Climatology of Westerly Wind Events in the Lee of the Sierra Nevada

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ABSTRACT

A 5-yr climatology of westerly wind events in Owens Valley, California, is derived from data measured by a mesoscale network of 16 automatic weather stations. Thermally driven up- and down-valley flows are found to account for a large part of the diurnal wind variability in this approximately north–south-oriented deep U-shaped valley. High-wind speed events at the western side of the valley deviate from this basic pattern by showing a higher percentage of westerly winds. In general, strong westerly winds in Owens Valley tend to be more persistent and to display higher sustained speeds than strong winds from other quadrants. The highest frequency of strong winds at the valley floor is found in the afternoon hours from April to September, pointing to thermal forcing as a plausible controlling mechanism. However, the most intense westerly wind events (westerly windstorms) can happen at any time of the day throughout the year. The temperature and humidity variations caused by westerly windstorms depend on the properties of the approaching air masses. In some cases, the windstorms lead to overall warming and drying of the valley atmosphere, similar to foehn or chinook intrusions. The key dynamical driver of westerly windstorms in Owens Valley is conjectured to be the downward penetration of momentum associated with mountain waves produced by the Sierra Nevada ridgeline to the west of the valley.

1. Introduction

Owens Valley is a narrow valley in eastern California, approximately north–south oriented and bounded by the highest portion of the Sierra Nevada (high sierra) to the west and by the White–Inyo Range to the east. Within such a valley, one expects different types of terrain-induced circulations (Whiteman 2000) to occur. Dynamically driven winds [e.g., topographically channeled flow and intense downslope winds related to mountain waves] are the result of the local modification of large-scale airflow by the underlying terrain (Jackson et al. 2013). Thermally driven winds are baroclinic circulations caused by diurnally reversing thermal gradients at distinct spatial scales, for instance, an individual slope or the whole length of a valley (Zardi and Whiteman 2013).

Strong dynamically driven winds, in particular, are a prominent feature of Owens Valley’s climate. In fact, there is an abundance of anecdotal evidence for the occurrence of downslope windstorms on the eastern slope of the Sierra Nevada. Indeed, Owens Valley has been the theater of two major research efforts and several field campaigns. These include the Sierra Wave Project and the Jet Stream Project in the 1950s (Holmboe and Klieforth 1954, 1957; Grubišić and Lewis 2004) and more recently the 2004 Sierra Rotors Project (SRP; Grubišić and Billings 2008b) and the 2006 Terrain-Induced Rotor Experiment (T-REX; Grubišić et al. 2008). The two latter projects focused on the characterization of atmospheric rotors, which may form when strong, dynamically forced downslope flow...
separates from the lee slope of the Sierra Nevada, that is, the western sidewall of Owens Valley.

The existing climatological studies of high-wind events in this area reveal hardly any signature of westerly winds. Strong winds are instead reported to blow preferentially from the north and south, reflecting the main orientation of the valley axis and most likely being a consequence of synoptic flow channeling (Zhong et al. 2008b). Nevertheless, it is a reasonable expectation that mountain-wave activity generated by westerly flow over the Sierra Nevada can lead to various degrees of penetration of momentum to the valley floor on the lee side (Jiang and Doyle 2008; Strauss et al. 2016).

Dynamical and thermal forcing mechanisms are known to cause downslope winds on the lower mountain slopes (or the valley bottom or leeside plains) at many locations around the world. The connection between large-amplitude mountain waves and strong downslope winds is well documented east of the Rocky Mountains in Colorado (Lilly 1978) and in Wyoming (French et al. 2015; Grubišić et al. 2015), as well as along the eastern Adriatic coast of the Mediterranean Sea (Smith 1987) and over the ice shelf east of the Antarctic Peninsula (Elvidge et al. 2015). In fact, flow acceleration beneath a large-amplitude mountain wave is the most widely accepted explanation for the onset of strong and potentially damaging downslope winds, including foehn, chinook, and bora (Smith 1985; Jackson et al. 2013).

However, the extent to which dynamically driven winds affect a topographically sheltered area is determined, more often than not, by local thermal forcing. For instance, McGowan et al. (2002) extensively discussed the role of convectively driven mixing in favoring the propagation of the foehn front down the Lake Tekapo valley in the Southern Alps of New Zealand. Also, two possible modes of interaction of chinook winds with leeside cold-air pools (viz., turbulent erosion of the cold air mass or flow over it, possibly with an undulating or nonstationary airmass boundary) have been documented along the eastern slope of the Rocky Mountains by Glenn (1961) and Beran (1967).

In this study, we show that similar processes occur relatively frequently in Owens Valley. We present a wind climatology, constructed using data from a mesonet of 16 automatic weather stations installed by the Desert Research Institute (DRI) for the SRP in 2004. As opposed to previous climatologies, which considered only measurements from stations located along Owens Valley’s main axis (e.g., Zhong et al. 2008b), we analyze data from stations distributed along several cross-valley transects, reaching a considerable distance up the valley’s western slope. The study complements an existing satellite-based climatology of Sierra Nevada mountain lee waves (Grubišić and Billings 2008a) and provides climatological context for several observational and numerical case studies that have emerged from the SRP and T-REX experiments (e.g., Grubišić and Billings 2007; Sheridan and Vosper 2012; Strauss et al. 2016). The material presented herein builds on a previous study by Grubišić et al. (2014).

2. instrumentation and data

The DRI long-term network of 16 automatic weather stations (AWSs) with telemetry was part of the core instrumentation deployed in the SRP and in T-REX. The network is located south of Independence, California, in the central portion of Owens Valley (Fig. 1) and provided data from the end of February 2004 to October 2013. The 16 AWSs of the DRI network are arranged in three roughly parallel rows oriented in the cross-valley direction. The station altitudes range from 1109 m MSL (station 16, valley bottom) to 1710 m MSL (station 1, eastern slope of the Sierra Nevada). This 600-m altitude range corresponds approximately to the lowest 20% of Owens Valley’s depth (see section 3 for further information). The area covered by the network is relatively small (approximately 150 km²), but the density of measurements is very high (on the order of 1 station every 10 km², the average separation between individual stations being approximately 3 km). As a term of comparison, a recent case study of Santa Ana winds in Southern California (Cao and Fovell 2016) used data from a mesonet with 140 stations covering an area of 11 700 km² (approximately 1 station every 80 km²).

Each station of the DRI network consists of a standard 10-m tower and of sensors for wind, temperature, relative humidity, and pressure (Fig. 2). To minimize errors in static pressure detection during high-wind speed events, sensors are equipped with static pressure heads (quad-disc probes; Nishiyama and Bedard 1991). Sensors are sampled every 3 s, and the data are temporally averaged over 30-s nonoverlapping intervals, saved on the stations’ dataloggers, and transmitted to DRI’s central repository through 900-MHz spread-spectrum wireless radios.

In this study, we use 5 yr of 30-s data from the time interval between 29 February 2004 and 28 February 2009. Data availability during this period is generally very good, but it deteriorates in 2009–13. Before 2009, the fraction of missing data remains below 5% for all stations except for station 13 (approximately 18% of missing data). The analyses presented in the following sections show either 10-min average (Figs. 3–5 and 8–15) or 1-h average (Figs. 6 and 7; Table 1) measurements of weather parameters.
3. Flow regimes

a. Valley geometry

Circulation patterns captured by the DRI station network are to a large degree determined by the valley geometry. Owens Valley is about 150 km long and 15–30 km wide and oriented approximately north (higher elevation) to south (lower elevation). The orientation of the valley axis is 15° west of north (Fig. 1). In the cross-axis direction, the valley has a wide U-shaped profile. The valley bottom (~15 km wide) has a flat central part that connects to a gentle alluvial slope (~5%) on the western side. The eastern slopes of the Sierra Nevada and the western Inyo slopes (both close to 30% steepness) rise steeply above the alluvial slope to the west and almost directly from the flat bottom on the east side of the valley. The ridge-to-ridge width of the valley is nearly double its width at the bottom.

The Sierra Nevada ridgeline reaches heights between 3600 and 4400 m MSL near Owens Valley, while the crest of the White–Inyo Range generally lies between 2300 and 4200 m MSL. Hence, the valley depth is approximately 3000 m. Heterogeneity of the Sierra Nevada terrain in the along-valley direction is small across the area of the network but is more pronounced within distances of less than 5 km to the north and 10 km to the south of the network location. For instance, the major incision of Kearsarge Pass (at 3600 m MSL in the main Sierra Nevada ridgeline) lies to the west of Independence, while several peaks exceeding 4000 m MSL, culminating in Mount Whitney at 4418 m MSL, rise to the southwest.

b. Diurnal circulation patterns

Thermally driven winds have a well-defined diurnal periodicity and therefore appear prominently in climatological analyses covering mountainous regions. Owens Valley is no exception: its two primary thermally driven circulation patterns are up-/downslope flow and up-/down-valley flow.

Figure 3 represents the typical diurnal variability of wind directions for wind speed $U > 1 \text{ ms}^{-1}$ at four selected stations of the network. The connection between the wind variability and the diurnal cycle should be interpreted considering that the astronomical sunrise and sunset times in Owens Valley range between 0530 and 0800 LST and 1740 and 2010 LST, respectively, and that the prominent topography reduces day length by induced shading.

Station 1 (Fig. 3a) is representative of five stations that lie on the alluvial slope (group 1: stations 1, 2, 7, 8, and 13; see Fig. 1, blue dots). During the night (from 1900 to 0700 LST), almost unidirectional southwest (240°–340°) downslope katabatic winds are clearly evident. No clear directional preference, however, is detectable during the day (from 0700 to 1900 LST). Rather, winds gradually turn from the east-northeast direction (60°–80°, upslope) during the morning hours to southeast (130°–150°, up valley) during the afternoon. In the early evening, the wind abruptly turns again to the downslope direction.

A typical pattern of up-valley flow during the day and down-valley flow during the night is most clearly evidenced at five stations located in the central flats of the valley (group 3: stations 5, 11, 12, 15, and 16; green dots in Fig. 1). Station 12, which is prototypical for this group (Fig. 3b), displays a predominance of south-southeast winds (~160°–180°) during the day and of north-northwest winds (~320°–340°) at night. The frequency of up-valley winds has a late afternoon maximum between 1600 and 1800 LST, and it is close to zero at night. In contrast, winds from the 320° to 340° range are observed even during daytime. While the larger frequency of down-valley winds at night is explained by
thermal forcing, daytime events are likely an evidence of the occurrence of channeled down-valley flow on strong gradient days. These winds can occur in principle at any time of day, so their hourly distribution is approximately uniform.

One additional cluster of data points is visible both at station 1 and at station 12, namely late afternoon (1700–1900 LST) westerly (270°) winds.

Stations on the border between the central flats and the alluvial slope (group 2: stations 3, 4, 9, 10, and 14; red dots in Fig. 1) are characterized by a transitional, hybrid pattern. Data from station 9 are represented in Fig. 3c. While the flow still blows preferentially up valley in the afternoon, large variability is found in the evening transition period and early at night (1700–2300 LST). Persistent nocturnal downslope (westerly) winds gradually turn down valley in the early morning. Presumably, this happens as the stable layer of cold air that flows down valley at night grows sufficiently thick to reach the altitude of group 2 stations under the influence of nocturnal cooling.

The circulation pattern at station 6 (Fig. 3d) resembles that of group 3. However, some relevant differences exist at night. The primary nocturnal wind direction is slightly northeasterly at station 6, as opposed to northwesterly at stations 9 (group 2) and 12 (group 3). In addition, nocturnal winds from the southeasterly quadrant are almost as frequent as northerly winds at station 6 (in particular between 0000 and 0800 LST), while they are rare at station 12 and absent at station 9. The larger nocturnal variability of wind directions is likely related to the position of station 6, a little up the Inyo Mountains slope and close to the exit of a tributary valley. The higher scatter of nighttime wind directions at this site, relative to the valley-bottom stations (group 3), has prompted us to place station 6 in a separate category.

The average strength of the thermal circulation in the valley depends on the station location and on the branch of the thermally driven flow (daytime vs nighttime). For example, the average speed of the nighttime downslope flow at stations in group 1 is about 4 m s⁻¹, except for station 7, where it is slightly below 3 m s⁻¹, reflecting the microscale characteristics of that location. The stations in group 2 show similar wind speed characteristics to those in group 1. For stations in group 3, where the up-valley–down-valley flow pattern is strongest, the difference in the strength of daytime and nighttime flows is more pronounced, with weaker nighttime winds (down valley ~3 m s⁻¹) and stronger daytime winds [up valley (5–7) m s⁻¹].

In Fig. 4, we illustrate some of this variability with histograms referred to only $U > 7$ m s⁻¹ at the same four stations shown in Fig. 3. The occurrence of strong up-valley flows in the afternoon is apparent at all stations. Strong down-valley flows, which can occur at any time of the day, especially at group 2 and group 3 stations, are most likely evidence of flow channeling under a variety of synoptic conditions (Whitman and Doran 1993; Kossmann and Sturman 2003; Zhong et al. 2008b).

Finally, a distinct cluster of westerly winds in the late afternoon is clearly visible in all panels of Fig. 4, in particular for stations in groups 1 and 2. At station 1 (group 1), these winds blow from a slightly different direction (270°) than nocturnal katabatic currents (240°; Fig. 3a). Actually, the signal of katabatic flow, which is dominant for $U > 1$ m s⁻¹ (Fig. 3a), entirely disappears for $U > 7$ m s⁻¹ (Fig. 4a). Therefore, the 7 m s⁻¹ wind speed threshold effectively separates two different classes of downslope flows: relatively weak nocturnal katabatic flow on one side and relatively strong downslope flow, more frequent in the late afternoon, on the other.

The focus of this study is the latter class of westerly winds, which is examined in detail in the next section.
c. Westerly wind events and westerly windstorms

From the preceding analysis, it is clear that westerly winds represent a distinct class of events at all stations of the network. Our definition of a westerly wind event (WWE) is determined by criteria of directionality (from the western direction), strength (strong enough to erode any valley inversion and protrude into the valley atmosphere), and spatial coherence (simultaneous detection at more than one measurement point). Accordingly, we define a WWE by 1-h-average wind direction $\theta > 255^\circ \pm 45^\circ$ and wind speed $>7\text{ m s}^{-1}$, simultaneously occurring at any two of stations 1, 2, and 7 of the network (all in group 1; at stations in groups 2 and 3, winds exceeding $7\text{ m s}^{-1}$ are very rarely westerly). The “westerly” quadrant is centered on $255^\circ$ to account for the offset of the valley axis direction from the north. The $7\text{ m s}^{-1}$ wind speed threshold has been used in previous studies to characterize high-wind events in the southwestern United States [e.g., Zhong et al. (2008a) and Smith et al. (2014), both referring to Washoe Valley in Nevada]. In the context of Owens Valley, this threshold defines winds strong enough to generate dust storms by picking up soil particles from the Owens Lake bed (Zhong et al. 2008b, and references therein). As shown in the previous section, it is also an effective filter for katabatic winds on the Sierra Nevada lee slope, which typically have lower wind speeds.

In Fig. 5, we provide an illustration of different periods of westerly winds and a detected WWE (the T-REX intensive observation period 6, described in further detail in section 4). The detected 2-h-long WWE starts at hour 6, with a pulse of strong westerly momentum reaching the upper parts of the alluvial fan. Strong winds are recorded at stations 1, 2, and 7 and last until hour 8. In two other periods, sustained westerly winds do not satisfy all three of our criteria. First, the period between hour 8 and hour 12 is characterized by wind speeds exceeding $7\text{ m s}^{-1}$ but also by large
variability of the wind direction among the stations. Second, the period between hours 30 and 42 features persistent and spatially coherent westerly winds, which are caused by nocturnal katabatic flow with rather low wind speed ($<5 \text{ m s}^{-1}$). Consequently, these two periods are not regarded as WWEs.

The frequency distribution of wind events with $U > 7 \text{ m s}^{-1}$ from all quadrants (the north, east, and south}
ones being centered, respectively, at $345^\circ \pm 45^\circ$, $75^\circ \pm 45^\circ$, and $165^\circ \pm 45^\circ$) as a function of their duration in hours is reported in Fig. 6. Apparently, winds with $U > 7 \text{ m s}^{-1}$ may occur from any direction but rather infrequently from east. In most cases, these events have a duration limited to one or two hours; however, one-third of the detected events lasted for three hours or longer. There is a clear tendency for WWEs to dominate the sample for the longest durations, although the low number of counts for events longer than 10 h does not allow us to draw firm conclusions. In the period under consideration, a few exceptionally long-lasting strong wind events (duration $\geq 15$ h) were observed, predominantly from the west quadrant.

Figure 7 displays the frequency distribution of strong wind events as a function of their mean (throughout the event) and maximum (1-h average) wind speeds at station 2 (stations 1 and 7 provide very similar results). Events with a mean wind speed of 7–8 m s$^{-1}$, corresponding to a moderate breeze, are predominantly southerly and most likely coincide with up-valley flows. The strongest events, that is, gale force winds with sustained wind speed $>15$ m s$^{-1}$ or maximum 1-h-averaged wind speeds up to 25 m s$^{-1}$, come exclusively from the west quadrant. The highest recorded hourly mean wind speeds in Owens Valley are comparable to those observed during severe windstorms along Colorado’s Front Range although generally lower and observed less frequently. In Owens Valley, the hourly mean 10-m wind speed during westerly windstorms equaled or exceeded 20 m s$^{-1}$ on 10 occasions in 2004–09 (Fig. 7). In comparison, Brinkmann (1974) reports observations of 20 downslope windstorms with maximum hourly wind speeds of order 20 m s$^{-1}$ (measured at 3.4 m AGL) during the three winter seasons from 1968 to 1971 in Boulder, Colorado.

Figure 7 also includes a diagram for the frequency distribution of wind events (with duration greater than 2 h) with respect to the maximum gust intensity. The great majority of events with gusts $> 25$ m s$^{-1}$ (up to 38 m s$^{-1}$) at station 2 corresponds to westerly wind cases. Besides causing more intense gusts, strong winds from the westerly quadrant also tend to have a higher gustiness factor than winds from other directions. For a given 10-min period with 30-s wind measurements, the gustiness factor is defined as $g_f = \frac{\max(u') - \min(u')}{U}$, where $U$ is the average wind speed and $u'$ is the perturbation with respect to the average. Values of $g_f$ for stations 1 (group 1) and 9 (group 2) are stratified according to wind speed and direction in Fig. 8. In accordance with the previous analyses, only wind speeds greater than 7 m s$^{-1}$ are considered. At station 1 (Fig. 8a), the enhanced gustiness of downslope westerly winds (for which $g_f$ occasionally reaches 1.3–1.4) in comparison with down-valley and up-valley flow is apparent. Farther down the slope, at group 2 stations (Fig. 8b, but similarly at all other mesonet stations at the valley bottom), $g_f$ decreases markedly for all directions (it never exceeds 0.6–0.7), and no enhancement is visible for westerly winds.

The distribution of WWEs in Owens Valley throughout the day and through the year (Fig. 9a) shows that they can occur in any season but most frequently in the period from April through September. In cold winter months (DJF), the frequency of these events is much
lower than at other times of the year. WWE during summer months (JJA) occur exclusively in the afternoon hours. In contrast, the very few ones occurring in the other months can extend late into the evening and throughout the night.

This pattern of diurnal and seasonal variability is not specific to westerly winds. Figure 9b shows that, at stations 1, 2, and 7, winds with $U > 7\text{ m s}^{-1}$ from all other directions also have relatively high frequency in the afternoon hours and in the summer months (the wind events in Fig. 9b are defined in the same way as WWE but dropping the directional criterion and excluding WWE). Interestingly, the afternoon frequency peak occurs between 1500 and 2000 LST for WWE, while it is shifted to earlier hours (1300–1800 LST) for strong winds from other directions. The ratio between WWE (3% of the dataset) and all strong wind events (6.9% of the dataset) amounts to 43% at stations 1, 2, and 7. This means that WWE events represent a large fraction of the strong wind events in that part of the DRI network. At stations closer to the valley axis (e.g., group 3 stations represented in Figs. 9c,d), the westerly fraction of strong wind events decreases to <10% (19% of the dataset displays strong winds in at least one station in group 3, but only 1.5% corresponds to westerly wind; therefore, the ratio decreases from 43% to 8%). However, the features of the diurnal and seasonal distribution of the strong wind events (i.e., the afternoon and summertime maxima) remain essentially unchanged.

Table 1 lists the westerly windstorms (WWSs) that occurred in the 2004–09 period. WWSs represent the strongest WWEs and are defined as periods with either sustained $U > 15\text{ m s}^{-1}$ (throughout the period) or maximum hourly $U > 20\text{ m s}^{-1}$. That is, WWSs correspond to the WWEs falling in the rightmost bins in Figs. 7a,b. In contrast to wind events with $U > 7\text{ m s}^{-1}$, windstorms are exclusively westerly, as discussed above. Also, they do not have any distinct tendency to occur preferentially in the warm season, as shown in Table 1: less than half of the listed events (5 out of 12) occurred between April and September.

A plausible explanation of these findings is that the primary forcing for WWSs in Owens Valley is partially different from that of other WWEs. Relatively strong winds ($U > 7\text{ m s}^{-1}$, including WWE) are known to
depend on synoptic flow channeling in Owens Valley, whatever their direction (Zhong et al. 2008b). Their seasonal and diurnal periodicity suggests that the diurnal and seasonal cycle of the convective boundary layer in the area serves as a modulating factor, capable under certain conditions of enhancing the turbulent entrainment of midtropospheric momentum from any directional quadrant toward the core of the valley [for the case of westerly winds, see Jiang and Doyle (2008)]. Conversely, a likely factor responsible for the strongest WWSs is mountain-wave activity, which causes strong near-surface winds from a very specific direction and does not require a well-developed convective boundary layer to be present in the valley. These concepts are explained in greater detail in the next two sections.

4. Mechanisms of westerly windstorms

The prototypical characteristics of WWSs in Owens Valley can be illustrated by examining and contrasting two intensive observation periods (IOPs) of the T-REX campaign, namely, IOP 3 (9 March 2006) and IOP 6 (25 March 2006).

IOP 3 is a moderately strong lee-wave and rotor event in a context of west-northwest flow over the Sierra Nevada, with some evidence of an eastward shift of the wave pattern in time (Cohn et al. 2011). During the most intense phase of IOP 3 (the 5 h from 0900 to 1400 LST), the mean and maximum hourly wind speeds at stations 1, 2, and 7 of the DRI network ranged, respectively, between 10 and 11 m s\(^{-1}\) and between 13 and 16 m s\(^{-1}\). Therefore, IOP 3 does not qualify as a WWS according to the criteria introduced in section 3c.

TABLE 1. WWSs, i.e., observed WWE with mean wind speed $\geq 15$ m s\(^{-1}\) or maximum hourly wind speed $\geq 20$ m s\(^{-1}\) in at least two of stations 1, 2, and 7 (group 1). Of these 12 events, 7 are outside the April–September WWE maximum, pointing to the role of dynamic forcing and mountain waves for their origin.

| Event | LST and date | Duration (h) | $U(1)$ | $U_{\text{max}}(1)$ | $U(2)$ | $U_{\text{max}}(2)$ | $U(7)$ | $U_{\text{max}}(7)$ |
|-------|--------------|--------------|--------|---------------------|--------|---------------------|--------|---------------------|
| 1     | 1300 2 Dec 2005 | 7            | 17.1   | 20.8                | 13.4   | 16.9                | 14.7   | 17.6                |
| 2     | 1800 25 Mar 2006 | 2            | 19.8   | 22.3                | 17.3   | 20.4                | 15.6   | 16.3                |
| 3     | 0800 15 Sep 2006 | 11           | 16.5   | 21.3                | 17.6   | 24.8                | 13.4   | 19.7                |
| 4     | 1200 16 Oct 2006 | 10           | 13.8   | 19.8                | 12.7   | 19.3                | 13.1   | 21.2                |
| 5     | 1600 10 Jan 2007 | 5            | 17.0   | 24.3                | 16.1   | 24.9                | 16.5   | 27.9                |
| 6     | 1000 25 Feb 2007 | 16           | 15.6   | 21.6                | 10.6   | 23.4                | 14.0   | 21.3                |
| 7     | 0900 26 Feb 2007 | 9            | 15.3   | 19.8                | 16.7   | 24.0                | 15.6   | 23.2                |
| 8     | 1300 5 Jun 2007 | 11           | 11.7   | 19.6                | 14.0   | 21.8                | 11.2   | 22.8                |
| 9     | 1100 19 Apr 2008 | 30           | 16.0   | 26.2                | 11.3   | 23.5                | 12.9   | 24.3                |
| 10    | 1700 29 Apr 2008 | 13           | 19.9   | 25.5                | 13.2   | 20.3                | 17.2   | 23.1                |
| 11    | 1000 31 Aug 2008 | 9            | 19.2   | 25.0                | 14.3   | 20.2                | 13.0   | 20.5                |
| 12    | 1600 13 Feb 2009 | 3            | 17.8   | 20.0                | 13.6   | 16.6                | 15.5   | 18.0                |

FIG. 8. Frequency of wind events with $U > 7$ m s\(^{-1}\) as a function of wind direction and gustiness factor at stations (a) 1 and (b) 9 of the DRI mesonet.
IOP 6 is, in contrast, related to a frontal passage (Sheridan and Vosper 2012; Kühnlein et al. 2013) associated with a short-wave trough, directing strong southwesterlies toward the mountain range (Jiang et al. 2011). As the trough transits over the barrier, large-amplitude waves develop and eventually break in its lee, with strong downslope winds ensuing right below the breaking region (Reinecke and Durran 2009). The windstorm terminates with the onset of channeled north-northwest flow in the valley after the front has passed the Sierra Nevada. IOP 6 is the strongest windstorm event observed during T-REX and one of the WWSs listed in Table 1.

Figures 10 and 11 present time series of wind speed, pressure perturbations, and alongslope pressure gradients along a cross-valley array of measurement stations (from 1 to 6) for the two IOPs. The 10-min block-averaged data are considered, and pressure measurements are reduced to the height of the lowermost station before computing perturbations.

During IOP 3, westerly flow is initially (0700–1000 LST) confined to the upper stretches of the Sierra Nevada slope, as shown in Fig. 10. Wind profiler measurements at two locations in Owens Valley provided evidence of persistent wave up- and downdrafts and of a well-developed rotor circulation in this early period (Cohn et al. 2011). A local pressure minimum exists at station 1, and the perturbation pressure gradient is positive (i.e., pressure increases in the cross-valley direction, corresponding to a heading of 15° north of east) between stations 1 and 2. The cross-valley wind detaches from the surface between stations 1 and 2. After 1000 LST, westerly momentum starts to protrude deeper into the valley. The transition lasts approximately until 1200 LST and corresponds to an eastward shift of the overlying wave pattern (Cohn et al. 2011). The pressure minimum on the lee slope moves from station 1 in the first phase (0700–1000 LST) to station 2 in the second phase (1200–1400 LST), which is the most intense in terms of
both the perturbation pressure gradient and the wind speed. Shortly after 1300 LST, $u_r$ (the wind component along the cross-valley direction) reaches 18 m s$^{-1}$ at station 2, where the local pressure minimum is located. The perturbation pressure gradient is negative (i.e., pressure decreases along the cross-valley direction) between stations 1 and 2 and positive between stations 2 and 3. Accordingly, the cross-valley wind accelerates upstream of station 2 and decelerates downstream of it. No detachment of the westerly flow from the ground is visible at this time.

The dynamics are similar during IOP 6, which is, however, much more intense (Fig. 11). While pressure perturbations amount to less than ±1 hPa throughout IOP 3, they exceed ±3 hPa during IOP 6, in connection with downslope winds in excess of 25 m s$^{-1}$. The penetration of

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**Fig. 10.** Hovmöller diagrams of (a) cross-valley wind speed $u_r$, (b) pressure perturbation $p'$, and (c) perturbation pressure gradient $dp'/dx$ in a cross-valley array comprising mesonet stations 1–6 during the T-REX IOP 3. Identical vectors in the two leftmost panels refer to the $u_r$ (positive from left to right) and $v_r$ (positive from bottom to top) components. The black square at 0920 LST and the black diamond at 1310 LST in (c) refer to the maximum observed alongslope pressure gradients in the two phases of the event. See text for details.
the westerly wind in the valley is limited by the very strong positive pressure gradient, exceeding 1 hPa km$^{-1}$ between stations 2 and 4 from 1800 to 1900 LST. Indeed, reverse flow (with a negative cross-valley component $u_x$) is apparent downstream of the windstorm front, indicative of the occurrence of wave-induced boundary layer separation (Doyle and Durran 2002; Jiang et al. 2007; Grubišić et al. 2015; Sachsperger et al. 2016) before north-northwest wind sets in with the frontal passage, at approximately 2000 LST.

The key feature of the dynamical evolution of the T-REX IOPs 3 and 6, that is, the relationship between the near-surface wind speed and pressure fields, can also be observed in a climatological sense. That is, it is possible to track certain features of the climatological correlation between the pressure and wind fields back to the occurrence of dynamically forced downslope windstorms.

The variability of the pressure difference ($\Delta p$) between stations 3 and 2 at the seasonal and diurnal scales is represented in Figs. 12a and 12b, respectively. The two

![Image](image-url)
locations are chosen because the maximum cross-valley pressure gradient occurs between them in both IOP 3 and IOP 6 (see Figs. 10 and 11 above). To remove the obvious altitudinal and thermal dependence of $\Delta p$, the pressure at station 2 was reduced to the level of station 3 (205 m below) using the hypsometric equation. The quantity $\Delta p$ is affected by a distinct diurnal variability (Fig. 12b), determined by the larger amplitude of the diurnal pressure cycle typically observed at station 3. The amplitude of the diurnal pressure cycle is known to increase from west to east on the western slope of Owens Valley, presumably as a consequence of the regular diurnal development of the convective boundary layer (Li et al. 2009). Since the mixing height tends to be horizontally uniform in Owens Valley, the boundary layer depth increases from west to east down the sierra slope and so do the amplitudes of the diurnal cycles of surface pressure and temperature.

The pressure imbalances implied by this periodicity are relatively weak, that is, mostly within $\pm 1$ hPa. However, values of $\Delta p$ in the range [1, 4] hPa, implying a relative pressure minimum at station 2, are not infrequent and indicate a weak afternoon maximum and a weak summertime minimum (Fig. 12). Values of $\Delta p < -1$ hPa are conversely very rare.

Figure 13 shows joint frequency distributions of $\Delta p$ versus $\Delta u$, between three pairs of stations. Considering stations 3 and 2 or stations 2 and 1 (Figs. 13a,b), strong positive values of $\Delta p$ apparently correlate with strong negative values of $\Delta u$. That is, when the upper station features a pressure minimum, it is generally affected also by relatively strong cross-valley winds. Cross-valley wind speed differences between the upper and the lower station can exceed 20 m s$^{-1}$ for pressure differences of 3 hPa or greater.

This finding is compatible with a scenario in which a windstorm occurs in Owens Valley but only affects the upper portion of the sierra lee slope. The upper station is affected by strong downslope winds and features a relative pressure minimum. At the same time, the lower station lies downstream of the point where the strong downslope flow separates from the ground, possibly in response to a strong adverse pressure gradient force. In fact, pressure is higher there, and the cross-valley wind component $u_r$ is close to zero. The stronger the adverse pressure gradient force (and therefore $\Delta p$) are, the stronger are the deceleration (and therefore $\Delta u$).

The condition $u_r = 0$ at the lower station does not necessarily imply that the wind speed is close to zero at that location. In fact, it is possible that $v_y$ (i.e., the along-valley wind component) is nonnegligible there. It is fairly common in Owens Valley to observe marked shear lines between dynamically accelerated downslope winds on the sierra slope and well-developed along-valley winds near the valley axis, possibly caused by pressure-driven channeling of the synoptic flow (Strauss et al. 2016).

The relationship between $\Delta p$ and $\Delta u$, between any two locations at the valley bottom, for example, stations 5 and 11, both located near the valley axis, does not reveal any meaningful pattern (Fig. 13c). Values of $\Delta p$ and $\Delta u_r$ are in fact more evenly distributed and much closer to zero in the great majority of cases. Hence, strong contrasts in pressure and cross-valley wind speed between neighboring stations appear to be a unique feature of the western portion of the DRI network.

The black square and diamond (for IOP 3) and inverted triangle (for IOP 6) in Figs. 13a and 13b represent the maximum $\Delta p$ and $\Delta u_r$ observed during those two IOPs. The square marker for IOP 3 (Fig. 13b) refers...
to the first phase of the event (0700–1000 LST, fully developed rotor, and maximum pressure perturbation observed between stations 1 and 2), whereas the diamond marker (Fig. 13a) refers to the second phase (1200–1400 LST, westerly winds at the valley floor, and maximum pressure perturbation observed between stations 2 and 3).

Clearly, IOP 6 ranks among the events with the strongest observed cross-valley pressure gradients. As shown before, the pressure difference of 4.1 hPa (corresponding to an alongslope pressure gradient of 1.2 hPa km\(^{-1}\)) is sufficient to cause the downslope windstorm to separate from the ground. This is consistent with the notion that a pressure gradient \(O(1\text{ hPa km}^{-1})\) is necessary for wave-induced boundary layer separation to occur in Owens Valley (Strauss et al. 2016). Figure 13a shows that such strong streamwise pressure gradients occur only rarely at this site (given the distance of 3.3 km between stations 2 and 3, a gradient of 1 hPa km\(^{-1}\) corresponds to a pressure difference of 3.3 hPa, indicated by the dashed horizontal line). This confirms the exceptionally severe character of T-REX IOP 6.

Conversely, the pressure and wind speed differences observed during IOP 3 do not appear to be particularly strong. The maximum pressure difference was observed during the second phase of the event (1.3 hPa). It corresponds to a pressure gradient \(\approx 0.4\text{ hPa km}^{-1}\), well below the 1 hPa km\(^{-1}\) threshold. Hence, the relatively strong surface winds at station 2 do not separate from the ground in this case, and the windstorm affects the whole area of the DRI network.

During the first phase of IOP 3, on the other hand, the westerly flow separates from the Sierra Nevada slope (Fig. 10a) even if the alongslope pressure gradient is rather weak (\(\Delta p \approx 0.8\text{ hPa}\), or \(\Delta p \approx 0.25\text{ hPa km}^{-1}\) at 0920 LST between stations 1 and 2; Fig. 10c). Figure 13b

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**Fig. 13.** Relationship between \(\Delta p\) and \(\Delta u_r\) for (a) station 3 minus station 2, (b) station 2 minus station 1, and (c) station 11 minus station 5. As in Fig. 12, the logarithmic color scale represents the population of bins. Black markers in (a) and (b) represent the observed pressure and wind speed differences in T-REX IOPs 3 and 6 (see also Figs. 10 and 11). The dashed horizontal line in (a) and (b) indicates the value of \(\Delta p\) that corresponds to a streamwise pressure gradient of 1 hPa km\(^{-1}\) (the distances between stations 3 and 2 between stations 2 and 1 are 3330 and 3130 m, respectively).
shows that pressure differences of this magnitude between stations 1 and 2 are most often associated with moderate cross-valley wind speed differences. Hence, the sharp contrast in cross-valley wind speed observed between stations 1 and 2 in this case suggests that flow separation might be favored by factors other than the wave-induced pressure perturbation.

A preliminary examination of radiosonde measurements taken within Owens Valley during the early phase of IOP 3 shows evidence of a surface-based stable layer with a depth of a few hundred meters (not shown). The upward buoyancy force acting on the westerly downslope wind as it encounters the stable layer is likely an important component in the balance of forces leading to flow separation in this case. In fact, the timing of the transition between phase 1 and phase 2 of IOP 3 (1000–1200 LST) is compatible with the late-morning breakup of the valley inversion. Strauss et al. (2016) conjectured that a similar process might be at play during other T-REX events, for example, the first phase of IOP 13. The detailed analysis of the dynamics of flow separation during IOP 3 is out of the scope of the present study and is left for future investigation.

To summarize, the contrasting evolution of the T-REX IOPs 3 and 6 illustrates the considerable complexity inherent to WWSs in Owens Valley. The most severe events (e.g., IOP 6) are associated with large-amplitude mountain waves, and the degree of intrusion of westerly winds to the valley floor in these cases is mostly controlled by wave-induced pressure perturbations. In weaker events (e.g., IOP 3), mountain-wave forcing is still the primary dynamical forcing, but other factors (e.g., the stratification of the valley atmosphere) might determine whether westerly winds are confined to the upper reaches of the western sidewall or can reach the valley floor.

Further analysis of the measurements from the DRI mesonet shows that WWEs, during which the cross-valley wind component \( u_x \) measured at station 2 and (a) specific humidity \( q \) measured at station 2 and (b) virtual potential temperature difference \( \Delta \theta_v \) between stations 3 and 2.

![Diagram](image)

**Fig. 14.** Relationship between the cross-valley wind component \( u_x \) measured at station 2 and (a) specific humidity \( q \) measured at station 2 and (b) virtual potential temperature difference \( \Delta \theta_v \) between stations 3 and 2.

**5. Discussion**

**a. Thermal versus dynamical forcing**

WWEs in Owens Valley occur most frequently in the period from April through September, predominantly in the afternoon hours. This points to thermal forcing as an important process at play during the warmer months.

If synoptic-scale westerlies are present, local thermal forcing can enable or enhance the penetration of westerly momentum into the valley. In fact, the intrusion of westerly momentum along the lee slope of the Sierra Nevada is favored when the convective boundary layer in the valley grows beyond the mountain tops, lifting the capping inversion that shelters the valley atmosphere (Jiang and Doyle 2008; Mayr and Armi 2010). This
The scenario is clearly more likely in the afternoons of fair-weather days, especially during the summer months. Conversely, when a layer of very stable air is present in the valley, strong westerly winds remain confined to the upper stretches of the sierra lee slope, possibly at altitudes greater than the range covered by the DRI network. This is more likely to happen in the wintertime, at night, or in the early morning, explaining the rare occurrence of WWEs at the valley floor at these times.

Furthermore, during summer, regional-scale thermal forcing can cause westerly flow on the eastern slope of the Sierra Nevada even in the absence of synoptic-scale westerlies. This is due to the thermally generated pressure gradient between a thermal low over the Great Basin and a high-pressure area over coastal California. North of Owens Valley, in particular in Washoe Valley in western Nevada, this mechanism is known to drive the Washoe Zephyr westerly wind (Twain 1871; Siscoe 1974; Kingsmill 2000; Clements and Zhong 2005; Zhong et al. 2008a).

We showed that, relative to strong winds from other directions, WWEs in Owens Valley have a tendency to be more persistent (occasionally even more than 15 h) and of stronger sustained wind speeds (>15 m s\(^{-1}\)). A large fraction of the most severe WWEs, which we referred to as WWSs, occur in the winter months. Clearly, in the absence of the thermal forcing factors discussed above, dynamical forcing related to Sierra Nevada mountain waves is left as the most likely primary driver of these events [for a comparative...
b. Comparison with other windstorm climatologies

Climatologies of downslope winds at different sites around the world have appeared in the scientific literature as early as the 1930s. See for instance the reviews by Atkinson (1981) and Barry (2008) and the references therein, as well as Richner and Hächler (2013). Most of these studies have dealt with warm and dry downslope winds (foehn in the European Alps and chinook in the western United States), but some have touched on cold downslope winds as well (e.g., bora along the east Adriatic coast in Europe).

These climatological investigations have used diverse approaches. In one case, a chinook climatology was assembled by meticulous scrutiny of newspaper accounts, spanning a period of over a century (Whiteman and Whiteman 1974). In most other studies, episode identification was based on instrumental data. A key component of such studies is a decision rule, that is, a criterion that defines what periods of a data record should be regarded as downslope windstorm episodes. Criteria adopted in the literature typically consider the wind, as well as variations in temperature and relative humidity. For instance, in the case of foehn, the wind should blow from the direction of a nearby mountain range, and it should cause warming and drying. It is known that consideration of these parameters alone may lead to a high rate of misclassifications (Brinkmann 1970). This led some to question the possibility of detecting foehn episodes based only on pointwise surface measurements and to develop criteria that also take into account the synoptic pressure field and the upper-level wind (e.g., Brinkmann 1971). Readers interested in the subtleties of foehn recognition may refer to the
extensive discussions by Atkinson (1981, p. 82) and Barry (2008, p. 173).

Whatever the parameters they consider, identification rules for downslope windstorms typically rely on subjective threshold values, and they are combined and weighted in a subjective manner. For example, westerly windstorms in the present study are identified solely from wind speed and direction. In contrast, much of the early research on foehn in the European Alps emphasized primarily the aspects of drying and warming at the onset of these winds without paying much attention to wind speed and gustiness (Atkinson, 1981). Similarly, Beran (1967) identified areas affected by chinook in the Colorado plains only from the dewpoint depression. Given the absence of a well-established and univocal criterion to identify downslope windstorm episodes, it is hard to relate the results of previous foehn or chinook climatologies to those presented in this paper.

Because we identified WWSs as periods with anomalously intense and persistent downslope winds, it would be misleading to label them indiscriminately as foehn or chinook episodes. This is well illustrated by Fig. 15, which displays the records of wind speed, wind direction, relative humidity, and virtual potential temperature $\theta_v$ during the 12 WWSs listed in Table 1.

In most cases, WWSs in Owens Valley occurred in connection with the displacement of preexisting air masses by new ones. This is evident in the events 1, 4, 5, and 7. In these cases, the onset of westerly flow in the three stations of group 1 coincided with a sharp drop in humidity and with rising $\theta_v$. Also, the initially stratified valley atmosphere became better mixed during the WWSs, as can be inferred by comparing the $\theta_v$ records from different stations. Before the WWSs, $\theta_v$ increased with height and distance upslope from station 12 to station 1, but differences in $\theta_v$ between the five stations were greatly reduced during the WWSs. Sometimes,
the airmass boundary related to the downslope wind front oscillated around a measurement station (e.g., 2 h into event 5, when weak, cold, and moist northerlies set in only for a short time at station 12). All of these are well-known features of foehn or chinook episodes (Glenn 1961).

On the other hand, many WWSs exhibited striking deviations from the prototypical behavior of a foehn intrusion.

- Several cases (e.g., 2, 6, 11, and 12) led to overall drying at their onset but no marked variation in $\theta_v$.
- In one case (number 3), $\theta_v$ increased, but relative humidity remained approximately constant.
- During case 6, differences in $\theta_v$ between the highest (1 and 7) and lowest (2, 9, and 12) stations appeared to grow in time after 2000 LST: this is likely indicative of westerly flow separating from the surface to the east of stations 1 and 7 and riding over a stable air mass that got cooler overnight and flowed down valley.
- In three springtime cases (8, 9, and 10), westerly winds hit the valley floor during the afternoon and persisted into the night, while surface temperature gradually decreased according to the normal diurnal cycle. The virtual potential temperature was spatially homogeneous across Owens Valley before these three events and for most of their duration, with the exception of cold-air pools developing at the valley bottom at night in cases 9 and 10 (other evidences of flow separation). Relative humidity did not undergo noticeable variations in these three events except for one (number 8) in which an overall moistening of the valley atmosphere occurred.

This overview shows that, clearly, the temperature and humidity tendencies brought about by the onset of westerly flow at the ground depend on how different these two quantities are in the intervening air masses. The many Owens Valley WWSs in which simultaneous drying and warming were not observed would not qualify as foehn episodes, despite the high and remarkably persistent wind speeds. Therefore, any comparison of the present results with foehn or chinook climatologies should be done with prudence.

Despite these notes of caution, some parallels can be drawn between downslope windstorms in Owens Valley and similar phenomena at other locations. Among the several published climatological studies on foehn and chinook, seemingly only a report by Riehl (1974), cited by Atkinson (1981), uses a methodology similar to ours. The report describes 87 chinook events observed in Fort Collins, Colorado, in 1964–71, which were identified from the peak wind speed. Speed thresholds of 11 and 15 m s$^{-1}$, respectively, in summer and in winter had to be exceeded for a given period to be regarded as a chinook occurrence. The key findings were that no clear seasonal pattern existed in the chinook occurrences; that their duration was mostly between 2 and 10 h (only 14% of the events, exclusively in winter–spring, exceeded this duration); and that peak gusts, generally weaker in summer, mostly reached between 16 and 33 m s$^{-1}$ and occasionally up to 43 m s$^{-1}$.

Similarly, WWSs in Owens Valley display an unclear seasonal distribution, have a similar fraction of events longer than 10 h (14 out of 148 events, i.e., 9%), and feature marginally lower peak gusts (reaching 36 m s$^{-1}$; Fig. 7).

The indefinite seasonal dependence of downslope windstorms evidenced in this study and in the work by Riehl (1974) is at odds with other analyses. For instance, Whiteman and Whiteman (1974) report a wintertime maximum of windstorm days in Boulder. Richner and Hächler (2013) give an account of spring and fall maxima in the foehn occurrence in Altdorf, Switzerland. Apparently, in addition to all the previously mentioned ambiguities related to foehn classification, the seasonality of downslope windstorms is heavily affected by site-specific factors such as the typical synoptic circulation patterns and their interactions with local circulation systems.

6. Summary and conclusions

Based on the analysis of five years of wind data from the DRI network in Owens Valley, we have shown that strong westerly wind events represent a distinctive class of phenomena, in particular at the measurement stations located at higher altitudes on the western valley slope. WWEs are identified using criteria based on the wind speed (persistently stronger than a threshold value of 7 m s$^{-1}$) and direction (steadily westerly at more than one measurement site). WWEs affect the upper elevations of the network on the western side of the valley but only rarely the flat valley bottom. For this reason, previous wind climatologies in this area, based only on measurements taken along the valley axis (Zhong et al. 2008b), could show only very little evidence of the phenomenon.

The results shown here provide a consistent climatological context to the individual observations of westerly winds during the Sierra Rotors Project and T-REX already documented in the literature (Grubišić and Billings 2007; Cohn et al. 2011; Sheridan and Vosper 2012). They support the notion that westerly windstorms (i.e., the most intense WWEs in Owens Valley) are primarily dynamically driven, that is, they respond to a localized pressure minimum related to mountain-wave activity in the lee (east) of the Sierra Nevada. While WWEs can be affected by thermal layering in the valley and thus exhibit a dependence on the time of day, WWSs are primarily dynamically determined and can occur at any time of the day.
The comparison with the results of other windstorm climatologies, developed mostly in the context of foehn and chinook research, is not straightforward. We show that the variations in air temperature and moisture content caused by the onset of WWSs in Owens Valley often deviate from the typical characteristics of foehn intrusions. That is, WWSs in Owens Valley are not necessarily related to warming and drying of the valley atmosphere. However, the typical intensity of sustained wind speeds and maximum gusts in WWSs in the lee of the Sierra Nevada are comparable to those reported in other wind climatologies, for example, in the lee of the Rocky Mountains.

Further investigations are planned in order to clarify what kind of synoptic conditions are responsible for WWSs in Owens Valley even in the absence of appreciable thermal forcing. To this end, data from the DRI mesonetwork will be complemented with upper-air observations and three-dimensional weather analyses.

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