Coevolution of Down-Valley Flow and the Nocturnal Boundary Layer in Complex Terrain

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ABSTRACT

An enhanced National Center for Atmospheric Research (NCAR) integrated sounding system (ISS) was deployed as part of the Vertical Transport and Mixing (VTMX) field experiment, which took place in October of 2000. The enhanced ISS was set up at the southern terminus of the Great Salt Lake Valley just north of a gap in the Traverse Range (TR), which separates the Great Salt Lake and Utah Lake basins. This location was chosen to sample the dynamic and thermodynamic properties of the flow as it passes over the TR separating the two basins. The enhanced ISS allowed for near-continuous sampling of the nocturnal boundary layer (NBL) and low-level winds associated with drainage flow through the gap in the TR. Diurnally varying winds were observed at the NCAR site on days characterized by weak synoptic forcing and limited cloud cover. A down-valley jet (DVJ) was observed on about 50% of the nights during VTMX, with the maximum winds usually occurring within 150 m of the surface. The DVJ was associated with abrupt warming at low levels as a result of downward mixing and vertical transport of warm air from the inversion layer above. Several processes were observed to contribute to vertical transport and mixing at the NCAR site. Pulses in the strength of the DVJ contributed to vertical transport by creating localized areas of low-level convergence. Gravity waves and Kelvin–Helmholtz waves, which facilitated vertical mixing near the surface and atop the DVJ, were observed with a sodar and an aerosol backscatter lidar that were deployed as part of the enhanced ISS. The nonlocal nature of the processes responsible for generating turbulence in strongly stratified surface layers in complex terrain confounds surface flux parameterizations typically used in mesoscale models that rely on Monin–Obukhov similarity theory. This finding has major implications for modeling NBL structure and drainage flows in regions of complex terrain.

1. Introduction

As populations continue to increase in the semiarid Intermountain West region, the need for improved forecasts of air stagnation and scouring events is magnified because of the susceptibility of these areas to prolonged pollution episodes affecting public health and periods of reduced visibility in more moist environments affecting the transportation industry (aviation, ground transportation, etc.). However, the prediction of cold-air-trapping events in valleys and basins has proven to be one of the most difficult forecast problems in mountain meteorology (Smith et al. 1997). A number of studies (Wakimoto and McElroy 1986; Neff 1989, 1997; Alexandrova et al. 2003) describe relationships between air quality and thermally driven circulations in basins throughout the Intermountain West region. Under conditions of weak synoptic forcing, cold dense air produced by strong radiative cooling at the surface flows down slopes and canyons and builds up in mountain valleys and basins during the night (e.g., Neff and King 1989). Depending on atmospheric conditions and details of the surrounding terrain features, this stable pool of cold air can persist for days in the winter (e.g., Wolyn and McKee 1989; Whiteman et al. 2001), resulting in the accumulation of unhealthful pollutants. This situation is exacerbated in parts of the Intermountain West region where the frequency of scouring precipitation events is limited.

A number of processes may affect the accumulation of cold air in concave terrain features. These processes may be divided into local-scale processes and synoptic-
scale processes. Local-scale processes that affect cold-pool development include those that modulate the surface energy budget (e.g., radiational cooling) or cause local areas of vertical transport and mixing (e.g., gravity waves, shear-induced turbulence associated with drainage flows, and vertical transport and mixing associated with hydraulic jump). Synoptic-scale processes that affect cold-pool development include frontal passage, mountain waves, and the passage of upper-level disturbances.

The Vertical Transport and Mixing (VTMX) field experiment was designed to study the local processes that affect the evolution of cold pools that occur in the Great Salt Lake Valley (GSLV) in autumn (Doran et al. 2002). During the field phase of VTMX, it was found that a down-valley jet (DVJ) developed over the western third of the GSLV on a number of nights (Banta et al. 2004). It will be shown that this DVJ played an important role in the evolution of the cold pool at the southern end of the GSLV. Observations obtained at the southern entry of the basin are used to characterize the vertical structure of the DVJ as it emerged from the Utah Lake basin. The impact that this DVJ had on turbulence and mixing near the surface and aloft as well as on the thermodynamic structure of the nocturnal boundary layer (NBL) is discussed.

2. Background

The contribution of thermally driven circulations to the development of a cold pool in a mountain basin has been well documented (e.g., Neff and King 1989; Allwine et al. 1992; Whitman et al. 1996; Fast et al. 1996; Clements et al. 2003). Circulations affecting mountain basins at night include slope flows, down-valley flows, and channeled slope winds or canyon flows (see Whitman 2000). The GSLV offers a rich environment for studying each of these circulations and their contributions to the evolution of a cold pool. Additional features that complicate the development of thermally driven circulations in the GSLV include the presence of the Great Salt Lake (GSL) to the north, its connection to other basins through a gap in the Traverse Range (TR) to the south, and thus impacts of the Utah Lake as well. Because land breezes (which occur at night when the lake is warmer than its surroundings) are typically shallow and weak (e.g., Atkinson 1981), circulations originating from Utah Lake and the Great Salt Lake are likely to have played only a minor role in determining the nighttime flow. The gap in the TR, however, allowed flow to pass readily between basins, and thus the connected basins appear to behave as a long valley, resulting in a large-scale diurnally varying valley flow (Banta et al. 2004).

Evidence for the existence of thermally driven circulations in wide valleys and basins has existed for many years. Hawkes (1947) and Stone and Hoard (1990a,b) observed diurnally varying winds occurring in the Salt Lake Valley and surrounding valleys, respectively. Stewart et al. (2002) confirmed that these diurnally varying winds were the result of thermally driven basin-scale circulations. They found similar thermally driven circulations in a number of basins throughout the Intermountain West region. Stewart et al. (2002) addressed neither the vertical structure of these circulations nor their influence on cold-pool development and vertical mixing.

The diurnal variation in flows observed in these wide valleys and basins may be likened to those observed in more detail in narrow valleys. Drainage winds that form in narrow valleys were the focal point of the Atmospheric Studies in Complex Terrain (ASCOT) Program, which had field phases in Brush Creek Valley of Colorado and the Big Sulfur Creek and Anderson Creek Valleys in California between 1979 and 1984 (e.g., Neff and King 1987; Clements et al. 1989). Neff and King (1987) related the depth and strength of down-valley flow in different valleys to the size of the drainage source regions and the phasing with other mesoscale circulations such as lake breezes.

Down-valley flow is often observed to pulse in strength (Fleagle 1950; Doran and Horst 1981; Neff and King 1987) and to be very turbulent. Pulses in the flow strength act to create local areas of convergence and divergence that can have a significant impact on the vertical transport of heat, momentum, and pollutants out of the boundary layer. This pulsing may be caused by the interaction of surface cooling, adiabatic heating, advection, and friction (e.g., Fleagle 1950) or through the formation of internal seiches, which can be described as tidal sloshing of a cold pool in a basin or valley (e.g., Neff and King 1987). Several investigators describe the ubiquitous presence of waves and turbulent mixing that develop through Kelvin–Helmholtz (KH) instability within and atop accelerating flows that occur in stable layers (e.g., Horst and Doran 1986; Neff and Coulter 1986; Neff 1988; Holden et al. 2000).

Recent mesoscale modeling studies have revealed the importance of improved parameterizations, grid resolution, and domain size for accurately simulating drainage flows and their attendant mixing in regions of complex terrain (Zhong and Fast 2003; Chen et al. 2004). Mesoscale models are able to capture the broad details of drainage flows, but they have difficulty in simulating the coevolution of the NBL (e.g., Zhong and Fast 2003). Zhong and Fast (2003) noted a tendency of mesoscale models to overpredict the depth of the cold
pool, resulting in a cold bias within the DVJ. High-resolution ($\Delta x = 250$ m and $\Delta z = 5$ m) simulations by Chen et al. (2004) reproduced the observed diurnal pattern in terrain-driven flow between the Utah Lake basin and GSLV, but their simulations also had a cold bias. They found that drainage flow over the TR creates a hydraulic jump in the flow that results in low-level warming in the lee of the TR to the east of Jordan Gap. Large-eddy-simulation studies of slope flows indicate that turbulence generated within the downslope flow through shear production is maximized in the surface layer, which may have great implications for surface flux parameterizations (Skyllingstad 2003).

As will be discussed below, a number of processes appear to influence the evolution of the DVJ and cold pool at the southern end of the GSLV. The instrumentation deployed at the southern terminus of the GSLV to characterize the evolution of the DVJ in the gap of the TR is described in section 3. The characteristics of the DVJ observed in the gap of the TR are described in section 4, followed by a description in section 5 of the impacts that the DVJ had on local thermodynamics. In section 6, the processes associated with the DVJ that contribute to vertical transport and mixing are described and implications for surface flux parameterizations are given. A brief summary of the results and conclusions is given in section 7.

3. Description of field experiment and instrumentation

The VTMX field experiment took place in the GSLV in October of 2000 (see Doran et al. 2002). One of the goals of this experiment was to observe how thermally driven circulations affect vertical transport and mixing of heat, momentum, and air pollutants in the NBL over complex terrain. To this end, 12 observation sites were strategically positioned throughout the GSLV (Fig. 1). In addition, the National Oceanic and Atmospheric Administration (NOAA) Environmental Technology Laboratory transverse-excited, atmospheric pressure, carbon dioxide (TEACO2) Doppler lidar was situated in the center of the valley to sample circulations throughout the GSLV (Banta et al. 2004).

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In this study, data collected at the site labeled “NCAR site” (for National Center for Atmospheric Research) in Fig. 1 are used to determine the dynamic and thermodynamic characteristics of the DVJ just downwind of the notch in the TR known as Jordan...
Gap. The Jordan Gap is a 6-km-wide notch in the TR through which flow between the Salt Lake Valley to the north and the Utah Lake Valley to the south may be channeled. Most of the instrumentation was located in a narrow valley near an active railroad track about 3 km south-southeast (SSE) of Bluffdale, Utah, at 40.463°N, 111.932°W. This location, termed the “Main” site, was bordered by a 30-m-high plateau to the east and a riverbed immediately to the west (Fig. 2). The site was situated roughly 2.4 km to the north of Jordan Gap. The height of the gap or notch in TR (not including Jordan Narrows) averages around 1450 m, or about 80 m above the Main site.

The Main site was instrumented with an array of in situ and remote sensors as part of an enhanced integrated sounding system (ISS) that was specifically designed for boundary layer studies. The standard NCAR ISS has been described by Parsons et al. (1994). It includes a wind profiler, a radiosonde sounding system, and a surface meteorological tower. A tethered balloon system, a minisodar, a backscatter lidar, and a second surface meteorological tower were enhancements chosen to provide a detailed look at the characteristics of the DVJ as it passed through the gap in the TR. Brown et al. (2004) describe the enhanced ISS in more detail and characterize its operating characteristics in the dry, stable environment encountered during VTMX.

The surface meteorological towers were instrumented to measure pressure, temperature, and humidity at 2 m and wind speed and direction at 10 m. The Main site was also instrumented with a sonic anemometer provided by the Pacific Northwest National Laboratory (PNNL; Doran 2002) to measure turbulent motions (variances and covariances) at 9 m. In addition, a radiation stand was set up at the Main site to measure downwelling shortwave and longwave radiation. A net radiometer was also mounted on the radiation stand to measure the total (shortwave + longwave) net (downwelling – upwelling) radiation. The second tower (labeled “Hill site” in Fig. 2) was erected on the plateau to the east.

During intensive observation periods (IOPs), soundings were typically released from the Main site every 2–3 h between 2200 of one day and 1500 UTC of the following day. The tethered balloon was deployed for up to 6 h at a time, with five tethersondes spaced about 100 m apart up to 600 m. It was operated mainly in “tower” mode (i.e., the length of the tether line is held constant). In tower mode, the tethersondes actually move up and down because of horizontal and vertical movement of the balloon caused by changes in balloon buoyancy and variations in the horizontal wind speed. As a result, the altitude of each tethersonde typically varied by ±30 m, resulting in a sequence of vertically spaced miniprofiles.

The minisodar deployed during VTMX was leased
from Meteorologische Messtechnik GmbH (METEK) through Aerovironment, Inc. This sodar (DSDPA.90-24) is a smaller version of the DSDPA.90-64 described in Engelbart et al. (1999). Raw vertical velocities obtained with the vertical beam were stored every 10 s. Horizontal winds were calculated using METEK software at 15-min intervals. The sodar was configured to have a vertical range resolution of 25 m and a maximum range of 375 m. Postprocessing revealed that the sodar tended to underestimate the horizontal wind speed. A relationship developed between the sodar wind speed at 50 m and Hill-site wind speed measured 40 m above the profiler (offset horizontally by ~100 m) was applied to all range gates. Independent comparisons between corrected sodar winds and sounding winds above 200 m confirmed a reduced bias in the sodar winds (not shown).

The scanning aerosol backscatter lidar (SABL) was deployed to monitor boundary layer structure. During VTMX, its viewing angle was fixed at zenith and operated at a more eye-safe mode of 1064 nm (though it can also be operated at 532 nm). During VTMX the laser pulse rate was set to 10 Hz and backscatter returns were averaged to 1 s, resulting in a vertical resolution of about 3 m. The system was operated continuously; however, its performance degraded with time during the experiment because of a gradual loss of laser power.

The data collected at the NCAR site are used to determine the characteristics of the DVJ as it passes through the gap in the TR. In the following sections, the characteristics of the flow through the gap and their temporal evolution are described. The impact of the DVJ on local thermodynamics and vertical transport and mixing at the southern end of the GSLV is then discussed.

4. Characteristics of flow through the gap

Doppler lidar data obtained during VTMX revealed that a southerly DVJ spanned the entire western half of the GSL basin on most nights characterized by strong surface radiative cooling and weak synoptic forcing (Banta et al. 2004). Radial winds from the terminal Doppler weather radar, surface MesoWest observations (Horel et al. 2002), and PNNL weather-station data (Allwine et al. 2002) indicate that this DVJ was present on a number of non-IOP days as well. Characteristics of the southerly flow through the gap in the TR at the southern end of the GSL basin are given in Table 1. The southerly flow typically commenced between 0400 and 0500 UTC, or about 3–4 h after sunset. This observation is similar to that found for the GSLV by Stewart et al. (2002). The cessation time of the flow through the gap was more variable and appeared to depend on a number of factors, including the passage of mesoscale fronts (e.g., 3 October), the synoptic weather pattern, and insolation.

The DVJ tended to end earlier on days for which the north–south pressure gradient was directed southward (i.e., positive values associated with high pressure to the north, e.g., 4, 7, and 8 October; see Table 2). The north–south pressure gradient is estimated using the difference in sea level pressure observed between Salt Lake City, Utah, and Pocatello, Idaho, located about 250 km to the north-northwest. On days for which the pressure gradient force is directed southward (i.e., positive), the DVJ dissipated just after sunrise, indicating that a combination of an opposing synoptic-scale pressure gradient and solar heating was required to interrupt the southerly flow. On days for which the pressure gradient was weak (<0.75 hPa) or directed to the north (i.e., negative values in Table 2), the DVJ tended to persist late into the afternoon before finally being interrupted by northerly flow associated with passage of the lake-breeze front (15–18 October; Tables 1 and 2).

Observations of the DVJ in the gap region of the TR indicated that it tended to pulse in strength. The amplitude and period of these pulses varied from event to event but were more-or-less constant for a given event. DVJs with large-amplitude pulses were observed on a number of days during the first half of the experiment (Table 1). The evolution of these pulses is characterized by the 7 October 2000 event shown in Fig. 3. Note the large-amplitude/long-period changes in wind speed over the entire depth of the DVJ. A second class of DVJ events was characterized by much smaller amplitude/high-frequency oscillations in wind speed, with pulses being limited in vertical extent as well (note little variation in sodar winds in Fig. 3). Monti et al. (2002) described similar high-frequency pulsing in the strength of downslope flow observed along the eastern slope of the Wasatch Mountain range and related it to alongslope internal-wave sloshing.

The evolution of the DVJ in the gap region of the TR can be described in terms of four stages: 1) decreasing up-valley flow, 2) transition to down-valley flow, 3) down-valley flow, and 4) cessation. The northerly up-valley flow diminished soon after sunset with winds at the southern end of the GSLV decreasing with time to below 2 m s\(^{-1}\) within 3–4 h of sunset. After a period of weak northerlies, an abrupt wind shift [from northerly (345°) to southerly (165°)—see Fig. 3] took place, indicating the influence of locally driven down-valley drainage flow. This abrupt wind shift marked the beginning of the transition period—during which time the 10-m winds were generally from the south and were less than 2.5 m s\(^{-1}\). The transition period was followed by onset
Table 1. Characteristics of the nocturnal jet observed at the NCAR site. Start time refers to time of initial wind shift to southerly. Boldface indicates that the period of variations in DVJ strength exceeded 2 h. If the end time is less than or equal to the start time then end time refers to the following day.

<table>
<thead>
<tr>
<th>Date (IOP)</th>
<th>Times: start-end (UTC)</th>
<th>Max flow depth (m)</th>
<th>Mean surface wind speed (m s(^{-1}))</th>
<th>No. of jet max in vertical direction</th>
<th>Max wind speed of lowest jet (m s(^{-1}))</th>
<th>Height of lowest max wind speed (m)</th>
<th>No. of pulses</th>
<th>Amplitude of pulse(s) (m s(^{-1}))</th>
<th>Transition period (h)</th>
</tr>
</thead>
<tbody>
<tr>
<td>3 Oct (1)</td>
<td>0430–0540</td>
<td>230</td>
<td>1.5</td>
<td>1</td>
<td>1.8</td>
<td>30</td>
<td>1</td>
<td>1.5</td>
<td>12</td>
</tr>
<tr>
<td>4 Oct</td>
<td>0400–1600</td>
<td>300+</td>
<td>2.0</td>
<td>—</td>
<td>9.0</td>
<td>150–300</td>
<td>3</td>
<td>1.5</td>
<td>15</td>
</tr>
<tr>
<td>6 Oct</td>
<td>0500–1930</td>
<td>300+</td>
<td>3.0</td>
<td>—</td>
<td>6.7</td>
<td>100</td>
<td>2</td>
<td>2–4</td>
<td>16</td>
</tr>
<tr>
<td>7 Oct (2)</td>
<td>0600–1500</td>
<td>400</td>
<td>1.2</td>
<td>2</td>
<td>6.7</td>
<td>100</td>
<td>2</td>
<td>5–6</td>
<td>16</td>
</tr>
<tr>
<td>8 Oct (3)</td>
<td>1015–1500</td>
<td>100</td>
<td>0.7</td>
<td>1</td>
<td>2.0</td>
<td>75</td>
<td>1</td>
<td>1.5</td>
<td>1.5</td>
</tr>
<tr>
<td>9 Oct (4)</td>
<td>0430–0000</td>
<td>1250</td>
<td>4.8</td>
<td>2</td>
<td>9.5</td>
<td>175–400</td>
<td>4</td>
<td>1–2</td>
<td>0.75</td>
</tr>
<tr>
<td>15 Oct (5)</td>
<td>0415–2215</td>
<td>940</td>
<td>3.3</td>
<td>1</td>
<td>7.0</td>
<td>175</td>
<td>2</td>
<td>4</td>
<td>0.75</td>
</tr>
<tr>
<td>16 Oct</td>
<td>0415–2230</td>
<td>400+</td>
<td>4.7</td>
<td>—</td>
<td>9.0</td>
<td>225</td>
<td>2</td>
<td>5</td>
<td>0.75</td>
</tr>
<tr>
<td>17 Oct (6)</td>
<td>0430–2200</td>
<td>580</td>
<td>4.0</td>
<td>2</td>
<td>8.0</td>
<td>160</td>
<td>6</td>
<td>1–2</td>
<td>1.75</td>
</tr>
<tr>
<td>18 Oct (7)</td>
<td>0400–0100</td>
<td>770</td>
<td>5.3</td>
<td>2</td>
<td>12.0</td>
<td>300</td>
<td>1</td>
<td>5</td>
<td>0.75</td>
</tr>
<tr>
<td>19 Oct</td>
<td>0400–1800</td>
<td>400+</td>
<td>5.6</td>
<td>—</td>
<td>12.0</td>
<td>300</td>
<td>6</td>
<td>1–3</td>
<td>0.75</td>
</tr>
<tr>
<td>20 Oct (8)</td>
<td>0415–1600+</td>
<td>810</td>
<td>1.8</td>
<td>1</td>
<td>8.0</td>
<td>130</td>
<td>5</td>
<td>2–4</td>
<td>1.75</td>
</tr>
<tr>
<td>24 Oct (9)</td>
<td>0430–0000</td>
<td>400+</td>
<td>4.5</td>
<td>—</td>
<td>13.0</td>
<td>300</td>
<td>3</td>
<td>4–6</td>
<td>2.0</td>
</tr>
<tr>
<td>26 Oct (10)</td>
<td>0000–0000</td>
<td>1250</td>
<td>3.3</td>
<td>2</td>
<td>12.5</td>
<td>400</td>
<td>5</td>
<td>2–5</td>
<td>0.75</td>
</tr>
<tr>
<td>Avg, or range,</td>
<td>0500–2015</td>
<td>100–1250</td>
<td>3.4</td>
<td>1–2</td>
<td>8.4</td>
<td>200</td>
<td>1–6</td>
<td>1–6</td>
<td>1.5</td>
</tr>
</tbody>
</table>

\(^a\) Start is the time of the initial shift in 10-m winds to southerly; end is the time at which 10-m winds shifted back to northerly.

\(^b\) Mean 10-m wind speed during down-valley flow event.

\(^c\) Transition period defined as period during which down-valley flow is less than 100 m deep.

\(^d\) Multiple jets present if wind maxima are separated by more than 300 m and wind speed is at least 33% greater than winds above and below.
and evolution of the DVJ. In this study, the DVJ is defined as a surface-based layer of southerly winds of greater than 2.5 m s⁻¹ at 10 m and greater than 100 m in depth. Table 1 indicates that the DVJ had usually developed within about 2 h of the initial abrupt wind shift. On a number of occasions, the DVJ deepened from the surface up, with a much weaker ambient flow present aloft. Examples of this type of flow evolution are given in Figs. 3 and 4 (e.g., 7 and 17 October). On these days, the DVJ can be directly related to the generation of a shallow thermally driven circulation because 1) flow aloft was weak and 2) the synoptic-scale pressure gradient was weak or opposed down-valley flow (Table 2). On other days the DVJ was apparently caused or strengthened by the downward mixing of southerly winds from aloft (see 9 and 24 October; Fig. 3). The DVJ often persisted well into the afternoon (Table 1), particularly when the synoptic-scale gradient favored a southerly component to the flow (see 18 October 2000; Table 2). On other days, when opposed by synoptic-scale gradient, the DVJ was disrupted much earlier (7 October 2000). In this case, the DVJ was also limited in horizontal extent—observations from the NOAA Doppler lidar showed no evidence of a jet on this day (Banta et al. 2004).

5. Impact on local thermodynamics

The characteristics of the flow during each stage of evolution in the DVJ have important consequences for the evolution of the cold pool and, thus, the accumulation and dispersion of pollutants near the surface and temporal variations in surface visibility and surface air temperature.

### 5.1 Surface-layer warming

Each occurrence of the DVJ was associated with a rapid warming in a shallow layer above the ground (Fig. 5). This warming always occurred during the initial onset of the DVJ and also when the DVJ restrengthened following a lull (Fig. 5a). The magnitude of the warming Δθ was much greater at the Main site than that observed at the Hill site. The differential warming of the surface layer as a function of terrain elevation results in potential temperatures θ that were nearly constant with terrain elevation as seen when the DVJ was well developed. LeMone et al. (2003) have shown that when the 2-m potential temperature does not change with elevation in the NBL, the air parcels must be moving along terrain surfaces (i.e., parcels move up and down with variations in terrain) with the ambient wind. Thus, the mesoscale DVJ effectively recouples the surface with small variations (<30 m) in terrain evolution with the overlying atmosphere (e.g., LeMone et al. 2003; Acevedo and Fitzjarrald 2003). Extending this finding to a larger range of terrain elevations is beyond the scope of this paper.

During VTMOV in the southern end of the GSLV, this recoupling took place when the 10-m wind speed exceeded about 2.5 m s⁻¹ (Table 3). This wind speed threshold, which can be used as a proxy for the strength of low-level nondirectional wind shear, is significantly greater than that (1.5 m s⁻¹) reported by Acevedo and Fitzjarrald (2003) for recoupling in areas characterized by more gently sloping terrain. Using data from the two surface meteorological towers at the NCAR site, we find an exponential relationship between low-level wind shear and the change in 2-m potential tempera-

### Table 2. Synoptic-scale conditions influencing down-valley flow. Ridge-top winds and direction are taken as averages between 2800 and 3000 m above mean sea level. Here NA means not available and VAR means directions varied by more than 90° in the layer.

<table>
<thead>
<tr>
<th>Date</th>
<th>Sounding times (UTC)</th>
<th>Ridge-top wind speed (m s⁻¹)</th>
<th>Ridge-top wind direction (°)</th>
<th>Pressure gradient* (hPa) 0000, 1200 UTC</th>
<th>Cloud cover</th>
</tr>
</thead>
<tbody>
<tr>
<td>3 Oct</td>
<td>0300, 0500, 0700</td>
<td>5.0, 4.5, 6.2</td>
<td>275, 275, 260</td>
<td>NA</td>
<td>Partly to mostly cloudy</td>
</tr>
<tr>
<td>4 Oct</td>
<td>0000, 1200**</td>
<td>10.0, NA</td>
<td>270, NA</td>
<td>2.7, 3.2</td>
<td>Clear</td>
</tr>
<tr>
<td>6 Oct</td>
<td>0000, 1200**</td>
<td>7.5, 5.5</td>
<td>280, 240</td>
<td>2.5, 3.4</td>
<td>Clear</td>
</tr>
<tr>
<td>7 Oct</td>
<td>0500, 0900, 1200, 1500</td>
<td>2.0, 2.0, 1.5, 1.0</td>
<td>030, 070, 215, 175</td>
<td>3.5, 5.0</td>
<td>Clear</td>
</tr>
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<td>290, 215, 245, 225</td>
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<td>Scattered</td>
</tr>
<tr>
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<td>235, 260, 245, 250</td>
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</tr>
<tr>
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<td>NA, 310</td>
<td>−0.2, 0.4</td>
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</tr>
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<td>VAR, 340, 010, 060</td>
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</tr>
<tr>
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<td>6.9, 6.7, 7.7, 6.4</td>
<td>145, 180, 190, 210</td>
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<td>Scattered to partly cloudy</td>
</tr>
<tr>
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<td>205, 190</td>
<td>0.9, 0.8</td>
<td>Clear</td>
</tr>
<tr>
<td>20 Oct</td>
<td>0300, 0700, 1200, 1500</td>
<td>7.3, 8.7, 1.3, 3.3</td>
<td>295, 335, 000, 200</td>
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<td>Clear</td>
</tr>
<tr>
<td>24 Oct</td>
<td>0000, 1200**</td>
<td>6.9, 6.6</td>
<td>115, 170</td>
<td>3.4, 1.3</td>
<td>Cloudy to clear</td>
</tr>
</tbody>
</table>

* Difference in sea level pressure between Pocatello, ID, and Salt Lake City, UT, (POC – SLC) at 0000 and 1200 UTC. Positive values indicate pressure gradients that oppose down-valley flow.

** Values obtained from SLC sounding.
ture with terrain height (Fig. 6). As low-level wind shear increases because of the strengthening DVJ, the change in 2-m potential temperature with terrain height decreases (i.e., potential temperature become horizontally uniform at a height 2 m above the surface), similar to that found in weaker drainage flows (e.g., LeMone et al. 2003). As the 10-m wind speed increases, it becomes increasingly likely that the 2-m potential temperature at the Hill site will exceed that observed at the Main site, indicating the occurrence of a superadiabatic warming.

Fig. 3. Data from METEK sodar and 10-m wind towers depicting the evolution of the nocturnal down-valley jet at the NCAR site on 4 days in October 2000. The top panel of each pair of figures shows a time–height plot of wind barbs (indicating wind speed and direction) from the METEK sodar located at the Main site. The bottom panel of each pair depicts 10-m wind speed [Main site (solid red line) and Hill site (dotted red line)], 10-m wind direction from the Hill site (blue dots), and downward solar radiation from the Main site (green curve). The horizontal line at 2.5 m s$^{-1}$ indicates the 10-m wind threshold that must be reached to be classified as a DVJ in this study.
(i.e., \( \theta \) at the Hill site minus \( \theta \) at the Main site is less than 0) that may have been produced by air being forced below its level of neutral buoyancy. These superadiabatic variations in 2-m potential temperature are fairly transient in nature (Fig. 5) but occur often (Fig. 6).

There is a great degree of scatter in the relationship between wind speed and the difference in 2-m potential temperature at two elevations (Fig. 6). This observation suggests that other processes are equally important in determining how the 2-m potential temperature varies with terrain height. For example, the amount of warming \( \Delta \theta \) seen in the 2-m potential temperature at the Main site is also proportional to the strength of the inversion layer below the layer of maximum DVJ winds (Fig. 7).

### b. Thermodynamic structure of the NBL

The evolution of the boundary layer’s thermodynamic structure followed a fairly consistent pattern for each DVJ event. Figure 8 depicts the typical evolution of potential temperature at the NCAR site, corresponding to the deepening DVJ observed on 17 October that is shown in Fig. 4. Soundings launched from the Main site at nominally 0000, 0500, 0700, 0900, and 1500 UTC demonstrate how the surface-based inversion layer develops quickly around sunset in response to a deficit in the surface energy budget (Fig. 9). Over the next 5 h, the surface-based inversion layer strengthens at a rate of nearly 10 K km\(^{-1}\) h\(^{-1}\) (Fig. 8). After onset of the DVJ (around 0600 UTC), the air adjacent to the surface rapidly warms, with the amount of warming decreasing with height. This warming coincides with strong cooling above the surface-based inversion layer, with cooling rates of up to 1 K h\(^{-1}\) between 80 and 300 m (Fig. 8). After this burst of low-level warming, the entire layer cools for the rest of the night (i.e., between 0900 and 1500 UTC).

Although the sounding temperature data are very accurate and provide good vertical resolution, the frequency of sounding launches was inadequate for resolving details of the thermodynamic processes that occur during the transition from up-valley flow to the DVJ.
Using a combination of sounding, tethered atmospheric observing system (TAOS), and surface tower data yields a more complete picture of how the warming associated with the DVJ unfolded. Successive soundings obtained about 2.5 h apart indicate that the warming associated with the DVJ on 9 October 2000 was confined below 50 m, with cooling aloft (Fig. 10). The higher-temporal-resolution observations available from TAOS indicate that the warming actually extended above 100 m for a short period of time (Fig. 10). Between 0500 and 0520 UTC, the potential temperature warmed by 1 K at 100 m. The Hill site also experienced warming beginning at 0500 UTC (Fig. 11). Warming observed at the Hill site lasted more than 1 h and was larger in magnitude (−2.5 K) than that observed at 100 m AGL with TAOS. The warming observed with TAOS and at the Hill site began 30 min before it was observed at the Main site. This time delay in warming indicates that the surface layer at the Main site was briefly sheltered from the warmer ambient flow just above.

Warming associated with the DVJ was shallow and short lived. Analysis of data from the surface meteorological towers for all of the jet-onset events (including redeveloping jets) showed that this warming period typically lasted less than 1 h (Table 3) and was followed by an extended period of cooling (e.g., Figs. 5 and 10). The transience of this warming is also evident in the TAOS data for 9 October, which depict a cooling of the profile between 0520 and 0600 UTC that exceeds the warming that took place minutes earlier (Fig. 10).

The warming seen at low levels is directly related to the onset of the DVJ. The net effect of the initial warming at low levels and cooling aloft is an overall weakening of the nocturnal inversion and a reduction in the low-level stability (Table 4). Such changes in the thermodynamic profile allow the propagation of waves at an increasing range of frequencies. The interaction and breaking of these waves generate turbulence, which enhances the vertical diffusion of pollutants out of the nocturnal boundary layer (see section 6 for evidence of this vertical diffusion in the SABL data).

c. Budget analyses

The temporal evolution of the potential temperature profile is a function of radiative cooling, turbulent heat flux divergence, and 3D advection. Heat budget pro-
files obtained for the DVJ that occurred on 17 October 2000 are shown in Fig. 12. The longwave radiative cooling rates are calculated using a two-stream radiative transfer model with 105 spectral bands in the longwave (Key 2001). Radiative cooling rates are obtained from the net flux divergence in each layer. The total tendencies are obtained by calculating the difference of potential temperature profiles from successive soundings. The residual term is the difference between the total tendency and the radiative cooling rate and represents contributions by turbulence and 3D advection.

The heat budget analysis shows the strong radiative cooling just above the surface followed by warming associated with the DVJ (Figs. 12a,b). Note that the largest radiative cooling (~50 K day$^{-1}$) takes place within

\begin{table}[h]
\centering
\begin{tabular}{|c|c|c|c|c|c|c|}
\hline
Date & Start–end time (UTC)$^a$ & Mean wind (shear) (m s$^{-1}$)$^b$ & Max $\theta$ deficit (K)$^c$ & Cold pool period (h) & Wind speed (shear) at onset of mixing (m s$^{-1}$)$^b$ & Length of mixing period $\tau_x$ (h)$^d$ \\
\hline
3 Oct & 0245–0535 & 1.5 (0.8) & 5.5 & 2.83 & 3.5 (1.0) & 1.00 \\
4 Oct & 0240–0650 & 1.8 (0.8) & 5.6 & 4.17 & 2.5 (2.5) & 0.75 \\
6 Oct & 0250–0740 & 1.7 (0.8) & 5.5 & 4.83 & 1.0 (2.9) & 1.20 \\
7 Oct & 0145–0730 & 1.7 (0.9) & 6.6 & 5.75 & 1.0 (2.8) & 2.50 \\
8 Oct & 0320–1415 & 1.0 (1.1) & 6.7 & 10.92 & — & — \\
9 Oct & 0200–0530 & 1.4 (0.7) & 6.3 & 3.50 & 1.4 (2.0) & 0.50 \\
15 Oct & 0220–0450 & 1.5 (0.6) & 2.8 & 2.50 & 1.0 (1.8) & 0.28 \\
16 Oct & 0230–0505 & 1.5 (0.3) & 4.4 & 2.58 & 2.6 (2.3) & 1.00 \\
17 Oct & 0240–0715 & 1.2 (0.8) & 5.3 & 4.58 & 2.3 (3.0) & 0.80 \\
18 Oct & 0000–0520 & 1.2 (1.2) & 5.7 & 5.33 & 2.2 (3.5) & 0.80 \\
19 Oct & 0115–0530 & 1.1 (1.1) & 7.5 & 4.25 & 2.5 (2.5) & 1.00 \\
20 Oct & 0200–0730 & 1.4 (0.9) & 6.8 & 5.50 & 3.1 (2.8) & 1.10 \\
24 Oct & 0215–0640 & 1.0 (0.8) & 5.4 & 4.42 & 2.5 (4.0) & 0.90 \\
26 Oct & 0100–0200 & 2.4 (1.8) & 2.4 & 1.00 & 1.7 (2.6) & 0.50 \\
Avg, $N = 14$ & 0205–0630 & 1.5 (0.8) & 5.5 & 4.4 & 2.1 (2.6) & 0.95 \\
\hline
\end{tabular}
\caption{Characteristics of cold pool at the NCAR site prior to jet formation.}
\end{table}

$^a$ Start and end times refer to uninterrupted periods during which the potential temperature difference exceeds 0.5 K.
$^b$ Shear between the Hill and Main sites is given in parentheses.
$^c$ Potential temperature deficit = $\theta$(Hill) − $\theta$(Main).
$^d$ Mixing period is defined to end when $\theta$(Hill) is within 0.25 K of $\theta$(Main).

Fig. 6. Difference (Hill site – Main site) in 2-m potential temperature as a function of 10-m wind speed at the Main site. Data are 5-min averages from all time periods prior to sunrise that experienced down-valley flow (i.e., 120° < wind direction < 180°).

Fig. 7. Warming in the 2-m potential temperature at the Main site $\Delta \theta$ as a function of the difference between the maximum potential temperature occurring in the profile below the maximum winds in the DVJ and the minimum 2-m potential temperature at the Main site. Profile data are instantaneous values from soundings obtained just prior to a warming event; the 2-m potential temperatures used to determine the amount of warming that occurred in the surface layer are 5-min averages. The dotted line shows the best fit to the data, with a correlation of 0.75.
a thin (<20 m thick) layer in the surface-based inversion. At first, much of the change in potential temperature can be explained by radiative cooling (Fig. 12a). The positive residuals seen above the surface and at 100 m may be indicative of the initial impact of the DVJ, which formed around 0600 UTC, midway through the first analysis period.

By 0907 UTC, warming evident in the residual extends from the surface up to 100 m (Fig. 12b). The warming is likely associated with turbulent mixing because the horizontal winds observed during that time are light (Fig. 3), making horizontal advection small.

During the final time period, the residual is negative throughout the lowest 500 m, with radiative cooling accounting for just 30% of the observed cooling. This excess cooling observed during the mature stage of the DVJ may be the result of advection (note: horizontal...
advection may be a factor now as winds become much stronger) and/or turbulent mixing; however, quantification of these components requires more detailed modeling studies, which is beyond the scope of this paper. It is clear that these nocturnal jet events played an important role in determining the local thermodynamics and vertical mixing within the Jordan Narrows region. The vertical mixing associated with the jet limits the minimum 2-m air temperature locally to values well above those that would have occurred through unhindered radiational cooling. The strong vertical mixing associated with the DVJ also reduces the low-level stability (Table 4). This decreased stability increases wave propagation and the likelihood of turbulence and mixing. Thus, increased turbulence caused by the nocturnal jet enhances ventilation of pollutants away from the surface. The mechanisms that drive the vertical transport and mixing within the DVJ are described in detail below.

6. Vertical transport and mixing associated with the nocturnal jet

As described above, the onset of the nocturnal jet and warming near the surface are nearly coincident in time. Budget studies suggest that this warming at low levels is caused by a combination of downward vertical advection and turbulent mixing of warm inversion-layer air into the shallow cold pool found at the southern terminus of the GSLV.

a. Vertical transport

Air flowing down the TR will warm adiabatically by about 0.8 K by the time it reaches the Main site (roughly 80 m below the mean height of the notch or gap in the TR). Analyses of sodar data reveal that the DVIJ has a mean downward motion of \(-0.1\) m s\(^{-1}\) in the lee of the TR (e.g., Fig. 13). Vertical advection of warm, inversion-layer air toward the surface results in

<table>
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<th>Date</th>
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<th>(\Delta \theta) (K)</th>
<th>(\Delta z) (m)</th>
<th>Time (UTC)</th>
<th>(\Delta \theta) (K)</th>
<th>(\Delta z) (m)</th>
</tr>
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<td>110</td>
<td>0900</td>
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<td>250</td>
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<td>60</td>
<td>0700</td>
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<td>80</td>
<td>0700</td>
<td>6.0</td>
<td>320</td>
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<td>110</td>
<td>0907</td>
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<td>400</td>
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<td>80</td>
<td>0700</td>
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<td>12.1</td>
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<td>10.5</td>
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<td>8.9</td>
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![Fig. 12. Potential temperature budget profiles obtained from sounding data during the 17 Oct 2000 DVJ event. Data have been interpolated to a grid that has 40-m spacing below 600 m (approximate depth of the mature DVJ layer). The total tendency (solid line) is obtained by taking the difference of successive soundings, radiative cooling rates (connected dots) are computed using a two-stream radiative transfer model, and the residual is the difference between the total tendency and the radiative cooling rate (i.e., total − radiative: dashed line). Calculations are for the (a) 0457–0657, (b) 0657–0907, and (c) 0907–1159 UTC time periods.](image-url)
significant warming in the middle of the inversion layer (~75 m), where the inversion may be as large as 100 K km$^{-1}$ (Table 4) and the mean downward vertical motion (~0.05 m s$^{-1}$) is still appreciable. Closer to the surface, the vertical velocity and, thus, the vertical advection approach zero.

Upward motion along the leading edge of pulses in the DVJ could also play an important role in the ventilation of the boundary layer. Evidence for this is shown Fig. 14. As the DVJ develops between 0600 and 0800 UTC, vertical motions become increasingly upward as might be expected in response to low-level convergence. The strongest upward motions observed between 0800 and 0900 UTC coincide with a deepening aerosol layer (Figs. 14a,c). As discussed below, it is believed that this deepening is produced through a combination of vertical transport, breaking waves, and turbulent mixing.

b. Turbulent mixing

Breaking waves and shear-driven turbulence associated with the pulsing DVJ are important factors in the erosion of the surface-based inversion layer or cold pool. The production of turbulence is a function of atmospheric stability and wind shear. The bulk Richardson number $\text{Ri}_B$ is used to assess the likelihood of KH waves and turbulent mixing in a layer using vertical gradients in mean properties across the layer. The $\text{Ri}_B$ is calculated as

$$ \text{Ri}_B = \frac{g \Delta \theta \Delta z}{\overline{\theta_v}} \left[ (\Delta u)^2 + (\Delta v)^2 \right]^{-1}, $$

where $\overline{\theta_v}$ is the mean virtual potential temperature of the layer $\Delta z$ and $u$ and $v$ are the vector wind components. Surface meteorological data collected at the Main and Hill sites are used to characterize the temporal variations in $\text{Ri}_B$ within the lower portion of the surface-based inversion layer as the DVJ evolves. Though not strictly valid, the temperature and wind speed and direction measured at the Hill site are assumed to be representative of the flow 40 m above the Main site to allow for rough approximation of the vertical gradients in wind and temperature over the Main site. The Hill site had good exposure to flow from the northwest and southeast (Fig. 2), the two predominant wind directions, and thus the atmospheric state observed at the Hill site was relatively uncontaminated by surface effects and local topography. Comparisons between sounding data and surface meteorological tower data at the hill site revealed that this approximation was reasonable (see Fig. 10 for a representative comparison of potential temperatures from the soundings and surface meteorological towers).

The temporal evolution of $\text{Ri}_B$ on nights with low winds and large net radiation deficits is depicted in Fig. 15a. Under the light-wind pre-DVJ conditions, the $\text{Ri}_B$ and its scatter increase with time as the surface cools radiatively. During this time, both the sensible heat flux and vertical velocity variance (Fig. 15b) are small, indicating a limited amount of waves and turbulence below 150 m. About 1.5 h after the wind shift, the $\text{Ri}_B$ begins to decrease as the low-level jet develops and low-level shear increases; however, the strong surface-based inversion is maintained as $\Delta \theta$ remains large during this time (Fig. 15a). The inversion layer does not start to erode until the $\text{Ri}_B$ is very close to 0.25 (see line marked as erosion in Fig. 15b). This erosion is coincident with increases in vertical velocity variance (evident in both sodar and sonic instrument data) and the downward surface turbulent heat flux near the surface (Fig. 15b). The surface-based inversion layer was observed to erode each night that a well-formed DVJ occurred, with the onset of jet-induced mixing at the Main site occurring at $\text{Ri}_B$ from 0.1 to 0.75.

c. Impact on surface fluxes

The temporal variations in surface sensible heat flux and momentum fluxes evident in Figs. 9 and 15 are characteristic of those observed during each DVJ event. The downward fluxes of heat and momentum evolved with the same periodic nature seen in the DVJ.
As the DVJ weakened, decreasing the shear instability that initially drove the turbulence, the turbulence would begin to decay. Often the jet would strengthen again, regenerating shear instability and turbulence near the surface, and the turbulent fluxes would increase once again. Overall, the DVJ elevated the downward sensible heat flux from a median pre-DVJ value of $-3.0 \, \text{W} \, \text{m}^{-2}$ [similar to that found by Doran et al. (2002)] to median post-DVJ values of $-11$ and $-51 \, \text{W} \, \text{m}^{-2}$ for pulsing and nonpulsing jets, respectively.

The performance of surface-layer parameterizations typically used in mesoscale models is likely to be poor in areas of complex terrain and drainage flows because of the nonlocal nature of the turbulence. That is, the amount of turbulence and turbulent heat flux observed locally is a function of what happened upwind. To demonstrate this problem, the bulk sensible heat flux is calculated from observations by following the parameterization of Delage (1997), described in Poulos and Burns (2003). This parameterization is a modified form of the Louis (1979) formulation and is used in several operational mesoscale models (e.g., Benoit et al. 1997).

Fig. 14. Time–height cross sections of (a) aerosol backscatter in the near-infrared from SABL and (b) horizontal wind velocity and (c) vertical velocity from sodar for 4 Oct 2000. The black square in (a) is shown in high resolution in Fig. 16, and the white square in (a) indicates the time period in which large-amplitude waves occurred. Horizontal wind speeds in (b) range from 1 (blue) to 8 (red) $\text{m} \, \text{s}^{-1}$. Vertical velocities contoured in (c) are 2-min averages, with 5-min averages at the 75-m range gate given by the solid line ($\text{m} \, \text{s}^{-1}$). The dotted line denotes vertical velocity of $0 \, \text{m} \, \text{s}^{-1}$ at 75 m; dashed lines denote $-0.1$ and $0.3 \, \text{m} \, \text{s}^{-1}$, as labeled.
The Delage (1997) formulation is a function of the roughness length, mean wind speed, surface–air temperature difference, and \( R_{ib} \). These quantities were observed with the meteorological towers and radiation stands, with \( R_{ib} \) calculated as discussed above. The observed momentum flux was used to determine that the surface roughness was approximately 0.1 m, following Stull (1988). The parameterized downward sensible heat flux plotted in Fig. 9 is small during the early stages of the cold-pool formation but becomes much too large after onset of the DVJ in comparison with observations. The large negative fluxes obtained with the parameterization resulted from a combination of the strong 10-m winds and reduced surface-layer static stability associated with the DVJ (Figs. 3 and 5b, respectively) and from the fact that the air at 2 m warms more rapidly than the ground, resulting in an increasing surface–air temperature difference (not shown).

d. Waves and turbulent mixing caused by DVJ

The strong shear associated with the strengthening DVJ and the large difference in densities between the surface-based inversion layer and the residual-layer air aloft resulted in conditions ideal for the formation and propagation of KH waves. Breaking KH waves are expected to occur when the \( R_{ib} \) falls below 0.25 (Miles 1961; Howard 1961). These conditions occurred often within the shear layers above and below the DVJ. Conditions for the production of breaking waves and turbulence are first met above the maximum winds in the DVJ where the static stability is weaker (see Fig. 8). It takes a while for these waves to generate turbulence within the surface-based inversion layer where stability is strongest, as seen in the delayed warming signal in the surface meteorological data (Figs. 10 and 15).

Analysis of the sodar and SABL data obtained at the Main site reveals the existence of waves on a number of nights. The waves were observed above and below the DVJ where maxima in the wind shear exist. On 4 October 2000, two episodes of KH wave activity were fortuitously observed in the aerosol backscatter obtained with SABL. The first occurrence of KH waves was between 0645 and 0725 UTC when the nocturnal jet was in its formative stage. These waves had a period of about 3 min and were evident in both the SABL and the sodar data (Fig. 16). It is noted that, as expected for gravity waves, the upward vertical velocity from the sodar and the crests of the waves seen in the SABL backscatter are roughly 90° out of phase (i.e., maximum upward motion occurs prior to wave crests). This wave packet had difficulty ventilating aerosols out of the surface-based inversion layer (see Fig. 14). Between 0800 and 0830 UTC, a packet of larger-amplitude KH waves, enclosed by a white box in Fig. 14, more effectively transports aerosol out of the stable boundary layer, significantly affecting their vertical distribution.

There is also evidence of waves occurring within the lower half of the DVJ in time–height plots of the sodar vertical velocity data (Fig. 17). These waves have a period of about 2.7 min (close to the intrinsic buoyancy frequency of the residual layer—calculated using the 0437 UTC profile shown in Fig. 10) and often appear to penetrate into the surface-based inversion layer. Because the Brunt–Väisälä (BV) period (\( BV = 2\pi/\sqrt{N^2} \)) of the surface-based inversion layer of 1.5 min is much less than the period of the observed wave, they can propagate vertically through this layer relatively unattenuated. This observed period of wave activity coincides
with decreasing $Ri_b$ but occurs prior to the period of enhanced turbulence. Thus, it appears that the waves observed aloft initiate the mixing that eventually makes it down to the surface, as indicated by dramatic increases in vertical velocity variance and turbulent heat flux (Fig. 15).

The nocturnal DVJ observed at the NCAR site appears to generate enough shear to create breaking KH waves that generate mixing both above and below the DVJ. Clear evidence of this elevated mixing is seen in the SABL data (Figs. 14 and 16) that illustrate how the layer of enhanced backscatter deepens with time in a region of strong shear in an intensifying DVJ. The DVJ produces surface-based and elevated layers of mixing (breaking waves and turbulence) that are instrumental in warming near the surface, eroding the cold pool and mixing pollutants out of the stable boundary layer. These analyses also indicate that wave instabilities and shear-induced mixing are strong enough to erode the surface-based inversion layer and to recouple the surface layer with the flow aloft. This recoupling causes the DVJ to “feel” the surface. The resulting frictional drag is a sink for momentum, which causes slower low-level wind speeds, a disruption of the jet, and an eventual decoupling from the surface once again. This suggests a slightly different explanation for the pulsing nature of the flow observed under conditions of weak synoptic forcing than those discussed previously in the literature (see section 1).

7. Discussion and conclusions

Analyses of data collected with the array of sensors deployed at the southern end of the GSLV provide a fairly clear picture of the structure and evolution of the DVJ as it passes over the TR and its thermodynamic...
characteristics. Data from soundings, sodar, SABL, tethered balloon, the PNNL sonic anemometer, and the surface meteorological towers revealed the complex co-evolution of the DVJ and its vertical structure and some of the local processes that contributed to vertical transport and mixing at this site.

As discussed in section 5, the evolution of the DVJ in the GSL valley can be divided into four stages based on the characteristics of the flow (Fig. 3). Each stage also has unique dynamic and thermodynamic characteristics. During the period of weakening up-valley flow, strong radiative cooling along the undulating valley floor results in the formation of shallow cold pools that preferentially fill the lowest points (denoted by smaller-scale terrain features in the 2D schematic in Fig. 18) in the valley (Fig. 18a). Strong stability associated with the cold pool results in decoupling of the surface layer from the flow aloft. The flow near the surface typically weakened to mean velocities of less than 2.5 m s$^{-1}$ at the Main site. During this period of light and variable winds (transition period), the surface-based temperature inversion often strengthened to over 100 K km$^{-1}$.

The onset of the DVJ is marked by rapidly increasing winds from the SSE that usually deepen up from the surface and tend to pulse in strength. The amplitude and period of the pulses in the DVJ were largest on days characterized by weak synoptic forcing. Upward vertical transport occurred at the leading edge of these pulses. The DVJ also promoted the production of shear-driven turbulence and waves near the surface and aloft. The downward mixing of heat and momentum toward the surface recouples the flow with the surface. Frictional drag would be expected to increase during this recoupling, which would tend to slow the DVJ, reducing turbulence, and to allow the cold pool to reform. This mechanism could explain the ebb and flow observed during several of the DVJs.

The nocturnal jet was observed to emerge fully formed from the Utah Lake basin on a number of days (Banta et al. 2004). Mesoscale modeling studies (e.g., Zhong and Fast 2003) and surface meteorological data from the MesoWest network indicate that the flow originated in the interconnected basins far to the south of the GSLV. Flow over the TR is downward into the GSLV, resulting in adiabatic warming and vertical advection. At the NCAR site, which was situated just downwind of the gap or notch in the TR, shear-driven turbulence and waves promoted rapid warming at the surface as the DVJ developed. At the same time, vertical transport along the leading edge of pulses in the DVJ strength along with elevated mixing layers atop the DVJ promoted ventilation of pollutants from the

Fig. 17. Time–height cross sections of sodar vertical velocity obtained on 9 Oct 2000. The filled circles and solid line indicate the location of the base of the low-level jet as determined from sodar and sounding horizontal wind data. The thin dashed line denotes the top of the surface-based inversion layer. Arrows indicate maxima in the sodar vertical velocity. Waves with a period (crest to crest) of about 2.7 min are evident just above the surface-based inversion layer between 0443 and 0515 UTC.
cold pool (Fig. 18b). The spatial extent of this mixing and the degree to which flow channeling through the notch accelerates the DVJ and enhances mixing are not clear. Mesoscale modeling studies could offer insight in these areas. However, the observations made at the NCAR site have shown that the evolution of the cold pool relies on a complex interaction of mesoscale dynamics, radiation, turbulence, and wave activity that is a function of the local terrain.

These aforementioned processes occur at a range of scales that is too broad for current operational mesoscale models to resolve. In addition, most surface flux parameterizations used in mesoscale models do not properly represent the turbulent heat fluxes in drainage-flow conditions because of the nonlocal nature of the turbulence generation. Zhong and Fast (2003) attributed cold biases in several mesoscale models to the poor simulation of the surface energy budget. Evidence presented in this study supports their conclusion. Feedbacks between errors in the surface energy budget and the simulated nocturnal boundary layer will have a deleterious impact on the strength and character of drainage flows.

Simulations of drainage flow run at very high, large-eddy-resolving, resolution have been shown to improve the simulation of turbulent drainage flows (e.g., Chen et al. 2004); however, these simulations still suffer from inaccurate representation of the surface fluxes. A modified approach to parameterizing surface fluxes that includes a term to characterize terrain effects as suggested by Poulos and Burns (2003) is needed to represent the nonlocal nature of turbulence in complex terrain. Better observations of the horizontal variations in the cold pool, the DVJ, and the mixing associated with it are needed to improve our understanding of cold-pool formation, maintenance, and dissipation and our ability to forecast these events to improve the quality of life in urban basins.

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**Fig. 18.** Schematic diagram representing (a) up-valley flow that is not coupled to the surface and (b) the pulsing down-valley nocturnal jet that is coupled to the surface. Location of radiosonde ascent is given by the line originating between two terrain features. The resulting potential temperature and wind profiles are given at the right of the figure. The magnitude of the flow is indicated by the length of the arrows. Turbulent mixing above and below the jet maximum is indicated by circular arrows. Vertical transport in areas of convergence is denoted by long upward-pointing curved arrows.
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