High-Resolution Simulations of Lee Waves and Downslope Winds over the Sierra Nevada during T-REX IOP 6

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ABSTRACT

The downslope windstorm during intensive observation period (IOP) 6 was the most severe that was detected during the Terrain-Induced Rotor Experiment (T-REX) in Owens Valley in the Sierra Nevada of California. Cross sections of vertical motion in the form of a composite constructed from aircraft data spanning the depth of the troposphere are used to link the winds experienced at the surface to the changing structure of the mountain-wave field aloft. Detailed analysis of other observations allows the role played by a passing occluded front, associated with the rapid intensification (and subsequent cessation) of the windstorm, to be studied. High-resolution, nested modeling using the Met Office Unified Model (MetUM) is used to study qualitative aspects of the flow and the influence of the front, and this modeling suggests that accurate forecasting of the timing and position of both the front and strong mountaintop winds is crucial to capture the wave dynamics and accompanying windstorm. Meanwhile, far ahead of the front, simulated downslope winds are shallow and foehnlike, driven by the thermal contrast between the upstream and valley air mass. The study also highlights the difficulties of capturing the detailed interaction of weather systems with large and complex orography in numerical weather prediction.

1. Introduction

The Sierra Nevada range is a well-known source of strong mountain waves, downslope windstorms, and turbulence associated with lee-wave rotors, which represent hazards to aviation, residents, and property and are difficult for forecasters to predict (Holmboe and Klieforth 1957; Grubisic and Lewis 2004). Continued increase in the resolution of operational numerical weather prediction (NWP) models is expected to improve forecasts as the phenomena become more explicitly resolved. Meanwhile, research into such hazards increasingly also utilizes high-resolution NWP models (e.g., Gohm et al. 2004; Belusic et al. 2007; Agustsson and Olafsson 2007; Jiang and Doyle 2008; Chan 2009; Reinecke and Durran 2009), many of which can be also be run in idealized configurations (Schmidli et al. 2011; Doyle et al. 2011). Thus, study of how high-resolution NWP models behave over the Sierra Nevada during strong mountain-wave events is valuable more generally.

Owens Valley, in the lee of the Sierra Nevada in California, was the site of the recent Terrain-Induced Rotor Experiment (T-REX), which involved a variety of ground-based and airborne in situ and remote sensing measurements that were focused on the period March–April 2006 (Grubisic et al. 2008). Figure 1 shows the geography of the region (Owens Valley is identifiable by the high, steep, quasi-two-dimensional slope of the sierra forming its west side), some of the T-REX instrument and sounding locations, research aircraft flight tracks, and research automobile tracks in and around Owens Valley.

The comprehensive dataset available from T-REX affords an opportunity for detailed comparison with model output. Strong wave events over Owens Valley are typically characterized by strong cross-ridge synoptic-scale winds and stability close to mountaintop. These features favor lee-wave trapping and low-level breaking of waves. During the earlier, historic Sierra Wave Project (Holmboe and Klieforth 1957), the synoptic situation conducive to the strongest wave events was found to include a prefrontal situation at Owens Valley. On some occasions, the generated waves may be large enough to control the winds occurring within the valley, with extensive strong downslope windstorms and possible formation of rotors. These may be referred to as dynamically forced downslope
wind systems. A number of authors have discussed moderate downslope winds in Owens Valley that are forced by a thermal mechanism whereby the upstream air mass becomes cooler than that in the valley (typically because of daytime solar warming within the valley) and, flowing over the Sierra Nevada, induces downslope winds by undercutting the valley atmosphere. Jiang and Doyle (2008), in a study of diurnal variation of downslope winds in Owens Valley, demonstrated this effect for the case of only moderate mountaintop winds by using observations and high-resolution modeling, terming the flow “in-valley westerly.” Mayr and Armi (2010) show further evidence from T-REX for flows of this kind, likening them to a shallow Alpine foehn, and a discussion by Raab and Mayr (2008) of measurements in the valley suggests that this mechanism may also play a part in initiation of downslope winds in dynamically forced cases. Jiang and Doyle (2008) also present observed cases of stronger dynamically forced waves in which the thermal mechanism was not expected to play a part (e.g., at night). Work by Billings and Grubisic (2008b) emphasizes the importance of dynamical forcing by the upstream wind and stability profile, and the wavelength of the resulting waves, in determining the degree of penetration of westerly flow and suggests that the thermal mechanism assists or controls westerly in-valley flow when dynamic forcing is insufficient to enable complete penetration across the valley floor (Billings and Grubisic 2008b,a).

Intensive observation period (IOP) 6 contained the most intense downslope windstorm observed by the valley...
instruments during T-REX, accompanied by large-amplitude lee waves (Grubisic et al. 2008). For this study, IOP 6 has been simulated using the Met Office Unified Model (MetUM) in a nested, high-resolution configuration. The aim is to describe the observed flow and to compare the model results to gain insight into the model’s ability to capture this extreme event and similar events more generally. Times and dates will be given throughout according to local standard time (LST), which is UTC – 8 h. Section 2 describes briefly the T-REX observations used and the geography of the region. Section 3 contains a description of the model and nested-simulation setup. The observed event is discussed in section 4, and the modeled flow is compared in section 5. Conclusions are summarized in section 6.

2. Observational data

During T-REX, a comprehensive set of measurements of the flow aloft of the Sierra Nevada and in the Owens Valley was obtained. Owens Valley is very straight, with little variation in width along its length. It is over 100 km long and 15 (30) km wide measured at the base (peaks) of the bounding terrain. The sierra terrain to the west rises roughly 3000 m above the valley floor, and occurrences are well documented of large-amplitude mountain waves induced by stable flow over the mountains, associated westerly downslope windstorms sweeping across the valley, or turbulent regions due to separation of such flows within the valley. An overview and details of the instrumentation are given by Grubisic and Billings (2007) and Grubisic et al. (2008). A subset of the deployed instrumentation is used in this study, with locations shown in Fig. 1. At the surface, an array of automatic weather stations (AWS) installed by the Desert Research Institute (DRI) and University of Leeds measured wind, temperature, and pressure at 10 m at locations spaced roughly 3–5 km apart close to the town of Independence, California, as shown in Fig. 1b. The DRI data were recorded as 30-s averages, and the Leeds data were recorded as 1-min averages. During IOPs, additional measurements were taken, including operation of the University of Innsbruck Weather Station on Wheels (WOW; Mayr et al. 2002; Raab and Mayr 2008), research-aircraft flights, and radiosonde releases west of the Sierra Nevada and from within the valley. WOW was an instrumented car equipped with GPS, operating over a broader area than the main array, measuring wind and other atmospheric variables continuously as 30-s averages along the routes shown in Fig. 1a. Also shown in Fig. 1a is an example of the typical cross-valley “racetrack” pattern flown by the Facility for Airborne Atmospheric Measurement (FAAM) BAe-146 aircraft during the IOPs (6, 8, and 10) discussed herein. Two other research aircraft [the National Science Foundation/National Center for Atmospheric Research (NCAR) High-Performance Instrumented Airborne Platform for Environmental Research (HIAPER) and University of Wyoming King Air] flew similar tracks, with cross-mountain legs roughly collocated in the horizontal plane with the BAe-146 northern leg for the three aircraft, but covering different height ranges, altogether encompassing about 2000–13 000 m MSL (roughly 6000–40 000 ft). The remaining markers in Fig. 1a indicate three radiosonde release stations west of the Sierra Nevada used during IOPs to obtain the upstream profile, including the Mobile GPS Advanced Upper Air Sounding System (MGAUS) platform, situated at Visalia in the Central Valley for IOPs 6, 8, and 10. Radiosondes were released from one of these sites every 1.5–3 h during IOPs. In addition to the above, satellite cloud images from the high-resolution National Oceanic and Atmospheric Administration Geostationary Operational Environmental Satellite-10 (GOES-10) are used to find evidence of mountain-wave activity and frontal zones. Also, data from the so-called T-REX composite (abbreviated to TRC in this paper) dataset, combining data from routinely operational sites in the region, run through a common quality-control procedure and averaged to 5-min intervals, are used to supplement the main T-REX surface data over a broader area.

3. Model simulations

The numerical simulations were performed using MetUM on nested grids, the outermost of which was that of the operational global NWP configuration of the model. The horizontal grid spacing of the global model used was 0.5625° and 0.375° in the zonal and meridional directions, respectively. Forecast data from the global model were used to initialize and drive (through lateral boundary conditions) the flow in the higher-resolution inner domains using a one-way nesting technique. The inner domains had horizontal resolutions of approximately 12 km, 4 km, 1 km, and 333 m. The approximate locations of these domains are shown in Fig. 2a.

The formulation of MetUM is described by Davies et al. (2005). The simulations presented here were conducted using version 6.1 of the model. MetUM is a finite-difference model on a latitude–longitude grid and uses a semi-implicit semi-Lagrangian two-time-level dynamical core. For limited-area simulations, the latitude–longitude grid is rotated to give a quasi-uniform grid across the
domain. The 1-km- and 333-m-resolution simulations were performed using a vertical grid that consisted of 76 levels with the model upper boundary at 39.2 km MSL. The lowest model level is placed at 10 m above the ground, and the grid is stretched so that the grid spacing increases smoothly with height. At a height of 500 m, for example, the grid spacing is approximately 110 m. At 2.5 km this increases to around 220 m. The coarser-resolution (global, 12 km, and 4 km) domains used a similar but somewhat-coarser-resolution vertical grid that consisted of 38 levels and had a grid spacing of approximately 2 times that used in the finer-resolution domains. The time-step lengths employed in the global, 12-km, 4-km, 1-km, and 333-m domains respectively were 15 min, 5 min, 30 s, 10 s, and 5 s.

The configuration of the model at the finest (1 km and 333 m) resolutions differed from that at the coarser resolutions in the following ways:

1) The semi-Lagrangian advection scheme for potential temperature involved a fully three-dimensional interpolation scheme in place of the standard scheme, which is noninterpolating in the vertical direction. Tests have shown that this approach gives more-accurate representation of internal gravity wave motions. This fully interpolating scheme requires a relatively short time step to retain numerical stability.

2) In the semi-implicit time-integration scheme, values of 0.6 were used for the off-centering time weights.\(^2\)

3) The column-based parameterizations of deep and shallow convection (Gregory and Rowntree 1990) applied at the coarser resolutions were not implemented because at these finer resolutions the convective motion is largely resolved.

4) The cloud microphysical parameterizations at all model resolutions follow those described by Wilson and Ballard (1999), but, for resolutions of 12 km or finer, the parameterization accounted for the effects of advection of precipitation between grid boxes by the three-dimensional wind field.

At all resolutions, the simulations were conducted with a one-dimensional (vertical) boundary layer turbulence parameterization (the model’s 8B scheme; Lock et al. 2000). For all but the finest-resolution (333 m) model domains, the model orography was extracted from the Global Land One-Kilometer Base Elevation (GLOBE) 30 arc s-resolution dataset (Hastings and Dunbar 1999), interpolated onto the model grid. For the global, 12-km, and 4-km configurations these data were then smoothed using a sixth-order low-pass implicit tangent filter similar to that described by Raymond (1988). At 1-km resolution no smoothing was applied. For the 333-m grid, better-resolution data were obtained from the Space Shuttle Radar Topography Mission (SRTM) dataset (Farr et al. 2007), whose horizontal resolution is 3 arc s, or approximately 90 m. These data were interpolated onto the

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\(^2\) Off-centering weights of 0.5 correspond to a centered-in-time scheme and imply formal second-order accuracy; values of 1 correspond to a first-order fully implicit scheme.
model grid without any smoothing, and the resulting orography across the 333-m domain is shown in Fig. 2b.

One-way nesting was employed, with lateral boundary conditions updated every 15 min for the innermost (333 m) domain, every 30 min for the 1- and 4-km domains, and every hour for the 12-km domain. The IOP-6 simulation was initialized with Met Office global analysis fields at 0100 LST 25 March 2006. As a basic test, two simulations were performed of the weak–moderate-strength waves occurring in southwesterly synoptic flow during IOP 8 (31 March–1 April 2006) and IOP 10 (7–8 April 2006). Strong downslope windstorms did not occur during these cases. The vertical velocities in the 333-m domain for these cases were compared with those measured by the FAAM BAe-146 aircraft. The model was found to perform well, given the inherent difficulty of representatively sampling the observed and modeled three-dimensional fields along one-dimensional tracks.

Examples are shown in Fig. 3 for southern legs of the racetrack pattern shown in Fig. 1, in which the model produces wave fields that are quantitatively similar to those measured by the aircraft.

Idealized MetUM simulations at 1-km resolution were also carried out as sensitivity tests to various aspects of the upstream profile of wind and temperature (discussed later). These used dynamical and physics settings and start times that are identical to those of the nested simulations but with simulation of only the 1-km-resolution domain. Also, rather than using data from the 4-km-resolution simulation to initialize the 1-km domain and update its lateral boundaries, the initial state and lateral boundaries were set using a single horizontally homogeneous profile. Several simulations were performed, each using a different profile, to represent different synoptic conditions as discussed within section 5.

4. Observed event during IOP 6

A strong wave/rotor event was documented during IOP 6, involving a severe downslope windstorm and separation of the flow from the valley floor, resulting in a turbulent rotor region in the flow (Grubisic et al. 2008). In this section, we present a summary of the observational evidence from IOP 6.

a. Synoptic evolution

Figures 4a–d show the synoptic development that occurred during IOP 6 in the form of mean sea level pressure, wind vectors at 5 km over the western United States, and low-level temperatures from the global MetUM forecast. Although some details may differ, the forecast may reasonably be expected to reproduce the actual broad synoptic situation well on the scale shown. Figure 4a indicates a depression moving inland at 0400 LST 25 March 2006, with an approaching region of strong west-southwesterly winds, which eventually passes over Owens Valley, and an associated front off the coast of California (discernible as a sharp gradient in the shaded temperature field to the south of the low center). Note that Steenburgh et al. (2009) performed a study that
focused on the front associated with this system, which is occluded as it encounters the terrain. Because of the terrain-following nature of the model level whose temperature is depicted, high terrain appears colder in the plot and the rough location of the Sierra Nevada can be discerned from a finger of cold temperatures in the western United States. At 1000 LST (Fig. 4b), the wind has strengthened over the sierra, and the front has made landfall. By 1600 LST (Fig. 4c), around the time of the windstorm, the front is by now interacting with the mountain topography of the region and is no longer easily discernible, the low center has become a trough over the Pacific coast to the northwest, and the winds over the sierra have begun to turn more westerly. Later by 2200 LST (Fig. 4d) the system has progressed inland, replaced by a ridge approaching from the west.

The synoptic flow is further illustrated by the wind and potential temperature profiles in Fig. 5. The thick solid black line represents a radiosonde ascent from upstream of the Sierra Nevada. This profile indicates a layer of enhanced stability close to the ridgetop, where strong west-southwest winds also occur, and strong positive vertical wind shear in the troposphere. A stable layer and northerly near-surface jet below 1 km MSL reflect the passage of the front at the launch location that occurred about an hour before the time of the sounding.

b. Gravity waves

Figures 6a–d show four GOES-10 visible satellite images over the western United States between 1030 and 1730 LST 25 March 2006. Well-developed lee waves are evident downstream of the coastal mountains and Sierra Nevada.

The three research aircraft detected gravity wave motion in the variation of vertical velocity. A method frequently used for studying gravity wave motion is to plot vertical velocities for individual flight legs during a wave
event. Here, measurements from legs at different heights have instead been interpolated to give a single composite vertical valley cross section for a given period. Interpolation onto a regular grid was performed by the "GETMAT" matrix function of the "DISLIN" graphics library. This function performs a weighted average of the observations on the basis of horizontal and vertical ranges of influence and a weighting factor. A small area of influence and high weighting factor emphasize local detail. The grid was chosen to reflect the relative density of data in the vertical (30 grid points) and horizontal (1000 grid points) directions, with area of influence being two points in the vertical direction and four points in the horizontal plane, using the default weighting factor. Artifacts exist beyond the outer (i.e., top and bottom) flight legs because of extrapolation. Boundary conditions, for instance of zero vertical velocity, were not used because they would artificially reduce contour values near the surface and be unrealistic at the upper boundary. This unique synthesis, possible only because of the simultaneous deployment of the three aircraft, provides a convenient visualization of the wave field and facilitates easy comparison with model cross sections. Northern flight legs, which were collocated in the horizontal plane for all three aircraft, have been used to create the composites. Figures 7a and 7b show composites that are based on data from a part of the IOP between 0820 and 1115 LST and from a later part between 1300 and 1630 LST 25 March, respectively. Note that these 3–3.5-h intervals are sufficiently long for the waves to have evolved somewhat between the initial and final legs used for the interpolation and that the composites cannot be considered to be true instantaneous snapshots of the flow. In Fig. 7a, the uppermost (HIAPER) measurement legs took place between 0945 and 1115 LST, the BAe-146 legs and the upper King Air legs above mountain crest took place between 0820 and 1115 LST, and the lower King Air legs took place between 1000 and 1100 LST. In Fig. 7b, the uppermost (HIAPER) measurement legs took place between 1300 and 1430 LST, the BAe-146 legs above mountain crest took place between 1440 and 1630 LST, and the shorter King Air legs took place.

FIG. 5. Comparison of profiles recorded by radiosondes and aircraft with profiles through the 1-km-resolution MetUM domain at similar locations: MetUM profiles at 1500 LST 25 Mar (thin solid lines) and 1700 LST (thin dashed lines), the profile detected by the radiosonde released from the MGAUS station at 1458 LST 25 Mar (thick solid lines), and the profile detected by the BAe-146 at 1731 LST 25 Mar (thick dashed lines). The aircraft track corresponding to the 1731 LST profile lies within the short flight legs, oriented roughly along the San Joaquin Valley, in the westernmost part of the flight track shown in Fig. 1a.
between 1520 and 1600 LST, the legs being flown in descending order for all three aircraft. Despite time differences, adjacent legs flown at different times by different aircraft seem to mesh reasonably well, except perhaps the King Air and BAe-146 legs in Fig. 7b, and this result may reflect the rapid evolution of this lower portion of the wave field at this time.

The aircraft composites provide a revealing depiction of the wave field structure, in which vertical motion is weak upstream of the mountains, with large up- and downdrafts downstream associated with the wave motion. The vertical orientation of up- and downdraft regions in both Figs. 7a and 7b would normally suggest that the wave field in the mid–low troposphere is (partially) trapped. In the earlier of the two aircraft composites (Fig. 7a), the waves are characterized by a strong downdraft into the Owens Valley followed by an updraft of similar strength, also over the valley, and then alternating up- and downdrafts of decaying strength farther downstream. In the later period (Fig. 7b) the waves are much stronger and the primary downdraft is shifted significantly downstream, beyond Independence, while the wavelength of the rest of the wave train is slightly longer so that the first updraft spreads beyond the peak of the Inyo Mountains on the east side of the valley. The shift of the primary downdraft also appears to be visible in the wave clouds in Fig. 6 using the Independence airfield as a reference, with notable development in Fig. 6c at 1530 LST: the initially straight edge of the “cap” cloud above the crest becomes disturbed, with parts extending farther out from the crest. This change becomes more pronounced by 1600 LST (not shown). The King Air tracks in the lower troposphere measured the largest (and most turbulent) motions.

The distribution of wave motion in the vertical direction also differs between the two panels of Fig. 7. The depth of atmosphere sampled appears to be split into two regions from this point of view: one close to the tropopause and the other below this level and occupying the remaining depth of the troposphere. Where waves are strongest in one of these regions, they appear weaker in the other region, above a given point in the horizontal plane. This would seem to reflect the degree to which waves are “ducted” within different layers. In Fig. 7a, the strong waves close to the tropopause appear like trapped waves, although they were measured over three HIAPER legs spaced over a somewhat narrow vertical range of around 1.4 km, making it difficult to be conclusive. There is a phase disconnect between these waves and those at

![Fig. 6. Images from the visible channel of GOES-10 at (a) 1030, (b) 1200, (c) 1530, and (d) 1730 LST 25 Mar 2006. Lines of latitude and longitude, coastlines, and state boundaries are marked in white. Black and white thick crosses denote the positions of the Fresno (west of the sierra) and Independence airfields, respectively.](image-url)
the lower levels. Smith et al. (2008) and Woods and Smith (2010) undertook detailed study of the HIAPER data during T-REX IOPs using Fourier and wavelet analysis methods, building a detailed picture of wave propagation. They discovered trapped waves of relatively short wavelength at the tropopause during IOP 13 (thought to be secondary waves generated locally by a wave-steepening mechanism) coexisting with longer-wavelength upward- and downward-propagating waves within the troposphere, but similar evidence was not presented for IOP 6. Instead, Woods and Smith (2010) discuss downward-propagating waves coexisting with the primary upward-propagating wave at a height of 13 km during IOP 6. Smith et al. (2008) and Woods and Smith (2010) do not draw conclusions concerning the wave behavior lower in the troposphere during IOP 6.

Reinecke and Durran (2009) used an ensemble of high-resolution nested simulations to model this flow, finding strong sensitivity to variation of synoptic conditions. In ensemble members containing weaker waves, the primary downdraft remained confined over the upper part of the sierra downslope (similar to Fig. 7a), whereas in the strongest members the wave downdraft was found to extend farther east (in common with Fig. 7b). In the latter cases, this structure was associated with wave breaking.

c. Flow within the valley

A time series of the wind speed and direction at DRI AWS 9 is shown in Fig. 8 (solid lines). Until 1530 LST 25 March, flow was variable and episodical, on the whole being either up valley (roughly southerly) or more westerly or southwesterly (cross valley), seldom increasing above 10 m s\(^{-1}\) in strength. The wind strengthened and steadied after this time. A particularly intense westerly downslope flow subsequently commenced after 1735 LST.
peaking at 24 m s$^{-1}$ and ending abruptly at 1851 LST in a transition to a northerly down-valley flow. This intense windstorm was observed throughout the Independence AWS array, including the easternmost stations at the foot of the Inyo Mountains, peaking at close to 30 m s$^{-1}$ at some sites. A snapshot of wind vectors from the most intense part of the windstorm is shown in Fig. 9a. The windstorm was observed (on the basis of the whole AWS array) to both grow and diminish starting in the east of the valley, suggesting that a thermally driven flow from passes within the sierra barrier was not its primary driver. The windstorm is presumably connected with the change in wave structure detected by the aircraft just beforehand (Fig. 7b), suggesting a primarily dynamically forced downslope-flow case. Note that, in the ensemble of high-resolution nested simulations of this flow of Reinecke and Durran (2009), strong downslope winds were directly linked with a breaking-wave structure aloft.

Of interest is that examination of TRC stations within the Owens Valley (see Fig. 1) and the WOW instrumented car data did not reveal such exceptionally intense, and consistently westerly, flow elsewhere in the valley (not shown). Only C1028, located within the Independence array, detected winds of exceptional strength, peaking at 29 m s$^{-1}$ at 1830 LST, while winds at the other TRC stations within the valley in Fig. 1 detected nothing over 15 m s$^{-1}$. Jiang et al. (2011) also report winds peaking over 25 m s$^{-1}$ west of Owens Lake around this time. This suggests that the main windstorm may have been relatively localized to these areas.

**d. Attribution**

Analysis of synoptic details and of the timing of the observed wave evolution and windstorm affords some insight into the possible precursors of such a dramatic event. As already mentioned, strong mountaintop wind and stability are expected to be important factors in the
development of strong, dynamically forced wave cases. In addition to the 1458 LST upwind balloon sounding shown in Fig. 5, a profile measured by the BAe-146 aircraft upstream of the sierra later, at 1731 LST, is also shown. Together these cover the period leading up to the onset of the observed windstorm. The mean wind between 3000 and 5000 m MSL (roughly mountaintop height for the Sierra Nevada crest) was strong in this period, taking values of 28.55 and 22.01 m s\(^{-1}\) at 1458 and 1731 LST, respectively. Meanwhile, by the time of the later aircraft sounding, the stability had increased significantly at \(\sim 4000\) m, within the same layer. The position of the front is discernible in the satellite images in Fig. 6; for instance, in Fig. 6b, to the west of the Sierra Nevada, it is evident as the boundary between the cloudier area (mostly midlevel stratiform with some high-level cloud) covering the sierra and most of the Central Valley to the east and a clearer region containing some streets of convective cloud to the west. The front is initially oriented roughly linearly in a south-southwest–north-northeast direction and progresses to the southeast. As it approaches the sierra, it distorts, with a portion in the Central Valley progressing with less impediment than the portion to the north that encounters the mountains, so that as it approaches the sierra foothills the front becomes more meridionally oriented (Fig. 6c) and eventually, by 1730 LST (Fig. 6d), lies along the sierra foothills roughly parallel to the ridge (north-northwest–south-southeast). A manual analysis of the progress of the front, carried out by examining time series of wind and temperature at the TRC stations marked in Fig. 1c, was found to be consistent with this visual inspection [an example time series is shown in Fig. 10, in which the front is indicated by a drop in temperature of about 2–3 K accompanied by a large change (here a reversal) in wind direction and speed]. Although Steenburgh et al. (2009) focus on the stronger part of the front farther north rather than on Owens Valley, their placement of the front at a given time is in keeping with our placement of its southern extension. They discuss the complexity of the front’s evolution in relation to the terrain, with only a trough in the surface winds detected at times. For simplicity, however, we will always refer to a “front.” Farther to the north and south, the front appears to retain its original south-southwest–north-northeast orientation (also see Steenburgh et al. 2009) so that its shape traces a zigzag wrapping around the higher part of the sierra.

The evidence suggests that the progress of the front over the Sierra Nevada range (accompanied by an apparent increase in mountaintop stability, possibly associated with the front) during a period of strong mountaintop wind plays a role in the severity of the windstorm event. A further coincident factor is the afternoon heating of the valley atmosphere. If the thermal contrast between cooler air upstream of the Sierra Nevada crest and warmer valley air has any implication in activating or assisting the development of the downslope windstorm, conditions for this situation also will be optimal.

Mechanisms by which a front may influence prefrontal gravity wave activity are not fully understood, although cases of prefrontal amplification of gravity waves have been documented elsewhere (e.g., in the Colorado Rockies; Darby and Poulos 2006). Ralph et al. (1997) discuss evolution of a lee-wave field—to be specific, its wavelength—in response to changing upstream vertical profiles in the region of a front, also in the Colorado Rockies. Neiman et al. (2001) discuss gravity waves generated by the motion of a front, behaving like a “moving obstacle,” in the same range. It is interesting that precisely 2 yr before the windstorm discussed here, at 1600 LST 25 March 2004 during IOP 8 of the preceding Sierra Rotors Project (SRP; a preliminary phase of T-REX), a synoptic situation occurred that was similar to that during IOP 6 of T-REX, with wind speed at 5000 m of 25 m s\(^{-1}\), strong stability close to mountaintop, and strong vertical shear in the troposphere ahead of an occluded front (Grubisic and Billings 2007). Only a much weaker windstorm developed in 2004, however, peaking in strength around 17 m s\(^{-1}\) within the DRI AWS array. One outstanding difference during SRP IOP 8 as compared with IOP 6 of T-REX was a much later frontal passage, between 0100 and 0400 LST the next day.
5. Modeling of IOP 6

a. Reproducing synoptic factors

Simulating the conjunction of the factors implicated in generating the windstorm proves difficult. The peak in wind activity in the model is somewhat weaker than is observed and occurs too early. The mean wind between 3000 and 5000 m decreases in strength approaching the time of the observed windstorm, taking values at 1500, 1600, 1700, 1800, and 1900 LST of 23.6, 22.6, 16.4, 13.4, and 13.0 m s$^{-1}$, respectively (also see Fig. 5). The profile of potential temperature at 1700 LST from the model shown in Fig. 5 also suggests underpredicted stabilization of the profile near the mountaintop.

By comparing time series at the TRC station locations in the 4-km model domain (and the 1-km domain where possible) with those observed, the front was found to approach the sierra and move down the Central Valley to the west generally 1–2 h late in the model (see Fig. 10; a result that is also supported by sounding comparisons such as that in Fig. 5 for the MGAUS location). Over the mountain area, by the time it reaches the main part of Owens Valley the modeled front “catches up” with the observations. The valley and mountain analyses taken together indicate that the modeled front’s shape must experience less of the distortion exhibited by the actual front, mostly retaining its original south-southwest–north-northeast, cross-ridge orientation. This becomes evident in high-frequency animations of snapshots of the wind and potential temperature fields from the model, most discernible as the progress of a line of coherent disturbance in the wind vectors. Because only still images may be included here, the position of the front is indicated by arrows in Figs. 12a and 12b (described below). The front is just touching the northern tip of Owens Valley and is still aligned roughly north-northeast–south-southwest. The front appears to manifest itself in Owens Valley by the onset of strong down-valley flow.

In summary, the simulation does not capture the precise conjunction of synoptic precursors of the windstorm at Independence, with the decay in synoptic wind occurring too early and the front arriving late (and in a different alignment). In light of this finding, two useful investigations may be performed: 1) an investigation into the performance of the model at Independence, taking advantage of the high density of observations, to see to what extent the high-resolution model gives a representative picture despite synoptic inaccuracies and 2) an investigation to examine the 333-m domain at the location where synoptic precursors do coincide. Both of these have implications for the interpretation of future operational forecasting systems as resolution increases to levels that are comparable to today’s research models.

b. Simulated waves near Independence

Figures 11a and 11b show vertical cross sections of the vertical velocity field from the 1-km-resolution domain of the nested MetUM simulations. As with the measured waves, the simulated waves are characterized by a strong downdraft and updraft over the Owens Valley, reaching their strongest by 1700 LST 25 March (Fig. 11b). The deep eastward spread of the primary downdraft beyond Independence detected by the aircraft around the time of the windstorm, however, is absent, with the modeled wave structure in Fig. 11b instead continuing to resemble that measured earlier on (Fig. 7a). The waves also take later to develop (cf. the times of Figs. 11a and 7a). The waves subsequently decay, changing phase and shortening in wavelength (not shown). Examination of the vertical velocity at 5-km altitude (not shown) confirms similar behavior along the length of the ridge.

If one leaves aside timing differences, the similarity in structure of the waves and distribution of wave amplitude in Fig. 11b to the aircraft-based composite in Fig. 7a is strong in, for example, the waves on the tropopause, the weakness of the second downdraft over Owens Valley at lower levels, and the simulation of the downdraft maximum immediately downstream of the Inyo Mountains. In Fig. 11b, phase lines do not appear to be vertical, slanting both upward and downward in different parts of the section—a result that suggests consistency with the description of superposed upward- and downward-propagating waves that was given by Woods and Smith (2010) for this case.

c. Simulated winds near Independence

Model winds at DRI AWS 9 are overlaid with the measured wind in Fig. 8. Before the westerly event, the flow is in a roughly up-valley regime. The strengthening of winds in the valley after 1500 LST is apparently captured by the model (although the simulated wind direction is more steadily westerly). Later, periods of stronger westerly or southwesterly flow, centered around 1700 and 1800 LST, occur in the model but generally stay below 16 m s$^{-1}$; these periods are part of a broader, pulsing outflow from Kearsarge Pass just west of Independence, which is most well developed just after 1700 LST. Numerous such flows in the valley can be seen developing at earlier times in Figs. 12a and 12b, which depict snapshots of the near-surface flow at 1507 and 1600 LST, respectively. If one allows time for advection from the MGAUS sounding location, the above flows coincide roughly with the period of peak mountaintop winds in the model. As the modeled synoptic winds subsequently drop, the windstorm observed at DRI AWS 9 between 1700 and 1900 LST is absent in the model, the
wind there instead diminishing and becoming variable before the transition to northerly down-valley flow around 1900 LST. The strong westerly flow at 10 m in the model near Independence at 1716 LST is shown in Fig. 9b for comparison with the observed windstorm. Although mistiming and underestimating the observed windstorm in strength and seemingly lacking the link to the wave motion aloft, the model appears to be representative in a more qualitative sense, with areas of slacker or reversed flow in both the model and the AWS winds at the edge of the flow and where it terminates in the east of the valley. From the perspective of future forecasting systems, it is encouraging that some useful guidance as to the occurrence of strong westerly flow within the valley at Independence (and strong wave motion aloft), at close to the right time, would be gained from the model. A similarly high-resolution nested simulation of IOP 6 over Owens Lake by Jiang et al. (2011) also significantly underpredicts downslope wind strength. They find, however, that the predicted winds are essentially sufficient for the practical purpose of predicting aerosol lofting.

To investigate the mechanism for the downslope flow in the model, profiles of the modeled potential temperature at Independence and above the ridge crest directly upwind (locations shown in Fig. 1) were plotted for different stages of the flow, as shown in Fig. 13. Before the period of sustained westerly and southwesterly flow in the model, Fig. 13a shows little difference between the two profiles. The onset of the flow occurs as the valley profile warms and becomes less stable in the afternoon, as shown in Fig. 13b (the crest profile also cools during this time). During the period of most intense westerly flow in the model, the base of the crest profile is now lower in potential temperature than is almost the entire depth of the valley atmosphere (Fig. 13c). The westerly flow is shallow and is accompanied by an area of reversed flow in the upper part of the valley atmosphere (not shown). It is clear that when the potential temperature...
at the crest becomes lower than that at some depth within the valley atmosphere the conditions favor penetration of the west-southwesterly flow aloft farther into the valley, from buoyancy considerations. As the flow subsequently dies, the temperature contrast also diminishes (Fig. 13d), now because of cooling and stabilization of the valley atmosphere in the evening. The above behavior mirrors the diurnal cycle discussed by Jiang and Doyle (2008). To make the connection between the ridge–valley air potential temperatures and development of the “in-valley westerly” flow more obvious, we define the depth of penetration of westerly flow into the valley by two measures: $h_{le}$, the height of the ground surface at the leading (eastern) edge of the modeled windstorm, and $x_{pen}$, the eastward extent of the windstorm at its leading edge with respect to the base of Kearsarge Pass, just west of Independence. Both of these measures are determined within the quadrilateral depicted in Fig. 9b, with $x_{pen}$ defined by the farthest point east where the zonal wind component exceeds 8 m s$^{-1}$. These are plotted as time series in Fig. 14a in addition to $h_{eq}$, the height below which the valley air (dotted lines in Fig. 13) is lower in potential temperature than that at the base of the ridgetop profile (dashed lines in Fig. 13). The decrease of $h_{eq}$ as the valley atmosphere warms (and the ridgetop profile cools) is coupled closely with $x_{pen}$. Meanwhile, $h_{le}$ decreases faster than might be expected from the decrease of $h_{eq}$. These are unlikely to follow each other exactly, however, since the momentum of the downslope wind will cause overshoot at the point of buoyant equilibrium, and mixing between the downslope flow and the valley atmosphere as a result of strong shear between them may also assist penetration by reducing the temperature gradient that inhibits descent.

The profile measured by a radiosonde launched from Independence at 1456 LST 25 March has been added to Fig. 13b. Assuming that the model represents the ridgetop profile sufficiently well for comparison (the upstream MGAUS profile is fairly well replicated at this time; see Fig. 5), it seems that at 1500 LST the profiles favor penetration of winds into the valley atmosphere as in the model, and this may explain why the model reproduces the initial strengthening of wind shown in Fig. 8 just after this time. This hints at the possibility of the thermal mechanism assisting within the development of the observed windstorm.

In light of the differences between the modeled and observed waves and downslope winds, and their attribution to synoptic factors, some idealized tests were performed (as described in section 3) to demonstrate the sensitivity of the flow to the inaccuracies in the conditions upwind of Independence that are fed through from the
driving simulation. Simulations were run for 36 h to allow
the wave field to develop, with the representative time
taken as when the strongest waves occur over Indepen-
dence. A control simulation was performed, driven using
the profile above the grid point closest to the MGAUS
release site from the 1-km domain of the nested simulation
at 1500 LST 25 March, shown in Fig. 11c. This produced
a primary downdraft similar to that in the nested simulation
shown in Fig. 11a, although with a broader subsequent
updraft over Owens Valley, spreading as far as the Inyo
crest. Because of this last difference we will focus on the
sensitivity of the primary downdraft (which in any case is
the principal feature affecting the flow in the valley) and
more generally on the wavelength and amplitude of the
waves. The results of three tests are shown in Figs. 11d–f:
1) replacing the winds in the control simulation below
roughly 7 km with those from the 1458 LST MGAUS
profile and using the MGAUS moisture profile (Fig. 11d),
2) additionally incorporating the temperature profile measured by the BAe-146 aircraft at 1730 LST
(Fig. 11e), and 3) repeating test 1 but with the radiation
parameterization disabled (Fig. 11f). This allows one to
assess the impact of 1) the underprediction of synoptic-
scale winds (and wind shear) at mountaintop level by the
model, 2) the underprediction of increased stability at
mountaintop level in the model (stability that we specu-
late may be related to the approaching front), and 3) af-
fternoon heating and mixing of the valley atmosphere at
the time of the windstorm. For the two alterations of the
driving profile it was possible to match the observational
data seamlessly into the model profile by appropriate
choice of the precise heights within which to insert the
data (see Fig. 5).

Figure 11d indicates that inserting the observed winds
below 7 km results in stronger waves and a more pene-
trative downdraft over the valley; the wavelength of the

![Fig. 13. Profiles of modeled potential temperature directly above Independence (dotted lines) and above the ridge
crest at the location shown in Fig. 1a directly upwind of Independence (dashed lines) at (a) 1200, (b) 1500, (c) 1700,
and (d) 1830 LST 25 Mar 2006. Profiles measured by radiosondes released from the MGAUS station (solid line) at
1458 LST and Independence airport (dot–dashed line) at 1456 LST have also been included in (b).]
waves is also slightly longer than in the control simulation, with the subsequent updraft spreading beyond the Inyo ridgetop. In Fig. 11c, adding the stronger stable layer from the BAe-146 profile run has no significant further impact on the amplitude of the waves over Owens Valley, but again penetration of the downdraft into the valley is greater, and this time with the strongest portion of the downdraft spreading deeper downslope. Figure 11f shows the importance of daytime heating of the valley atmosphere: without it, a stable boundary layer is present, limiting the downdraft to a relatively shallow penetration into the valley, resulting in much weaker waves, and sheltering of the valley bottom from strong winds. While the precise effect of the observed front, and perhaps cyclone structure too, is unlikely to be replicated by tests 1 and 2 since they employ horizontally homogeneous conditions, the above idealized tests demonstrate sensitivity to changes in upstream conditions associated with these synoptic factors, which is consistent with the differences seen between the observations and the nested simulation. Furthermore, the importance of the valley temperature structure with regard to the response of the wave field to the underlying topography, whose double-ridge shape has been shown to be of importance to wave structure at this location (Grubisic and Stiperski 2009; Stiperski and Grubisic 2011). To probe the importance of nonlinear processes in this case, however, a further idealized test was performed in which the amplitude of the mountains was reduced by scaling down height variations below the mountain crest by a factor of 10 (effectively “filling in” valleys but retaining the spectrum of horizontal orographic scales) to reduce nonlinear effects, with driving using the control profile. This was found to produce waves with about 1/10 of the original amplitude and to increase the wavelength of the waves, spreading the downdraft farther over the valley (not shown), which is more consistent with the observed waves (although note that to assume that this is a pure test of excluding nonlinear processes assumes that other factors such as vertical scales, e.g., the depth of the valley atmosphere, are unimportant). Also, given the possible implication of wave breaking in the rapid evolution of the wave field [as highlighted here and by Reinecke and Durran (2009)], it seems equally likely that nonlinear processes are crucial.

d. Simulated flow near the front

It seems likely that the differences between the modeled and observed wave and windstorm evolution, and the mechanisms behind them, are a manifestation of the shortcomings of the model’s prediction of the conjunction of synoptic precursors highlighted earlier—namely, the peak in synoptic winds and the approach of the front toward Owens Valley. To further this argument, the flow just in advance of the front is inspected at an earlier time,
during the period of strong mountaintop wind. The propagation of the front along the sierra ridge in the 333-m domain is accompanied by brief (1–2 h) strong downslope flows breaking out from the ridge just ahead of the front. An example can be seen for instance in Figs. 12a and 12b along the straight thick white line in the north of the domain, in an area lying just north of Bishop, California (roughly 80 km north of Independence). The flows at 10 m are stronger than that simulated at Independence and are qualitatively more similar to the observed Independence windstorm. Cross sections of the flow along the thick line shown in Fig. 12 have been studied for comparison with the modeled and observed flows at Independence. Figures 15a and 15c depict the vertical velocity and zonal wind along the above cross section at 1430 LST, and Figs. 15b and 15d depict the same at 1530 LST. The valley here is wider (and the projection of the cross sections enhances the appearance of this), with some significant terrain between the Sierra Nevada and the White Mountains to the east, but otherwise similar to the Independence area of Owens Valley in terms of the height and steepness of the two barriers. Figure 15a shows a typical lee-wave pattern of consecutive up- and downdrafts, with two downdrafts and an updraft over the valley, and another at the White Mountains, akin to the observed arrangement in Fig. 7a. The wave structure an hour later in Fig. 15b is spread eastward, with a broader downdraft and the first updraft displaced far to the east toward the downstream orography, with a less organized structure; the stronger part of the updraft over the White Mountain upslope is confined below 6 km. This occurs just as the windstorm sweeps out across the lower ground east of the main sierra ridge toward the White Mountains (see Fig. 12b), and strongly resembles the change seen in the later aircraft composite Fig. 7b at the time of the observed windstorm at Independence. The vertical cross
sections of zonal velocity in Figs. 15c and 15d depict how the strong cross-valley flow sweeps rapidly into the valley, displacing the up-valley flow (negative zonal velocity) that existed before 1430 LST to its eastern edge. This cross-valley flow fills the depth of the valley, unlike that simulated at Independence. The horizontal wavelength of the waves subsequently rapidly contracts as the synoptic wind strength diminishes (not shown).

To compare the mechanism for the modeled flow north of Bishop with that simulated at Independence, profiles at the ridgetop and in the middle of the valley in the plane shown in Fig. 15 were compared as in Fig. 13 (not shown). Unlike in Fig. 13, there was little contrast in potential temperature at a given height (typically no more than 1–2 K). A figure analogous to Fig. 14a is shown in Fig. 14b for this northern part of the ridge. The orography present within the lower ground downstream of the Sierra Nevada here complicates the determination of the penetration depth of the downslope flow into the valley, since wind speeds are elevated over the high point around x = 33 km in Fig. 15. Therefore, the “sweeping out” of the valley by the downslope flow, quantified by $x_{\text{pen}}$ in Fig. 14a, is instead represented by the number of model grid points $n_{12}$ at which wind speed at 10 m is greater than 12 m $\text{s}^{-1}$, within an area analogous to the quadrilateral shown in Fig. 9. Meanwhile, $h_{\text{lw}}$ represents the surface height at the lowest point within the above area where wind speed at 10 m is greater than 12 m $\text{s}^{-1}$. Plotted in Fig. 14b, $n_{12}$ and $h_{\text{lw}}$ appear to be similar to $x_{\text{pen}}$ and $h_{\text{le}}$ for Independence in Fig. 14a. Meanwhile, the similarity of the profiles above ridgetop and valley bottom in this part of ridge mean that the thermal mechanism does not have the same influence here; $h_{\text{eq}}$ probes some lower levels only during the onset of the downslope flow. Possibly the mechanism assists in the initiation of the flow.

Both the evolution of the modeled lee waves and downslope flow north of Bishop qualitatively resemble those measured in the Independence area during the windstorm, much more than the flow simulated close to Independence itself. This supports the argument that the peak in synoptic wind and the approach of the front are the precursors of the event observed at Independence and that the more bland behavior in the model at Independence is due to the model not capturing these features with the appropriate timing there. It is clear that what would be judged, on a larger scale, to be fairly minor synoptic inaccuracies may result in significant errors in the downslope winds predicted at a given location and time when modeling complex-terrain flow at high resolution. It is encouraging, however, that a flow similar to that observed at Independence appears to be simulated within the model domain around the same time where the appropriate synoptic factors coincide, suggesting that 1) a forecast system that is based on a model such as this one would be able, given an accurate synoptic forecast, to produce a realistic and representative downslope windstorm at Independence and 2) even given synoptic inaccuracies, the forecast produced would give some warning of severe windstorms along the ridge. Also, the association with the front would mean that surface observations indicating the front’s progress could be used by forecasters to modify predictions of where and when the severest winds would occur.

6. Conclusions

High-resolution simulations were compared with observational evidence for the downslope windstorm detected at Independence during IOP 6 of T-REX. Aircraft measurements reveal that the windstorm is associated with large change of phase, wavelength, and amplitude in the wave system aloft through the depth of the troposphere, coinciding with the arrival of a frontal system over the sierra during the period of strongest synoptic winds. Synoptic winds peak earlier in the model, at which point the front (which lags the observed front) lies a large distance north of Independence. Downslope winds are simulated at Independence, but they are weaker and shallow foehnlike winds, driven by thermal contrast between upstream and valley air, and the observed changes of wave structure are also absent there. Meanwhile, to the north, windstorms emanate from the ridge across the valley floor just ahead of the front, linked to a rapid evolution of the wave structure throughout the depth of the troposphere, in common with the flow observed at Independence. It is possible, though, that thermal contrast between the valley and the ridgetop is still a factor involved in initiating or assisting the development of such stronger, dynamically forced cases of downslope winds that occur during the day, such as IOP 6. It is encouraging from the point of view of future NWP forecasting systems that the model correctly predicts strong waves and downslope winds over Independence while apparently being sensitive to the mechanisms behind the intense IOP-6 windstorm where the appropriate synoptic conditions are in play. Nevertheless, accurate prediction of the progress of weather systems and fronts through large, complex orography such as that of the western United States is challenging even for numerical models with very high resolution. Further research is therefore desirable to understand better the interaction of weather systems and fronts with significant orography. Meanwhile, research investigating the interaction of idealized baroclinic systems with both idealized and realistic terrain would be valuable in establishing the connection
between the particular synoptic features highlighted here and the intensification of mountain waves, rotors, and windstorms.

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