Diurnal Variation of Downslope Winds in Owens Valley during the Sierra Rotor Experiment

QINGFANG JIANG
University Corporation for Atmospheric Research Visiting Scientist Program, Naval Research Laboratory, Monterey, California

JAMES D. DOYLE
Naval Research Laboratory, Monterey, California

(Manuscript received 29 November 2007, in final form 28 January 2008)

ABSTRACT

The impact of diurnal forcing on a downslope wind event that occurred in Owens Valley in California during the Sierra Rotors Project (SRP) in the spring of 2004 has been examined based on observational analysis and diagnosis of numerical simulations. The observations indicate that while the upstream flow was characterized by persistent westerlies at and above the mountaintop level the cross-valley winds in Owens Valley exhibited strong diurnal variation. The numerical simulations using the Coupled Ocean–Atmosphere Mesoscale Prediction System (COAMPS) capture many of the observed salient features and indicate that the in-valley flow evolved among three states during a diurnal cycle. Before sunrise, moderate downslope winds were confined to the western slope of Owens Valley (shallow penetration state). Surface heating after sunrise weakened the downslope winds and mountain waves and eventually led to the decoupling of the well-mixed valley air from the westerlies aloft around local noon (decoupled state). The westerlies plunged into the valley in the afternoon and propagated across the valley floor (in-valley westerly state). After sunset, the westerlies within the valley retreated toward the western slope, where the downslope winds persisted throughout the night.

1. Introduction

Owens Valley is a quasi-two-dimensional arid valley located in southeastern California and bounded by the Sierra Nevada ridge to the west and the Inyo and White Mountains to the east (Fig. 1). The valley floor elevation is approximately 1200 m above sea level (MSL) with an average width of about 15 km. While the mountains on either side of the valley reach above 4 km MSL in elevation, the crest-to-crest distance is only 30 km or less, making Owens Valley one of the steepest in the United States. The valley is notorious for downslope windstorms, wave clouds, and rotors, which are typically associated with strong westerly winds impinging on the nearly north–south-oriented Sierra Nevada ridge. One of the first modern U.S. multiagency meteorological field programs, the Sierra Wave Project (SWP; Holmboe and Klieforth 1957; Grubišić and Lewis 2004) took place in the vicinity of Owens Valley in the early 1950s, during which a B-29 documented vertical velocities up to $\pm 20$ m s$^{-1}$ in a mountain lee wave. Five decades later, researchers’ interest in waves and rotors over Owens Valley was refueled by recent advances in numerical modeling of terrain-induced rotor circulations (e.g., Doyle and Durran 2002; Hertenstein and Kuettner, 2005). Two field campaigns have been conducted recently over Owens Valley—namely, the exploratory Sierra Rotor Project (SRP) in the spring of 2004 and the more extensive Terrain-Induced Rotor Experiment (T-REX) in the spring of 2006. This paper presents a case study of an afternoon downslope westerly wind event documented during the intensive observational period (IOP) 12 of SRP, which took place in Owens Valley from 13 to 17 April 2004. The objective of this study is to examine the diurnal variation of flows in Owens Valley, which involves multiscale interactions among the large-scale westerlies, mountain waves, and differential heating within the boundary layer (BL).
Over the past few decades, downslope windstorms and large-amplitude waves aloft in the lee of idealized topography have been the subject of many theoretical studies, mostly motivated by observations of the famous 11 January 1972 Boulder, Colorado, windstorm (Lilly and Zipser 1972; Lilly and Kennedy 1973). These studies include nonlinear numerical simulations (e.g., Clark and Peltier 1977; Klemp and Lilly 1978; Peltier and Clark 1979; Doyle et al. 2000) and analytical studies of nonbreaking finite-amplitude waves using Long’s equation (Long 1953; Huppert and Miles 1969; Smith 1985). In addition, it has been demonstrated in a number of studies that hydraulic theory can often provide theoretical framework for downslope windstorms (e.g., Houghton and Isaacson 1968; Smith 1985; Jiang and Doyle 2004). The application of hydraulic theory to the continuously stratified atmosphere has been discussed by Durrant (1986) and Jiang and Smith (2003). For simplification, most theoretical studies of mountain waves make use of simple idealized terrain and a free-slip bottom boundary condition. However, a few recent model-based studies have shown that bottom friction may significantly weaken downslope windstorms (Richard et al. 1989) and modify the mountain-wave drag (Olafilesson and Bougeault 1997; Peng and Thompson 2003). Moreover, the formation of leeside rotors critically depends on BL separation induced by lee waves aloft (Doyle and Durrant 2002, 2004, 2007; Vosper et al. 2006; Jiang et al. 2007). Observations obtained from a recent major field campaign, the Mesoscale Alpine Programme (MAP; Bougeault et al. 2001), also highlighted the importance of the BL impact and windward blocking in modulating waves excited by complex terrain (Smith et al. 2002; Jiang and Doyle 2005). Motivated by MAP observations, the impact and dynamics of wave–BL interactions have been further examined by several authors (Jiang et al. 2006; Smith et al. 2006; Smith 2007; Jiang et al. 2008), who showed that the atmospheric BL could substantially weaken propagating waves and cause rapid decay of trapped waves through partial absorption of downward-propagating wave beams. In addition, it has been shown that the interaction between waves and the BL can be modulated by surface heating and cooling effects; specifically, surface heating or cooling may influence the strength of downslope winds (Doyle et al. 2005), the depth and strength of leeside rotors (Kuettner 1959), BL absorption efficiency (Jiang et al. 2006), and BL separation (Jiang et al. 2007). It is noteworthy that after investigation of 20 Boulder windstorms, Brinkmann (1974) found that strong surface winds and high gusts occur most frequently between about midnight and 1000 mountain standard time (MST) and least frequently in the late morning and afternoon. 

![Fig. 1](image-url)
evening. Little is known regarding the underlying dynamics of the diurnal control of surface windstorms. There have been numerous studies of in-valley winds induced by differential heating over valley slopes and neighboring plateaus, most of which assume “undisturbed” conditions; that is, the large-scale flow is relatively weak (e.g., Gleeson 1951; Whiteman 1990; Rampollini et al. 2004), with a few exceptions. For example, the relationship between valley winds and the synoptic-scale flow has been examined by Whiteman and Doran (1993) using a numerical model, and mechanisms including thermal forcing, downward mixing of momentum, valley channeling, and pressure-driven channeling have been addressed.

The remainder of the paper is organized as follows. The numerical model is described in section 2. The observations are presented in section 3. Section 4 provides a model evaluation. The diurnal variation of downslope winds in Owens Valley is illustrated in section 5 through the diagnosis of a real-data simulation. The relevant physics and dynamics are discussed in section 6 based on idealized simulations, and the results are summarized in section 7.

2. Numerical aspects

The atmospheric component of the Coupled Ocean–Atmospheric Mesoscale Prediction System (COAMPS1; Hodur 1997) is used for this study. COAMPS is a fully compressible, nonhydrostatic terrain-following mesoscale model. The finite-difference schemes are of second-order accuracy in time and space in this application. The BL and free-atmospheric turbulent mixing and diffusion are represented using a prognostic equation for the turbulence kinetic energy (TKE) budget (Mellor and Yamada 1974):

\[
\frac{D e}{D t} = \frac{\partial}{\partial z} \left( l_m e^{1/2} S_e \frac{\partial e}{\partial z} \right) - u'w' \frac{\partial U}{\partial z} - v'w' \frac{\partial V}{\partial z} + g \beta w' \theta' - \frac{\Gamma}{l_m} e^{3/2} + D_e,
\]

where \( e = (u'^2 + v'^2 + w'^2)/2 \) is the turbulence kinetic energy per unit mass; \((U, V)\) and \((u', v', w')\) denote grid-scale horizontal wind vectors and ensemble turbulent wind fluctuations, respectively; \( \beta \) is the thermal expansion coefficient; \( S_e = 0.5 \) and \( \Gamma = 0.17 \) are constants; \( l_m \) is the mixing length formulated based on Mellor and Yamada (1974) and Thompson and Burk (1991); and \( D_e \) represents the subgrid-scale TKE mixing. The subgrid-scale mixing of momentum and heat fluxes is parameterized as

\[
(u'w', v'w') = -K_M(\partial U/\partial z, \partial V/\partial z) \quad \text{and} \quad w'\theta' = -K_H(\partial \theta/\partial z),
\]

where \( K_M \) and \( K_H \) are eddy mixing coefficients of momentum and heat fluxes given by \( K_{M,H} = S_{M,H} l_m (2e)^{1/2} \), and \( S_{M,H} \) are constants. The surface heat and momentum fluxes are computed following the Louis (1979) and Louis et al. (1982) formulation. The grid-scale evolution of the moist processes is explicitly predicted from budget equations for cloud water, cloud ice, rainwater, snowflakes, graupel, and water vapor (Rutledge and Hobbs 1983) and the subgrid-scale moist convective processes are parameterized using an approach following Kain and Fritsch (1993). The short- and longwave radiation processes are parameterized following Harshvardhan et al. (1987).

The initial fields for the model are created from multivariate optimum interpolation analysis of upper-air sounding, surface, commercial aircraft, and satellite data that are quality controlled and blended with the 12-h COAMPS forecast fields. Lateral boundary conditions for the outermost grid mesh are derived from Navy Operational Global Analysis and Prediction System (NOGAPS) forecast fields. The computational domain contains four horizontally nested grid meshes of 91 × 91, 131 × 131, 157 × 157, and 256 × 256 grid points, and the corresponding horizontal grid spacings are 27, 9, 3, and 1 km, respectively. There are 55 levels in the vertical on a nonuniform sigma grid with finer spacing in the lower troposphere. The model top is located approximately at 30 km MSL and a sponge boundary condition is applied to the upper one-third of the domain to reduce the reflection of gravity waves. The terrain data is based on the Global Land One-Kilometer Base Elevation (GLOBE) dataset and the 1-km mesh terrain adequately represents the Sierra and Owens Valley as apparent in Fig. 1b. The model was initialized at 0000 UTC 13 April and integrated over 36 h.

In addition to the real-data simulation, a series of two-dimensional simulations have been performed using idealized topography, a single sounding, and specified surface heat fluxes to further examine the impact of the diurnal variation of surface heat fluxes on waves and valley flow interaction over Owens Valley. The configuration for the idealized simulations will be described in section 6.

3. Observations

The instrumentation deployed in Owens Valley during the SRP special observing period (SOP) include a surface network of 16 automatic weather stations (AWS) from Desert Research Institute (DRI) and two Integrated Sounding Systems (ISS) from the National Center for Atmospheric Research (NCAR). A mul-

---

1 COAMPS is a registered trademark of the Naval Research Laboratory.
tiple-antenna profiler radar (MAPR) was deployed as a part of one of the NCAR ISS and was located 8 km south of Independence in Owens Valley. The other ISS, a mobile ISS (MISS), which includes a 915-MHz wind profiler, was located near the Independence Airport (Fig. 1). During each IOP, radiosondes were launched from both the western side of the Sierra Nevada ridge [i.e., either Lemoore, California, or the NCAR Mobile GPS Advanced Upper-Air Sounding System (MGAUS) site at Fresno, California, alternatively; see Fig. 1 for locations], and Owens Valley (i.e., MISS site near Independence). In this section, we will first illustrate some general characteristics of westerly events documented during the 2-month-long SOP and then focus on a prototypical westerly wind event that occurred on 13 April.

a. Overview of the westerly events during SRP

Shown in Fig. 2 are the time series of the westerly component documented by AWS station 1 for all 16 IOP days characterized by prevailing westerlies during the 2-month intensive observational period of SRP. It is evident that late afternoon westerly events occurred quite frequently in Owens Valley. The time period between late morning (1000 Pacific standard time (PST)) and early afternoon (i.e., 1200–1500 PST) appeared to be relatively calm most days with four exceptions, namely, 26 March and 14, 21, and 28 April; the late morning calm period was absent on 21 April and 26 March and the afternoon westerlies started before local noon on 14 and 28 April. It is noteworthy that the analysis of data from other AWS stations indicates similar diurnal variations. Figure 3 shows the frequency of occurrence of the in-valley westerly events as a function of time computed using data from AWS station 1 for all 16 westerly IOP days. A westerly event is defined as the surface cross-valley wind component (i.e., southwesterly, \(~245^\circ\) from the north) greater than 5 m s\(^{-1}\). For most westerly IOP days the surface westerlies exhibited a similar diurnal variation characterized by a
calm period in the late mornings, which persisted through early afternoon in some cases with strong west-erlies in the late afternoons.

The westerly wind component at 700 hPa, derived from upstream radiosondes, is listed in Table 1 for each westerly IOP day. Except for 26 March and 14, 21, and 28 April, the westerly component at 700 hPa was less than 10 m s$^{-1}$. The four days with stronger westerly winds at 700 hPa correspond to the four outliers without well-defined afternoon westerlies or with an early onset of the westerlies within the valley (Fig. 2). Based on this observation, we suggest that the ridgetop westerly component could be a useful index for the prediction of afternoon westerly flow in Owens Valley: with strong westerly winds aloft, the surface westerlies will be present in the morning, and when the prevailing westerlies above the ridgetop are relatively weak, the surface westerlies most likely reach the valley in the afternoon. The connection between prevailing wind speed above the ridgetop and the onset of surface westerlies in the valley will be further discussed in section 6.

In the remainder of the section, we focus on the westerly event that took place on 13 April 2007.

b. The 13 April westerly event

1) SYNOPTIC CONDITION

The IOP 12 of SRP, which lasted from 13 to 17 April 2004, was characterized by persistent westerlies across the Sierra Nevada mountain range at and above the mountaintop level. The synoptic condition during IOP 12 can be seen in Fig. 4, which shows the COAMPS 12-h forecast of winds and geopotential height fields at 500 hPa valid at 1200 UTC 13 April 2004. A developing trough was located off the west coast of North America and a weak pressure ridge was positioned over the central United States on 13 April, which resulted in relatively strong southwesterly flow impinging on the Sierra Nevada ridge. The general wind directions were approximately perpendicular to the main Sierra Nevada crest and Owens Valley. This is known to be favorable for strong wave generation (e.g., Durran 1990). The geopotential height patterns from the COAMPS 18-h forecast and corresponding National Centers for Environmental Prediction (NCEP) GFS analysis show satisfactory agreement, especially in terms of the location and strength of the main trough (not shown). During the next few days, the trough progressed slowly westward while amplifying, and it directed persistent flow with a predominant westerly component across Owens Valley. The conditions were clear in Owens Valley during IOP 12 with a few wave clouds present in the afternoon. Satellite imagery (not shown) indicates that, during the daytime, scattered shallow cumulus clouds were present over the windward (i.e., western) slope of the Sierra ridge. While the prevailing westerlies persisted throughout the entire intensive observational period, this study will focus on the first day of the event, that is, 13 April 2004.

2) RADIOSONDE ANALYSIS

Figure 5 shows the profiles of the cross-valley wind component and potential temperature, derived from two pairs of radiosondes launched at 1800 UTC 13 April and 0000 UTC 14 April from the upstream side (i.e., Fresno or Lemoore) and the MISS site in Owens Valley, respectively. The upstream soundings indicate that the cross-valley winds were fairly weak in the low-
The wind speed was around 10 m s\(^{-1}\) near the ridgetop level and generally increased with the altitude between 3 and 12 km (note that the 0000 UTC Lemoore sounding only reached 7.5 km). The atmosphere was more stable between 3 and 5 km with an average buoyancy frequency (i.e., \(N = \sqrt{\frac{g \, \theta'_y}{\theta'_z}}\), where \(\theta\) is potential temperature and \(g\) is the acceleration of gravity) of 0.012 s\(^{-1}\). Between 5 km MSL and the tropopause, located approximately at 11 km MSL, the stratification was relatively uniform with an average buoyancy frequency of 0.009 s\(^{-1}\). In general, the winds in the troposphere were relatively steady throughout the observational period except for the lowest 3 km. In the lowest 3 km, the stratification evolved with time because of the diurnal variation of surface heating. A shallow mixed layer was present at 1800 UTC, which became deeper by 0000 UTC 14 April. It is interesting to compare the upstream profiles with those observed in Owens Valley. The wind profiles derived from the two in-valley soundings are similar to their upstream counterparts except that the 0000 UTC 14 April sounding reveals an approximately 1.5-km-deep westerly jet above the surface.

### 3) Surface observations

The 16 automatic weather stations, each of which consists of a standard 10-m meteorological tower and sensors for wind, temperature, relative humidity, and pressure measurements, were deployed near Independence approximately along three linear segments across the valley floor (Fig. 1c). For the convenience of description, these stations are numbered sequentially, with station 1 located in the northwest corner. The cross-valley distance between two adjacent stations is approximately 3 km. Shown in Fig. 6 are time–distance sections of the cross-valley wind component and temperature constructed using the 30-s data from stations 1–6 (Figs. 6a,b) for the 24-h period of 13 April in PST. It should be pointed out that, throughout this paper, we use the terms westerlies, downslope winds, and in-valley flow to loosely refer to the cross-valley wind component. During the observational period, the along-valley wind component was much weaker. A digital filter has been applied to filter out high-frequency perturbations. The diurnal variation of the surface westerlies across the valley floor is evident. Before 1400 PST (i.e., 2200 UTC), it was calm or easterly over the valley floor. A sudden transition occurred around 1400 PST when strong westerlies propagated across the valley floor from stations 1–6 within approximately 70 min, that is, an average speed of about 4 m s\(^{-1}\). In general, the surface westerlies were stronger over the western side of the valley. The in-valley westerly event lasted for approximately 5 h. After sunset (~1900 PST), the leading edge of the westerlies started slowly retreating (~1 m s\(^{-1}\)) westward. The identical time–distance section of the cross-valley wind component constructed using data from stations 7–12 exhibits similar diurnal variation (not shown), implying that these observed features are quasi-two-dimensional near Independence. As expected, the surface temperature exhibited strong diurnal variation with a maximum at 1400 PST and a minimum around 0500 PST respectively. The largest diurnal variation (~25 K) was observed at station 5 (Fig. 6c) where the elevation is the lowest; much less variation was observed over the western side of the valley (e.g., ~15 K observed at station 1). Similar diurnal variation was evident in the surface pressure fields as well (not shown), which were approximately in opposite phase with the temperature variation (i.e., minimum in early afternoon and maximum around midnight). The maximum diurnal variation of surface pressure around 5 hPa was recorded at station 5.

### 4) Wind profiler

Figure 7a shows a time–height plot of the cross-valley wind component derived from MISS on 13 April. Consistent with surface network observations, MISS indicates that the westerlies reached the observation site...
around 1400 PST, before which the air was nearly stagnant in Owens Valley. The depth of the westerly jet was between 1 and 2 km, which was consistent with the 0000 UTC 14 April MISS sounding. Above the surface westerlies, a layer of stagnant or weakly reversed flow was present, separating the surface westerlies from the persistent westerly winds above the mountaintop level. The surface westerlies at the MISS site ceased around 2000 PST, which was consistent with the surface network observations.

4. Model evaluation
To evaluate the real-data COAMPS simulation, the simulated fields are interpolated to the corresponding

---

**FIG. 5.** Profiles of (a) cross-barrier wind component and (b) potential temperature derived from a pair of upstream soundings launched at 1800 UTC 13 Apr from MGAUS (M) and at 0000 UTC 14 Apr from Lemoore (L). The corresponding COAMPS (C) simulated profiles (from the 1-km grid) are shown as curves for comparison. (c), (d) Same as (a) and (b) but derived from two in-valley soundings launched from the MISS site (MS).
observation locations and times, and are shown in Figs. 5–7. Overall, the agreement between the available observations and the COAMPS simulation is encouraging. The simulated upstream soundings show a weak cross-valley wind component below 3 km MSL and stronger winds above the ridgetop (Fig. 5b), in reasonable agreement with the observations (Fig. 5a). In terms of stratification, there is a stable layer between 3 and 5 km with a diurnal variation apparent in the lowest 3 km. The upper troposphere (i.e., from 5 km to tropopause) is characterized by a nearly linear increase of potential temperature with an average buoyancy frequency comparable to the observation (i.e., 0.009 s\(^{-1}\)). The simulated altitude of the tropopause and average buoyancy frequency in the lower stratosphere are in good agreement with the radiosonde observations as well. In Owens Valley, COAMPS indicates weak cross-valley winds below the mountaintop level at 1800 UTC and the presence of a shallow westerly jet above the valley floor around 0000 UTC 14 April as observed. The strength of the simulated westerly jet is in good agreement with the observations (Fig. 5c). As expected, the observed wind profiles show more small-scale features than the simulated ones.

The diurnal variation of the in-valley westerlies and temperature near the surface is well captured by the COAMPS simulation (Figs. 6b,d). According to COAMPS, the air in Owens Valley was nearly stagnant in the morning except for over the western valley slope where moderate westerlies were present. Around 1400
PST, a sudden transition occurred and strong westerlies propagated across the valley floor in a time span of approximately 1 h. The depth and duration of the westerly jet are comparable to the MISS observations. Again, in general, the simulated jet is stronger than observed. There are also notable discrepancies between MISS observations and the COAMPS simulation above 2 km AGL; the observations clearly exhibit more temporal variability. Further diagnosis of the real-data simulation indicates that these discrepancies were likely due to the highly three-dimensional characteristics of the wind field above 2 km AGL in the vicinity of MISS location, which was strongly modulated by gravity waves aloft. In addition, it should be noted that the signal-to-noise ratio is much lower above 3 km AGL.

In summary, the COAMPS real-data simulation captures many of the observed salient features of the westerly event, including relatively weak cross-valley flow below the mountaintop level, stronger persistent cross-valley flow above the crest, the diurnal variation of the surface westerlies in Owens Valley, and the presence of weakly reversed flow above the shallow westerly jet during the surface westerly event. Considering the highly nonlinear nature of the in-valley westerly event, it is especially encouraging that the timing, duration, and propagation speeds of the simulated in-valley westerly event are in quantitative agreement with the observations, indicative of the skill of the simulation. It is noteworthy that we also simulated the 15 April event using the identical model configuration and COAMPS captured the diurnal evolution of the in-valley westerlies reasonably well.

5. Diurnal variation of in-valley flow

The strong diurnal variation of the westerlies in Owens Valley was observed by both the AWS and MISS and was well captured by the real-data simulation. This allows us to further examine the three-dimensional structure and the evolution of in-valley westerlies and the associated wave response through diagnosis of the numerical simulation.

For a given domain, a useful measure of the domain-wide wave activity is the vertical flux of the horizontal momentum components, which are defined as

\[ M_x(z, t) = \overline{p}_h(z) \int_A u' w' \, dx \, dy / A \quad \text{and} \]

\[ M_y(z, t) = \overline{p}_h(z) \int_A v' w' \, dx \, dy / A, \]

where

\[ u' \]

\[ w' \]

are the horizontal and vertical components of the wind perturbation, respectively, and \( \overline{p}_h(z) \) is the average pressure. These fluxes provide a measure of the vertical transfer of horizontal momentum and can be used to diagnose the wave activity in the simulations.
where the primes denote deviations from the domain averages, $\rho(z)$ is the mean air density, and $A$ is the domain area. The time–height section of the zonal momentum flux, computed using Eq. (2) from the simulation is shown in Fig. 8. The meridional component of the momentum flux (not shown) shows similar diurnal variation with a much smaller amplitude. A few aspects of the momentum flux evolution deserve mention. First, consistent with the onset of the in-valley westerlies observed near Independence, the momentum flux amplitude clearly indicates a primary maximum in the late afternoon, a secondary maximum during the previous night and a minimum between the sunrise and local noon. The strong diurnal variation in the momentum fluxes implies that the observed diurnal variation is likely widespread in Owens Valley rather than localized to the Independence area. Secondly, the amplitude of the momentum flux decreased rapidly with increasing altitude, and was substantially smaller in the upper troposphere. The maximum momentum flux convergence occurred just above the mountaintop level, indicating that most of the wave momentum flux was deposited into the lower troposphere through low-level wave breaking or hydraulic jumps.

Shown in Figs. 9a,b are plan views of the zonal wind component and wind vectors at the surface valid at 1800 UTC 13 April (i.e., 1000 PST) and 0000 UTC 14 April (i.e., 1600 PST), respectively. At 1800 UTC, the surface winds were characterized by a maximum in the vicinity of the Sierra Nevada crest and were fairly weak within Owens Valley (Fig. 9a). Over the windward slopes of the Sierra Nevada ridge the surface winds were fairly weak, likely due to windward blocking. At 0000 UTC 14 April, the surface winds were substantially stronger over the western valley slope and across the valley floor (Fig. 9b), which is consistent with the AWS observations. Although the wind speed varied considerably along the valley, the strong surface westerlies were widespread over the lee side of the Sierra Nevada ridge. At the mountaintop level (i.e., 3.5 km MSL), the upstream winds were southwesterly and much stronger than the surface winds (Figs. 9c,d). The vertical velocity field at the 5-km level shows a primary wave located along the Sierra Nevada crest, characterized by flow descent immediately over the lee slope and ascent slightly downstream. Trapped wave patterns extended further downstream. It is noteworthy that the vertical motion at 0000 UTC is significantly stronger than that at 1800 UTC, while the mountaintop-level winds show little change during the same time period.

Analysis of a series of vertical cross sections indicates the existence of three distinctive in-valley flow states (Fig. 10). During the evening, downslope winds of moderate strength were confined over the western slope of the valley and reached only partially into the valley, seemingly lacking sufficient energy to flush out the nearly stagnant cold valley air. Instead, the isentropes ascended sharply associated with a sudden decrease of the cross-valley wind speed, the structure of which resembled an internal hydraulic jump (Fig. 10a, referred to as the shallow penetration state). After sunrise, a well-mixed layer began to develop in the valley associated with surface warming. The downslope flow weakened with time and its leading edge slowly retreated toward the western valley slope. By 2000 UTC (i.e., 1200 PST), the well-mixed valley air was nearly decoupled from the westerlies above (Fig. 10b, referred to as the decoupled state), and accordingly, the wave amplitude above the mountaintop was significantly weaker than before sunrise. At approximately 2200 UTC, the well-mixed layer extended beyond the mountaintop level and the incoming air near the western ridge crest became slightly colder than the valley air at the same level (Fig. 10c). The westerlies aloft abruptly plunged into the valley along the western slope, the head of which resembled a gravity current. As its leading edge propagated across the valley floor, the relatively thin surface westerly flow (1–1.5 km) accelerated and subsequently lifted the stagnant valley air up and away from the valley floor. Aloft, the valley air flowed westward toward the Sierra Nevada crest similar to the
structure of a classical exchange flow (Baines 1995). At 0100 UTC 14 April (1700 PST), the surface westerlies were well developed and extended over the peak of the Inyo Mountains (Fig. 10d) with a layer of nearly stagnant or weakly reversed air located between the surface westerlies and prevailing westerlies aloft (referred to as in-valley westerly state). During the course of the development of the surface westerlies, the waves aloft gradually became stronger. After sunset, the leading edge of the surface westerlies retreated toward the western slope of the valley, while cold stagnant air accumulated over the eastern side of the valley (Fig. 10e). Between the late evening and sunrise, the westerlies were confined over the western slope, while a cold pool established over the valley floor; that is, the in-valley flow returned to the shallow penetration state (Fig. 10f).

It should be pointed out that the flow states and transitions described above were consistent with observations. For example, the nearly calm condition in the valley before 1400 PST associated with the decoupled state, the sudden transition from the decoupled state to the in-valley westerly event around 1400 PST, and the slow retreat of the in-valley westerlies after sunset were documented by the AWS network (Fig. 6). The presence of the stagnant layer between the surface westerlies and the prevailing westerlies aloft was observed by both the wind profiler and radiosondes (Figs. 5 and 7).
The nocturnal shallow penetration state likely corresponded to the secondary momentum flux amplitude maximum shown in Fig. 8. The westerlies did not reach the valley floor during the night, in agreement with the AWS network and MISS observations.

The sudden transition from the decoupled state to the in-valley westerly state requires downward momentum transport through processes such as vertical momentum mixing (Whiteman and Doran 1993) or a density-driven current (Benjamin 1968). An examination

**Fig. 10.** Vertical cross sections of $u$ (grayscale, increment: 2 m s$^{-1}$) and isentropes (contour, interval: 1 K) oriented approximately along stations 1–6. Only the portion between 1 and 7 km MSL is shown. Areas with the eddy viscosity $K_m$ greater than 100 m$^2$s$^{-1}$ are hatched. The thick and dashed curves correspond to $u = 4$ and 0 m s$^{-1}$, respectively. (a) 1200, (b) 2000, and (c) 2200 UTC 13 Apr and (d) 0100, (e) 0400, and (f) 1200 UTC 14 Apr. The corresponding PST time is labeled in each panel.
of the potential temperature and buoyancy $B$, defined as $B = g\theta'/\overline{\theta}(z)$, where $\theta' = \theta - \overline{\theta}$ and $\overline{\theta}$ is the horizontal average potential temperature] fields immediately prior to the time the in-valley westerlies reached the valley floor indicates that the valley air was warm and convectively unstable (Fig. 11a). Over the upper portion of the western slope, the air above the thin surface layer was slightly colder than the valley air at the same level, which is consistent with the density-driven current mechanism. The temperature difference was of the order of 0.5 K, and the resulting gravity current speed can be estimated using $C = \sqrt{g' H}$, where $g' = g\theta'/\overline{\theta}$. Assuming a current depth $H \sim 1000$ m, $\theta'/\overline{\theta} \sim 0.5/300$, we obtain $C \sim 4$ m s$^{-1}$, which is comparable to the observed propagation speed of the in-valley westerlies (section 3b). Diagnosis of the momentum flux induced by subgrid turbulence mixing [i.e., $K_m(\partial u_c/\partial z)$, where $u_c$ is the cross-valley wind component] indicates that substantial downward momentum mixing occurred over the upper portion of the western slopes before the plunging of the westerlies occurred, associated with strong turbulence induced by the vertical wind shear and thermal effects (Fig. 11b). Therefore, both momentum mixing and the gravity current may have contributed to the triggering of the afternoon in-valley westerlies. This issue will be explored further in the following section.

In summary, the real-data simulation indicates that surface westerlies in Owens Valley and wave response aloft evolved among three states during a diurnal cycle: (i) a nocturnal shallow penetration state, characterized by moderate downslope winds over the western slope, cold stagnant air over the valley floor, and moderate wave response aloft (Fig. 12a); (ii) a morning decoupled state, characterized by well-mixed stagnant air in the valley and weak wave response aloft (Fig. 12b); and (iii) an afternoon in-valley westerly state, characterized by westerlies within the valley and a strong wave response aloft (Fig. 12c). The “shooting flow” state (e.g., Smith 1985; Durran 1986), which is not present in this simulation, is included in Fig. 12d for comparison purposes. In the in-valley westerly and shooting-flow states, the presence of the stagnant air layer between the surface and the prevailing westerlies is likely crucial for the existence of the surface westerlies (Smith 1985). However, the well-mixed stagnant layer in the shooting-flow state is induced by vigorous low-level wave breaking and typically requires strong low-level winds and stability (Durran 1986). In the westerly state, the unstratified stagnant layer forms during the decoupled phase and its existence in the valley likely requires the bounding effect of the eastern valley slope. In addition, secondary waves are often generated at the leading edge of the shooting flow (Durran 1986), which was not observed in the in-westerly state. It is also noteworthy that similar diurnal evolution among these three flow states in Owens Valley was also evident in the observations and real-data simulation of the 15 April event.

6. **Idealized simulations**

The diurnal variation of the westerlies in Owens Valley illustrated in the last section likely involves nonlinear interactions among large-scale flow, gravity waves, the atmospheric BL, and surface heating. A systematic
investigation of all these mechanisms is beyond the scope of this paper. The objective of this section is to seek a consistent interpretation of the underlying dynamics that governs the diurnal variation of flows over Owens Valley. Specifically, we address the following questions: (i) in the nocturnal shallow penetration state, what prevents the downslope flow from further penetrating into the valley; (ii) what is the dynamic mechanism that governs the slow transition from the shallow penetration state to the decoupled state; (iii) what process triggers the abrupt transition from the decoupled state to the in-valley westerly state; and (iv) what determines the timing of the onset of afternoon westerlies in Owens Valley?

To address these issues, we carry out a series of two-dimensional simulations with idealized topography and a single sounding. The model domain comprises 1001 grid points in horizontal with a grid spacing of 1 km and 90 vertical levels with vertical grid spacings varying from 30 m in the BL to 700 m near the model top. The model top is located at 21 km and Rayleigh damping is applied to the top 7 km. The model is initialized using the 1800 UTC 13 April MGAUS sounding. Open boundary conditions are applied along the lateral boundaries. For valley simulations (Table 2), an idealized valley is located at the domain center between two sinusoidal ridges described by

![FIG. 12. Schematics depicting four mountain-valley flow states: (a) shallow penetration, (b) decoupled, (c) in-valley westerlies, and (d) shooting flow.](image)

<table>
<thead>
<tr>
<th>Acronym</th>
<th>Description</th>
<th>Surface heat flux</th>
<th>Idealized terrain</th>
</tr>
</thead>
<tbody>
<tr>
<td>CPV</td>
<td>Constant positive flux</td>
<td>$F_s = 300 \text{ W m}^{-2}$</td>
<td>Valley</td>
</tr>
<tr>
<td>CNV</td>
<td>Constant negative flux</td>
<td>$F_s = -100 \text{ W m}^{-2}$</td>
<td>Valley</td>
</tr>
<tr>
<td>CRefV</td>
<td>Reference: no heat flux</td>
<td>$F_s = 0$</td>
<td>Valley</td>
</tr>
<tr>
<td>CPR</td>
<td>Constant positive flux</td>
<td>$F_s = 300 \text{ W m}^{-2}$</td>
<td>Isolated ridge</td>
</tr>
<tr>
<td>CNR</td>
<td>Constant negative flux</td>
<td>$F_s = -100 \text{ W m}^{-2}$</td>
<td>Isolated ridge</td>
</tr>
<tr>
<td>CRefR</td>
<td>Reference: no heat flux</td>
<td>Reference, $F_s = 0$</td>
<td>Isolated ridge</td>
</tr>
<tr>
<td>DRefV</td>
<td>Diurnal: control</td>
<td>from $-100$ to $500 \text{ W m}^{-2}$</td>
<td>Valley</td>
</tr>
<tr>
<td>DRefR</td>
<td>Diurnal: control</td>
<td>from $-100$ to $500 \text{ W m}^{-2}$</td>
<td>Isolated ridge</td>
</tr>
<tr>
<td>DRHV</td>
<td>Diurnal: reduced heating</td>
<td>from $-100$ to $250 \text{ W m}^{-2}$</td>
<td>Valley</td>
</tr>
<tr>
<td>DRMV</td>
<td>Diurnal: reduced mixing</td>
<td>from $-100$ to $500 \text{ W m}^{-2}$</td>
<td>Valley</td>
</tr>
</tbody>
</table>
where the maximum heat flux \( F_m \) is the ridge crest heat flux to mimic the diurnal variation, constant heat flux, or a piecewise sinusoidal surface flux before it passes the ridge crest. As listed in Table 2, in the diurnal variation of in-valley westerlies, a few additional simulations have been carried out with an isolated ridge by setting \( h(x) = 0 \) for \( x > 0 \) in Eq. (3).

The surface stress is computed following Louis (1979) and Louis et al. (1982) using a constant surface roughness length \( z_0 = 0.1 \) cm. The surface heat flux is specified only over the valley (i.e., \( -a < x < a \)), and is zero over the rest of the domain. This is partially justified by the severe blocking effect of the Sierra Nevada ridge as apparent from the weak cross-valley winds below the mountain top level. The flow over the ridge crest has an origin from above the BL and is much less influenced by the diurnal variation of the surface heat flux before it passes the ridge crest. As listed in Table 2, two types of simulations have been carried out with a constant heat flux, or a piecewise sinusoidal surface heat flux to mimic the diurnal variation,

\[
F_h(t) = F_m \sin[2\pi(t - T)/T_0],
\]

where the maximum heat flux \( F_m \) is zero for the spinup period \( T_1 = 3 \) h, \(-100 \) W m\(^{-2}\) for the next 12 h nocturnal cooling (i.e., \( 3 + n T_0 \leq t < 15 + n T_0 \)) and \(-500 \) W m\(^{-2}\) for the 12-h daytime heating (\( 15 + n T_0 \leq t < 27 + n T_0, n = 0, 1, 2, \ldots \)), respectively. The diurnal period is \( T_0 = 24 \) h. Similarly, the surface heat flux is set to zero for the first 3-h spinup period in the constant heat flux simulations as well.

We start with three pairs of simulations with a constant heat flux \( F_h = 0, -100, \) and \( 300 \) W m\(^{-2}\), respectively (i.e., CRefV/CRefR, CNV/CNR, and CPV/CPR in Table 2), to illustrate the response of the in-valley flow to the imposed surface heating or cooling. The simulations are carried out for 24 h, and a quasi steady state is reached in the lower troposphere in approximately 8 h for most simulations. The vertical cross sections of the \( u \) component and isentropes corresponding to \( t = 12 \) h from the three pairs of simulations are shown in Fig. 13. In the absence of surface heating or cooling, the quasi-steady-state solution is characterized by a moderate downslope flow over the western valley slope and nearly stagnant air over the valley floor, which is qualitatively similar to the shallow penetration state identified in the real-data simulation (Fig. 13a). In the absence of the second ridge, both the downslope flow and gravity waves become substantially stronger (Fig. 13b). With surface cooling imposed over the valley, the flow structure is still qualitatively similar to the shallow penetration state (Fig. 13c). However, surface cooling significantly enhances downslope winds and allows the downstream flow to further penetrate into the valley. Over the valley floor, a relatively shallow layer of cold air is evident. When removing the second ridge, the strong downslope flow extends all the way to the mountain base with wave breaking occurring above it (Fig. 13d). Relative to the corresponding reference ridge run (i.e., CRefR), the leeside cooling dramatically strengthens the downslope winds and increases the amplitude of gravity waves. The enhancement of the downslope flow by surface cooling over the valley is consistent with Doyle et al. (2005) and Jiang and Doyle (2008). Figure 13d also suggests that a cold pool over the flat surface downstream of the terrain tends to inhibit further development and propagation of the downslope flow, which is consistent with the finding by Lee et al. (1989). For the CPV run, approximately 4 h after surface heating is initiated (i.e., \( t = 7 \) h), the flow state over the valley transits rapidly from the decoupled state to the in-valley westerly state (Fig. 13e). The simulation with an isolated ridge and surface heating (i.e., CPR) reveals a slow transition between \( t = 6 \) and 14 h from the shallow penetration state to a shooting-flow state as schematically shown in Fig. 12f. In the shooting-flow state, the downstream flow is stronger than the corresponding in-valley westerly state (Fig. 13f) and extends farther downstream with nearly stagnant air above it associated with low-level wave breaking. The leading edge of the shooting flow terminates at the downstream edge of the heating zone (i.e., \( x = a \)), where the shooting flow decelerates and flow depth expands through a hydraulic jump. Relative to the corresponding reference run, CRefR, the surface heating over the lee slope of the ridge significantly enhances the downslope winds and results in low-level wave breaking (Fig. 13f). According to Jiang and Doyle (2008), uniform surface heating tends to weaken the stratification and enhance the turbulence in the BL, and therefore to weaken mountain waves. In the CPR and CPV simulations, the surface heat flux is applied only in the lee, which further reduces the low-level pressure in the lee side and significantly enhances the downslope flow. Shown in Fig. 14 are results from a 42-h simulation with a specified diurnal variation of the heat flux (e.g., DRefV). The two-dimensional simulation reproduces most salient features observed in the valley, namely, the nocturnal shallow penetration state (Figs. 14 and 15a),
FIG. 13. Vertical cross sections of the $u$ component (grayscale, increment: 3 m s$^{-1}$) and potential temperature (contour, increment: 2 K) from the set of idealized simulation with a fixed surface heat flux for (a) valley and $F_s = 0$ (CRefV), (b) ridge and $F_s = 0$ (CRefR), (c) valley and $F_s = -100$ W m$^{-2}$ (CNV), (d) ridge and $F_s = -100$ W m$^{-2}$ (CNR), (e) valley and $F_s = 300$ W m$^{-2}$ (CPV), and (f) ridge and $F_s = 300$ W m$^{-2}$ (CPR). The thick and dashed curves correspond to $u = 4$ and 0 m s$^{-1}$. Areas with eddy mixing coefficient larger than 100 m$^2$ s$^{-1}$ are hatched.
parameters to the control simulation except for temperature profile. As an example, assuming the level can be estimated using a heat budget equation, depth of the warm and turbulent layer in the valley to momentum mixing and the gravity current, require the westerly state discussed in section 5, downward momentum mixing may only play a secondary role in this state resulting density current, and the downward momentum mixing is significantly smaller. Regardless of the variation in vertical momentum mixing, the simulated surface winds show a very similar diurnal variation to the reference simulation (i.e., DReFV) and the onset of the in-valley westerlies still occurs around the time when the surface heat flux reaches the maximum (Fig. 16b). Therefore, the afternoon plunging of the prevailing westerlies into the valley was primarily caused by cold air advection over the Sierra ridge crest and the resulting density current, and the downward momentum mixing may only play a secondary role in this state transition.

When removing the second ridge, the surface westerlies are much stronger during both the cooling and heating phases relative to the reference run (i.e., DReFV), implying that the role of the second ridge is to weaken and confine the surface westerlies (Fig. 16c). For $t > 18$ h, the surface westerlies extend downstream to the location of the second ridge crest. Wave breaking occurs aloft and creates a stagnant layer above the surface jet, resembling the shooting flow state depicted in Fig. 12d.

7. Summary

This paper presents the first documentation of the diurnal variation of downslope winds in Owens Valley...
using observations obtained through the SRP and a high-resolution real-data numerical simulation. During the study period on 13 April 2004, the westerlies over the valley floor exhibited strong diurnal variation in response to the diurnal cycle of the surface heat flux, despite the persistent and fairly steady large-scale westerlies at and above the Sierra Nevada mountaintop level throughout the observational period. The real-data numerical simulation captures many of the salient features observed during this westerly event. According to the observations and the real-data simulation, during a diurnal cycle the in-valley flow evolved among three states: the shallow penetration, decoupled, and in-valley westerly states. From the previous late evening to early morning time period, downslope winds were confined over the western valley slope. Surface heating after sunrise gradually weakened the downslope winds through destratifying the downslope flow and eventually led to the decoupling of the valley air from the westerlies above the mountaintop before local noon. During the decoupled period, the well-mixed layer in the valley became deeper with time and extended beyond the ridge crest in the afternoon, which triggered an abrupt transition from the decoupled state to in-valley westerly state, characterized by a layer of stagnant air separating the surface westerlies over the valley floor from the prevailing mountaintop westerlies. After sunset, the surface westerlies began to retreat slowly toward the western slope, and eventually returned to the shallow penetration state.

FIG. 15. Vertical cross sections of the $u$ component (grayscale, increment: 3 m s$^{-1}$) and potential temperature (contour, increment: 2 K) from idealized simulation DRefV with piecewise sinusoidal surface heat fluxes for (a) $t = 9$ h, (b) $t = 18$ h, (c) $t = 24$ h, and (d) $t = 32$ h. The thick and dashed curves correspond to $u = 4$ and 0 m s$^{-1}$. Areas with eddy mixing coefficient larger than 100 m$^2$ s$^{-1}$ are hatched.
A series of idealized simulations indicate that the shallow penetration state prevails in the valley in the absence of surface heating. Surface cooling over the western valley slope tends to enhance the downslope flow, in contrast to the cooling over the valley floor, which tends to weaken the westerly flow. Surface heating in the lee plays dual role in the development of the downslope flow; it tends to weaken the downslope flow by reducing its stratification and to enhance it through creating a leeside surface low. The former role helps to decouple the valley air from the prevailing westerlies aloft, and the later promotes the afternoon in-valley westerlies. The abrupt transition from the decoupled state to the afternoon in-valley westerly state appears to be triggered by a gravity current developed along the western valley slope associated with the growth of the well-mixed layer in the valley.

Examination of other westerly IOPs documented during the 2-month-long field observational period of SRP indicates that similar diurnal variation of the in-valley westerlies occurred during most of the westerly IOP days with a few exceptions. It has been further demonstrated that all the exceptions correspond to stronger westerlies at 700 hPa upstream of the ridge. Dynamically, stronger mountain-level winds may enhance the downslope flow and force it into the shooting flow state without going through the decoupled state, and, therefore, disrupt the typical diurnal variation of the valley flow.

It is interesting to compare the diurnal variations of the downslope winds in Owens Valley with the famous Boulder windstorms that form in the lee of the Front Range of the Rocky Mountains. Both the Boulder and Owens Valley downslope winds tend to weaken after sunrise and usually cease after 1000 local time, likely because of the weakening of stratification in the down-
slopes. The most remarkable difference between the two is that the afternoon westerlies that are frequently observed in Owens Valley are not a typical characteristic of the Boulder windstorm. As discussed in section 6, the development of the afternoon downslope winds in Owens Valley requires severe upstream blocking as well as a deep well-mixed layer in the lee. The upwind slope of Front Range is much gentler, and the low-level air is subjected to surface heating over the long windward slope during daytime. Consequently, the airflow passing the crest is potentially warmer than the air in the lee side, which precludes the development of a density current. Accordingly, surface heating mostly tends to weaken waves through destratification of the low-level flow.

In summary, a few elements that are likely conducive to the three-state diurnal variation of the flow in Owens Valley have been identified based on numerical modeling and theoretical considerations. These factors include weak flow below the mountaintop level, which promotes severe windward blocking, moderate mountaintop cross-valley winds, which favors a shallow penetration state rather than a shooting-flow state during the night, and a deep valley that traps the decoupled valley air. In addition, the combination of relatively small heat capacity of the arid soil in Owens Valley and the sparse cloud coverage during the observational period results in a strong diurnal variation of the air temperature in the valley and contributes to the formation of the deep well-mixed layer in the lee.

Acknowledgments. This research was supported by the Office of Naval Research (ONR) program element 0601153N. The data were collected in a joint effort by scientists and staff from the Desert Research Institute, National Center for Atmospheric Research, University of Washington, and Naval Research Laboratory. We thank Dr. Vanda Grubišić for providing us the AWS data and Dr. Ron Smith for helpful comments on an early draft of the manuscript. The simulations were made using the Coupled Ocean–Atmospheric Mesoscale Prediction System (COAMPS) developed by the U.S. Naval Research Laboratory.

REFERENCES


Kain, J. S., and J. M. Fritsch, 1993: Convective parameterization


