1. Introduction

With cross-barrier density differences and westerly flow across the Sierra Nevada mountain range in California, a stratified flow over nearly two-dimensional topography is established. Over Owens Valley, this flow adjusts to conditions downstream through a complex dynamical process. The case analyzed here is from 9–10 April 2006. The Sierra Nevada in California is a mountain range that is attractive for the study of stratified flows because of its quasi-two-dimensional shape (Fig. 1). The western slope of the Sierra Nevada rises gently to approximately 4 km MSL at its crest from where it drops steeply by almost 3 km into Owens Valley. At the valley floor, Owens Valley is approximately 15 km wide. At its eastern side lies another mountain range made up of the Inyos to the south and the higher Whites to the north, separated by the gap of Westgard Pass (2230 m MSL). The crest-to-crest distance across Owens Valley is almost 30 km.

This paper examines the complicated interaction of stably stratified air overflowing the passes and crest of the Sierra Nevada with the flow in the deep Owens Valley. We will identify density differences upstream and downstream of the Sierra Nevada to have been crucial for the development of the descending flow. This cross-barrier density-driven approach for flow over mountains was pioneered by Prandtl (1942) and developed further for foehn in the central Alps by Schweitzer (1953) and recently by Mayr and Armi (2008). The underlying theoretical framework is that of continuously stratified hydraulics. Frequently only the special case for one neutrally stratified layer of constant potential temperature is used. These concepts have been termed “hydraulic theory” and “reduced-gravity shallow-water theory,” respectively. An important connection between single-layer reduced-gravity hydraulics and continuously stratified hydraulics can be found in Durran and Klemp (1987) for downslope flows.

The stratified hydraulics situations of interest to us are asymmetric flows, which accelerate as the so-called control, typically formed by a crest or a gap, is approached. The air continues to accelerate down the lee side where it eventually adjusts to the reservoir/airmass conditions encountered there, in a hydraulic jump. The situation here is complex because of continuously stratified layers upstream, each separated by a potential temperature step, and an existing different air mass in the valley downstream into which the downslope flow descends. We will show that the “resonant amplification theory” (cf. Lin 2007, section 5.3.2, for a recent review), which is characterized by
a wave-breaking region aloft and is often invoked to explain downslope windstorms, is not applicable in this case.

During March and April of 2006 this region hosted a large field campaign, the Terrain-Induced Rotor Experiment (T-REX), for which an overview is given by Grubišić et al. (2008) and a special collection of papers is kept by the journals of the American Meteorological Society. The observations including those of the internal hydraulic jump discussed here were made during T-REX. The Sierra Wave Project, the first large field
campaign in Owens Valley (Holmboe and Klieforth 1954, 1957) took place a half century prior to T-REX. Photographs from the Sierra Wave Project showed internal hydraulic jumps aligned almost in parallel with the major terrain features (Kuettner 1958, 1959), but their position varied with the time of day, being farthest downstream in late afternoon. Mobbs et al. (2005) conducted a detailed study of rotor flows in the lee of the Falkland Islands. A similar although simpler oceanic example can be seen in the internal hydraulic jump of the Mediterranean Outflow studied by Armi and Farmer (1988, their Fig. 2.5, section 12) and Wesson and Gregg (1994). In the case studied here, coherent vortex structures originating at the sheared face of the internal hydraulic jump were observed to entrain and mix in the downstream direction. Vortex pairing and collapse were observed as the flow evolved downstream.

Similar to the central Alps (cf. Armi and Mayr 2007; Mayr and Armi 2008), the Sierra Nevada also has embedded gaps through which a relatively small volume of colder air flows. For the Alps, the downstream effects of the gap flows are called foehn streaks (in German: Föhnstriche) and are also encountered in the Sierra Nevada downstream of passes. Gap flows were recently reviewed by Mayr et al. (2007).

Descending downslope flow or foehn in Owens Valley depends not only on the existence of southwesterly flow but also on the potential temperature of the upstream air mass being dense or cold enough to overflow and penetrate the valley cold pool. A numerical study of the importance of cold pools downstream of mountain barriers was conducted by Lee et al. (1989), and an oceanographic example of this phenomenon for stratified flow over an inlet sill was observed by Klymak and Gregg (2003). The existence of a cross-barrier density difference is a necessary condition for the downslope flow to descend over a significant depth as was already seen in Mayr and Armi (2010).

2. Observational overview

We take a stratified-hydraulics approach to interpret the 9 April 2006 foehn case. Such an approach requires knowledge not only of upstream conditions but also of downstream conditions. The locations of measurements used here are shown in Fig. 1. A radiosounding at Lemoore, California, (marked with a “U” in the top panel) measured vertical profiles of the upstream conditions, and a radiosounding at Independence (California) Airport (shown with a “D” in both panels) provided downstream data. In situ and cloud radar data from the University of Wyoming King Air aircraft (Kelly et al. 1992; Damiani and Haimov 2006) fill the gap between the radiosoundings. The aircraft flew a cross-barrier track from the Sierra Nevada to the Inyos and along-barrier tracks covering altitudes from the upper troposphere to a few hundred meters above the Owens Valley floor. Thick white lines in Fig. 1 mark typical tracks.

Automatic weather stations just downstream of the gap formed by Kearsarge Pass and at the Integrated Sounding System 2 (ISS2) facility of the National Center for Atmospheric Research on the gentle slope west of the valley center were used to determine conditions close to the surface and to study details of the gap flow descending from Kearsarge Pass. An instrumented car (weather station on wheels, or WOW; Raab and Mayr 2008) measured horizontal wind speed and direction, temperature, and dewpoint along the road (thin lines in the bottom panel) from Kearsarge Pass trailhead (shown with an open square) to the eastern end of the valley. All results are shown in UTC. Note that local time (LST) is 8 h behind UTC; when diurnal effects are important, LST will also be given. All data were obtained from the T-REX data archive (online at http://www.eol.ucar.edu/projects/TREX/). In addition, visible satellite images at 1-km resolution and ECMWF analyses are shown.

Figure 2a, a photograph taken from the King Air at the location marked by the black triangle in Fig. 1, provides a visual overview of the fully established foehn late in the afternoon (0036 UTC 10 April; i.e., 1636 LST 9 April). Midlevel clouds covered the region upstream, or west, of the Sierra crest. An upstream cloud deck during a foehn event is a common feature also in other foehn regions of the world (Mayr and Armi 2008; Seibert 1990). As the air descends in the lee of the Sierra Nevada, the clouds evaporate to form a so-called foehn gap. A photograph of the evaporating clouds taken from the ground is shown in Fig. 2b. Where the air rebounds, the clouds reform as a crest-parallel “jump cloud.” It will be shown later in the paper that this rebound occurs as an internal hydraulic jump. On top of the jump cloud, pileus clouds formed at several locations. The downward-pointing King Air cloud radar provides a view of the inside of these (Fig. 3). Visible portions of the cloud are depicted in warm colors. The upstream cloud tops rise toward the crest to about 5 km MSL. The cloud radar also shows the subsequent descent and evaporation. The jump-cloud top is about the same altitude as the upstream cloud. The pileus cloud extends almost 1 km higher.

3. Establishment and evolution of the foehn

A combination of changes on the large scale and the local scale made it possible for foehn to penetrate nearly 3 km down the Sierra Nevada slopes to the floor of
Owens Valley. Figure 4 tracks the large-scale evolution at approximate crest height of the model Sierra Nevada in the European Centre for Medium-Range Weather Forecasts (ECMWF) global analyses in 6-hourly intervals from 1800 UTC through 1200 UTC (1000–0400 LST). Throughout this 18-h period, the large-scale flow upstream and across the sierras remains southwesterly as shown by the isohypses (gray lines). Between 1800 and 0000 UTC, large-scale cross-barrier inflow weakened somewhat as inferred from the increasing distance between the isohypses, which is detrimental for the onset of foehn. If one were to judge the likelihood of foehn from mountain height nondimensionalized by the ratio of buoyancy frequency to impinging cross-barrier wind component, this likelihood would decrease with the observed decrease of the cross-barrier flow [for a review of

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**Fig. 2.** (a) Photograph taken from a King Air flying at 6.5 km MSL at 0036 UTC while looking north along the Sierra Nevada. (b) Photograph taken while looking south from the road to Kearsarge Pass at 2344 UTC showing the moist overflow evaporating as it descends down the eastern sierra slope and recondensing to form the upstream edge on the internal hydraulic jump.
flow-regime diagrams, see Schär (2002)]. The crucial large-scale change, however, is the approach of a cold front on the upstream side of the Sierra Nevada. At 0000 UTC the strong temperature gradient shows that the cold air mass has reached the Sierra Nevada, putting colder air with lower potential temperatures on the upstream side than on the downstream side of the Sierra Nevada—at least at crest level.

On the local scale, during the same period of midmorning to midafternoon, diurnal heating has increased potential temperatures in the valley. Both changes, the advection of cold air to the upstream side and heating in the valley, are reflected in the time series of potential temperatures (Fig. 5a) of the weather stations high up on the slopes close to Kearsarge Pass and in the valley. In the morning at 1500 UTC (0700 LST) potential temperatures are lower in the valley than near the pass (thick line) as a result of the nocturnal cooling during the previous night. On this east-facing slope of the Sierra Nevada, wind directions are upslope near the pass and have a combined upslope/upvalley direction at the valley station, as expected for a thermally induced valley wind system (Fig. 5b). Although the valley potential temperature follows the diurnal course and increases, the large-scale cold-air advection is strong enough to eventually arrest the normal diurnal increase near the pass: early in the afternoon at approximately 2200 UTC (1400 LST), when potential temperatures at both the trailhead and valley location reached the same value of almost 304 K, wind direction at trailhead switched to downslope, counter to the slope wind direction expected at this time in the diurnal cycle. Because potential temperature values start to simultaneously decrease, we interpret this as the beginning of the overflow of the colder air from upstream through Kearsarge Pass. When the potential temperature difference between the trailhead and the valley reached about 2 K, the gap overflow rapidly increased to 15 m s$^{-1}$ at trailhead and reached the valley floor, where the winds also picked up and flipped direction to downslope—the onset of gap flow in the valley. A similar concerted action

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**FIG. 3.** King Air cloud radar reflectivity measured from the flight leg at 6.4 km MSL between 2306 and 2310 UTC.

**FIG. 4.** The 700-hPa (approximately at model crest height) temperature (black lines; 2°C interval) and geopotential (gray lines; 200 m s$^{-2}$ interval) at 6-hourly intervals from 1800 UTC 9 Apr to 1200 UTC 10 Apr. The study region is highlighted by the white ellipsis within the gray-shaded region.
of large-scale cold advection reaching close to crest level on the upstream side and diurnal heating of the valley atmosphere was found to be needed for the 2 March T-REX gap-flow case discussed in Mayr and Armi (2010). This mechanism seems to be general; it was also seen in foehn flows in the Alps (cf. Mayr and Armi 2008).

The existence of a road between the trailhead weather station and the eastern end of the valley allowed the WOW to provide high spatial resolution (better than 30 m) in addition to the high temporal resolution of the weather stations. The first WOW traverse in Fig. 6a taken between 2202 and 2238 UTC shows upvalley flow across the whole valley. Over the stretch marked by the gray bar in Fig. 6d, some very small scale topographical features affect the wind direction as the road winds in and out of a canyon in the steep face of the Sierra Nevada (cf. Raab and Mayr 2008). The next traverse is from just after the onset of gap flow in the valley (Fig. 6b). Westerly flow extends across the valley floor to \( x = +6 \) km where it encounters the southerly upvalley flow. Potential temperature increases down the slope and across the western part of the valley. Potential temperatures high on the sierra slopes are 301–302 K. They increase by almost 2 K through the canyon part of the road, with zero being at Independence. All data within the town of Independence are masked.
gap overflow, which was stably stratified on the upstream side (cf. Fig. 4). At $x = -10$ km, where the WOW reaches the smooth gentler slopes of Owens Valley, the flow accelerates slightly. From $x = -9$ km on, potential temperature increases steadily to eventually reach 304 K. We explain this increase simply through mixing of the cold gap overflow with the much larger air mass of the warmer valley flow.

While WOW provides spatial details across the valley, the King Air aircraft leg in Fig. 7 provides the spatial details along the slope of the Sierra Nevada. In this pseudo-3D depiction, the three-dimensional wind vectors are shown as cones. The length of the cones indicates (3D) wind speed, but because of the perspective view a cone of fixed wind speed will appear smaller the farther away it is. Because of the vertical exaggeration of the topography, the vertical component of each wind cone is also exaggerated by a factor of 2.5. Cones are additionally color coded with the vertical velocity component. With stable stratification on the upstream side, the lower the altitude is from which the air originates the more likely it is that its potential temperature is low enough to descend all the way to the Owens Valley floor. Figure 7 shows the vertical flow component along the sierra slope to alternate between strongly downslope to strongly upslope. The downslope regions are numbered 1–3. Region 1 is downslope of Kearsarge Pass through which gap flow occurs. Because potential temperatures of the air overflowing the crest on either side of the pass are higher, their equilibrium buoyancy level is higher on the slope and they rebound strongly already on the upper of the two King Air legs as seen by the dark-red upward-pointing cones in the figure. Region 2 is downstream of another pass of similar altitude to Kearsarge Pass so that strong gap flow can also be seen on both flight legs again with a strong rebound from higher-potential-temperature air coming over the higher crest on both sides of the pass. The altitude of the pass of region 3, on the other hand, is higher so that potential temperatures are only low enough to descend past the upper flight leg. At the lower leg, the air already flows horizontally.

The rebound of the cold air overflowing the crest profoundly influenced clouds downstream of the Owens Valley. A sequence of satellite images in the visible spectrum shown in Fig. 8 depicts the cloud evolution in 15-min intervals starting at 2230 UTC (1430 LST) when the advected upstream cold air has already become deep enough to start overflowing (cf. Fig. 5). The top of the cloud-filled moist jump layer is evident upstream of the sierras and evaporates as it descends across the sierra crest. Downstream in the lee, only from 2315 UTC on is a sharp-edged signature of the rebound of the cold air overflowing the crest evident as part of the evaporated cloud recondenses. This rebound takes place as an internal hydraulic jump, discussed in the next section, that extends north and south over Owens Valley to the limits of the image area shown. This quasi-2D behavior justifies using a single stack section for the description of the flow above the gap overflow. A satellite image at 0045 UTC is also included in Fig. 8 to show the persistence and stationarity of the internal hydraulic jump throughout the King Air flight.
Of the two processes that have made potential temperatures of the overflowing air low enough to reach the valley floor, only one persists after the last satellite image shown at 0045 UTC (1645 LST): synoptic cold-air advection at pass/crest level upstream of the Sierra Nevada as seen in the ECMWF analysis at 0600 UTC in Fig. 4 and in the continuing fall of potential temperatures at the trailhead weather station in Fig. 5a. Diurnal heating of the valley, on the other hand, switches sign to nocturnal cooling, causing potential temperatures close to the valley floor to fall rapidly. At local midnight (0800 UTC) potential temperatures at trailhead have become higher than at the valley station. Wind speeds decrease as gap overflow is undercut by air flowing down the slopes and down Owens Valley. The strong nocturnal radiative cooling in the valley finally provides an air mass that is colder than the synoptically advected air mass flowing through the gap.

WOW data from around 2130 LST (0530 UTC) in Fig. 6c again resolve the spatial details of the flow near the surface. In the valley, flow has switched direction to northerly downvalley flow independent of the southwesterly flow aloft. Since by this time the gap overflow has higher potential temperatures than the valley air, the gap flow from Kearsarge Pass must have flowed over the lowest part of the valley air, as opposed to underflowing it during the afternoon. The lift-off location is at $x \approx -6$ km where wind changes from westerly to northerly. In a similar situation in the Alps, gap overflow from the Brenner Pass frequently flows over the nocturnally cooled air in the Inn Valley and foehn is not experienced in Innsbruck at the valley floor.

By 0400 LST (1200 UTC 10 April) in Fig. 4 (rightmost panel), the temperature contrast across the sierra crest has almost disappeared. At sunrise, indicated by the pronounced temperature increase at both stations in Fig. 5, potential temperature at trailhead is far too warm to penetrate down to the valley station. A few hours later, gap overflow stops also at the trailhead station and the regular diurnal upvalley and upslope wind system becomes reestablished.

Although we have primarily focused in this section on the ground observations of the gap overflow, without the strong synoptically caused cooling at crest height gap overflow would not have been possible since the potential

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Fig. 8. Visible satellite sequence spanning the onset and development of internal hydraulic response. The resolution is 1 km, and each image is $125 \times 125$ km$^2$, with Independence airport K207 marked (site of downstream sounding D in Fig. 1). The Sierra Nevada crest runs approximately along the western edge of the foehn gap.
temperatures at that altitude would have been too warm to descend more than 2 km down to the Owens Valley floor. Note that onset and demise of gap overflow occurred under nearly constant cross-barrier flow at crest height (cf. spacing of the isohyposes in Fig. 4)! As is evident in the sequence of visible satellite images (Fig. 8), the start of the gap overflow occurs simultaneously with the overflow along the entire Sierra Nevada crest west of Owens Valley.

4. The leeside response and internal hydraulic jump

The onset of the gap overflow at the trailhead was seen to occur between 2300 and 2330 UTC (Fig. 5). The visible satellite sequence (Fig. 8) also shows the structure of the foehn gap and well-formed downstream cloud to the east, extending over Owens Valley beginning at 2315 UTC. In this section we will discuss detailed observations made during the 2-h time frame from 2306 to 0120 UTC of the established downslope flow and internal hydraulic jump.

Upstream and downstream soundings are shown near the start of the event in Fig. 9. The important features of the flow are labeled on the soundings, which provide high vertical resolution. The upstream sounding occupies the left part of the figure, and the downstream sounding takes up the right portion; between them in the center, potential temperatures from upstream and downstream soundings are shown together. Features are also labeled on the vertical velocity/turbulence (Fig. 10) and speed/direction (Fig. 11) sections from the King Air. Detailed thermodynamic sections of potential temperature and mixing ratio in Owens Valley are shown in Figs. 12a and 12b. The vertical extent of the clouds can be seen in Fig. 3; because all of the flight legs had to stay clear of clouds, the vertical spacing of the flight legs was not uniform.

A section from the King Air in Fig. 10 shows the vertical velocity and turbulent dissipation rate color coded. At 2246 UTC the wave response along the 7.7-km MSL flight leg shows only a weak response. This leg was the first flown, and the downslope flow as discussed in the previous section was not yet established. In contrast, the repeat leg at 0051 UTC at the same level shows a strong 5 m s⁻¹ wave response, yet the speed and direction of the southwesterly flow aloft are the same for both of these flights, as can be seen in Fig. 11. During the 2-h time frame discussed in this section the established westerly flow was steady, as can be seen from the

FIG. 9. Annotated radiosoundings. (two left panels) Upstream relative humidity (%), mixing ratio (g kg⁻¹), and cross-barrier wind component (m s⁻¹); (center panel) upstream (thin line) and downstream (thick line) potential temperature (K); (two right panels) downstream cross-barrier wind component, relative humidity, and mixing ratio. The upstream sounding at LeMoore was released at 2323 UTC; the downstream sounding at Independence was released at 2306 UTC 9 Apr 2006.
repeat flight legs at an elevation of 6.5 km at 2312 and 0055 UTC.

a. Structure of the flow in the lee

The lowest layer to flow across the Sierra Nevada was seen in the previous section as a gap overflow. In the upstream sounding of Fig. 9 it is just above the elevation of Kearsarge Pass and is capped by a 2.5-K potential temperature step. This “descending gap flow” is labeled on the upwind potential temperature sounding and can be seen in Fig. 11 on the deepest flight level across Owens Valley with its distinctive westerly direction. It is also appears as the cold moist westerly gap underflow in Figs. 11 and 12. Beneath the gap-flow layer, the stratification upstream is blocked.

On the downwind sounding the deepest air mass is the “relatively cool valley air mass,” which is very weakly stratified. It has no cross-barrier component in the downwind sounding and can be seen flowing up valley in the speed/direction section from the King Air (Fig. 11). This upvalley flow dominates the flow below crest level east of −3 km along the 3-km MSL flight leg. With the exception of the deepest flight leg in the gap underflow, the upvalley flow dominates the flow below crest level in the eastern half of the valley. The valley air is drier (Fig. 12) than the moist gap and descending flow to the west.

The layer in the upwind sounding extending from 4.5 to 6.5 km will be seen to dominate the internal hydraulic jump over Owens Valley. Upstream it flows westerly at 20 m s\(^{-1}\) throughout its 2-km vertical extent. This “jump layer” has moist and dry sublayers each with 1-km vertical extent. The lower “moist” portion was cloud filled, as can be seen in Figs. 2 and 3 as well as the satellite images of Fig. 8. The “cloud tops” are at the top of this sublayer and are labeled in both the upwind and downwind soundings. The “dry” sublayer with a 0.2 g kg\(^{-1}\) mixing ratio upstream is mixed with moist air from the moist sublayer below; downstream the mixing ratio has increased to 0.3 g kg\(^{-1}\), sufficient to allow formation of the pileus cloud seen above the jump in Figs. 2 and 3. Note that in the downstream sounding

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**Fig. 10.** Vertical cross section of King Air aircraft data approximately from west to east: vertical wind speed (m s\(^{-1}\)) is shown as lines oscillating around aircraft tracks and color coded with the \(u\) component of the turbulent dissipation rate [color bar: (m\(^2\) s\(^{-3}\))\(^{1/3}\)]. The early leg at 7.7 km is dashed for clarity.
the jump layer is nearly linearly stratified in both potential temperature and mixing ratio. It is also sheared, in contrast to being unsheared in the upstream sounding. Downstream, the jump layer is bounded below by the valley air and above by the “flow aloft.”

b. Details of the internal hydraulic jumps

Details of the structure of the internal hydraulic jump were visualized in unprecedented detail with the Wyoming cloud radar flown on the King Air. Figure 13 shows an image produced from the aircraft flying at 3.1 km MSL with the radar looking upward beneath the internal hydraulic jump. Upstream (west) is to the left, and the start of the cloud near the beginning of the jump is 3.5 km west of Independence. The advantage of the radar over visual images is that radar can penetrate and reflect from within the entire cloud layer. Figure 13 has a 1:1 aspect ratio to facilitate a discussion of the turbulence elements seen in it. The wind was from 210° at the levels shown in the image, but the flight track was upwind along 245°; hence the image appears to be lengthened by about 20% from a cut at 210°. There is a compensating “focal plane shutter”–like shortening of about 10% for this image because of the ratio of the flow speed to the aircraft speed.

The coherent structures visible in the image contain the strongest reflectors (yellow and red in the image) and originate to the west at the upstream face of the internal hydraulic jump. They have the distinctive signature of finite-amplitude shear-layer instabilities. These shear-layer coherent structures have been seen many times such as in the radar images of Atlas et al. (1970) of shear-layer instabilities. Similar “subrotor vortices” can be seen in the numerical simulations of Doyle et al. (2009, their Figs. 17, 19, and 21) along the ascending branch of the lee wave in the region of strong vertical wind shear. They have also been observed in laboratory experiments [see Pawlak and Armi (1998) for a review of some of these]. The coherent structures are responsible for the initial mixing of the jump layers.

The first imaged vortex marked with an arrow in Fig. 13 centered 1.5 km west of Independence has
entrained high-reflectivity (4 dBZ) moist air from below and wrapped it around its core. The next core centered 0.5 km west of Independence is less distinct, but the vortex centered at 0.5 km east of Independence has a clear mass of high reflectivity (5 dBZ) mixed throughout its core. The next two vortices downstream at 1.5 and 2.5 km east of Independence appear to be pairing, a process first clearly identified by Winant and Browand (1974) and Brown and Roshko (1974) as the mechanism for growth of a shear layer. With the pairing, additional entrainment of dry air from above and moist reflective air from below can be seen in the paired vortex that is 2 times the size of the upstream vortices.

From 4 km east of Independence to the end of the reflective image, shown here at 8 km, the spacing of the vortex cores has doubled to 2 km, which is consistent with the vortex pairing seen upstream. The finite-amplitude Kelvin–Helmholtz eddies are at 1-km spacing initially and grow by pairing to 2-km spacing. They start at the upwind jump face and collapse after only one pairing downstream.

The laboratory experiments of Koop and Browand (1979) for growth and collapse of a stratified shear layer can be used to interpret the result shown here and the initial instability. The sounding from Independence crossed the shear region at 4.9 km MSL at 2318 UTC, only 15 min earlier than the radar image shown. The sounding crossed the shear about 3 km north of the aircraft track, and the projection of its path on the image is shown in Fig. 13. The Richardson number computed
from the sounding was $\text{Ri} \approx 0.5$ in the shear as shown on Fig. 9. This implies the shear in the portion showing the collapse of the coherent structures was just stable. Upstream in the sheared portion of the flow in which the instabilities develop, however, the $\text{Ri} \approx 0.25$ since each pairing doubles the Richardson number. This is consistent with the observed instabilities forming as a result of shear at the upstream face of the internal hydraulic jump acting on a marginally stable flow, causing it to go unstable. The Richardson number associated with the instability is then close to one-quarter, which after only a single pairing results in a stable flow again downstream.

We defined the length of the hydraulic jump from the location of the face where shear and upward vertical velocity increase at $x = -2.5$ km in Fig. 13 to the end of the mixing region, where the turbulence collapses, at $x = 8$ km. In situ data from the aircraft track beneath, at the altitude from which the image was collected (3.1 km), are available and are also shown in Fig. 13. A large vertical velocity of 7 m s$^{-1}$ was observed beneath the upstream part of the image; a similar large $\pm 0.5$-K fluctuation of potential temperature is observed out of phase with the large velocity fluctuation. These fluctuations die out after the collapse of the instabilities around 3 km east of Independence. Similar results were found for the growth and collapse in the laboratory experiments of Koop and Browand (1979) although at a much lower Reynolds numbers ($10^4$ vs the $10^5$ here).

The intensity of the turbulence and the amplitude of the vertical velocity oscillations decay over about 10 km as also seen in Fig. 10. The 0120 UTC track at 4.7 km MSL just beneath the cloud measured strong descent ($-5$ m s$^{-1}$; continuous line) in the part closest to the crest. The descending part is nearly laminar, with low turbulence levels. When the aircraft flew through the upstream face of the internal hydraulic jump it encountered a nearly laminar outer “sheet” of the jump with still very low turbulence levels. Turbulence increased to the east as the aircraft entered the turbulent part of the jump and diminished gradually across the valley toward the foot of the Inyos. The large amplitudes of vertical velocity in rapid up–down sequence have already decayed by midvalley. The valley air mass is turbulent, with the highest dissipation rates of $\varepsilon^{1/3}$ of about 0.35 (m$^2$ s$^{-3}$)$^{1/3}$ closest to the Sierra Nevada, decaying toward the center of the valley but remaining fairly high with dissipation rates of about 0.2 (m$^2$ s$^{-3}$)$^{1/3}$.

c. The flow response aloft

The flow aloft seen on flight legs at 6.1 km MSL and above in Figs. 10 and 11 is laminar. Also note that the cross-barrier flow at these altitudes is nearly constant along each flight leg. Vertical velocities aloft do not exactly follow the layer underneath. Instead, the smooth vertical velocity changes follow the envelope around the strong up and down oscillations in the jump layer seen at 4.6 km MSL. These correspond to the wavelength of the single wave above 6.1 km MSL.

The flow response aloft does not appear as a classical mountain wave (e.g., Queney 1948). The wavelength of a classical linear mountain wave would be comparable to the scale of the whole mountain range and the wave would be above it. Here (Fig. 10), however, the wavelength is much shorter and the wave starts at the sierra crest. We hypothesize that the flow aloft does not respond directly to the sierras. Instead it responds to the position of the potential temperature step capping the jump layer seen at 6.4 km MSL in the downstream sounding and 6.6 km in the upstream sounding. The crucial horizontal scale is the size of the foehn gap,
which is approximately 9 km. As such, the pattern of the wave follows more closely the one found for a valley cut into a plateau (Baines 1995, section 5.1.4.1), where the valley width is equivalent to the foehn gap length. The undulation aloft is a response of the flow in that layer to a change of the shape at its base, in this case a potential temperature step of 6 K. Only when the length of the bottom shape divided by the wind speed reaches a scale similar to the buoyancy period in that layer, can an undulation or standing wave exist.

The amplitude of the wave aloft correlates with depth of the descent of the downslope flow. Mayr and Armi (2010) show this for morning and afternoon flights on 2 March 2006. For the 9 April case here, the strength of the wave aloft is weak along the 7.6-km MSL flight leg at 2246 UTC (Fig. 10) before the establishment of the downslope flow. In contrast the flight leg at the same altitude after establishment encountered strong ($w > 5 \text{ m s}^{-1}$) vertical velocities. The speed and direction for both of these flight legs were nearly identical as seen in Fig. 11. The 700-hPa analysis, approximately at crest height, is shown for the 5 h prior to the event in Fig. 4. This southwesterly flow had been flowing for more than 24 h.

5. Discussion and conclusions

The observations of the components of the descending stratified flow across the Sierra Nevada into the valley air mass in Owens Valley are summarized with the aid of the sketch shown in Fig. 14. The lowest layer to cross the Sierra Nevada did so only through the passes. The gap overflow can be delineated in the upstream sounding (Fig. 9) by the increase of the cross-barrier wind component, and a capping potential temperature step at 4.2 km MSL, a layer of increased stability just above the level of the sierra crest. Above the gap overflow layer, the jump layer with two sublayers can be distinguished in the soundings. The moist sublayer on the upstream side was cloud filled. The cloud layer evaporated in the descending part of the flow and condensed over the ascending part of the jump layers at the western side of Owens Valley. The clouds and the foehn gap were seen in the reflectivity from the aircraft cloud radar (Fig. 3), photographs (Fig. 2), and visible satellite images (Fig. 8).

The jump, in which the mixing took place, is not the classical hydraulic jump. Across a classical hydraulic jump or a heated jump (Kuettner 1958), mass and momentum are conserved but energy is not. For this particular jump, mass was not conserved because the jump layer flowed into a different air mass and partially mixed with it. The upper and westernmost parts of the valley air mass were modified by the moist jump layer. The dry sublayer of the jump layer was cloud free and the mixing ratio was significantly less than in the lower cloud-filled jump layer. Downstream of the hydraulic jump, the jump sublayers are not distinct but the stratification of the layer was continuous and mixing ratio nearly linear in its decrease with altitude (Fig. 9).

The base of the layer aloft was demarcated upstream of the Sierra Nevada by a potential temperature step at approximately 7 km MSL, which also formed the top of the jump layer. Cross-barrier flow was higher ($\sim 35 \text{ m s}^{-1}$) and roughly uniform with altitude. This layer was very dry both in absolute and relative terms. The base of the wave response aloft followed the underlying layer in its descent across the sierras and was about 0.5 km lower in the downstream sounding. Because of the strength of the potential temperature step, the response aloft was directly to the shape of the underlying jump layer as if the shape of the strong step capping the dry jump layer were the (invisible) topography.

On the basis of the soundings in Fig. 9, the jump-layer internal Froude number above the sierra crest,

$$F_i^2 = \frac{u^2}{(g'\delta)}, \quad (1)$$

is critical, $F_i^2 \approx 1$. Here, $u$ is the layer speed ($\sim 20 \text{ m s}^{-1}$), $g'$ is reduced gravity based on the strong potential temperature step of 6 K and $\delta$ is the thickness of each layer ($\sim 2 \text{ km}$). The situation here is complex because the jump layer is nonuniformly stratified and moving beneath a deep stratified layer above. Although Wood (1968) gives an analytical solution for a flow of arbitrary stratification through a contraction, we are not aware of an analytical hydraulic treatment for this complex situation for flow over a crest. On the basis of the asymmetry of the flow as it accelerates and descends across the sierras,
however, the jump layer is most likely controlled in the neighborhood of the crest.

T-REX took place seven years after another large-scale field campaign, the Mesoscale Alpine Programme, which studied foehn and gap flows in the Alps [see the reviews of Mayr et al. (2007) and Drobinski et al. (2007)]. There are notable similarities but also differences between the downslope flows found in both locations. A crucial difference stems from the larger and steeper terrain drop of almost 3 km for the Sierra Nevada as compared with 1–2 km for the Alps. The sierras also lack channeling through glacial cuts as provided, for example, by the Wipp Valley and the Rhine Valley in the Alps. Foehn-like winds are relatively rare in the Owens Valley, because seldom are potential temperatures in the valley warm enough to allow the flow to descend to the valley floor, as explained by Armi and Mayr (2007) and Mayr and Armi (2008) for the Alps. The larger terrain drop combined with a stably stratified layer on the upstream side rarely allows the air at the valley floor to have warmer potential temperatures than air coming across the crest. Air coming across gaps can more easily and frequently have colder potential temperatures relative to temperatures in the valley than air coming across the crest. Most cases of strong winds in the Owens Valley result from flow channeling, that is, have northerly or southerly directions, whereas winds shooting down the Sierra Nevada slope are much rarer (Zhong et al. 2008).

For the Alpine cases (Armi and Mayr 2007; Mayr and Armi 2008), stable stratification on the far upstream side resulted in a Wood’s solution (Wood 1968) on the near-upstream side with isolating, nearly neutrally stratified and stationary layers both above and below the flowing layer and a linear stable stratification with a parabolic-like velocity profile in the flowing layer. Downstream, the flowing layer became mixed by turbulence at the ground and in hydraulic jumps into a single neutrally stratified layer. The mixing was either complete or covered only the lower part in the case of very deep flowing layers. The isolating layer prevented significant undulations aloft. A similar isolating layer was invoked by Smith (1985) in his steady-state solution. No such isolating layer was observed here, presumably because of a preexisting potential temperature step upstream at 7 km MSL above the jump layer and below the flow aloft. The absence of an isolating layer aloft excludes the resonant amplification theory [see Lin (2007), section 5.3.2, for a recent review] as a mechanism for the formation of the downslope flow in this case.

The most important similarity between Sierra Nevada and Alpine foehns lies in the cross-barrier density difference. The key condition for foehn to occur downstream of the Sierra Nevada into the deeply cut Owens Valley was colder potential temperatures (higher density) in the flowing layer on the upstream side than near the slope or valley surface on the downstream side. This confirms similar conclusions drawn in Mayr and Armi (2010) for the T-REX case of 2 March 2006 and with a more limited dataset from four cases during the pilot phase of T-REX in 2004 (Raab and Mayr 2008). In agreement with Mayr and Armi (2008) for the Alps, a strong cross-barrier component by itself was not sufficient to cause air to flow down toward the valley on the lee side. The cross-barrier wind component at crest level remained nearly constant at 20 m s$^{-1}$ before, during, and after the foehn event.

Often the occurrence of downslope windstorms, which are also known as deep foehns, is predicted by using a simple nondimensional mountain height computed from upstream values of a layer-averaged buoyancy frequency and cross-barrier component. As the 9–10 April 2006 case, the 2 March case (Mayr and Armi 2010), and the above-cited Alpine cases have shown, upstream and downstream densities or potential temperatures need to be known. In addition, stratification and its change with height are important. The cross-barrier density difference is a necessary condition for a variety of overflows such as foehn, bora, chinook, gap flow, and downslope windstorms. To forecast the occurrence of any of them, the potential temperature at the upstream crest/gap height and at the downstream valley floor need to be forecast. Crest-level winds are insufficient as seen in Fig. 4.

In summary, the downstream adjustment to the flow descending from the nearly two-dimensional Sierra Nevada to Owens Valley on 9 April 2006 included a complicated internal hydraulic jump involving a westerly overflow layer with moist and dry sublayers and an upvalley-flowing valley air mass. The westerly-flowing jump layer could only flow into and displace valley air in the lee of the sierras when its potential temperature had fallen to match that of the valley air mass. This cooling was clearly visible in the temperature and velocity records from the automated weather station at trailhead (Fig. 5), and the associated onset above can be seen in the visible satellite images (Fig. 8). These also show the two-dimensionality of the flow above crest height.

The two jump sublayers acted dynamically as a single layer but were thermodynamically distinct upstream by stratification and mixing ratios. They mixed internally but also entrained some air from the existing valley air mass and the waving flow aloft. Beneath the lowest jump layer an even colder gap overflow layer plunged under the existing upvalley-flowing air mass in Owens Valley. Mixing in the internal hydraulic jump can be clearly associated with shear-layer instabilities that form initially on the increased shear at the upwind jump face. These were visualized with the airborne Wyoming cloud
radar shown in Fig. 13. The Kelvin–Helmholtz instabilities first grow to finite amplitude and pair as they are advected downstream. After one pairing the stratification and shear have spread sufficiently that the flow is stable to structures at the 1-km vertical scale of the paired vortices and the structures are seen to collapse. Aloft, a stably stratified layer with a strong cross-barrier component of about 35 m s⁻¹ responded to the large-scale undulations of the potential temperature step at the top of the descending jump layer and internal hydraulic jump. The layer aloft did not respond directly to the terrain but indirectly through the jump layer.

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