima Valley, Washington. Interestingly, we did not observe enhanced u_* during strong, cloudy PCAPs compared with weak PCAPs. One possible explanation is that the boundary layer cloud impacts are associated with cloud-top radiative cooling, where the cooling causes an instability at the top of the PBL and induces turbulent mixing by modulating the vertical thermal profile, but impacts the surface momentum transfer less, therefore having a lag effect on u_* . This is one case study and supports the top-down mixing found in Holmes et al. (2015). However, the boundary layer cloud impacts on the surface turbulence during PCAPs need further study in future investigations where turbulence datasets are available with more spatial and temporal coverage.

Spatial variations of H are related to varied C_h at different sites. In surface-layer schemes, the roughness lengths for heat (z_{0h}) and momentum (z_{0m}) are the two crucial elements that depend on land-use types in the C_h parameterizations. The seasonal values of z_{0m} are a function of the green vegetation fraction (GVF) in the Noah LSM, which is one of the most commonly used LSMs in the numerical modeling community. z_{0h} is calculated from z_{0m} with an empirical coefficient determined

FIG. 9. Distributions of (a) surface sensible heat flux (H , W m^{-2}), (b) surface latent heat flux (LE , W m^{-2}), (c) net radiation (R_n , W m^{-2}), and (d) energy balance ratio (EBR) as a function of stability parameter (ξ). The upper limit of each stability bin is given along the x axis. The lower limit stability value is the preceding value.

by the canopy height (Chen and Zhang 2009). However, it has been suggested that the static GVF dataset used in the Noah LSM cannot represent the vegetation growth well and thus hampers the realistic simulation of precipitation over the central United States (Xiaoyan et al. 2009). Zheng et al. (2014) reported that using 10-day normalized difference vegetation index (NDVI) products from satellite data to derive real-time GVF in Noah LSM gave satisfactory z_{0m} values compared with observations collected at the Tibetan Plateau. Chen and Zhang (2009) found that using realistic canopy heights improved the simulated C_h in the Noah model. This suggests the importance of including dynamic land-use datasets in numerical models to get more realistic C_h .

To investigate the impact of atmospheric surface stability on surface turbulent fluxes, which is parameterized in surface-layer schemes in numerical models, the observations were stratified by the M-O stability parameter (ξ), as shown in Fig. 9. We found that the ξ during PCAPs were not higher than non-PCAPs, that is, a high bulk atmospheric stability ($H22$) does not correspond to a high surface stability (ξ). Here we include the whole wintertime observation data to discuss the surface atmospheric stability impact on turbulent fluxes. Using these box and whisker plots we can see that the maximum (90th percentile) H and LE occurred in the range of $-0.05 < \xi \leq -0.02$ under slightly unstable conditions. The large variability of R_n during periods of unstable ABL ($\xi \leq -0.02$) suggests the presence of boundary

layer clouds. Interestingly, there is a weak variation of H with increasing instability. This is suspected to be related to the boundary layer cloud impact, which leads to enhanced H with reduced R_n in the presence of boundary layer clouds. This suggests that both CAP scenarios (PCAP and non-PCAP) and CAP types (cloudy and noncloudy) impact the surface turbulent fluxes.

Uncertainties of the EC method to measure surface turbulent fluxes exist. This is illustrated through analysis of the EBR (Fig. 9). The median EBR reached a maximum of 0.56 under slightly unstable conditions for $-0.05 < \xi \leq -0.02$, which was near the neutral stability range ($-0.02 < \xi \leq 0.025$). As the atmosphere became more stable or more unstable, the median EBR became smaller. Similar EBR variation with surface stability conditions were also found by Franssen et al. (2010) for European Flux Network (FLUXNET) stations and Barr et al. (2006) for boreal forest sites. Under very unstable conditions, larger eddies with lower-frequency turbulence may occur and lead to degraded EBR (Franssen et al. 2010). Under very stable conditions, the 30-min Reynolds averaging time may not be appropriate for capturing the intermittent turbulence that occurs under very stable conditions (Mahrt 2010).

5. Conclusions

In this study we presented the results of the surface meteorological variables, surface turbulent fluxes,

and radiation fluxes from the wintertime Persistent Cold-Air Pool Study in the Salt Lake Valley, Utah. The cold-air pool (CAP) events were associated with low near-surface temperature, weak wind speed, and stable ABL periods (high valley heat deficit). Strong winds, induced by synoptic systems and/or mountain waves (e.g., foehn wind), can enhance the mechanical mixing and preclude or weaken the CAP. Snow cover was observed in three of the four strong persistent CAPs (PCAPs) during the 2010/11 winter field campaign. We found that the snow-cover impact, accompanied with reduced surface available energy, can sometimes be offset by strong winds from the valley top and produce a well-mixed boundary layer through turbulent erosion mechanisms without CAP formation.

Quantification of the surface turbulent fluxes during PCAPs is important since they can directly modulate the boundary layer structure and impact PCAP evolution. We found that the daytime surface turbulent fluxes were lower under strong PCAPs compared with non-PCAPs. The lower ratio of H/R_n during PCAPs also indicates suppressed surface turbulence. The higher flux ratios during strong PCAPs compared with weak PCAP are suspected to be related to boundary layer clouds with a reduced R_n and increased H at the same time.

Variations of the surface energy balance terms with stability indicate that the largest median surface turbulent fluxes occurred under slightly unstable conditions, close to the neutral stability range ($-0.02 < \xi \leq 0.025$). The relatively flat variation of H and LE with increased instability under unstable conditions indicates that there are other processes that impact the surface turbulence, like the cloud top-down mixing, that can potentially question the M-O similarity theory employed in numerical models. The energy balance closure (EBR) was highest near the neutral stability range. Unstable conditions had higher EBR than stable conditions.

The coupling strength between the surface and atmosphere, which can be represented by the surface bulk transfer coefficient, depends on the land-use type. We found a slightly higher median C_h over bare land site compared with the short vegetation site during wintertime. This suggests that dynamic land-use datasets are desired in numerical models considering the seasonality of the surface exchange coefficients, especially over surfaces with vegetation. A follow-on study will determine uncertainties in surface flux estimates from numerical model simulations due to this parameter for the PCAPS time period.

Future work includes evaluation of numerical weather prediction model capabilities in capturing the surface turbulent flux variations during PCAPs, focusing on the M-O similarity theory applicability and the surface

exchange coefficient. The numerical model evaluation will be the second part of this research devoted to understanding the land-atmosphere exchange during PCAPs. Future work will also be done to investigate the numerical weather prediction model uncertainties on air quality modeling during PCAP events.

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REFERENCES

- Arya, P. S., 2001: *Introduction to Micrometeorology*. International Geophysics Series, Vol. 79, Academic Press, 420 pp.
- Barr, A. G., K. Morgenstern, T. A. Black, J. H. McCaughey, and Z. Nescic, 2006: Surface energy balance closure by the eddy-covariance method above three boreal forest stands and implications for the measurement of the CO_2 flux. *Agric. For. Meteorol.*, **140**, 322–337, <https://doi.org/10.1016/j.agrformet.2006.08.007>.
- Bornstein, R. D., 1968: Observations of the urban heat island effect in New York City. *J. Appl. Meteorol.*, **7**, 575–582, [https://doi.org/10.1175/1520-0450\(1968\)007<0575:OOTUHI>2.0.CO;2](https://doi.org/10.1175/1520-0450(1968)007<0575:OOTUHI>2.0.CO;2).
- Charuchittipan, D., W. Babel, M. Mauder, J.-P. Leps, and T. Foken, 2014: Extension of the averaging time in eddy-covariance measurements and its effect on the energy balance closure. *Bound.-Layer Meteorol.*, **152**, 303–327, <https://doi.org/10.1007/s10546-014-9922-6>.
- Chen, F., and Y. Zhang, 2009: On the coupling strength between the land surface and the atmosphere: From viewpoint of surface exchange coefficients. *Geophys. Res. Lett.*, **36**, L10404, <https://doi.org/10.1029/2009GL037980>.
- Crosman, E. T., and J. D. Horel, 2010: Sea and lake breezes: A review of numerical studies. *Bound.-Layer Meteorol.*, **137**, 1–29, <https://doi.org/10.1007/s10546-010-9517-9>.
- , and —, 2017: Large-eddy simulations of a Salt Lake Valley cold-air pool. *Atmos. Res.*, **193**, 10–25, <https://doi.org/10.1016/j.atmosres.2017.04.010>.
- Dirmeyer, P. A., and Coauthors, 2018: Verification of land-atmosphere coupling in forecast models, reanalyses, and land surface models using flux site observations. *J. Hydrometeorol.*, **19**, 375–392, <https://doi.org/10.1175/JHM-D-17-0152.1>.
- Foken, T., F. Wimmer, M. Mauder, C. Thomas, and C. Liebethal, 2006: Some aspects of the energy balance closure problem. *Atmos. Chem. Phys.*, **6**, 4395–4402, <https://doi.org/10.5194/acp-6-4395-2006>.
- Foster, C. S., 2015: Simulation of a cold-air pool in Utah's Salt Lake Valley. M.S. thesis, Department of Atmospheric Sciences, University of Utah, 78 pp.
- Franssen, H. J. H., R. Stöckli, I. Lehner, E. Rotenberg, and S. I. Seneviratne, 2010: Energy balance closure of eddy-covariance data: A multisite analysis for European FLUXNET stations. *Agric. For. Meteorol.*, **150**, 1553–1567, <https://doi.org/10.1016/j.agrformet.2010.08.005>.
- Hojstrup, J., 1993: A statistical data screening procedure. *Meas. Sci. Technol.*, **4**, 153, <https://doi.org/10.1088/0957-0233/4/2/003>.

- Holmes, H. A., J. K. Sriramasamudram, E. R. Pardyjak, and C. D. Whiteman, 2015: Turbulent fluxes and pollutant mixing during wintertime air pollution episodes in complex terrain. *Environ. Sci. Technol.*, **49**, 13 206–13 214, <https://doi.org/10.1021/acs.est.5b02616>.
- Holt, T. R., D. Niyogi, F. Chen, K. Manning, M. A. LeMone, and A. Qureshi, 2006: Effect of land–atmosphere interactions on the IHOP 24–25 May 2002 convection case. *Mon. Wea. Rev.*, **134**, 113–133, <https://doi.org/10.1175/MWR3057.1>.
- Horst, T., and D. Lenschow, 2009: Attenuation of scalar fluxes measured with spatially-displaced sensors. *Bound.-Layer Meteor.*, **130**, 275–300, <https://doi.org/10.1007/s10546-008-9348-0>.
- Kalverla, P. C., G.-J. Duine, G.-J. Steeneveld, and T. Hedde, 2016: Evaluation of the Weather Research and Forecasting Model in the Durance Valley complex terrain during the KASCADE field campaign. *J. Appl. Meteor. Climatol.*, **55**, 861–882, <https://doi.org/10.1175/JAMC-D-15-0258.1>.
- Karsisto, P., and Coauthors, 2016: Seasonal surface urban energy balance and wintertime stability simulated using three land-surface models in the high-latitude city Helsinki. *Quart. J. Roy. Meteor. Soc.*, **142**, 401–417, <https://doi.org/10.1002/qj.2659>.
- Kohsiek, W., C. Liebenthal, T. Foken, R. Vogt, S. Oncley, C. Bernhofer, and H. Debruin, 2007: The Energy Balance Experiment EBEX-2000. Part III: Behaviour and quality of the radiation measurements. *Bound.-Layer Meteor.*, **123**, 55–75, <https://doi.org/10.1007/s10546-006-9135-8>.
- Lareau, N. P., and J. D. Horel, 2015: Turbulent erosion of persistent cold-air pools: Numerical simulations. *J. Atmos. Sci.*, **72**, 1409–1427, <https://doi.org/10.1175/JAS-D-14-0173.1>.
- , E. Crosman, C. D. Whiteman, J. D. Horel, S. W. Hoch, W. O. J. Brown, and T. W. Horst, 2013: The Persistent Cold-Air Pool Study. *Bull. Amer. Meteor. Soc.*, **94**, 51–63, <https://doi.org/10.1175/BAMS-D-11-00255.1>.
- Lee, S. H., and Coauthors, 2011: Evaluation of urban surface parameterizations in the WRF model using measurements during the Texas Air Quality Study 2006 field campaign. *Atmos. Chem. Phys.*, **11**, 2127–2143, <https://doi.org/10.5194/acp-11-2127-2011>.
- Lehner, M., and Coauthors, 2016: The METCRAX II field experiment: A study of downslope windstorm-type flows in Arizona's Meteor Crater. *Bull. Amer. Meteor. Soc.*, **97**, 217–235, <https://doi.org/10.1175/BAMS-D-14-00238.1>.
- Loridan, T., and C. S. B. Grimmond, 2012: Characterization of energy flux partitioning in urban environments: Links with surface seasonal properties. *J. Appl. Meteor. Climatol.*, **51**, 219–241, <https://doi.org/10.1175/JAMC-D-11-038.1>.
- Lu, W., and S. Zhong, 2014: A numerical study of a persistent cold air pool episode in the Salt Lake Valley, Utah. *J. Geophys. Res.*, **119**, 1733–1752, <https://doi.org/10.1002/2013JD020410>.
- Mahrt, L., 2010: Computing turbulent fluxes near the surface: Needed improvements. *Agric. For. Meteorol.*, **150**, 501–509, <https://doi.org/10.1016/j.agrformet.2010.01.015>.
- Maronga, B., and S. Raasch, 2013: Large-eddy simulations of surface heterogeneity effects on the convective boundary layer during the LITFASS-2003 experiment. *Bound.-Layer Meteor.*, **146**, 17–44, <https://doi.org/10.1007/s10546-012-9748-z>.
- Massey, J. D., W. J. Steenburgh, S. W. Hoch, and D. D. Jensen, 2017: Simulated and observed surface energy fluxes and resulting playa breezes during the MATERHORN field campaigns. *J. Appl. Meteor. Climatol.*, **56**, 915–935, <https://doi.org/10.1175/JAMC-D-16-0161.1>.
- Medeiros, L. E., and D. R. Fitzjarrald, 2014: Stable boundary layer in complex terrain. Part I: Linking fluxes and intermittency to an average stability index. *J. Appl. Meteor. Climatol.*, **53**, 2196–2215, <https://doi.org/10.1175/JAMC-D-13-0345.1>.
- Monin, A., and A. Obukhov, 1954: Basic laws of turbulent mixing in the surface layer of the atmosphere. *Contrib. Geophys. Inst. Acad. Sci. USSR*, **151**, 163–187.
- Oncley, S. P., and Coauthors, 2007: The energy balance experiment EBEX-2000. Part I: overview and energy balance. *Bound.-Layer Meteor.*, **123**, 1–28, <https://doi.org/10.1007/s10546-007-9161-1>.
- Pagès, M., N. Pepin, and J. R. Miró, 2017: Measurement and modelling of temperature cold pools in the Cerdanya Valley (Pyrenees), Spain. *Meteor. Appl.*, **24**, 290–302, <https://doi.org/10.1002/met.1630>.
- Pielke, R. A., 2001: Influence of the spatial distribution of vegetation and soils on the prediction of cumulus convective rainfall. *Rev. Geophys.*, **39**, 151–177, <https://doi.org/10.1029/1999RG000072>.
- Rhodes, M. E., and J. K. Lundquist, 2013: The effect of wind-turbine wakes on summertime US Midwest atmospheric wind profiles as observed with ground-based Doppler lidar. *Bound.-Layer Meteor.*, **149**, 85–103, <https://doi.org/10.1007/s10546-013-9834-x>.
- Schotanus, P., F. T. M. Nieuwstadt, and H. A. R. De Bruin, 1983: Temperature measurement with a sonic anemometer and its application to heat and moisture fluxes. *Bound.-Layer Meteor.*, **26**, 81–93, <https://doi.org/10.1007/BF00164332>.
- Sheridan, P. F., S. B. Vosper, and A. R. Brown, 2014: Characteristics of cold pools observed in narrow valleys and dependence on external conditions. *Quart. J. Roy. Meteor. Soc.*, **140**, 715–728, <https://doi.org/10.1002/qj.2159>.
- Sicart, J. E., J. W. Pomeroy, R. L. H. Essery, and D. Bewley, 2006: Incoming longwave radiation to melting snow: Observations, sensitivity and estimation in Northern environments. *Hydrol. Processes*, **20**, 3697–3708, <https://doi.org/10.1002/hyp.6383>.
- Silcox, G. D., K. E. Kelly, E. T. Crosman, C. D. Whiteman, and B. L. Allen, 2012: Wintertime PM_{2.5} concentrations during persistent, multi-day cold-air pools in a mountain valley. *Atmos. Environ.*, **46**, 17–24, <https://doi.org/10.1016/j.atmosenv.2011.10.041>.
- Sun, J., S. P. Burns, A. C. Delany, S. P. Oncley, T. W. Horst, and D. H. Lenschow, 2003: Heat balance in the nocturnal boundary layer during CASES-99. *J. Appl. Meteor.*, **42**, 1649–1666, [https://doi.org/10.1175/1520-0450\(2003\)042<1649:HBITNB>2.0.CO;2](https://doi.org/10.1175/1520-0450(2003)042<1649:HBITNB>2.0.CO;2).
- Sun, X., H. A. Holmes, O. O. Osibanjo, Y. Sun, and C. E. Ivey, 2017: Evaluation of surface fluxes in the WRF Model: Case study for farmland in rolling terrain. *Atmosphere*, **8**, 197, <https://doi.org/10.3390/atmos8100197>.
- Tokinaga, H., and Coauthors, 2006: Atmospheric sounding over the winter Kuroshio Extension: Effect of surface stability on atmospheric boundary layer structure. *Geophys. Res. Lett.*, **33**, L04703, <https://doi.org/10.1029/2005GL025102>.
- UCAR/NCAR–Earth Observing Laboratory, 1990: NCAR Integrated Surface Flux System (ISFS). UCAR/NCAR–Earth Observing Laboratory, <https://doi.org/10.5065/D6ZC80XJ>.
- Van Dijk, A., W. Kohsiek, and H. A. R. de Bruin, 2003: Oxygen sensitivity of krypton and Lyman- α hygrometers. *J. Atmos. Oceanic Technol.*, **20**, 143–151, [https://doi.org/10.1175/1520-0426\(2003\)020<0143:OSOKAL>2.0.CO;2](https://doi.org/10.1175/1520-0426(2003)020<0143:OSOKAL>2.0.CO;2).
- VanReken, T. M., and Coauthors, 2017: Role of persistent low-level clouds in mitigating air quality impacts of wintertime cold pool conditions. *Atmos. Environ.*, **154**, 236–246, <https://doi.org/10.1016/j.atmosenv.2017.01.043>.

- Vosper, S. B., J. K. Hughes, A. P. Lock, P. F. Sheridan, A. N. Ross, B. Jemmett-Smith, and A. R. Brown, 2014: Cold-pool formation in a narrow valley. *Quart. J. Roy. Meteor. Soc.*, **140**, 699–714, <https://doi.org/10.1002/qj.2160>.
- Webb, E. K., G. I. Pearman, and R. Leuning, 1980: Correction of flux measurements for density effects due to heat and water vapour transfer. *Quart. J. Roy. Meteor. Soc.*, **106**, 85–100, <https://doi.org/10.1002/qj.49710644707>.
- Whiteman, C. D., S. Zhong, W. J. Shaw, J. M. Hubbe, X. Bian, and J. Mittelstadt, 2001: Cold pools in the Columbia Basin. *Wea. Forecasting*, **16**, 432–447, [https://doi.org/10.1175/1520-0434\(2001\)016<0432:CPITCB>2.0.CO;2](https://doi.org/10.1175/1520-0434(2001)016<0432:CPITCB>2.0.CO;2).
- , S. W. Hoch, J. D. Horel, and A. Charland, 2014: Relationship between particulate air pollution and meteorological variables in Utah's Salt Lake Valley. *Atmos. Environ.*, **94**, 742–753, <https://doi.org/10.1016/j.atmosenv.2014.06.012>.
- Wilczak, J. M., S. P. Oncley, and S. A. Stage, 2001: Sonic anemometer tilt correction algorithms. *Bound.-Layer Meteor.*, **99**, 127–150, <https://doi.org/10.1023/A:1018966204465>.
- Xiao, W., S. Liu, W. Wang, D. Yang, J. Xu, C. Cao, H. Li, and X. Lee, 2013: Transfer coefficients of momentum, heat and water vapour in the atmospheric surface layer of a large freshwater lake. *Bound.-Layer Meteor.*, **148**, 479–494, <https://doi.org/10.1007/s10546-013-9827-9>.
- Xiaoyan, J., N. Guo-Yue, and Y. Zong-Liang, 2009: Impacts of vegetation and groundwater dynamics on warm season precipitation over the central United States. *J. Geophys. Res.*, **114**, D06109, <https://doi.org/10.1029/2008JD010756>.
- Yuan, G., L. Zhang, J. Liang, X. Cao, H. Liu, and Z. Yang, 2017: Understanding the partitioning of the available energy over the semi-arid areas of the Loess Plateau, China. *Atmosphere*, **8**, 87, <https://doi.org/10.3390/atmos8050087>.
- Zängl, G., 2005: Formation of extreme cold-air pools in elevated sinkholes: An idealized numerical process study. *Mon. Wea. Rev.*, **133**, 925–941, <https://doi.org/10.1175/MWR2895.1>.
- Zhang, T., 2005: Influence of the seasonal snow cover on the ground thermal regime: An overview. *Rev. Geophys.*, **43**, RG4002, <https://doi.org/10.1029/2004RG000157>.
- Zheng, D., R. van der Velde, Z. Su, M. J. Booij, A. Y. Hoekstra, and J. Wen, 2014: Assessment of roughness length schemes implemented within the Noah land surface model for high-altitude regions. *J. Hydrometeor.*, **15**, 921–937, <https://doi.org/10.1175/JHM-D-13-0102.1>.
- Zolina, O., and S. K. Gulev, 2003: Synoptic variability of ocean-atmosphere turbulent fluxes associated with atmospheric cyclones. *J. Climate*, **16**, 2717–2734, [https://doi.org/10.1175/1520-0442\(2003\)016<2717:SVOOTF>2.0.CO;2](https://doi.org/10.1175/1520-0442(2003)016<2717:SVOOTF>2.0.CO;2).