Using Surface Pressure Variations to Categorize Diurnal Valley Circulations:
Experiments in Owens Valley

YANPING LI AND RONALD B. SMITH
Yale University, New Haven, Connecticut
VANDA GRUBIŠIĆ
Desert Research Institute, Reno, Nevada

(Manuscript received 17 December 2007, in final form 30 October 2008)

ABSTRACT

Harmonic analysis has been applied to data from nearly 1000 Automatic Surface Observation System (ASOS) stations over the United States to extract diurnal pressure signals. The largest diurnal pressure amplitudes (~200 Pa) and the earliest phases (~0600 LST for surface pressure maximum) were found for stations located within deep mountain valleys in the western United States. The origin of these unique characteristics of valley pressure signals is examined with a detailed study of Owens Valley, California. Analysis of observational data from the Terrain-Induced Rotor Experiment (T-REX) project shows that the ratio of the valley surface pressure to temperature amplitude can be used to estimate the daily maximum mixed-layer depth $H$. On days with strong westerly winds above the valley, the mixed layer is found to be shallower than on quiescent days because of a flushing effect in the upper parts of the valley. Idealized two-dimensional Weather Research and Forecasting Model simulations were used to explain the pressure signal. In agreement with observations, the simulations show a 3-h difference between the occurrence of a surface pressure minimum (1800 LST) and a surface temperature maximum (1500 LST). The resolved energy budget analysis reveals that this time lag is caused by the persistence of subsidence warming in the upper part of the valley after the surface begins to cool. Sensitivity tests for different valley depths and seasons show that the relative height of the mixed-layer depth with respect to the valley depth, along with the valley width-to-depth ratio, determine whether the diurnal valley circulation is a “confined” system or an “open” system. The open system has a smaller pressure amplitude and an earlier pressure phase.

1. Introduction

Diurnal solar atmospheric tides are excited primarily by the absorption of solar radiation by water vapor in the troposphere and ozone in the stratosphere, as well as by the turbulent heat transfer near the ground (Forbes and Garrett 1979). Tides can be easily detected in surface pressure observations. In a hydrostatic atmosphere, surface pressure variations result from the integral effect in the vertical direction of density and temperature perturbations above the ground. For diurnal heating that is uniformly distributed along a latitude circle, a diurnal surface pressure signal is zonally homogenous and propagates westward, synchronously with the sun—the so-called migrating tide. Theory and observation of this tide indicate a surface pressure phase of about 105° (corresponding to a surface pressure maximum at 0700 LT) and an amplitude of approximately 20 Pa in the midlatitudes of the Northern Hemisphere (Chapman and Lindzen 1970). At most locations, however, the detected surface pressure signal is a combination of the diurnal migrating tide and local diurnal signals generated by mesoscale circulations such as sea-breeze, mountain–plain, or mountain–valley circulations (Mass et al. 1991). These mesoscale signals cause a longitudinal inhomogeneity in the diurnal tide.

For mountain–valley sites, the contribution of local valley circulations to the surface pressure signal typically dominates over the contribution by the diurnal atmospheric tide. Two factors account for the strong valley signal. First, the valley atmosphere is isolated...
from the free atmosphere by mountain massifs that are acting as walls (Vergeiner and Dreiseitl 1987). Sensible heat warms the valley atmosphere without pressure adjustment resulting from gravity wave generation (Banta 1984; Mohr 2004; Canavero and Einaudi 1987). Second, the heating of the upper slopes causes large temperature increases in the middle of the valley. Because the pressure variation at the valley floor is the result of vertical mean temperature variation in the valley, the diurnal pressure variation at the valley floor is significantly greater than that over the plain (Brehm and Freytag 1982), where heating of the planetary boundary layer (PBL) occurs only from below, and where the daily temperature range decreases rapidly upward.

Several field studies of diurnal valley circulations have been carried out in valleys around the world, such as the Brush Creek Valley in western Colorado in the United States (Whiteman 1982; Whiteman et al. 1989a,b), the Aizu Basin and Ina Valley in Japan (Kuwagata and Kimura 1995), and the Inn Valley in the European Alps (Vergeiner and Dreiseitl 1987). In the Inn Valley, a diurnal surface pressure variation exceeding 160 Pa in September was detected. The pressure maximum in this valley occurs around 0600 LT at the innermost station, about an hour earlier than at the station at the valley exit. The daily vertically averaged temperature range is about 2.2 times as large at Innsbruck, Austria, in the middle of the valley, than at Munich, Germany, located north of the valley exit.

In this paper, we investigate the relationship between diurnal surface pressure and temperature variations in Owens Valley, California, and seek to explain that relationship by examining the valley heating mechanisms. Section 2 introduces the method used to extract the diurnal component from surface observations. Section 3 gives the detailed analysis of Owens Valley observations from the Terrain-Induced Rotor Experiment (T-REX) project. In section 4 we propose a novel approach for estimating the valley daily maximum mixed-layer depth using surface pressure and temperature amplitudes. Idealized 2D Weather Research and Forecasting Model (WRF) simulations are described in section 5, together with a discussion of the physical processes involved in producing surface pressure and temperature variations in mountain valleys. Section 6 gives the results of the sensitivity experiments for different valley depths and seasons. Conclusions are presented in section 7.

2. Data analysis method

A convenient way to monitor and classify the diurnal circulations is by extracting the diurnal component from the observational data using harmonic analysis. The diurnal component is represented with an amplitude and phase:

\[ P_n(t) = C_n \cos \left( \frac{2\pi(t - \psi_n)}{24} \right) + e. \]

Here, \( P_n(t) \) is the surface observation at local solar time (LST) \( t (\text{h}) \) at station \( n \). The \( \psi_n \) is the calculated phase angle (°) of the diurnal component for \( P_n(t) \). It is related to the time \( t_{\text{max}} \) at which the diurnal part reaches its maximum: \( t_{\text{max}} = \psi_n(24/360°) \) (15° is equivalent to 1 h). Quantity \( C_n \) is the amplitude. As an illustration, in Fig. 1 we show hourly surface pressure observations from Bishop, California, for the month of August for the period 2001–04. The averaged hourly pressure data shows a strong diurnal cycle. Harmonic analysis provides a sinusoidal curve fit to that data. The calculated pressure amplitude shows that the diurnal component contributes about 86% of the total diurnal surface pressure variation, whereas the semidiurnal component contributes the remaining 13%. The phase is \( \psi = 84° \), which translates to 0536 LST, and is the time at which the diurnal component of surface pressure reaches its maximum.

3. Observations

a. Contiguous United States

Harmonic analysis has been applied to the hourly surface pressure observations for June–August of 1997–2006 from nearly 1000 Automatic Surface Observation System (ASOS) stations over the contiguous United States (CONUS; Fig. 2). The distribution of amplitude and phase for these stations shows some interesting patterns (Fig. 3a). Stations from the Florida peninsula have relatively small amplitudes and late phases (100° < \( \psi < 160° \)), which could be associated with the sea breeze. Stations from the Great Plains and Midwest show a “continentally enhanced tide” pattern, with phases similar to that of the global diurnal atmospheric tide but with much larger amplitudes (Tsuda and Kato 1989). Stations from dry western valleys have the largest amplitudes among the CONUS stations and have relatively early phases around 90° (0600 LST). Mass et al. (1991), using an earlier dataset, also detected the inhomogeneous distributions of the amplitude and phase of the diurnal surface pressure variations among the CONUS stations.

The seasonal cycles of the diurnal pressure amplitude/phase at three valley stations: Bishop (Owens Valley), Winnemucca, Nevada (Paradise Valley), and Omak,
Washington (Okanogan Valley) are presented in Fig. 3b. In general, the amplitudes are larger in the warm season and smaller in wintertime. The phases become later in wintertime for some stations. The global diurnal tide is difficult to discern from the valley-station observations because the local topographic effect is too strong.

b. Owens Valley

The unique diurnal pressure amplitude and phase in the western valleys drew our attention to the T-REX project held from 1 March to 30 April 2006 in Owens Valley in California (Grubišić et al. 2008). In association with the T-REX project, 16 Automated Weather Stations (AWS) were installed and maintained by the Desert Research Institute (DRI; Grubišić et al. 2005). The stations in the western part of the valley are distributed on a gentle slope, and stations on the eastern part are located on the flat valley bottom (Fig. 4). The fine spatial (about 3 km) and temporal resolution (30 s) of these AWS are very helpful for the study of the extreme diurnal surface pressure signals in Owens Valley.

The amplitude and phase of the diurnal component of surface pressure, temperature, and the along- and cross-valley wind components (for 10-m wind) for all stations are presented in Fig. 5. In general, the surface pressure amplitude increases and its phase decreases from west to east (Figs. 5a,b). The pressure phase is about 85° at the valley floor. This means that the diurnal component of surface pressure reaches its minimum there at around 1740 LST. The temperature amplitude and phase both increase from west to east (Figs. 5c,d). The temperature phase at the valley floor is about 224°. This means that the diurnal component of surface temperature reaches its maximum at about 1500 LST. The later temperature...
phase in the east is due in part to the effect of sunrise and sunset on valley illumination.

In this analysis of data from Owens Valley, we have stratified days according to the strength of westerly winds at the approximate height of the Sierra Nevada ridgeline (∼700 hPa), which was obtained from the Oakland and Independence, California, and Reno, Nevada, soundings. A day is classified as a “westerly” day if the westerly wind at 700 hPa exceeds 10 m s⁻¹ in the 0000 UTC [1600 Pacific standard time (PST)] sounding; otherwise it is classified as a “quiescent” day. For the 61-day period of the T-REX project, this leads to 29 westerly and 32 quiescent days.

The ridge-top wind has a detectable effect on the surface pressure and temperature (Fig. 5). The pressure amplitude is lower by about 20% (from 125 to 100 Pa) for the westerly days relative to that for the quiescent days (Fig. 5a), whereas the temperature amplitude is lower by about 0.5 K (Fig. 5c). The surface pressure phase is later (larger) and the temperature phase is earlier (smaller) for the westerly days when compared with the quiescent days.

The amplitudes and phases of the diurnal variation of the cross-valley wind component $U$ and the along-valley wind component $V$ are shown in Figs. 5e,f. In a symmetric valley, the phase of the $U$ wind is expected to change by 180° across the valley centerline. For the Owens Valley network, the phase difference between the easternmost (phase 220°) and the westernmost station (phase 350°) is less than 180°, due in part to the asymmetry of the valley slopes and in part to valley illumination effects.

The diurnal component of the along-valley wind $V$ gives an early afternoon 210° phase (i.e., 1400 LST peak). This phase for $V$ is similar to the surface temperature phase (224°) but much different than the surface pressure phase (Fig. 5). This raises the issue of whether temperature or pressure drives the axial valley–plain wind. According to Whiteman (1990), the axial temperature gradient between the valley axis and the valley exit triggers the up-valley wind.

c. Soundings

The rawinsonde data from the T-REX period used in our analysis were collected by the University of Leeds at Independence Airport (just north of Independence in
The 60-day-averaged potential temperature profile derived from the sounding data is shown in Fig. 6a. In the 0600 PST sounding, a deep nocturnal inversion is present with a shallow, strong inversion layer about 100 m thick near the ground. In the 1500 PST sounding, the temperature lapse rate is almost adiabatic below the ridge top. For quiescent days, the mixed-layer top in Owens Valley is about 3000 m, almost reaching the ridge-top level. For westerly days, the mixed-layer top is about 2500 m.

### 4. Mixed-layer depth

By applying the equation of hydrostatic equilibrium to the valley atmosphere, we can see that the surface pressure approximates the integral of the column air mass above the ground and is controlled by the vertical temperature profile. A schematic diagram of the diurnal variation of the vertical temperature profile is shown in Fig. 7. At night, an inversion forms in the valley with a stronger shallow inversion near the ground. In the morning and early-evening transitional periods, the vertical temperature profile approximates a standard lapse rate. In the late afternoon, the valley atmosphere is well mixed and the vertical temperature profile is close to an adiabatic lapse rate. The triangular area bounded by the deep inversion and the adiabatic temperature profile is related to the sensible heat stored in the valley. At each height, the width of the triangle...
represents the diurnal temperature range at that height. The lower boundary of the triangle represents the diurnal surface temperature change, whereas the height of the triangle corresponds to the afternoon mixed-layer depth. Thus, if the diurnal variations of surface pressure and surface temperature are known, the valley maximum mixed-layer depth could be easily estimated.

Our method for estimating the mixed-layer depth requires two assumptions. First, it is assumed that the diurnal temperature and pressure variations are negligible above the top of the mixed layer. Second, we assume that the shallow nocturnal inversion on the valley floor (about 100 m for Owens Valley) does not significantly affect the surface pressure. By eliminating the contribution of the latter to the diurnal surface temperature variation (2.5°C), we get an equivalent diurnal surface temperature variation, which is about 75% of the total diurnal surface temperature variation (10°C).

The pressure perturbation below a thin layer $dz$ is

$$dp'(z) = \rho'(z)gdz.$$  

(1)

According to the perfect gas law, $p(z) = \rho(z)RT(z)$, with

$$\frac{p'(z)}{p_b(z)} = \frac{\rho'(z)}{\rho_b(z)} + \frac{T'(z)}{T_b(z)}.$$  

(2)

Substituting $\rho'(z)$ into (1), we get

$$dp'(z) \approx -\rho_b(z)\frac{T'(z)}{T_b(z)}dz$$  

and

$$p'(z = 0) \approx -\int_0^z \rho_b(z)\frac{T'(z)}{T_b(z)}dz \approx -\frac{\rho_b(z)}{T_b(z)}\int_0^z T'(z)dz,$$

(3)

where $z = H$ is the top of the mixed layer. Here $\rho_b$ and $T_b$ are the column-averaged daily mean air density and air temperature, respectively, with

$$\rho'(z) \approx -\frac{T'(z)}{T_b(z)}.$$  

FIG. 6. Vertical potential temperature profile in the valley center for (a) 60-day averages of quiescent/westerly days at 0600/1500 PST in Owens Valley (derived from University of Leeds Independence Airport rawinsonde data). (b) WRF simulation at $t = 0300$, 0600, 0900, 1200, 1500, 1800, 2100, and 0000 LST.

FIG. 7. Schematic diagram of the diurnal evolution of the vertical temperature profile. The dashed line represents the strong nighttime inversion near the ground. Profiles at night ($t_N$), morning/evening transition ($t_S$), and afternoon ($t_D$) are labeled. Here $H$ is the depth of the mixed layer.
\[
\overline{p}_b = \frac{1}{H} \int_{z=0}^{z=H} \rho_b(z) \, dz;
\]

\(\overline{T}_b/\overline{p}_b\) could be replaced by the daily mean ground value of \(T_b(z = 0)/\rho_b(z = 0)\) when upper-air observations are not available. For Owens Valley,

\[
\frac{T_b(z = 0)}{\rho_b(z = 0)} \approx 0.9 \frac{\overline{T}_b}{\overline{p}_b}.
\]

Using

\[
T'(z) = \frac{H - z}{H} T'(z = 0)
\]

with (3) we obtain

\[
|\overline{P}| \approx \frac{\overline{p}_b g}{\overline{T}_b} |\overline{T}_{\text{EFF}}| H.
\]

An estimate of the daily maximum mixed-layer depth is then

\[
H = 2 \frac{\overline{T}_b}{\overline{p}_b g} \frac{|\overline{P}|}{|\overline{T}_{\text{EFF}}|}.
\]

Here, \(|\overline{P}|\) is the amplitude of the diurnal component of the surface pressure variations. We correct the surface temperature amplitude \(|\overline{T}|\) by introducing the equivalent surface temperature amplitude \(|\overline{T}_{\text{EFF}}|\) for which the nighttime shallow strong inversion near the ground has been eliminated; \(|\overline{T}_{\text{EFF}}| = 0.75|\overline{T}|\) for Owens Valley.

The mixed-layer depth \(H\) for each station in Owens Valley for the T-REX period was calculated using (5) (Fig. 8). For quiescent days, the daytime maximum mixed layer almost reaches the ridge-top level. For westerly days, the mixed-layer height is about 500 m lower. The results derived here from the ground DRI AWS observations (Fig. 8) are consistent with the mixed-layer heights from the Independence sounding data (Fig. 6a). It seems that the synoptic wind near the ridge-top level has a ventilation effect that carries away the heat and causes the daily mixed-layer depth to decrease.

If the amplitude of diurnal surface pressure and temperature variations can be used to estimate the daily maximum mixed-layer depth in the valley, then how about the phases? According to the schematic diagram (Fig. 7), the surface pressure perturbation is the accumulation of the temperature perturbation along the whole vertical air column. The time for the surface pressure to reach its minimum should be almost the time for the column-averaged air temperature to reach its maximum. Thus, the phase difference between surface temperature and surface pressure is the time lag between the ground temperature and the vertically integrated temperature. For example, as shown in Figs. 5b,d, the surface pressure phase is around 90° (maximum at 0600 LST and minimum at 1800 LST) and the temperature phase is around 224° (maximum at 1500 LST). This 3-h time lag between the pressure minimum and temperature maximum is an indication of the valley heat transport mechanism.

5. Numerical simulations

a. The setup of idealized 2D simulations

To explain the unique phases of the surface pressure observations from the valley stations, the WRF, version 2.1.2, is employed to simulate an idealized diurnal valley circulation and the characteristics of the valley thermal forcing. The model setup is similar to Rampanelli et al. (2004) and Robinson and Sherwood (2006). The 2D computational domain is 240 km in the \(x\) direction (across the valley) and 17 km in the \(z\) direction. The horizontal and vertical spatial resolutions are 1 km and 80 m. A periodic boundary condition is chosen at the lateral boundary of the domain. At the upper boundary of the domain, a rigid lid \((w = 0)\) is employed with a 5-km Rayleigh damping layer near the top. The time step is 10 s for the advection terms and 1 s for the acoustic modes. A third-order-accurate Runge–Kutta scheme is used for the time integration. Fifth- and third-order-accurate spatial discretization schemes are used for the horizontal and vertical advection schemes, respectively. The National Centers for Environmental Prediction Global Forecast System PBL scheme is chosen to represent the effects of convective heat transfer in the vertical direction, together with a five-layer thermal...
The heating is applied over the surface for the whole domain. The lateral boundary is at \(-120\) and \(120\) km (not shown).

diffusion scheme for the land surface scheme. Cumulus and microphysical parameterizations are not activated. The Coriolis term is set to zero. The model is integrated for 2 days, and the first 24 h are used as a spinup for the whole system. The diurnal circulation of the second day is treated as the representative of the diurnal valley circulation. The initial sounding is derived from an averaged Reno 0400 PST sounding for April but with no water vapor.

The idealized valley in this study is the area between two bell-shaped mountains (Fig. 9). The analytical expression is

\[
h = 1000 + h_m \left[ 1 + \left( \frac{x + x_c}{a_s} \right)^2 \right] + h_m \left[ 1 + \left( \frac{x - x_c}{a_s} \right)^2 \right].
\]

where \(h_m = 2000\) m is the valley depth. The height of the surrounding plain in the model domain is 1 km MSL, with \(x_c = 15\) km and \(a_s = 3.5\) km to make our idealized valley similar to the geography of Owens Valley (30-km ridge-to-ridge separation; about 15-km valley floor width). Because the lateral boundary condition is periodic, the two plains outside the hills actually create one very broad valley. Relative to the narrow valley between the hills, the width ratio is 7:1 (210:30 km).

The imposed heating function is also highly idealized. The daytime incoming shortwave radiation and nighttime outgoing longwave radiation are represented by a diurnal heating function applied at the ground \(z = h(x)\). The heating is uniform over the whole domain. The surface sensible heat flux has a sinusoidal shape during the day and is nearly constant at night (Fig. 10a), similar to the 5-min flux data from the National Center for Atmospheric Research (NCAR) Integrated Surface Flux Facility (ISFF) averaged over the 60-day T-REX period (Fig. 10b):

\[
\hat{Q} = \begin{cases} 
-160 \cos \frac{2\pi(t - 0.67)}{24} & 0600\text{ LST} \leq t \leq 1800\text{ LST}, \\
-80 & t < 0600\text{ LST}, t > 1800\text{ LST}
\end{cases}
\]

The value \(Q_{\text{max}} = 160\) W \(\text{m}^{-2}\) we chose for our idealized simulation is smaller than that obtained from observations in Owens Valley but is representative for a number of other valleys. For example, a noontime sensible heat flux of 140 W \(\text{m}^{-2}\) was measured at the valley floor in Innsbruck (47.2°N) in September (Vergeiner and Dreiseitl 1987). Our specified strong nighttime cooling (\(-80\) W \(\text{m}^{-2}\)) prevents the atmosphere in the lower part of the valley from getting warmer day by day. Furthermore, we neglect the transfer of heat to the valley atmosphere by longwave radiation from the surface.

In the model, the sensible heat put into the atmosphere reaches its peak earlier than the peak time of the input heating because the downward ground heat flux into the soil carries away more heat in the afternoon (Fig. 10a). To make sure that the diurnal sensible heat curve is comparable to the Owens Valley observations in which the sensible heat reaches its peak at 1220 LST, we shift the maximum of the input heating to 1240 LST instead of noon (7).

b. Diurnal valley circulation under quiescent conditions

1) Diurnal circulation

The diurnal valley circulation under quiescent conditions was simulated by assuming that the background atmosphere is motionless \((U = V = W = 0)\). Results are shown in Figs. 6b and 11. In qualitative terms, the results support the existing picture of diurnal valley circulations (Whiteman 1982, 1990; Vergeiner and Dreiseitl 1987). Between 0300 and 0600 LST, the valley atmosphere shows a strong nocturnal inversion (Fig. 6b), with a thin layer of downslope wind on both sides of the valley. In the center of the valley, some weak upward compensating motions exist. After sunrise, the upslope winds begin. Near the ridge top, they converge and descend at the center of the valley. The stable core in the valley descends and the nighttime inversion disappears (Fig. 6b). At 1700 LST, the upslope winds and descending motion strengthen and the valley atmosphere becomes mixed (Figs. 6b and 11). After sunset, the slope winds switch to the downward direction and the valley surface starts to cool. The valley atmosphere returns to the nighttime situation.

The model-simulated diurnal surface pressure and temperature variations at the valley bottom \((x = 0)\) are shown in Fig. 12 together with their extracted harmonic components. The calculated phase for surface pressure...
is 83° (i.e., minimum at 1732 LST). The phase for surface temperature is 227° (i.e., maximum at 1508 LST). The phases calculated from the observations and the WRF simulations are consistent. Both have a time lag of nearly 3 h between the surface temperature maximum and the pressure minimum.

2) ENERGY BUDGET

To identify the dominant heating process in different parts of the valley, a heat budget analysis was carried out. According to the first law of thermodynamics, the energy budget for incompressible dry air is
The terms are identified as follows: I is the local tendency, II is the imposed heating, III is the horizontal advection, IV is the horizontal resolved eddies, V is the vertical advection, VI is the vertical resolved eddies, and VII is the subgridscale turbulent diffusion. In (8), \( u(x, z, t) \) is the potential temperature evaluated at each grid point. It can be expressed by a temporal mean \( \bar{\theta} \) and a perturbation part \( \theta' \), with \( \bar{\theta} = \bar{\theta} + \theta' \). The overbar denotes a time average over 1 h. The calculated energy budget terms at every grid point are then averaged over a 4-km width centered at \( x = 0 \). The lateral averaging improves the statistics and captures the off-axis contribution from horizontal advection. Here, \( K_r \) is the eddy diffusivity for heat. Turbulent diffusion has terms for local and nonlocal mixing (Troen and Mahrt 1986; Hong and Pan 1996). Imposed heating \( \dot{Q}/(\rho C_p) \) is zero everywhere, except at the ground where it represents the prescribed sensible heating.

The time evolution for each term of the energy budget equation at the centerline of the valley (\( x = 0 \); Fig. 9) is shown in Fig. 13. In this diagram, we extend the previous studies of valley heat budgets (e.g., De Wekker et al. 1998; Rampanelli et al. 2004; Ye et al. 1987; Zhong and Doran 1995; Zhong et al. 2001) by displaying the full time–height variation in heating mechanisms.

The daily evolution of the valley-center energy budget for the quiescent case (Fig. 13a) shows a dominance of vertical advection over other mechanisms of heat transfer, with horizontal and vertical eddies, and vertical turbulent diffusion playing supporting roles. After sunrise (0600 LST), the vertical turbulent diffusion transports heat up to about 300 m, above which subsidence warming dominates up to 2800 m. At 1500 LST, the diurnal component of surface temperature reaches its maximum, but subsidence warming continues strongly aloft. At 1600 LST, horizontal advection (cooling aloft and warming below) reduces the static stability in the valley. By 1600 LST, the valley atmosphere is nearly well mixed at the center (Fig. 6b), and therefore the heating contributed by subsidence is reduced. At 1800 LST, the diurnal component of surface pressure reaches its minimum. The temperature tendency is negative near the ground but is still positive in most of the valley, with

\[
\frac{\partial \bar{\theta}}{\partial t} = \frac{\dot{Q}}{\rho C_p} - \bar{u} \frac{\partial \bar{\theta}}{\partial x} - \bar{w} \frac{\partial \bar{\theta}}{\partial z} - \frac{\partial}{\partial x} \left( \bar{u} \theta' \right) - \frac{\partial}{\partial z} \left( \bar{w} \theta' \right) + K_r \left( \frac{\partial \theta}{\partial z} - \gamma \right)
\]

(8)

for the whole air column.

During the evening transition from 1800 to 2000 LST, competing processes are at work. In the lowest 300 m, a warm converging slope wind loses heat to the surface with turbulent diffusion. Aloft, up to 1 km, cooling by ascent wins out over the upward heat flux in small eddies above the slope wind.

The downward phase tilt above 3000 m (Fig. 13a) is mainly caused by the transient gravity waves, with \( W \) perturbations with amplitude less than 0.08 m s\(^{-1}\). These gravity waves are generated by the heating in the valley, and are reflected by the valley walls toward the center line of the valley. The wave transport wave energy upward and their phase lines move downward, with dominant period around 1 h (Lane and Clark 2002). They cancel each other vertically and make little contribution to the surface pressure variation. In other circumstances (e.g., shallow wide valleys), such gravity waves would equilibrate pressure between the valley and the larger environment.
FIG. 13. The diurnal evolution for each term in the heat budget analysis at the center of the valley (along the vertical line $x = 0$ in Fig. 9) for the WRF simulation for (a) quiescent valley, (b) valley with ridge-top westerly ($U = 15 \text{ m s}^{-1}$), and (c) shallow valley ($h_m = 700 \text{ m}$). The dashed line indicates the mountain top. The solid contour is the zero line. Note that 0600 LST is the sunrise time and 1800 LST is the sunset time.
The energy budget analysis here makes it clear that our defined afternoon “mixed-layer” depth in section 3c includes both the convective boundary layer (CBL—lower sublayer, with warming by vertical turbulent diffusion) and the slightly stable upper layer (upper sublayer, with warming by subsidence). This upper layer is a transition zone between the CBL and the free atmosphere. It may not merge into the lower turbulent mixed layer even in the late afternoon (Bader et al. 1987; Brehm and Freytag 1982; Kondo et al. 1989; Kuwagata and Kimura 1995; Whiteman 1982). Above this upper layer, the diurnal pressure and temperature are almost invariant with time.

c. Comparison between model simulations and observations

Comparisons between Owens Valley observations and the idealized WRF simulation show that, although surface temperature amplitudes are comparable (both about 7.5 K), the surface pressure amplitudes and midlevel temperature amplitudes are much smaller for the simulation results (120 vs 30 Pa). Also, the simulated nighttime shallow inversion near the ground is too strong in the model and the nighttime surface temperature drop is overestimated (Fig. 6b). The nighttime shallow strong inversion near the ground contributes about 5 K for the diurnal surface temperature amplitude.

There are several causes for these discrepancies. First, our imposed idealized midday sensible heat is less than one-half of the Owens Valley estimates. Also, we have neglected the additional longwave heating of the valley atmosphere from the valley floor. In addition, we have neglected the diurnal along-valley winds that are evident in Owens Valley. For the problem with the nocturnal inversion strength, we suspect fundamental problems with the model subgrid-scale mixing under stable conditions (Moeng and Wyngaard 1989).

d. The effect of ridge-top wind

With the inclusion of a ridge-top wind, the cross-valley circulation is no longer symmetric and the daytime subsidence at the valley center is significantly modified (Fig. 14). The environmental wind can intrude into the valley and generate a convergence zone on the lee side of the upstream mountain barrier (Banta 1984, 1986). Previous numerical studies simulated the turbulent mixing in Owens Valley caused by the ridge-top wind shear (Jiang and Doyle 2008). In their severe shear cases (i.e., ridge-top wind speeds $U > 15 \text{ m s}^{-1}$), the intrusion of the ridge-top wind almost destroyed the cross-valley circulations.

In the WRF simulations, we examine only weak or moderate winds, with a specified wind of $U_1 = 0$ at $z = 3000 \text{ m}$, $U_2 > 0$ at $z > 3000 \text{ m}$, and $V = 0$, with

$$U_2 = 9.5 \text{ m s}^{-1} \text{ for the weak case and } U_2 = 15 \text{ m s}^{-1} \text{ for the moderate-ridge-top-wind case. A forward trajectory analysis (e.g., Bao et al. 2006; Reap 1972; Robinson and Sherwood 2006) was performed using model output to determine the hourly location of an air parcel starting upstream. When the ridge-top wind is weak (Fig. 14a), upstream flow will flush out the upper part of the valley and modify the valley circulation. In moderate-wind cases (Fig. 14b), upstream flow will intrude into the valley and may reach the valley bottom. For both situations a westerly wind is detected at the valley floor in late afternoon, but the origin of that westerly flow is different. In the weak-wind case, the westerly wind at the valley bottom is a modified valley circulation; in the moderate-wind case, the westerly wind is imported air from outside the valley.}$$

Unauthenticated | Downloaded 01/01/24 09:14 AM UTC
The ridge-top wind modifies or even destroys the descending motion in the valley in late afternoon (Fig. 14). The energy budget analysis (Fig. 13b) shows that in the upper part of the valley the vertical subsidence heating is weakened after 1500 LST, preventing further development of the mixed-layer top. Figure 13b also shows the influence of a strong quasi-stationary mountain wave at and above the mountain-top level. From 0600 to 1400 LST, the position of this wave causes upward motion on the valley centerline. The dominant heat budget is the familiar isentropic one for mountain waves, that is,

$$U \frac{\partial \theta}{\partial x} + W \frac{\partial \theta}{\partial z} = 0,$$

with vertical and horizontal advection cancelling. In a sudden way, at 1400 LST the position of the wave shifts, giving subsidence so that both advection terms change sign. Note also the turbulent heat transport near the valley top during the evening transition.

Besides the strength of the ridge-top wind, the diurnally varying stability of the valley atmosphere controls the intrusion effect (Jiang and Doyle 2008; Doyle and Durran 2002). Penetration can only occur late in the day when the stability is decreased sufficiently to allow turbulent interaction between the valley air and the overlying atmosphere (Figs. 13b, 14).

The daily maximum mixed-layer depth was calculated according to (5) for the three WRF simulations (Figs. 11, 14). This gives $H \sim 1600 \text{ m}$ and $H \sim 1400 \text{ m}$ for the weak and moderate westerly cases as compared with $H \sim 1800 \text{ m}$ (with mixed-layer top at 2800 m) for the quiescent case. This agrees with the observational results shown in Fig. 8.

6. The valley depth and seasonal effects

To extend our results to other valleys and seasons, we carried out WRF simulations with varying valley depth and insolation. From these simulations, the concept of a critical valley depth arises.

a. Depth effect

To study the effect of the valley depth on the diurnal valley circulation, a series of idealized diurnal valley simulations were carried out with a fixed surface sensible heat flux $Q_{\text{max}} = 160 \text{ W m}^{-2}$ [(7)], but with modified surrounding mountain height $h_m$ from 600 to 2400 m [(6)]. The surface temperature amplitude increases only slightly for the deeper valleys but the pressure amplitude doubles or triples (Fig. 15a). The explanation for the increased pressure amplitude is the “volume effect.” As the mountain is amplified, the rising slopes reduce the volume and mass of the air in the valley. With less mass to heat, the prescribed sensible heat warms the air more. As shown in Fig. 15b, for deeper valleys ($h_m > 1800 \text{ m}$), the surface pressure and temperature phases converge to 84° and 227°, respectively.

The estimated maximum mixed-layer depths for valleys with different depths are shown in Fig. 15c. According to the baseline simulation, the daily maximum mixed-layer depth for heating strength $Q_{\text{max}} = 160 \text{ W m}^{-2}$ is about 1800 m. For valleys deeper than 1800 m, the mixed-layer depth increases as the valley depth is increased but the valley mixed layer always remains below the ridge-top height ($H < h_m$). The surrounding mountains in this case act like rigid walls to prevent pressure equilibration between the valley atmosphere and the broader atmosphere. For valleys shallower than 1800 m, these trends persist until the valley depth drops to about 1200 m. At this point, the valley circulation is no longer confined to the valley and the mixed-layer depth is similar to what it would be over a broad plain.

For valley depths of less than 1200 m, for our fixed valley width of 30 km, the unique characteristics of the valley atmosphere begin to fade. The amplitude of the surface pressure levels off and the phase of the surface pressure drops significantly, while the amplitude and phase of the surface temperature are still not affected (Figs. 15a,b). These changes in pressure indicate the change from a “closed” to an “open” valley system. As the valley system opens, pressure equilibration between the valley and the broader environment begins.

This transition from confined to an open system is also nicely seen in the energy budget analysis (Fig. 13c). Until noon, when the subsidence layer depth approaches the valley depth, the deep and shallow valley circulations are not too different (i.e., Fig. 13a vs Fig. 13c). After noon, the depth of the heated layer continues to grow by two new mechanisms. Plumes, rising off the ridge tops (Fig. 16), cause converging flow high over the valley center so that subsidence warming continues aloft. In addition, conventional thermal convection develops above the valley, transporting heat up to 2500 m. Some related ascent and, later, descent cause heating anomalies even farther aloft, near 3000 m. By 1500 LST (Fig. 16), as gravity waves attempt to equilibrate the pressure, a “plain to valley” wind develops at ridge-top level. The trajectory analysis shows that an intrusion of cold air overwhelms the valley upslope wind and intrudes into the valley. Thus, the morning circulation resembles that in a deep valley but the afternoon circulation does not.

The decrease in pressure phase angle (Fig. 15b) can be put into a larger context. When air is mostly confined...
in a valley, the mean air temperature in the valley (and thus the surface pressure) will lag a sinusoidal heat input rate by \( \frac{1}{4} \) cycle, as expected for any periodically heated object. In contrast, for heat-generated unbounded gravity waves, the temperature will be in phase with the sinusoidal heating input. An in-phase relationship between heating and temperature is necessary for the heat to generate wave energy. While the phase reduction in Fig. 15b is not that dramatic (i.e., not a full \( \frac{1}{4} \) cycle), the principle is the same. For a shallower valley, with the gravity wave pressure adjustment occurring, the pressure phase will be earlier than in a deep confined valley.

A potential confusion is that the data in Fig. 3a show that the pressure phase in flatter terrain is later than that in valleys; contradicting the above point. In the real world, as the valley depth decreases, the smaller amplitude of the valley pressure signal is overpowered by the continental tide, with a later phase. We simulate this effect by vectorially adding the valley pressure signal to a small continental tide (e.g., amplitude 15 Pa and phase 105° for this special case). The pressure phase for the sum gets later for small valleys (Fig. 15b), in agreement with Fig. 3a.

### b. Seasonal effect

The most variable factor for a valley of fixed geometry is the change in surface sensible heat flux with season. The seasonal change in solar zenith angle is not the only contributor. In wintertime, snow cover increases surface albedo, sharply decreasing the ground sensible heating. In summertime, reduced soil moisture increases the Bowen ratio and the proportion of net radiation converted to sensible heating. To study the seasonal character of the diurnal valley circulation, a series of WRF idealized simulations have been done with a fixed valley depth (\( h_m = 2000 \) m) but with modified sensible heating.

As heating is increased from 0 to 300 W m\(^{-2}\), two distinct regimes are seen, with breakpoint at about 160 W m\(^{-2}\). For weak heating, the pressure amplitude increases linearly with heating rate while the pressure phase stays constant at 275°. This is the confined system behavior in which the mixed layer does not reach the valley top (Fig. 17).

When the heating rate exceeds 160 W m\(^{-2}\), the harmonic pressure behavior changes abruptly. The pressure
amplitude remains fixed at 33 Pa while the phase becomes earlier. As shown earlier, when the mixed layer extends above the valley top, the pressure begins to equilibrate with the broader environment, capping its amplitude (Fig. 17a). As the equilibration proceeds, gravity wave dynamics shifts the phase earlier. The analogy between shallower valley and stronger heating is a useful one. Either a shallow valley or strong heating can make the system open (Fig. 17b).

7. Conclusions

The amplitude and phase distributions of the diurnal component of surface pressure and temperature were calculated for 16 DRI stations in Owens Valley for the March–April 2006 T-REX period. In general, the pressure phases are around 85° (i.e., minimum around 1740 LST) while the temperature phases are around 224° (i.e., maximum around 1450 LST). The energy...
budget analysis for the WRF simulation shows that this 3-h lag between the surface temperature maximum and pressure minimum arises from persistence of subsidence warming aloft beyond the time of maximum surface temperature.

The maximum mixed-layer depth in the valley can be estimated by the ratio of surface pressure to temperature amplitude. This ratio can be used to detect wind flushing of valley air. For weak synoptic westerly days, the ridge-top windflushes out the upper part of the valley and weakens the adiabatic subsidence in the valley. The valley mixed layer becomes thinner in the afternoon. WRF simulations show that for synoptic westerly days the detected westerly wind at the valley bottom in the afternoon may be caused by the intrusion of the ridge-top wind but may also be the result of the modified valley circulation, depending on wind strength and the valley stability.

The amplitude and phase of the surface pressure and temperature at valley bottom provide a new way to categorize the diurnal circulation for valleys with different depths and in different seasons, even without any upper-air observation (Table 1). The distinction between open and closed valley circulation is most helpful in this regard. Open valley circulations can occur in two ways. First, if the mixed layer exceeds the valley depth, horizontal pressure equilibration and plain-to-valley winds can occur. Second, if the valley width-to-depth ratio exceeds a certain value, slantwise pressure equilibration can occur through gravity wave generation. Open valleys can be detected by smaller pressure amplitudes and earlier phases (when tidal effects are removed). Another clue is that the mixed-layer depth, estimated from the ratio of pressure to temperature amplitude, approaches the valley depth.

A closed valley circulation occurs when the valley walls help to heat the upper parts of the valley and prevent pressure equilibration with the larger environment. It can be detected by the large pressure amplitude and late phase. The mixed-layer depth, estimated from the ratio of pressure to temperature amplitude, is less than the valley depth.

Acknowledgments. This research was supported by the National Science Foundation, Division of Atmospheric Sciences, by the following grants: ATM-0531212 to Yale University and ATM-0524891 to the Desert Research Institute. The Independence Airport radiosonde data were collected by scientists at the Institute for Atmospheric Science, University of Leeds, United Kingdom, and were funded by the Natural Environment Research Council (NERC), United Kingdom. The flux tower data were collected and provided by the NCAR ISFF team. The radiosonde and flux data and the DRI AWS surface observations were gathered as part of the Terrain-Induced Rotor Experiment (T-REX). The primary sponsor of T-REX was the U.S. National Science Foundation. We thank Dr. Frank Robinson for his advice on an early draft of the manuscript. We are also grateful for the helpful comments from two reviewers.

REFERENCES

<table>
<thead>
<tr>
<th>Category</th>
<th>Shallow valley in summer</th>
<th>Deep valley in summer</th>
</tr>
</thead>
<tbody>
<tr>
<td>Valley circulation</td>
<td>Fig. 16</td>
<td>Fig. 11</td>
</tr>
<tr>
<td>Pressure amplitude</td>
<td>Smaller</td>
<td>Large</td>
</tr>
<tr>
<td>Pressure phase</td>
<td>Early, ~70°</td>
<td>~83°</td>
</tr>
<tr>
<td>Min pressure</td>
<td>~1640 LST</td>
<td>~1730 LST</td>
</tr>
<tr>
<td>Daily max mixed-layer depth</td>
<td>$H \left( \frac{\text{pressure amplitude/temperature amplitude}}{\text{(5)}} \right) &gt; h_m$</td>
<td>$H \left( \frac{\text{pressure amplitude/temperature amplitude}}{\text{(5)}} \right) \leq h_m$</td>
</tr>
<tr>
<td>Dominant heating mechanism</td>
<td>Vertical advection (a.m.)/vertical resolved eddies (p.m.) + turbulent diffusion</td>
<td>Vertical advection + turbulent diffusion</td>
</tr>
<tr>
<td>Ridge-top plain-to-valley wind</td>
<td>Yes</td>
<td>No</td>
</tr>
<tr>
<td>System</td>
<td>Open</td>
<td>Confined</td>
</tr>
</tbody>
</table>

* If considering continental tide effect, the pressure phase will be later, around 105°.

TABLE 1. Using surface pressure variations to categorize the diurnal valley circulations.


