This numerical study investigates the nighttime flow dynamics in Owens Valley, California. Nested high-resolution large-eddy simulation (LES) is used to resolve stable boundary layer flows within the valley. On 17 April during the 2006 Terrain-Induced Rotor Experiment, the valley atmosphere experiences weak synoptic forcings and is largely dominated by buoyancy-driven downslope and down-valley flows. Tower instruments on the valley floor record a continuous decrease in temperature after sunset, except for a brief warming episode. This transient warming event is modeled with good magnitude and temporal precision with LES. Analysis of the LES flow field confirms the event to be the result of a slope to valley flow transition, as previously suggested by researchers based on field observations. On the same night, a northerly cold airflow from the Great Basin is channeled through a pass on the eastern valley sidewall. The current plunges into the stable valley atmosphere, overshooting the altitude of its neutral buoyancy, and generating a large-scale oscillatory motion. The resulting cross-valley flow creates strong vertical shear with the down-valley flow in the lower layers of the atmosphere. A portion of the cross-valley flow is captured by a scanning lidar. The nested LES is in good agreement with the lidar-recorded radial velocity. Furthermore, the LES is able to resolve Kelvin–Helmholtz waves, and ejection and sweep events at the two-layer interface, which lead to top-down vertical mixing.

1. Introduction

The classical description of the nighttime atmospheric boundary layer (Nieuwstadt 1984) is set in a horizontally homogeneous environment over flat terrain, under quasi-steady-state conditions. Turbulence is generated at the surface and transported upward. Within the stable boundary layer (SBL), the production of turbulence by mean shear dominates over the destruction by buoyancy, such that turbulence is temporally continuous. The continuously turbulent SBL usually occurs under conditions of strong winds and/or large cloud cover, which leads to reduced net radiative surface cooling (Van de Wiel et al. 2003). On clear nights with weak winds, the SBL cools rapidly. Turbulent motions are strongly damped by buoyancy stratification. The SBL goes into a quiescent state, where turbulence is suppressed over prolonged periods greater than the time scale of the dominant eddies (Nakamura and Mahrt 2005).

In the intermittently turbulent SBL, energetic mixing events known as turbulent bursts can occur over relatively short periods and are usually responsible for the majority of the upward–downward transport of heat–momentum (Coulter and Doran 2002). In addition, another frequent signature of bursting is warming below the elevation of such events (e.g., Whiteman et al. 2009, hereafter WHP09). This is because intense vertical turbulent turbulence tends to mix down warm (in terms of potential temperature) air from aloft. This should be differentiated from propagating gravity currents, when warming associated with wave troughs (downward-curving potential temperature isentropes) vanishes immediately after the wave passes. The latter usually occur on a much shorter time scale $\sim O(1$ min).

Intermittent turbulence has been observed during field experiments since the last decade, yet its origin is not well understood. A few known mechanisms include passing density currents (Sun et al. 2002), breaking shear-instability waves (Newsom and Banta 2003), solitary waves and downward-propagating gravity waves.
(Sun et al. 2004), turbulence and mean shear interactions (Nakamura and Mahrt 2005), and slope and valley flow transitions (WHP09). Note that except for the last reference, all mechanisms are derived from the Cooperative Atmospheric-Surface Exchange Study -1999 (CASES-99) over nearly flat terrain.

In reality, SBL flows are usually affected by the complex land surface. Large-scale topographic features such as mountains and valleys have pronounced effects in stratified flows (Baines 1997). Heterogeneous land cover introduces additional variability into the flow (Derbyshire 1995). In this paper, we study the SBL at a site with highly complex terrain, focusing on terrain-induced turbulent events in a slope-valley system. We investigate one of the intermittently turbulent nights described in WHP09 over Owens Valley, California, a steep valley studied during the Terrain-Induced Rotor Experiment in 2006. With a numerical modeling approach, we reproduce the observed nighttime transient warming episode at the valley floor. The model also uncovers a new terrain-induced mechanism responsible for upside-down turbulent mixing. The elevated source of turbulence is an intruding cold-air drainage current into the stratified valley atmosphere.

The numerical approach used in this study is large-eddy simulation (LES). To conduct LES of realistic SBL flow over complex terrain, a grid-nesting approach is adopted. Grid nesting takes information from the coarse grid and passes it to the fine grid as lateral boundary conditions. This procedure can be repeated over several nests until the desired grid resolution, which resolves the motion of interest, is reached (Zhong and Chow 2012). Nested high-resolution LES is an emerging numerical tool to study turbulent boundary layer flows over complex terrain. In this paper, we show the feasibility and usefulness of nested LES in studying complex flows.

2. Description of site and instrumentation

The field site is Owens Valley in California, where the 2006 Terrain-Induced Rotor Experiment (T-REX) was conducted. Owens Valley is about 150 km long, 15–30 km wide, and oriented approximately north to south (see Fig. 1a). The Sierra Nevada mountain range west of the valley is at an elevation of ~4 km above mean sea level (MSL). To the east of the valley are the Inyo and White Mountain chains at ~3 km MSL. The average elevation change between the Sierra Nevada crest and the valley floor is nearly 3 km. Both valley sidewalls are steep, especially the east side. The valley has a semiarid climate with low, sparse vegetation. Some evergreen trees are found at the upper elevations of both valley sidewalls.

A comprehensive description of the field instrumentation can be found in Grubisic et al. (2008). As presented in Fig. 1b, this paper uses observation data from the National Center for Atmospheric Research (NCAR) integrated surface flux facility (ISFF) towers and the German Aerospace Center (DLR) Doppler lidar. The 34-m ISFF central tower had high rate sonic anemometers and temperature sensors installed at multiple elevations. The DLR lidar performed range–height indicator
(RHI) scans in the cross-valley direction (80° azimuthal angle to the right and 260° to the left), and recorded radial velocities.

3. Model configuration and description

The Advanced Regional Prediction System (ARPS) was used for the simulations. ARPS is developed at the Center for Analysis and Prediction of Storms at the University of Oklahoma (Xue et al. 2000, 2001). It is a nonhydrostatic mesoscale and small-scale finite-difference numerical weather prediction model. ARPS uses a generalized terrain-following coordinate on an Arakawa C-grid. A mode-splitting time integration scheme is employed (Klemp and Wilhelmson 1978). This technique divides a big integration step (Δtbig) into a number of computationally inexpensive small steps (Δtsmall) and updates the acoustically active terms, while all other terms are computed once every Δtbig.

The length of the simulations is 8 h starting from late afternoon (1600 LST/0000 UTC) to midnight (2400 LST/0800 UTC) on 17 April 2006. The model domains are centered at the ISFF central tower (36.801 52”N, 118.160 02”W). Three one-way-nested domains (Fig. 1a) are used to zoom into the region of interest in both horizontal and vertical directions. The ratio of horizontal domain lengths between successive nests is 5 to 1. Simulations are first performed on a 2400-m horizontal domain that covers the entire valley and the mountain ranges. Realistic initial and lateral boundary conditions are obtained from the North American Mesoscale (NAM) reanalysis dataset (Rogers et al. 2009). Results from the 2400-m grid are used to drive lateral boundaries of the 240-m grid, which are fed into the 50-m grid. The lateral boundaries are updated at a constant time interval (ΔTb) set by the user and are linearly interpolated in between. Following the recommendation from the sensitivity study in Michioka and Chow (2008), frequent updates are used. For the 50-m grid, lateral boundaries are updated every 180 s. Vertical grid stretching is applied to better resolve the surface layer. On the finest grid, the vertical resolution is 20 m with 5-m near-surface spacing. The land surface is represented with high-resolution terrain (10 m) and land cover (30 m) from the U.S. Geological Survey. A buffer zone of 10 grid points is used to merge topography between adjacent nests. A Rayleigh-damping layer is set for the top 33% of the domain for the 2400- and 240-m grid and top 20% for the 50-m grid. Some key model parameters are given in Table 1.

On the 2400-m grid, ARPS is run in mesoscale mode with boundary layer parameterizations (Sun and Chang 1986). On the inner (240 and 50 m) grids, LES is performed to resolve turbulent boundary layer flows. The LES governing equations for the resolved fields are the momentum [Eq. (1)], continuity [Eq. (2)], and scalar transport [Eq. (3)] equations:

$$\frac{\partial p u_i}{\partial t} + \frac{\partial p n_i u_i}{\partial x_j} = -\frac{\partial p}{\partial x_j} - p g \delta_{i3} + p c_{inm} f n_m - \frac{\partial p r_{ij}}{\partial x_j}, \quad (1)$$

$$\frac{\partial p u_i}{\partial t} + \frac{\partial p n_i}{\partial x_i} = 0, \quad (2)$$

$$\frac{\partial \sigma}{\partial t} + \frac{\partial p n \sigma}{\partial x_i} = -\frac{\partial p \chi_i}{\partial x_i} + S, \quad (3)$$

where the spatial filter is denoted by an overbar. The velocity components are $u_i$, $\overline{p}$ is the pressure, $\overline{\sigma}$ is the density, $f$ is the Coriolis parameter, $\tau$ is a scalar (e.g., temperature, moisture), and $S$ is a generic source–sink term. The filter can be considered a density-weighted Favre filter $\overline{\phi} = \rho \phi / \rho$, where $\phi$ is generic variable (Favre 1983). The turbulent stresses $\tau_{ij}$ and scalar fluxes $\chi_i$ are represented by the 1.5-order turbulent kinetic energy–based closure (TKE-1.5) of Moeng (1984). More details on the turbulence model are given in Xue et al. (2000).

4. Results and discussion

a. Transient warming

Figure 2 presents observed and modeled times series of temperature, wind speed, and direction at the ISFF central tower at 30 m above ground level (AGL). The observed temperature drops rapidly past 1900 LST on 17 April 2006 at a rate of nearly 3°C h⁻¹. A transient warming event takes place around 2040 LST when the temperature starts to increase. Over 1°C of warming is
achieved during the following half-hour period. The cause of this nighttime warming, as explained in HWP09, is a transition from downslope to down-valley flows. Briefly, after sunset downslope winds are initiated through horizontal thermal gradients resulting from cooling of the slope surfaces (Fleagle 1950). The downslope flows then transition to down-valley flows (from valley to plain), which continue until morning (Defant 1949).

Winds at the central tower start from the down-valley direction after sunset as shown in Fig. 2b. This is due to the residual daytime flows that are channeled along the valley by the northwesterly synoptic winds. Because of the background down-valley winds, the flow exhibits a combined downslope and down-valley direction (∼300°) after the downslope winds are initiated. Shortly after 2000 LST, faster down-valley winds begin to dominate. The increased wind speed leads to stronger turbulent mixing that brings down potentially warmer air from aloft and hence the observed warming signal. As shown in Fig. 3, simulated surface streamlines confirm this transition from the downslope influenced to the down-valley-dominated flow around the central tower.

In Fig. 2a, good agreement of modeled and observed temperature is achieved from 1700 to 2000 LST. The temperature rise during the warming event is captured by the 50-m LES with good magnitude and temporal precision. In comparison, the temperature predicted on the 240-m grid is relatively constant. Similarly, both wind speed and direction before the warming episode are well modeled, as shown in Fig. 2b. The transition to down-valley flow and the associated increase in wind speed are reflected on the both the 50- and 240-m LES grid, however, to a reduced magnitude. Model performance is less satisfactory after 2110 LST. On the LES grid, temperature is underpredicted by ∼2°C, wind speed by ∼5 m s⁻¹, and direction by 20° with a maximum deviation of 45° from the observation at 2124 LST. Despite the deviations, surface streamlines still indicate a down-valley-flow-dominated region around the central tower (not shown). The general conclusion of downslope to down-valley flow transition still holds in the model.

Inspection of the horizontal velocity components reveals that the model is underpredicting the north-south velocity component \( v \) at 30 m AGL. To attempt to
correct for this and increase the surface wind speeds, the aerodynamic roughness length for the dominant land cover type in the valley, which is the shrub–scrub category in the USGS national land cover dataset, is reduced from 6.5 to 1 cm. The latter value was used in the T-REX study by Doyle and Durran (2007). Given the same synoptic-scale pressure gradients, a reduced roughness length leads to a shallower boundary layer, therefore increasing surface wind speeds. In Figs. 2c and 2d, modeled temperature and wind speeds improve significantly in the last two hours of the simulation on the LES grid, however, the warming event is no longer predicted.

In Fig. 2d, the modeled wind direction is in between the observed downslope and down-valley directions, and nearly constant from 1900 LST onward. At the same time, the rapid cooling between 1900 to 2030 LST is absent (Fig. 2c). Surface streamlines show that downslope flows do develop in the simulations, but the cold drainage flows from the western slope are not able to reach the central tower location (figures are not shown, but are qualitatively similar to Fig. 3b). It is likely that the failure to predict the warming event is due to the absence of the downslope to down-valley flow transition.

Finally, note that around 1930 LST, temperature observations show a downward spike, that is, a rapid decrease and increase within \( \sim 20 \) min (see Fig. 2a). The same oscillation pattern is also found in the wind direction and speed. This could be caused by the spatial oscillations of the boundary between the competing downslope and down-valley wind regimes. At 1930 LST when the temperature drops, the central tower could be sensing more of the downslope flow. Another possibility for such short-term oscillations is a passing gravity wave. Unfortunately, the model fails to capture this small-scale oscillation.

### b. Cold-air intrusion

The DRL lidar scan reveals an interesting three-layer flow structure in the cross-valley direction at 1716 LST 17 April. The top panel of Fig. 4 presents the RHI lidar scan in the \( 80^\circ -260^\circ \) azimuthal directions. The axes are adjusted such that the lidar is located at the origin. Since the lidar records radial velocities toward and away from it, changes in the sign of velocities horizontally across the origin indicate flow in one coherent direction. Within the lidar scan plane, a westerly flow component is present in the bottom \( \sim 1.5 \) km and above \( \sim 2.5 \) km from the valley floor. An easterly flow component about 1 km in depth is found in between. Simulations capture the three-layer flow structure on both the 240-m (Fig. 4a) and the 50-m (Fig. 4b) grids.

As shown in section 4a, westerly flow in the bottom layer is a result of the downslope flow from the western sidewall. The drainage flow extends east of the lidar, to nearly the bottom of the eastern sidewall (see Fig. 3a). The jet-like structure of the drainage flow can be seen on the bottom panel of Fig. 4. The westerly flow in the top layer is a component of the synoptic flow, which is in the

![Instantaneous streamlines of surface wind at (a) 1909 and (b) 2109 LST from the 240-m grid.](image-url)
northwest direction \( \sim 300^\circ \). This synoptic flow is responsible for the daytime down-valley flow because of the channeling effect of the valley (WHP09). The source of the easterly flow is unclear because of the limited horizontal scan range of the lidar. Radiosondes could have captured this flow, but none were launched that night. The 3D simulations, on the other hand, have the advantage of complete spatial and temporal coverage. Once validated against observations, we can use the model results to investigate the source of the flow.

In Fig. 4, the flow is roughly homogeneous in the horizontal direction. Therefore, to quantitatively assess model results against lidar observations, horizontal averaging is performed along the cross-valley direction to obtain a vertical profile of the mean wind speed \( V_H \) on the scanning plane. For the lidar data, \( V_H \) is obtained by taking the horizontal component of the radial velocity vectors. Significant errors in \( V_H \) are expected in the region directly above the lidar, as a result of the large angles between the horizontal and radial directions. Therefore data within 45° of the vertical axis are not used in averaging. For the model data, \( V_H \) is computed by projecting horizontal wind vectors onto the scanning plane.

Figure 5 presents both \( V_H \) and its standard deviation \( \sigma_{V_H} \) along the horizontal direction. The easterly flow spans from 1.5 to 2.4 km AGL, with a maximum speed of \( \sim 2 \text{ m s}^{-1} \). The westerly drainage flow zone is roughly 1.5 km deep and less than 1 m s\(^{-1}\). The 240-m grid results overpredict wind speeds in both flow zones. In comparison, better agreement in terms of wind speeds is achieved on the 50-m grid. However, both grids produce
a ∼400-m deeper (shallower) easterly (westerly) flow zone than what the observations show. Here, \( \sigma_{V_H} \) can be considered a measure of turbulence intensity given the approximate horizontal homogeneity of the flow. Both the lidar and the 50-m LES data have \( \sigma_{V_H} \) decreasing slightly with height. The 50-m LES shows spikes in \( \sigma_{V_H} \) at the intersections between the easterly flow and the synoptic-westerly flow at \( \sim 2.3 \) km and between the easterly flow and the downslope westerly flow at \( \sim 1.1 \) km AGL. The 240-m simulation shows similar local maxima at the intersections, however, with a much larger magnitude. This model overprediction is a result of the large-scale wave motion on the 240-m grid. As we show in the following subsections, turbulent processes including shear-instability waves, ejection, and sweep events occur in the flow transition zones. These processes have comparable length scales to the 240-m grid spacing, and hence are not well resolved. The subgrid representation of these processes is less efficient in breaking down the large-scale waves, leading to an overprediction of \( \sigma_{V_H} \) on the 240-m grid. Overall, Fig. 5 shows that elevated sources of turbulence, indicated by peaks in \( \sigma_{V_H} \), are present at the flow transition zones.

1) SLOPE REGION

The source of the easterly cross-valley winds is a synoptic-scale northeasterly flow from the Great Basin northeast of the Owens Valley (see Fig. 1a). This is shown on the horizontal plane at 2.5 km MSL (∼1.5 km AGL) on Fig. 6a from the 2400-m grid. The northeast flow is channeled through the topographic “gaps” between the White and the Inyo Mountains at \( Y \sim 40 \) km and within the Inyo Mountains at \( Y \sim 5 \) km. The 240 m grid on Fig. 6b shows the clockwise turning of the synoptic flow into the valley. The cross-valley flow is energetic and nearly reaches the western sidewall, disrupting the down-valley flow between −10 and 10 km in the \( Y \) direction.

The potential temperature contours in Figs. 6a and 6b show that the inflow from the Great Basin is colder than...
the valley air at the same elevation. Upon entrance into the valley, the cold air sinks along the sidewall as a downslope current, plunging to an altitude of neutral density as shown in Fig. 6c. Moreover, the excess momentum gained during the katabatic descent leads to the cold-air current overshooting its equilibrium altitude, leaving the vicinity of the downslope flow, and returning to it in a large-scale cross-valley oscillatory current. This process, associated with down-slope dense fluid intrusion into a stratified environment, is termed "springback" motion by Baines (2005) and "stratified flow response" by Mayr and Armi (2010). Many natural phenomena are attributed to this process such as powder snow avalanches, dense overflows in the ocean, and cold river flow into lakes (Baines 2001). A 3D description of the cold-air intrusion is presented in Fig. 7 in the form of an isosurface intrusion for potential temperature at 295 K. The density driven flow resembles a waterfall coming over the topographic gap over the eastern sidewall. Ripples form at the end of the downslope flow and are advected down valley.

Similar flows are also observed at Owens Valley on the evening of 29 March during enhanced observing period 2 (EOP-2) of the T-REX campaign (Daniels 2010).

It is important to note that cold-air-induced downslope flows observed at the Owens Valley are different from classic drainage flows. The latter are initiated by a horizontal thermal gradient between the radiatively cooled slope surface and the warmer adjacent air (Fleagle 1950). The classic Boussinesq slope flow formulation is usually described in a rotated slope coordinate \((n, s)\) (Haiden 2003)

\[
\frac{du}{dt} = -\frac{1}{\rho_0} \frac{\partial p}{\partial s} + \frac{\theta}{\theta_0} \sin \alpha, \tag{4}
\]

\[
\frac{dw}{dt} = -\frac{1}{\rho_0} \frac{\partial p}{\partial n} + \frac{\theta}{\theta_0} \cos \alpha, \tag{5}
\]

where \(u_s\) and \(w_n\) are the along-slope and slope-normal velocities, \(\alpha\) is the slope angle, and \(\rho_0\) and \(\theta_0\) are the reference density and potential temperature. Since the depth of the drainage flow is usually shallow compared to the downslope length scale, quasi-hydrostatic equilibrium \(1/\rho_0 \partial p/\partial n = g \theta/\theta_0 \cos \alpha\) is often assumed such that flow is parallel and attached to the slope surface, that is, \(w_n = 0\) (Mahrt 1982). The motion of an air parcel at the slope surface is briefly described as follows. At sunset, a parcel at rest \((u_s = 0)\) cools radiatively. Driven by its negative buoyancy, it accelerates down the slope. As it approaches the elevation of its neutral buoyancy, the excess momentum gained from the downslope acceleration allows the parcel to continue its downslope motion into a colder environment.

In comparison, the cold-air-induced downslope flow has a different set of initial conditions. At the slope top, the air parcel is already colder (by \(\sim 2\,K\), as shown in Fig. 8) than the valley atmosphere at the same elevation. Furthermore, it has a nonzero initial momentum \((u_s \sim 3\,m/s^{-1}, \text{see Fig. 9a})\) as a result of the synoptic forcing. As the parcel plunges down-slope, it is accelerated by greater buoyancy forcing as a result of its initial temperature deficit. When it reaches the elevation of its neutral buoyancy, the excess momentum gained from the downslope acceleration, together with its initial momentum, allows the parcel to continue its downslope motion into a colder environment. At that point,
the buoyancy forcing in the slope-normal direction becomes positive. With both $w_n$ and $u_s > 0$, the parcel is lifted off from the slope into the valley atmosphere.

To investigate the characteristics of the density current, we first look at a snapshot of the flow at 1830 LST in Fig. 8, when the downslope flow descends deep into the valley. The flow is deep at the top of the slope, and quickly plunges, turning into a thin, fast current. It reaches its neutrally buoyant elevation at around $X \approx 9$ km, and spreads out over a deep layer into the valley atmosphere. A linear fit calculates a slope angle of $19^\circ$ between $X$ from 9 to 13 km. The wind vectors are then rotated into the slope coordinates and are presented in Fig. 9. For ease of comparison, the $X$ axis uses the original $X$ coordinate rather than the slope coordinate $s$. The density current is parallel to the slope, as indicated by the near-zero values of $w_n$ in Fig. 9a. Large positive $w_n$ and corresponding decreases in $u_s$ are observed at distances $X < 8.88$ km, indicating the springback motion.

The depth of the flow $h$ is estimated at the height of $|u_s|_{\text{min}}$. After an initial decrease from the slope top following the plunge, $h$ increases steadily down the slope from 250 m at 11.8 km to 330 m at 9.12 km. The growth rate of the cold airflow is $|dh/ds| \approx 0.04$ or $2.1 \times 10^{-3} \alpha$ normalized by the slope angle. This is larger than, yet in order of magnitude agreement with, the values of $1.1 \times 10^{-3} \alpha$ obtained from the laboratory studies of Ellison and Turner (1959). The growth of the cold-airflow is primarily due to turbulent entrainment from above the current (Turner 1986). Baines (2005) further showed that entrainment processes are more likely to

![Image of Fig. 9](https://example.com/fig9.png)
occur over steep slopes > 20°. The effect of entrainment of surrounding fluids is a net inflow of mass indicated by the increased depth of the downslope flow in Fig. 9b. Furthermore, since the valley air at the same elevation is potentially warmer than the cold-air current, downward turbulent mixing leads to warming within the current. Note that the difference in $u$ reverses sign above 170 m and approaches 1 K. This is an artifact of the background stratification and elevation difference between the two slope locations.

To quantify the extent of entrainment, the dimensionless entrainment coefficient $E$ is computed from a mass balance:

$$\frac{dQ}{ds} = EU_s = \frac{E}{h} Q,$$

where $Q = \int_0^h u_s \, dn$ (m$^2$ s$^{-1}$) is the 2D volumetric flow rate, and $U_s$ is the depth-averaged wind speed. Based on the original laboratory work of Ellison and Turner (1959), and modifications by Turner (1986), the following empirical relationship evaluates $E$ based on the gradient Richardson number $Ri$:

$$E = \begin{cases} 0.08 & Ri \leq 0 \\ \frac{0.08 - 0.1Ri}{1 + 5Ri} & 0 < Ri \leq 0.8 \\ 0 & Ri > 0.8 \end{cases},$$

where $Ri = N^2/S^2$ is the ratio of the buoyancy frequency ($N$) over vertical shear ($S$) squared. Negative–positive $Ri$ indicates unstable–stable conditions. Since entrainment is fundamentally a turbulent mixing process between the environmental fluid and the current, its negative correlation with $Ri$ in the above formulation is expected. As stratification increases, turbulent mixing is suppressed, hence $E$ decreases. Above a certain threshold, turbulence is nearly damped out such that turbulent entrainment no longer occurs.

A linear fit using data east of 9.84 km estimates $|dQ/dx|$ to be around 236 m$^2$ s$^{-1}$ km$^{-1}$. Variations in $U_s$ (5.35 ± 0.33 m s$^{-1}$) are small, since $Q$ and $h$ both increase down slope. An along-slope-averaged $U_s$ is used to compute $E$. Since the background stratification is not uniform along the valley slope, we estimate $Ri$ in a bulk measure from the surface to the elevation of maximum $u_s$, $n_{max}$:

$$Ri_b = \frac{g}{\theta} \left[ \frac{\theta(n = n_{max}) - \theta_1}{\theta} \right] \left[ u_s(n = n_{max}) - u_{s,1} \right]^2 \cos \alpha,$$

where $Ri_b$ is a bulk Richardson number, subscript 1 stands for the elevation of first grid point above the wall, and $\cos \alpha$ is a converting factor from slope to $X$–$Z$ coordinates. Figure 10 is obtained by repeating the same procedure from 1730 to 1900 LST during the cold-air intrusion. A considerable spread (0.045–0.09) exists in model data within the unstable regime ($Ri < 0$) early on during the evening transition. Data seem to cluster...
around $E = 0.06$ rather than 0.08. The agreement in the stable regime is fairly good, although most model data are obtained under moderately stratified conditions $\text{Ri} < 0.5$.

2) VALLEY REGION

Lidar scans in Fig. 4 reveal the cross-valley flow induced by the cold-air intrusion on the eastern sidewall. This elevated cross-valley current has a significant impact on turbulent mixing in the valley boundary layer during the evening transition. Figure 11 shows an $X$–$Z$ slice of $\theta$ contours on the 50-m LES grid. The presence of a large-scale wave with a peak at $X \sim 0.58 \text{ km}$ and trough at $X \sim -1.7 \text{ km}$ is indicated by the elevation and depression of $\theta$ isentropes. The wave is at the interface of the easterly cross-valley flow above and westerly drainage flow below (see also Figs. 4b and 5). Its amplitude is roughly 200 m and it is propagating from east to west. The horizontal extent of the wave is visualized by contours of $\theta$ and $w$ in Fig. 12 at the layer interface. Its peak and trough are indicated by zero $w$, which is $90^\circ$ out of phase with $\theta$. A phase speed of $-4.0 \text{ m s}^{-1}$ is measured by tracking the maximum $w$ through consecutive frames.

Figure 11 shows overturning Kelvin–Helmholtz (K–H) billows at the trough ahead of the cross-valley gravity current. The K–H waves have a much smaller wavelength $\lambda_{\text{KH}} \sim 500 \text{ m}$ and are propagating to the northwest in Fig. 12a. Given the 50-m horizontal grid spacing ($\lambda_{\text{KH}} \approx 10\Delta x$), the LES representation of K–H waves should be acceptable. They do not show up in the lidar scan, so we cannot confirm their presence with observational evidence. However, the existence of these K–H waves is made plausible through the action of a shear-instability mechanism at the two-layer interface. Figure 13 presents vertical profiles of key meteorological variables at the domain origin at 1700 LST right before the breaking of the K–H waves. The two-layer structure is indicated by opposing $u$ velocities at $-2 \text{ km MSL}$, separating the cross-valley flow ($80^\circ$) above and down-valley flow ($350^\circ$) below. The north–south velocity component $v$ varies only slightly from 4 to 2 m s$^{-1}$ over a depth of 1.5 km.
The two-layer flows introduce an elevated peak in shear in addition to the surface maximum. The vorticity thickness $\delta_u$, representative of the depth of the mixing layer, is approximately 100 m based on the vertical extent of the shear layer. The wavelength $\lambda_{KH}$ is about 5 times larger than $\delta_u$. Linear instability analysis predicts the $\lambda_{KH}/\delta_u$ ratio to be between 3.5 (Moore and Saffman 1975) and 7.0 (Michalke 1964). Briefly, if the average $\lambda_{KH}$ between vortices is closer than 3.5$\delta_u$, a vortex will be torn apart by the strain field of the neighboring vortices, hence the lower bound. The upper limit represents the most spatially-amplified unstable mode. Laboratory experiments of Dimotakis and Brown (1976) and direction numerical simulation of Rogers and Moser (1994) measured $\lambda_{KH}/\delta_u$ in the range of 3.5 to 5.0. Our simulation results fall within this general range. To explain the direction of the K–H waves, the flow is projected onto a vertical plane of maximum shear aligned in the 125°–305° wind directions, represented by the solid white line in Fig. 12a. This converts the 3D K–H shear-instability into a 2D phenomenon with 180° opposing flows. The K–H billows form in the direction of the projection plane and are advected perpendicular to it. The vertical shear is rapidly reduced after the passage of the K–H billows.

More energetic cross-valley-flow-induced mixing events occur later, particularly on the eastward side of the valley. After 1800 LST, the cold-air intrusion penetrates deeper into the valley, along with the cross-valley springback motion. The shear-instability process continues, resulting in the formation of K–H billows, whose subsequent destruction leads to further top-down turbulent mixing. For example, three wave periods are present in the highlighted region in Fig. 14a, directly under a large-scale wave trough at 2.5 km MSL. The wave amplitudes increase from east to west. Besides K–H waves,
sporadic upward ejections in Fig. 14b and downward-sweep events in Fig. 14c also contribute to turbulent mixing of the valley SBL.

Through K–H billows, ejection and sweep events, the elevated shear layer acts as a source of turbulence during the evening transition in the valley. Figure 15 presents time series of vertical velocity $w$ and turbulent kinetic energy (TKE) within the valley boundary layer. While the former represents resolved turbulence, the latter is a measure of the SGS mixing. Significant amounts of elevated turbulence are simulated at 2 km from 1730 LST and extending up and down to a thicker layer between 1.5 and 3 km MSL after 1830 LST.

5. Summary and conclusions

High-resolution LES has been used to simulate the nighttime SBL in Owens Valley. This study validates the terrain-induced intermittency mechanism based on slope-valley flow transitions and uncovers a novel top-down turbulent mixing mechanism resulting from valley cold-air intrusions. Numerically, this study demonstrates the feasibility and usefulness of nested LES of SBL flows over complex terrain. Simulations are driven by realistic initial and boundary conditions and nested sequentially from the NAM reanalysis field to a microscale LES domain where turbulence is largely resolved. Model results are validated with tower measurements and lidar scans. Good agreement is achieved without modeling tuning.

The transient nighttime warming event documented in WHP09 is reproduced with LES to good magnitude and temporal precision, although the model results deviate from the observations after the event. The 3D simulation confirms the warming mechanism proposed in WHP09. Surface warming is a result of the transition from cold west-slope drainage flow to the warmer down-valley flow about 4 h after sunset. Decreasing the surface roughness in the model improves predictions past the warming event; however, the resulting down-valley flow dominates over the downslope flow too early in the evening such that the warming event is absent. The sensitivity study shows that rapid cooling associated with the drainage flow at the onset of the evening transition is an important precursor to the warming event.

On the eastern slope, a synoptic-scale cold airflow from the Great Basin to the northeast of Owens Valley is advected into the valley through topographic “gaps” along the ridge line. The cold-air intrusion resembles the laboratory study of dense slope flows into a stratified environment (Baines 2005). Our simulations of the atmospheric flow show good qualitative agreement with the laboratory-scale salt-water current. Through the topographic gaps, the dense cold-air current plunges down the steep ($\sim 19^\circ$) eastern slope, overshoots the altitude corresponding to its neutral buoyancy, takes off from the slope because of its excess momentum and eventually returns to its equilibrium elevation inside the

FIG. 15. Time–height contours of (top) vertical velocity and (bottom) TKE. Contour intervals are 1 m s$^{-1}$ and 0.1 s$^{-2}$, respectively. Data are sampled every minute.
valley. The cold-air current entrains potentially warmer valley air aloft, leading to a linear increase of volumetric flow rate down the slope. The empirical formulation of Turner (1986) from laboratory studies is used to relate turbulent entrainment to the Richardson number. A good empirical fit is found for \(0 < R_I < 0.5\). A relatively wide spread in the entrainment coefficient \(E\) is found for near-neutral conditions during early evening. LES data suggest a smaller value of 0.06 compared to the published value of 0.08. In the future, idealized LES of cold-air currents on uniform slopes should be performed to better characterize turbulent entrainment processes in the slope SBL.

Finally, the impact of the cold-air-induced cross-valley current on turbulent mixing in the valley atmosphere is investigated. Directional shear between the elevated easterly cross-valley flow and the westerly drainage flow is a source of shear instability and leads to the development of K–H billows along the two-layer interface. Besides breaking K–H waves, energetic ejection and sweep events are present around the shear layer, further enhancing top-down turbulent mixing. This elevated source of turbulence leads to a nonclassic top-down transfer of TKE and contributes to a deeper valley SBL.

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CORRIGENDUM

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The title of Zhou and Chow (2013) was published incorrectly when the paper originally appeared. The correct title, since fixed online, is “Nighttime Turbulent Events in a Steep Valley: A Nested Large-Eddy Simulation Study.”

The staff of the Journal of the Atmospheric Sciences regrets any inconvenience this error may have caused.

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