

**AIRBORNE DOPPLER LIDAR MEASUREMENTS OF VALLEY FLOWS IN COMPLEX COASTAL
TERRAIN**

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Abstract

Three-dimensional winds obtained with an airborne Doppler lidar are used to investigate the spatial structure of topographically driven flows in complex coastal terrain in southern California. The airborne Doppler lidar collected four hours of data between the surface and 3000 m MSL along a 40 km segment of the Salinas Valley during the afternoon of 12 November 2007. The airborne lidar measurements, obtained at a horizontal and vertical resolution of approximately 1500 m and 50 m, respectively, reveal a detailed spatial structure of the atmospheric flows within the valley and their associated aerosol features. Clear skies prevailed on the flight day with synoptic northwesterly flows around 10 m s^{-1} . The data document a shallow sea breeze transitioning into an upvalley flow in the Salinas Valley that accelerates in the upvalley direction. Along with the acceleration of the upvalley wind, enhanced sinking motions are observed. No return flows associated with the sea-breeze or upvalley flows are observed but a distinct wind minimum is present at heights around 400 m MSL near the coast and 600 m MSL up the valley. While synoptic flows are aligned along the valley axis in the upvalley direction, lidar data indicate the presence of a northerly cross-valley flow around the height of the surrounding ridges. This flow intrudes into the valley atmosphere and induces, along with thermally-driven slope flows on the sunlit valley sidewall, a cross-valley circulation that causes an asymmetric distribution of the aerosols. This study demonstrates the large potential of airborne Doppler lidar data in describing flows in complex terrain.

1. Introduction

The temporal evolution of thermally-driven valley winds has been well investigated and is described in several reviews (e.g. Whiteman 1990). Valley flows occur when horizontal temperature differences along the valley floor produce a horizontal pressure gradient resulting in upvalley flows during daytime and downvalley flows during nighttime (e.g., Wagner, 1938; Whiteman 1990, 2000). The spatial structure of valley flows is often less understood with flow structure affected by factors such as the presence of ambient synoptic-scale winds, tributary valleys, and adjacent oceans. These factors can make the spatial flow structure much more complicated. For example, in the case of an adjacent ocean, sea breezes can interact with valley flows to generate an extended sea breeze (e.g. Kondo, 1990). Synoptic flows can be subject to channeling into valleys (Whiteman, 2000) and upvalley flows have been observed to accelerate along the valley floor (Rucker et al., 2008). Across the valley, flow heterogeneities have been observed due to dynamical effects caused by bends in the valley (De Wekker et al., 2005; Weigel et al., 2006) and by thermally-driven cross-valley circulations (Bader and Whiteman, 1989). Furthermore, ridge-top level winds may modify the subsidence over the valley center during daytime and produce asymmetries in the cross-valley circulation (Zardi and Whiteman, 2011). Documenting and investigating spatial variability in wind structure caused by the interaction between valley flows, sea breezes, and ambient flows is important especially for transport studies where deviations from the mean wind structure can lead to large errors in air pollution dispersion.

Studies in complex coastal terrain have mostly focused on the influence of coastal orography on the structure of the sea breeze (e.g. Banta et al., 1995; Drobinski et al. 2006; Bastin et al. 2005; 2006). In the US, this influence has been investigated intensively at Monterey Bay and the adjacent Salinas Valley in California, particularly during the LASBEX (Land And Sea Breeze EXperiment) project (Intrieri et al. 1990). Ground based Doppler lidar measurements during LASBEX indicated that the sea-breeze consisted of two scales of flow in the vertical and in time: a shallow (a few hundred meters deep), strong ($\sim 6 \text{ m s}^{-1}$) sea breeze with a horizontal scale of 30 to 40 km starting between 0900 and 1000 LST and a weaker, deeper sea breeze with a horizontal scale of 100 km forming later in the day around noon (Banta et al. 1993). These two flows eventually merged into a well-blended layer of onshore flow about 1 km deep. Other investigators (e.g., Fosberg and Schroeder 1966; Johnson and O'Brien 1973) have also observed this dual structure of the sea breeze along the west coast of the United States. Darby et al. (2002) investigated the two scales of flow using two-dimensional model simulations and concluded that the land–water contrast is responsible for the shallow sea breeze, and that the onset of deep slope flow associated with the coastal mountains is responsible for the weaker, deeper (up to $\sim 1500 \text{ m}$) onshore flow. The winds at the shore, below 1500 m, were most affected by the land–sea contrast and the coastal mountain range. Above 1500 m, winds at the shore were affected by the Sierra Nevada range located about 300 km away from the coastline. Slope flows generated by nearby sloping terrain affect the time of onset and the depth of the sea breeze (Darby et al., 2002). Also, while studies where sea-breezes have been studied in isolation have clearly

demonstrated the presence of return flows, these return flows have not been shown in the combined sea-breeze/valley wind system at Monterey Bay (Banta et al., 1995).

An investigation of valley flows and the influence from sea breezes and ambient flows requires making three-dimensional observations of the spatial structure of flows in complex terrain at the local to meso-scale (1 to 100 km). Making these observations is an important but difficult challenge. Much progress has been made in obtaining observational information on the spatial wind structure for meteorological and environmental applications by Doppler lidars that were developed in the late 1960s (see Huffaker and Hardesty 1996 for a review). Ground-based Doppler lidars have provided the research community with important new insights of thermally-driven valley flows (Post and Neff, 1986; Banta et al., 1999), sea-breezes (Banta et al., 1993; Darby et al., 2002), and nocturnal jets (Banta et al., 2004), among other flow structures. While a single ground-based scanning Doppler lidar obtains spatial wind structure information through the analysis of the measured radial velocity, the three dimensional wind vector can only be obtained at one single location. To avoid this limitation, scanning Doppler lidars have also been mounted on airborne platforms since the late 1970s (Bilbro et al. 1978; Bilbro et al. 1984; Bilbro and Emmitt, 1984) enabling the documentation of atmospheric flows at high spatial resolution for larger spatial scales than for ground-based Doppler lidars. Mountain-plain flows (Reitebuch et al., 2003; Weissmann et al. 2005) and sea breezes (Carroll 1989; Bastin et al. 2005; 2006, Drobinski et al. 2006) are among the studied flows using an airborne Doppler lidar. These previous studies investigated flow structure at a spatial resolution of about 10 km or larger which is too

coarse for studying across and along valley flow structure in many valleys. With increasing resolution of weather forecasting and air quality models, the documentation of flows at high spatial resolution (~ 1 km) becomes increasingly important so that these models can be evaluated and improved.

Over the past 10-15 years, the Integrated Program Office (IPO) of NPOESS (National Polar-orbiting Operational Environmental Satellite System), with an eye to eventual deployment in space, has funded the development of an airborne Doppler lidar initially mounted in a Navy Twin Otter aircraft to conduct a variety of investigations. Since 2002, under the direction of Simpson Weather Associates (SWA) and operated by the Navy's Center for Interdisciplinary Remotely Piloted Research Studies (CIRPAS), the Twin Otter Doppler Wind Lidar (TODWL) has flown more than 125 hours of atmospheric missions. The TODWL has mostly obtained data at high spatial resolution over the Pacific Ocean including complex coastal terrain of California with the main purpose of retrieving winds over water surfaces and determining ocean wave characteristics. However, a number of these flights were performed over the coastal mountains near Monterey Bay.

In this study, we use existing TODWL data to document and investigate the flow structure at high spatial resolution in the Salinas valley of California and adjacent mountainous terrain. We focus our discussion on data collected during the TODWL flight on 12 November 2007, a fair weather day with weak ambient winds. In contrast to previous measurements in the area, data were collected over a larger area and at much

higher resolution. Such measurements allow the investigation of interacting flows at multiple scales. Furthermore, observational studies in the area have mostly been conducted in late spring to early fall when thermally-driven flows are prominent. While surface heating is reduced relative to summer, winds in the Salinas valley can still be influenced by both the mountain-induced flows and the sea breeze systems. A complex wind pattern can therefore be expected with potential additional influences of other mesoscale flows and synoptic background flows. We will present some details of the topography of the investigation area and the collected data in section 2. In section 3, we will present and discuss the observed wind and aerosol structure on 12 November 2007 followed by conclusions and outlook in section 4.

2. Data and data domain

All data shown in this study were collected in the complex coastal terrain near Monterey Bay. The investigation area consists of complex shoreline, a coastal region with several parallel mountain ranges oriented north-northwest to south-southeast, and the interior valleys of California to the east (Fig. 1). The Salinas Valley is bounded by the Sierra de Salinas to the west and the Gabilan and Diablo Ranges to the east. The adjacent mountain ranges have a maximum elevation of about 1000 m MSL. The western sidewall of the Salinas Valley is more uniform and steeper than the eastern sidewall which is characterized by the presence of many side canyons and tributary valleys. The Salinas Valley is about 20 km wide at its mouth and extends about 130 km to the southeast at an orientation of 140° from north. The valley floor is rather flat with a height increase of

about 160 m from Monterey Bay to King City, located 100 km southeast from Monterey Bay. Land use in the Salinas Valley is dominated by agriculture, in particular irrigated crops. Details of the Twin Otter Doppler Wind Lidar and other data collected on 12 November are presented next.

2.1 Twin Otter Doppler Wind lidar

The TODWL is a 2 μm coherent system built by Coherent Technologies, Inc (now Lockheed Martin Coherent Technologies). Table 1 summarizes some of the technical details of the lidar. Issues related to the accuracy of the system are discussed in Appendix A. A defining capability of the TODWL is its ability to profile above, on and below the flight level. With its side door mounted, bi-axis scanner the beam can be adaptively directed in a variety of scan patterns including conical, nadir stares, and flight level stares.

To measure the wind velocity and wind direction with a Doppler lidar, conical scanning of the laser beam around the vertical axis is commonly used. Such a measurement method has traditionally been referred to in the Doppler radar community as the velocity azimuth display (VAD; Lhermitte and Atlas 1961, Browning and Wexler 1968). From the return signals measured at various azimuth angles, the Doppler frequency shifts are estimated which are proportional to the radial wind (or line of sight, LOS) velocity. From the measured dependence of the radial wind velocity versus azimuth angle, the wind velocity vector is retrieved with a height interval defined by the processing routine.

During the TODWL flights in the current study, twelve point step-stare scans were performed. During these scans, the azimuth angle changes in steps of 30 degrees, so that twelve LOS profiles are obtained during each conical scan. These twelve LOS profiles are used for processing into the wind retrieval algorithm to produce an individual vertical wind profile. The typical dwell time at each stare is about 1 s during which 500 pulses are transmitted into the atmosphere. With the transition from one stare to another stare of about 1 second, the total time it takes to complete a full rotation of the scanner is about 20-30s. The rotating lidar is projected 20 degrees off nadir and aircraft attitude corrections are made in the post-processing. The ground speed of the aircraft is 50-60 m s⁻¹ so that during that time a distance of about 1500 m is covered. In other words, a vertical profile of the horizontal and vertical wind component is obtained about every 1500 m. The selected vertical resolution (or, more accurately, grid spacing) of the wind profile is 50 m up to a height of about 2500 m which is 500 m below the aircraft altitude of 3000 m. The vertical resolution of a Doppler lidar measurement is dependent on the pulse length of 320ns which is, in the case of the TODWL, approximately 48m long at its full width half max. Since the raw data is processed using 96 meter gates, there is an overlap of photon returns between sequential 50 m resolution range gates. Note that the horizontal and vertical resolution obtained by TODWL is higher than the resolution obtained in previous airborne Doppler lidar studies (e.g. Weissmann et al., 2005). These other studies had different aircraft speeds and scanning patterns.

The wind velocity vector is retrieved in this study by fitting the radial velocities to a sine wave. An implicit assumption is that the flow is homogeneous with uncorrelated wind variability between the twelve azimuth angles in both aerosol distributions and velocity. More specifically, the sine wave is assumed to be the best model for computing a three-dimensional wind vector within a volume that is defined by the aircraft speed ($\sim 50 \text{ ms}^{-1}$), the nadir angle (20 degrees), the time to complete a twelve point step stare scan (20-30 s) and a processing LOS grid resolution (~ 50 meters). In this study, the amplitude, the average offset from zero, and the phase of the sine wave are proportional to the horizontal wind speed, the vertical wind speed, and the horizontal wind direction, respectively. The “goodness” of the sine fit to the 12 LOS observations and thus the deviations from the assumed homogeneity of the sampled atmospheric volume depends on instrumental, platform, and atmospheric parameters. For airborne systems, corrections need to be made for the speed, direction and attitude of the aircraft. A dedicated GPS/INS together with surface returns were used in the processing of TODWL data.(see Appendix A).

The retrieval of winds from airborne Doppler lidar over complex terrain is not straightforward. Major challenges exist, for example, in the retrieval of winds in the lowest 300 m AGL and over sloping terrain. The fundamental issue of coherent Doppler wind lidar velocity measurement precision and accuracy has been addressed for ground and space-based applications in various papers (e.g. Frehlich, 2001; Frehlich, 2004; Frehlich and Sharman, 2005). The implicit assumption of homogeneity in the flow and aerosol distributions within the twelve point scan volume of TODWL raises issues when applied to the Salinas Valley and adjacent mountain flows. Appendix A contains some

results of simulations have been conducted to put some bounds on the contributions of LOS scale turbulence and full scan scale coherent wind structures.

A total of six cross-valley flight legs were flown across the Salinas Valley on 12 November 2007. The location of the flight legs is illustrated in Fig. 1. Two of the legs (first and third) followed similar paths and were flown two hours apart. Each of the legs covered a distance of up to about 60 km and were oriented in a NE-SW direction perpendicular to the orientation of the valley. The legs were separated by about 8 km in the along-valley direction. The flights took place between 1200 and 1530 LST. The start and end times of each flight leg are given in Table 2. During each leg, up to about 50 downward conical scans were performed resulting in a vertical wind profile about every 1500 m (see section 2).

2.2. Other data

Data from five surface stations and an operational wind profiler in the Salinas Valley/Monterey Bay area are used to describe the temporal changes in the wind. The surface stations are part of the California Irrigation Management Information System (CIMIS) network. The Castroville station is located closest to the coast line and the wind profiler location. The north and south Salinas stations are within the investigation area while Arroyo Seco and King City are located further up the Salinas Valley (see Table 3 and Fig. 1). Data from the CIMIS network include temperature, winds, and net radiation at 1-hour intervals. Vertical wind profiles at half-hourly interval were obtained from the

Fort Ord 915 MHz boundary layer wind profiler which is part of the coastal profiling network operated by the Naval Postgraduate School. The profiler is operated in two modes resulting in vertical profiles with a resolution and range of approximately 50 m and 1 km in the fine mode and 200 m and 3 km in the coarse mode. The closest National Weather Service rawinsonde location is in Oakland, CA, which is more than 100 km north of the investigation area. These data are of limited usefulness for the investigation of mesoscale flows in the Salinas Valley. In-situ meteorological data on the Twin Otter aircraft were not recorded on 12 November.

3. Observed flows in and around the Salinas Valley

In this section, we will first describe the general atmospheric conditions on 12 November 2007, followed by a discussion of the data collected during the TODWL flight on that day.

a. General atmospheric conditions on 12 November 2007

Clear skies prevailed in the investigation area on 12 November 2007 with weak north-northwesterly synoptic flows around $5\text{--}10\text{ m s}^{-1}$ at 850 mb (Fig. 2; note that LST = UTC – 8 h for the Salinas Valley). These weak synoptic flows promoted the development of thermally-driven flows in the area. Data from all five surface stations show a shift from southeasterly (downvalley or offshore) to northwesterly (upvalley or onshore) flows around noon LST with hourly mean wind speeds at 2 m AGL of up to 5 m s^{-1} during the

various TODWL flights between 1200 and 1530 LST (Fig. 3). The shift from downvalley to upvalley flows occurs latest ($\sim$ noon LST) at King City which is located furthest up the valley. Going from the northerly to the southerly stations, afternoon winds turn from westerly to northerly winds as the westerly sea breeze channels into the northwesterly aligned Salinas Valley. Maximum wind speeds increase in the upvalley direction. Average wind conditions for the month of November for the 2003-2007 time period indicate that the diurnal variability of wind and temperature on 12 November follow the climatological average (not shown). Associated with the turning of the wind from offshore to onshore, a decrease in temperature occurs at Castroville and Salinas North. Further up in the Salinas Valley at Salinas South, this decrease in temperature occurs later in the afternoon around 1430 LST consistent with the propagation of a sea breeze front at about 4 m s^{-1} . Note the increase in the diurnal temperature range up the valley which is typical of valley atmospheres where valley volume decreases up the valley while heat input approximately remains the same (e.g. Whiteman, 2000). Maximum temperatures range from 17°C at Castroville to 23°C at King City. Data from nearby buoys indicate a sea water temperature around 14°C .

Data from the Fort Ord wind profiler provide detailed information on the temporal evolution of the vertical wind structure near Monterey Bay (Fig. 4). Near the surface (below 300 m), southeasterly (offshore) winds of about $6 - 9 \text{ m s}^{-1}$ prevail during the night and in the early morning up to about 0800 LST. Later in the morning, surface winds first turn to easterly/northeasterly direction while decreasing in speed, followed by a change to stronger (up to 6 m s^{-1}) northwesterly winds in the afternoon. The turning from

easterly to northeasterly flows in the morning up to about 600 m is likely due to thermally driven upslope flows initiated on the east-facing sidewalls of the coastal mountains. Westerly and northwesterly flows commence shortly after noon indicating the onset of the sea breeze. At altitudes above 600 m, winds gradually turn to the geostrophic northwesterly winds. From mid-morning onwards, winds are weaker than 10 m s^{-1} up to a height of about 2000 m except for a couple of hours in the afternoon when a distinct wind speed maximum of just over 10 m s^{-1} occurs at heights around 1300 m. The timing of this wind speed maximum coincides with the timing of the TODWL flight legs, indicated by the vertical arrows in Fig. 4. While this wind speed maximum appears of short duration, its spatial extent is large and obvious from the TODWL data, as will be discussed in the next section. The wind speed maximum can be seen more clearly in the vertical profiles in Fig. 5 which also shows that the wind direction is persistently from the northwest and that the wind speed maximum disappears later in the afternoon. The origin of the wind speed maximum of just over 10 m s^{-1} between about 1000 and 1500 m is unclear from the vertical profile data alone. Fig. 5 also shows a vertical profile of the Doppler lidar derived horizontal wind speed and direction, and the vertical wind speed at 1529 LST at about 8.5 km from the Fort Ord wind profiler. The lidar and wind profiler winds compare well and show that the TODWL data captures the wind speed maximum well and in more detail (i.e., at higher vertical resolution) than the wind profiler data. While there is generally a good agreement between the radar wind profiler and airborne Doppler lidar derived winds, we should recognize that the averaging time and the sampled volume is different for the two observing systems. This could explain, for example, the worse agreement with the 1600 LST radar profiler data. The strength of the TODWL data is in capturing

the wind structure at high spatial resolution which we will discuss in the next section. Note that directly above the shallow westerly sea breeze layer of a few hundred meters, winds are very weak ($< 2 \text{ m s}^{-1}$) and from northerly directions. Boundary layer heights determined from the wind profiler data using the signal-to-noise ratio indicate a maximum height of around 600 m MSL. The onset of the sea breeze is much delayed compared to observations made previously in summertime during which the onset ranged from 0830 to 1030 LST (Fagan, 1988; Banta et al., 1993). In November, the land heats up more slowly and less intensely and the sea-breeze onset is therefore expected to be delayed.

b. Doppler lidar data results

We will first discuss the spatial variability of the wind and aerosol structure in the entire flight area, followed by a detailed analysis of the flows in and at the exit of the Salinas valley. The temporal evolution along the Salinas Valley cross section that was covered by two flight legs (1 and 3) made two hours apart is also discussed. This entire flight took place in the afternoon, shortly after the surface winds changed from offshore/downvalley to onshore/upvalley direction.

1) SPATIAL WIND STRUCTURE IN THE ENTIRE FLIGHT AREA

Doppler lidars not only provide the radial velocities but also the signal-to-noise ratio (SNR) for each range gate. As the SNR is related to both the aerosol composition and

concentrations, we can use the SNR as a first order visualization of the spatial aerosol structure. The temperature and aerosol structure are often closely related. For example, in a convective boundary layer with vigorous mixing, one might expect a homogeneous distribution of aerosols with a sudden increase at the height of an elevated inversion where stability is large and mixing inhibited. Consequently, the aerosol structure provides some guidance on the presence of distinct atmospheric layers (e.g. De Wekker et al., 2004).

Figure 6 shows the spatial structure of the horizontal winds and the SNR for flight legs 2-6. The horizontal wind vectors are only shown every 200 m in the vertical. Three layers are clearly present, each of which has an increasing SNR going from the top to the bottom layer. A layer with the lowest SNR and northwesterly winds larger than 10 m s^{-1} is present above $\sim 2000 \text{ m}$. Winds in this layer, which we identify as the ‘free atmosphere’ and which represent the synoptic flow, are spatially homogeneous across the coastal mountainous terrain. The transition to synoptic flows above 2000 m was also seen in Figs. 4 and 5.

Up to a height of $\sim 1300 \text{ m}$, a layer with the largest SNR is found. This surface-based layer becomes somewhat shallower with a more heterogeneous structure as the terrain elevation increases. This layer does not appear to follow the terrain but there is a tendency for a deeper and more concentrated aerosol layer over the eastern sidewalls of the Salinas Valley. A possible explanation for this behavior of the aerosol layer is

provided later. Weak flows ($<6 \text{ m s}^{-1}$) dominate in this layer with increasing speeds up the valley (compare wind speeds in Salinas Valley between Figs 6 c,e,g).

An interfacial layer is present between the surface based layer and the free atmosphere. Winds in this layer and in the surface-based layer are generally between 4 and 10 m s^{-1} . Note however the larger windspeeds at the interface of these layers at about 1300 m where the sudden decrease in SNR indicates a change in stability. Furthermore, there is a tendency for stronger winds in this interfacial layer closer to the coastline than further inland contrary to the flows near the surface.

It is well known that close to the coastline, the northwesterly flow is enhanced by coastal baroclinicity and often results in a coastal jet near the strong inversion capping the marine boundary layer (Zemba and Friehe 1987; Beardsley et al. 1987). Operational model output from the COAMPS model at 3 km grid spacing does not clearly show evidence of a strong temperature inversion and/or a wind speed maximum. However, based on previous studies, it is expected that a strong temperature inversion between 1000 and 1500 m is present coinciding with the observed wind speed maxima. An indication of the presence of a strong temperature inversion is provided by in-situ flight data collected in the evening of 11 November 2007. Due to the diurnal variability of the coastal baroclinicity, the coastal jet has a diurnal cycle with the most pronounced and strongest jet close to the coast during mid- to late afternoon (Cui et al. 1998). The relative maximum of the winds in a layer between 1000 and 1500 m that is particularly evident in the northwestern part of the flight domain could be the result of this coastal jet. We also

notice that in legs 4-6, the core of this wind maximum extends to the east and corresponds with the wind maximum found at about 1300 m in the wind profiler data at Fort Ord (Figs. 3 and 4). The TODWL data therefore indicates the presence of a coastal wind maximum that affects the wind fields not only over the water but also further inland. It should be noted that the height of a coastal wind maximum is usually observed closer to the surface at around 500 m. Those observations are made in the summertime when subsidence associated with the Pacific High is stronger and suppresses the depth of the marine boundary layer relative to the late fall period in which our observations were made.

Besides an increase in the winds above the western part of the Salinas Valley that appears to be an extension of the coastal jet, there also is an apparent wind maximum at and above ridge height in the eastern part of the valley which again is most obvious in the northernmost legs. In the next section, we will discuss in more detail these and other flow features observed in the Salinas Valley.

2) SPATIAL WIND STRUCTURE IN THE SALINAS VALLEY

The acceleration of the along-valley flow in the Salinas Valley and the wind maximum in the eastern part of the domain identified previously become more obvious when considering a horizontal cross section of the flows at various heights as shown in Fig. 7. Decreasing in elevation from 2000 m MSL to 1000 m MSL (Figs. 7a-c), the approximate maximum height of the surrounding ridges, winds increase in strength and change

towards northerly and even towards more easterly directions. These northerly and easterly flows are also apparent at lower elevations (800 m and 600 m) while at 400 m MSL the flows are clearly aligned along the valley axis at 400 m MSL. These along valley flows increase in speed with distance upvalley. Also note the relative wind maxima in the western part of the valley at 1500 m (Fig. 7b) and in the eastern part of the valley at 1000 m (Fig. 7c), as mentioned in the previous section.

A closer look at the vertical profiles of wind speed and direction (Fig. 8) reveals a similarity of the wind structure among the different locations along the valley with relative wind speed minima and maxima at around 500 m and 1000-1500 m MSL, respectively. Below 500 m MSL, wind speeds increase and a wind speed maximum is anticipated a few hundred meters above the surface. Wind speeds at 300 m AGL correspond well with the wind speed measurements at the CIMIS surface stations (Fig. 3). Unfortunately, winds below 300 m AGL are difficult to retrieve from airborne lidar since the ground return from the emitted lidar pulse intensifies and overwhelms the aerosol signal near the surface. Efforts to improve the retrieval of these winds are ongoing and involve a better identification of the aerosol signal. These efforts have resulted in some cases in a retrieval of an extra 100 m and indeed show a further increase of the wind speeds near the surface. Note that below 500 m MSL, winds near the coast are westerly (onshore) indicating a clear influence from the sea breeze.

Previous studies of the sea breeze at Monterey Bay have indicated similar depths of the sea breeze (on the order of a few hundred meters) and a similar structure including the

absence of a return flow. Return flows related to the sea-breeze circulation are commonly observed over flat terrain but have not been observed in isolation in the Monterey Bay area in this and in previous studies (Intrieri et al., 1990; Banta et al., 1993). Possible reasons for the absence of a return circulation are the strong synoptic forcing and complex inland topography cause the return flow to be either incorporated into the slope/valley wind system or absorbed into the deep inland boundary layer (Banta et al., 1993). While return flows are absent, a clear minimum in the onshore wind component is present between about 300 and 500 m (Figs 4 and 8) where a return flow may be expected (e.g. Simpson, 1994). The height of the wind speed minimum also corresponds with the typical height of the internal boundary layer along the coast.

In the upvalley direction, the wind speed minima and maxima become less distinct. Valley averaged winds between 350 and 500 m MSL increase with distance up the valley from 1 m s^{-1} to 5 m s^{-1} (Fig. 9a) This increase is still present for winds averaged between 550 and 700 m (Fig. 9b) but disappears above 750 m (Fig. 9c). The vertical wind shear below 1000 m MSL is thus strongest near the coast and at the valley exit. In leg 2, the southernmost leg, the mean vertical wind shear (in speed and direction) and the wind speed maximum at $\sim 1000 \text{ m MSL}$ are smallest which suggests increased vertical mixing within the valley atmosphere. While wind speed variability decreases in the vertical with increasing distance up the valley, the wind speed variability increases across the valley (Fig. 8) especially below ridgetop level with the largest wind speeds usually occurring near the center of the valley (Fig. 7).

An acceleration of the along-valley flow has been observed in other valleys (e.g. Rucker et al. 2008; Whiteman, 2000). Such an acceleration is counterintuitive because one might expect a loss of valley air mass through the tributary valleys and side canyons (Zardi and Whiteman, 2011). An acceleration is normally attributed to the increase of the along-valley pressure gradient due to stronger warming of the valley atmosphere in the upvalley direction. Dynamic channeling of the sea breeze could be another reason for the wind speed increase in the Salinas Valley. As the valley becomes narrower in the upvalley direction, as observed in many valleys including the Salinas Valley, both the warming and dynamical channeling are enhanced, resulting in a wind speed increase.

The along-valley flow divergence should lead to sinking motions in the valley. The vertical wind speeds from the Doppler lidar (Fig. 9 d,e,f) suggests that this is indeed the case with valley-averaged sinking motions of about 0.2 m s^{-1} in flight leg 4. This flight leg is in that part of the along-valley segment where the strongest acceleration in wind speed took place (see also Fig. 7). An estimation of the vertical wind speed needed to maintain mass conservation, assuming that the observed flow divergence ($\Delta u / \Delta x = 4 \text{ ms}^{-1} / 14.5 \text{ km}$) takes place over a depth of 500 m, results in a value of about 0.1 m s^{-1} . Interestingly, while the acceleration of the along-valley flow decreases with height (compare Figs 9 a,b,c), sinking motions are still relatively large for flight leg 4 compared to the other flight legs in all height intervals (Figs 9 d,e,f). A detailed analysis of the corrections that are made to the raw data to account for aircraft motion and attitude revealed no apparent bias that could have caused this particular behavior for flight leg 4. While the magnitude of this sinking motion is comparable to the uncertainty indicated in

Table 1, the spatial coherence of the individual estimates shown in Figure 9 supports the claim that there is organized subsidence to meet the mass divergence within the valley. As pointed out in Appendix A, divergence of the flow of the observed magnitude across or along the flight track is a potential alternative explanation for a vertical velocity of about 0.05 ms^{-1} . The TODWL derived vertical velocity estimates are larger than this value. We furthermore observed in the aerosol structure of leg 4 (Fig. 6f) a decreased height of the aerosol layer top around 1000 m relative to the adjacent flight legs indicating enhanced sinking motions over the Salinas Valley along leg 4. We therefore conclude that the sinking motions are real. Possible reasons for enhanced sinking motions along a particular valley segment in addition to an along-valley flow divergence include a removal of airmass by upslope flows and flow diversion into the side canyons and tributary valleys. It is not clear why this would be more prominent along flight leg 4 than along the other legs within the Salinas Valley. Unfortunately, missing lidar data below 300 m AGL does not allow us to investigate this further.

Another aspect of the flow that was noted before (and seen in Figs. 6,7, and 8) and deserves some attention is the wind speed maximum above the eastern part of the Salinas Valley just above 1000 m MSL and particularly in leg 5. This maximum can partly be explained by a jet-like structure in the cross-valley component of the wind as shown in Fig. 10. The jet-like structure is apparent across the entire valley however, and not just on the eastern part of the valley. It is interesting that despite the approximate alignment of the synoptic flows along the valley, such a prominent cross-valley wind structure is observed. The jet-like structure is present in all legs but becomes less distinct in the

southernmost leg (leg 2). The cross-valley winds also appear to intrude at lower elevations in the Salinas Valley along flight legs 4 and 3. With a near surface (300-600 m MSL) across-valley wind component having the opposite direction, there is some indication of a cross-valley circulation in legs 2-4. The weaker easterly cross valley component and stronger westerly cross-valley component near the surface in flight leg 2 compared to the other legs cause aerosols in flight leg 2 to appear concentrated on the eastern part of the Salinas Valley as shown in Fig. 6j. Convection and upslope flows on the heated eastern sidewall probably also play a factor. As noted before, the current inability of retrieving winds from Doppler lidar data below 300 m AGL does not allow an investigation of upslope flows.

The origin of the cross-valley winds is unclear at this point. Because of the presence of these winds in all flight legs and at elevations above the surrounding mountains, it appears that the flows are generated at a regional scale. Previous studies (e.g. Darby et al. 2002) have indicated the influence of the Sierra Nevada mountains on flows observed over the investigation area. However, during daytime, the effect would be a stronger westerly component of the flow, not an easterly component. While the easterly flows appear to originate at a regional scale, local scale effects such as those caused by the aforementioned return flows associated with upslope and upcanyon winds along the eastern sidewall cannot be excluded. Another possibility includes a dynamical effect related to the specific shape of the valley sidewall and particularly at its exit (e.g. the pass towards San Joaquin Valley). These type of effects have been observed in smaller scale valleys (Weigel et al., 2006).

Two identical flight legs that were flown at approximately 1200 and 1400 LST allow an investigation of the temporal wind and SNR changes in the Salinas Valley. The first flight leg on November 12 took place shortly after a wind reversal at the surface from downvalley to upvalley direction (Fig. 3). Upvalley flows are present over the entire depth of the valley (Fig. 11a). Comparing leg 1 with leg 3, valley-averaged winds have increased in speed from approximately 4 to 7 m s⁻¹. The maximum in the cross-valley wind component that was evident in Fig. 10 was also present during flight leg 1 but at altitudes that were about 500 m higher than during flight leg 3 (not shown). Both at 1200 and 1400 LST, winds are generally weakest in the eastern part of the valley atmosphere. If we assume that upslope flows are present in lowest 300 m AGL over the sunlit slope, these flows could interact with the upvalley flows and reduce the wind speed. Some evidence of the presence of upslope flows is given by the asymmetric distribution of SNR with higher SNR over the heated eastern sidewall at both times.

4. Conclusions and outlook

An airborne Doppler lidar was operated in the Salinas Valley and surrounding mountains as part of a field experiment in November 2007. While the major goal of this experiment was related to the support of the development of a space borne Doppler lidar, the flights provided detailed information on the complex spatial structure of afternoon flows in the coastal mountains of California.

The Doppler lidar, flown at an altitude of 3 km MSL and operating in a twelve point, 20 degree off nadir step stare conical scanning mode, generated data sets from which horizontal and vertical velocities were derived with 50 m vertical resolution and an estimated accuracy of the horizontal wind components of $< 30 \text{ cm s}^{-1}$. Since the Doppler lidar uses backscatter from aerosols for its signal, the mapping of aerosol layers and visualization of organized aerosol laden structures yielded further information on flow stratification and dynamic processes.

To begin the mining of the Doppler lidar data sets, data from a flight on 12 November 2007 were used to develop methods of analyses and interpretive skills for a relatively new perspective on spatial wind structure in a valley. A thermally-driven sea breeze and upvalley flow developed around noon on this clear day on which northwesterly synoptic flows prevailed. The sea breeze onset at noon is delayed compared to previous studies in the area that were mostly conducted in summer and reported onset times about three hours earlier. The sea breeze started as a westerly flow near the coast and turned into northwesterly upvalley flows in the Salinas Valley increasing in speed from 2 to 6 m s^{-1} and in depth from about 400 to 600 m. Consistent with previous studies, no return flow was observed in the Salinas Valley. While the winds were predominantly from northwesterly directions, a cross-valley component of the wind with a jet-like structure was present at heights between 1000 and 1500 m MSL and mostly on the eastern side of the valley. The origin of this cross-valley flow and its interaction with the upvalley flow within the Salinas Valley remain unclear at the moment. Another distinct wind maximum was found between 1000 and 1500 m MSL in the western part of the Salinas Valley. This

wind maximum appears as an easterly extension of the wind maximum that was observed over the coastal waters.

The acceleration of the along valley flow resulted in vertical sinking motions on the order of 0.2 m s^{-1} , a value derived from both mass continuity considerations and the Doppler lidar data. Furthermore, while the vertical wind shear decreased in the upvalley direction, the horizontal wind shear increased with some indication of the presence of a cross-valley circulation. These processes resulted in an asymmetric distribution of the Doppler lidar mapped aerosols across the Salinas Valley with an elevated concentration of aerosols over the eastern sidewall.

In summary, the general picture of the flows in the Salinas Valley consists of a sea breeze transitioning into an upvalley flow which accelerates in the upvalley direction. Influences from synoptic flows, coastal jets, cross-valley flows, and slope flows cause a large spatial variability in the upvalley flow with wind speed maxima observed at and above ridge top level and sinking motions within the valley. While certain aspects of the flow features in the Salinas Valley have been observed in other valleys before, such as a transition from sea-breeze to upvalley flow, an acceleration of the upvalley flow and the presence of cross-valley flows, the detailed measurements of the three-dimensional winds are unprecedented and provide information of flows from the local to the synoptic scale.

It is important to emphasize that the insight gained from airborne Doppler lidar is greater when coordinated with ground-based instrumentation to more completely characterize

meteorological conditions and reduce uncertainties in the interpretation of the airborne lidar data. A radar wind profiler located near Monterey Bay and a network of surface stations along the Salinas Valley provided the temporal information that aided the interpretation of the Doppler lidar data. An investigation of the stability structure and its role in the observed flows and aerosol structure was hampered by the lack of temperature structure data and observations along the valley sidewalls. Since winds could not be faithfully retrieved from Doppler lidar below 300 m AGL, an investigation of the sea breeze and upvalley wind maxima that occur in this layer and the upslope flows over the sidewalls was not possible. However, efforts are ongoing to improve the retrieval of winds in this layer near the surface (Godwin 2011).

This paper demonstrated the ability of airborne Doppler lidar data to produce a detailed and comprehensive three-dimensional picture of atmospheric features and structure. The measurements allow an investigation of the integrated effects of multi-scale flows on atmospheric transport, and provide an important data set for the evaluation of numerical models. This evaluation and the investigation of underlying processes explaining the observed flows will be performed with a mesoscale numerical model in a separate paper. Such an investigation will hopefully clarify the mechanisms underlying the various wind maxima in and above the Salinas Valley, the along-valley acceleration of the winds, the sinking motions, and the asymmetric aerosol structure.

Due to recent developments in electronics at eyesafe wavelengths and lower cost of off-the-shelf components, it is expected that more airborne Doppler lidar systems will

become available in the near future. Lighter weight hardware components will also facilitate the mounting of Doppler lidar systems on aircraft with smaller payload resulting in cheaper and therefore potentially more frequent and widespread operations. Research on the processing and interpretation of airborne Doppler lidar data, and the combined use of the data with numerical model output is therefore important.

Acknowledgments

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Appendix A: Accuracy of the airborne Doppler lidar measurements

Table 1 in section 2.1 summarizes the TODWL instrument and the estimated accuracies. These accuracies are considered estimates since it is challenging to find truth data sets for the step-stare sampling pattern from a moving platform. We rely in large part on work by Frehlich (2001,2004) and Frehlich and Sherman (2005) who developed much of the signal processing and error characterization for ground-based coherent Doppler lidar observations. LOS accuracies are dependent upon many factors such as aerosol distributions within the pulse length, wind turbulence within the illuminated volume of the accumulated pulses (for a single LOS velocity estimate), aircraft motions, wind field characteristics, and signal processing related uncertainties (e.g. spectral resolution). Consequently, there is no single accuracy value that applies to all situations. The values reported here are based upon data taken during the period and location of study and may change depending upon sampling conditions. In the following, we briefly address the effect of aircraft motions and wind field variability on the accuracy of airborne DWL measurements. In our considerations of the accuracy, we use a 12 point step-stare conical scan with a 20 degree off nadir angle which was used extensively with the TODWL during the 12 November 2007 flights.

Effect of aircraft motion

To achieve the airborne wind lidar measurement accuracies claimed in Table 1, corrections must be made for the frequency jitter of the laser, physical misalignments of the beam relative to the aircraft's major axes, beam wandering within the scanner steering optics, and aircraft attitude changes during scanning. As part of our data processing

routines, we developed an algorithm to use ground returns of LOS velocity as the aircraft flew reasonably straight and level and the beam was conically scanned downward. Using several scans composed of twelve stares with 30 degree azimuth spacing, physical offsets (from those reported by the GPS/INS) in roll, pitch and yaw were quantified as well as any beam direction variation as a function of scanner azimuth. The net result of this calibration process was a LOS RMSE (Root Mean Squared Error) less than $.07 \text{ ms}^{-1}$.

Effect of wind field variability

We address two basic sources of wind variability that can contribute to errors in the representation of the wind vector components derived from a sine fit to conically scanned LOSs: (a) random deviations from the mean flow on scales an order of magnitude smaller than the volume dimensions; and (b) a divergent wind field on scales of and exceeding the full twelve point step-stare scan volume dimensions.

Rather than simulating atmospheric variability on the many scales that would influence the LOS velocity estimate, we calculated the deviations from a sine fit (12 point) found in real TODWL data for all data points in the 274 soundings on 12 November 2007. The histogram of these deviations is shown in Figure 12. The most frequent deviation is $\sim .1 \text{ ms}^{-1}$. Since it is possible for these deviations to be distributed within the sine fit in many different sequences, we ran a Monte Carlo simulation by using a weighted random number generator to select random deviations from homogeneous flow to apply to each of the 12 azimuth angles. The weights were derived from the real distributions in Figure A1. After 10,000 simulated deviation sequences, the average error for both the u and v

components was calculated to be $.32 \text{ ms}^{-1}$. Assuming that the “zero offset” of the sine fit represents the average contribution of vertical velocities (at the 12 locations), the error in the w estimate was calculated to be $.08 \text{ ms}^{-1}$.

TODWL measurements between 350 and 500 meters along the Salinas Valley floor were used to provide a realistic value of divergence within the scan dimensions. As was shown in Fig. 9, the along valley wind speed difference between legs 3 and 5, which were flown parallel to each other at a distance of 14.5 km, was approximately 4 ms^{-1} (divergence = $2.8 \times 10^{-4} \text{ s}^{-1}$). In the height interval between 350 and 500 m, the conical scanning width is $\sim 1.9 \text{ km}$. Therefore the velocity difference across the scanning area is approximately $.5 \text{ m/s}$. Using this value, two different scenarios were evaluated to test the effects of the background divergence in the horizontal wind field on individual wind components. The two scenarios consisted of a wind speed gradient along and across the flight path (Fig. 13). The along flight path wind speed began at 1 ms^{-1} at an aircraft relative azimuth angle of 0 degrees and concluded with the wind speed at 1.5 ms^{-1} at an aircraft relative azimuth angle of 330 degrees (Fig. 13a). For this simulation we assume there is no real average vertical motion over the scan domain. The resulting deviations from the real u , v , and w wind components were: $-.02 \text{ ms}^{-1}$, $.03 \text{ ms}^{-1}$, and $-.06 \text{ ms}^{-1}$, respectively. The total wind speed was calculated to be 1.23 ms^{-1} compared to the correct value of 1.25 ms^{-1} . The cross flight path wind speed also began at 1 ms^{-1} and concluded at 1.5 ms^{-1} (Fig. 13b). For the cross flight path gradient, the resulting deviations from the real u , v , and w wind components were: 0.00 ms^{-1} , 0.02 ms^{-1} , and 0.04 ms^{-1} . The wind speed was determined to be 1.25 ms^{-1} .

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Wavelength	2 microns (eye-safe)
Energy per pulse	2 milli-Joule
Pulse length (temporal; spatial)	320 ns; 48 meters (full-width-half-max)
Pulse Repetition Frequency	100 Hz (500 Hz maximum capability)
Data processing range gate	96 meters
Profile range gate resolution	50 meters (this results in some photon sharing between adjacent data processing gates)
LOS measurement accuracy ¹	< 0.2 m/s
Wind component accuracy ²	u and v component < 0.3 m s ⁻¹ w component ~0.1 m s ⁻¹

Table 1: Characteristics of the Twin Otter Doppler Wind Lidar (TODWL) system.

¹ The value used in this table has been estimated assuming an accumulation of 100 shots at a fixed dwell angle and a ground return based calibration to account for aircraft motion (see Appendix A).

² These values are estimated based on considerations using 12 LOS perspectives in a 20 degree off nadir scan. These values are dependent upon the specific atmospheric conditions and various other factors (see Appendix A).

Leg	# conical scans	Mean Profile separation (km)	Total Distance (km)	Begin Time (LST)	End Time (LST)
1	45	1.48	66.73	1216	1231
2	38	1.58	59.95	1400	1410
3	50	1.31	65.54	1419	1434
4	50	1.28	64.11	1441	1451
5	58	1.08	62.86	1459	1514
6	44	1.32	58.28	1520	1530

Table 2: Characteristics of the six flight legs flown on 12 November 2007.

Station ID	Station Name	Latitude	Longitude	Elevation (m MSL)	Distance from coastline (km)
CS	Castroville	36.77	-121.77	3	2.5
SN	Salinas North	36.72	-121.69	18	10.5
SS	Salinas South	36.61	-121.53	36	26.5
AR	Arroyo Seco	36.36	-121.29	70	58.5
KC	King City	36.12	-121.08	162	90.5
WP	Fort Ord	36.70	-121.76	48	4.5

Table 3: Name and location of the five surface weather stations from the CIMIS network and the 915 MHz radar wind profiler.

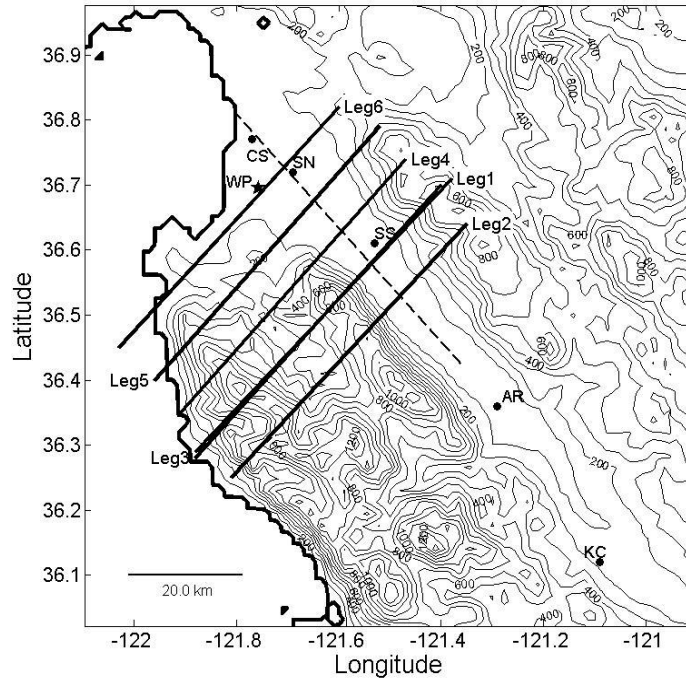


Figure 1: Map of the investigation area including the Monterey Bay, Salinas Valley and surrounding mountains. Contour lines show height in meters. The flight legs are shown with the solid black lines with the flight leg number indicated at the starting point of each flight leg. The dashed line indicates the along-valley direction ($\sim 140^\circ$ off north). Locations of the CIMIS surface stations (black dots) and wind profiler (black star), listed in Table 3, are also shown.

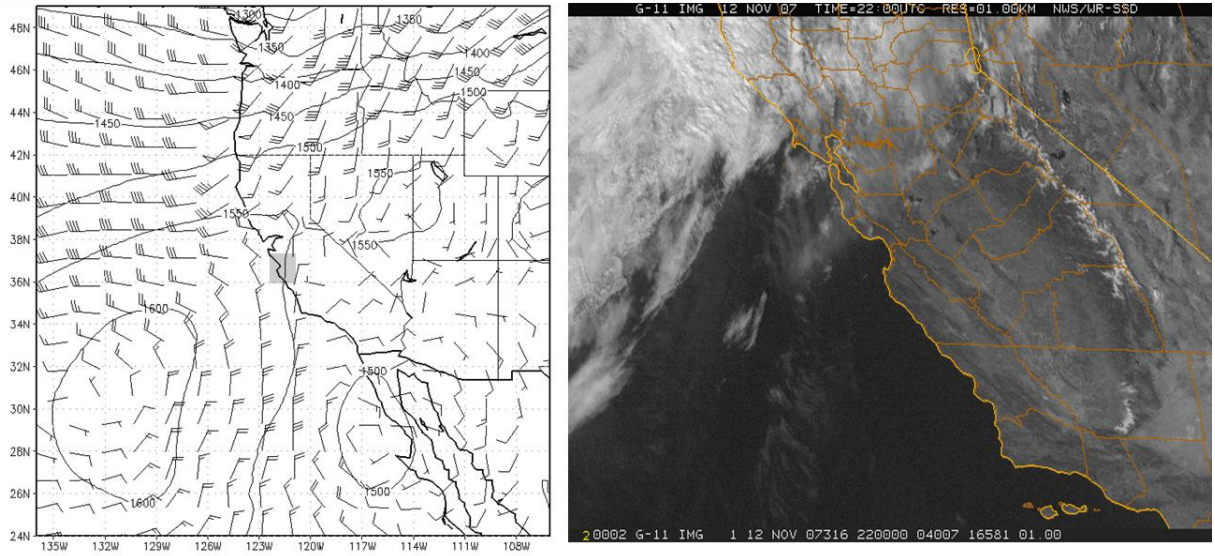


Figure 2: (a) Geopotential height (m) and wind at 850 mb on 12 November 2007 at 2100 UTC (LST = UTC – 8 h). A full barb denotes 10 knots. The area from Figure 1 is indicated with the transparent box (b) corresponding GOES satellite image (visible channel) for 12 November 2200 UTC. Note that the scale in the two figures is different. The investigation area is approximately in the center of both figures.

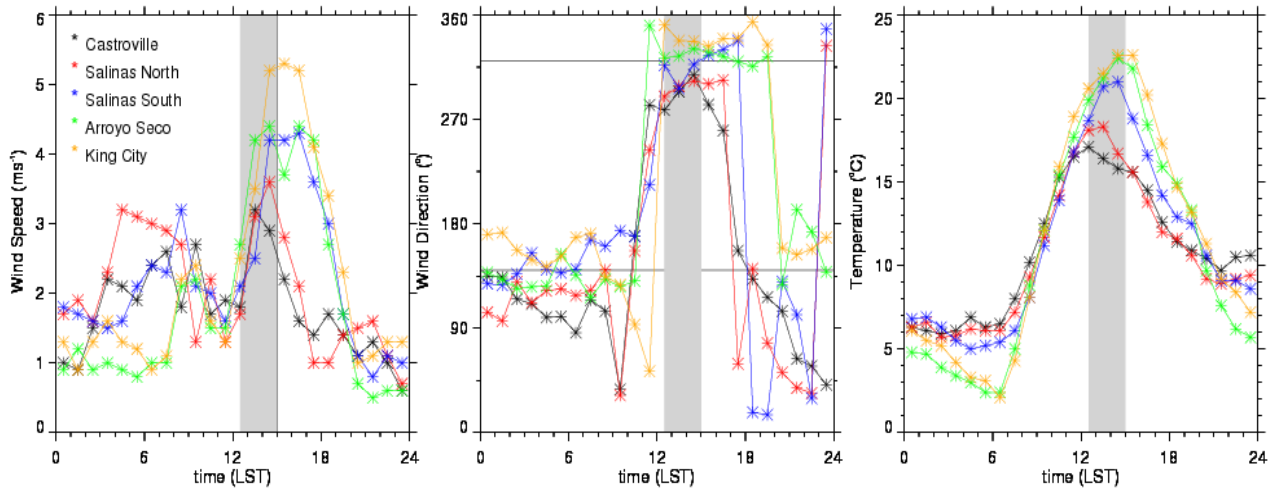


Figure 3: : Diurnal evolution of (a) wind speed, (b) wind direction, and (c) temperature at the five CIMIS network surface stations on 12 November 2007. Wind and temperature are measured at 2 m and 1.5 m AGL, respectively. The grey rectangle in the figures denotes the approximate time period during which the flights took place, the horizontal lines in (b) denote the upvalley wind direction (320°) and downvalley wind direction (140°).

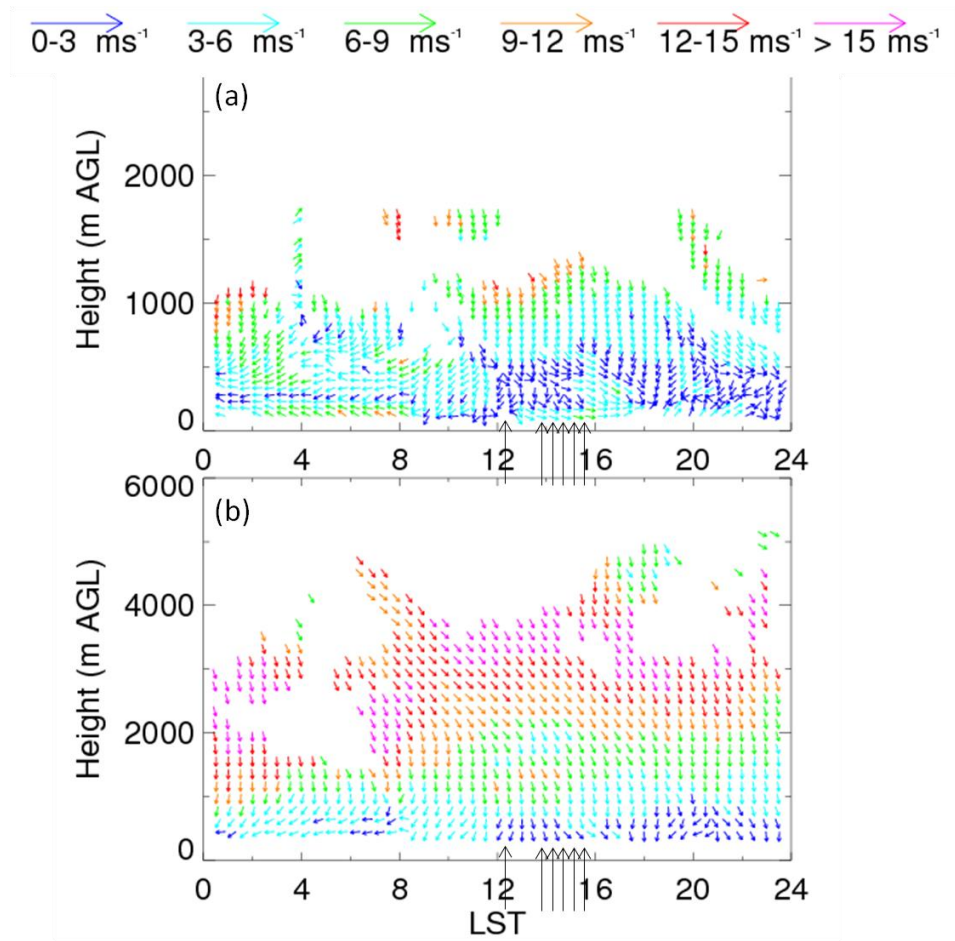


Figure 4: Time height cross section of wind speed and direction at Fort Ord from (a) coarse mode and (b) fine mode wind profiler measurements. The vertical arrows indicate the time of TODWL flight legs 1-6.

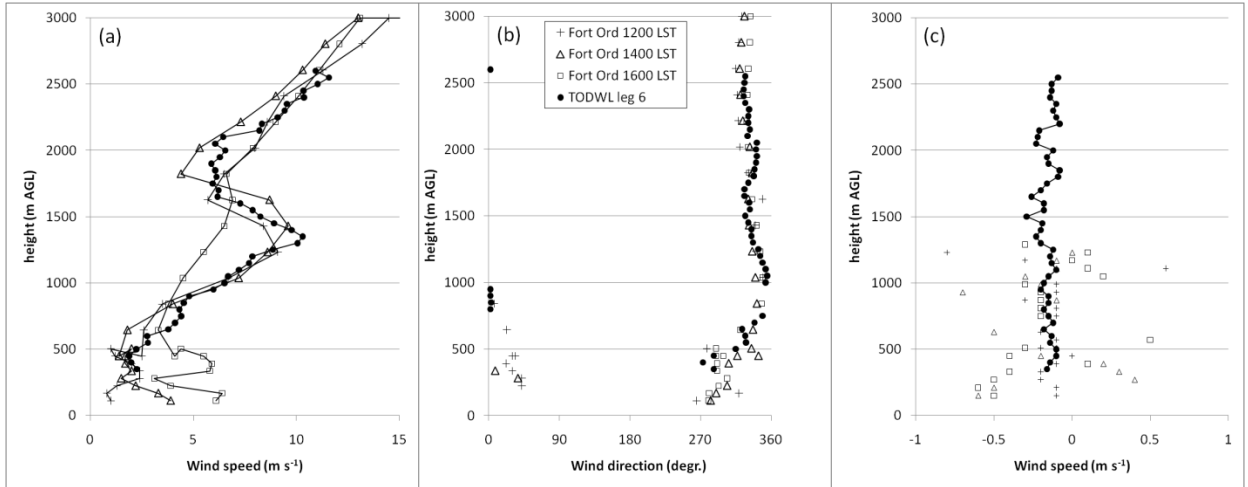


Figure 5: Vertical profiles of (a) horizontal wind speed and (b) wind direction and (c) vertical wind speed from wind profiler measurements at Fort Ord at 1200, 1400, and 1600 LST, and from TODWL near the location of Fort Ord (36.70 N, -121.76 W) at 1529 LST during flight leg 6. Wind profiler measurements are from the fine mode under 500 m and from the coarse mode above 500 m. Negative vertical wind speeds denote downward motions.

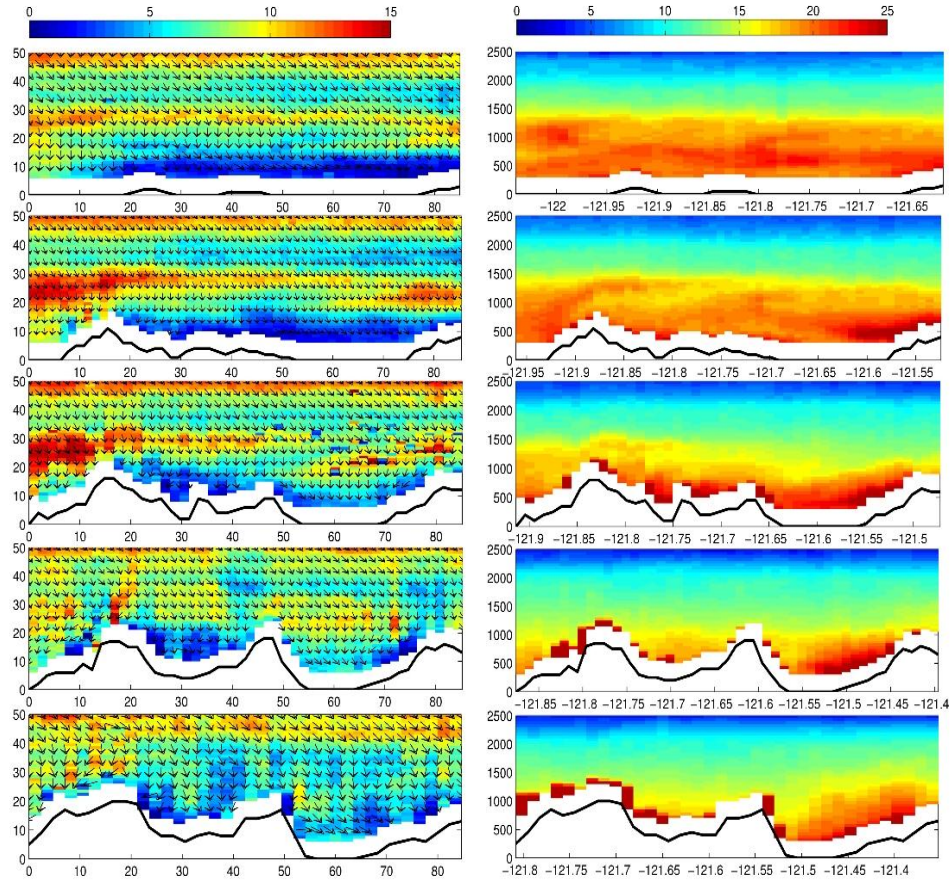


Figure 6: Vertical cross section of the TODWL derived horizontal wind and SNR for (a,b) flight leg 6, (c,d) flight leg 5, (e,f) flight leg 4, (g,h) flight leg 3, and (i,j) flight leg 2.

Wind vectors are only shown at every 200 m height level.

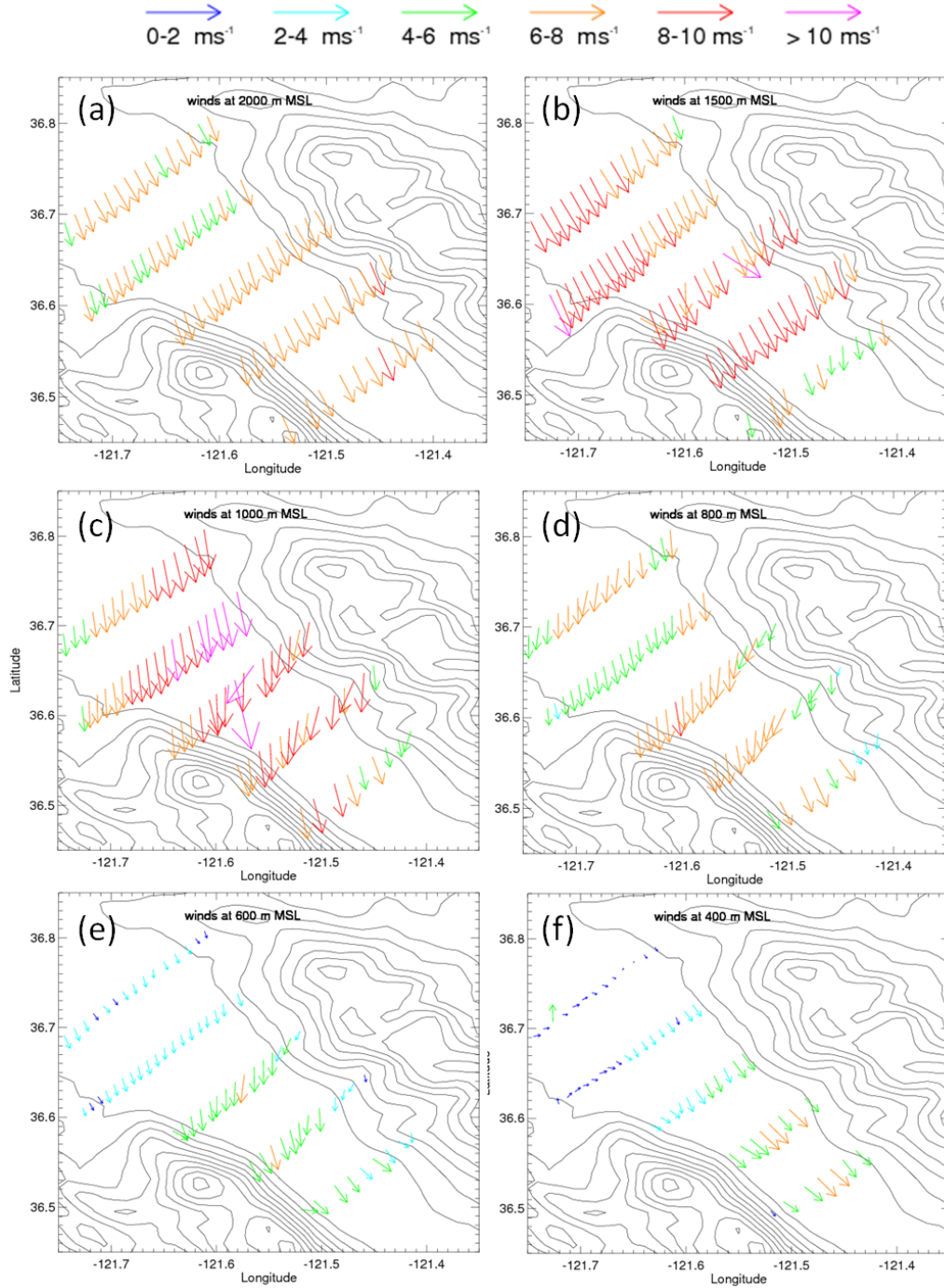


Figure 7: Horizontal cross sections of the TODWL derived winds over the Salinas Valley at (a) 2000 m, (b) 1500 m, (c) 1000 m, (d) 800 m, (e) 600 m, and (f) 400 m MSL.

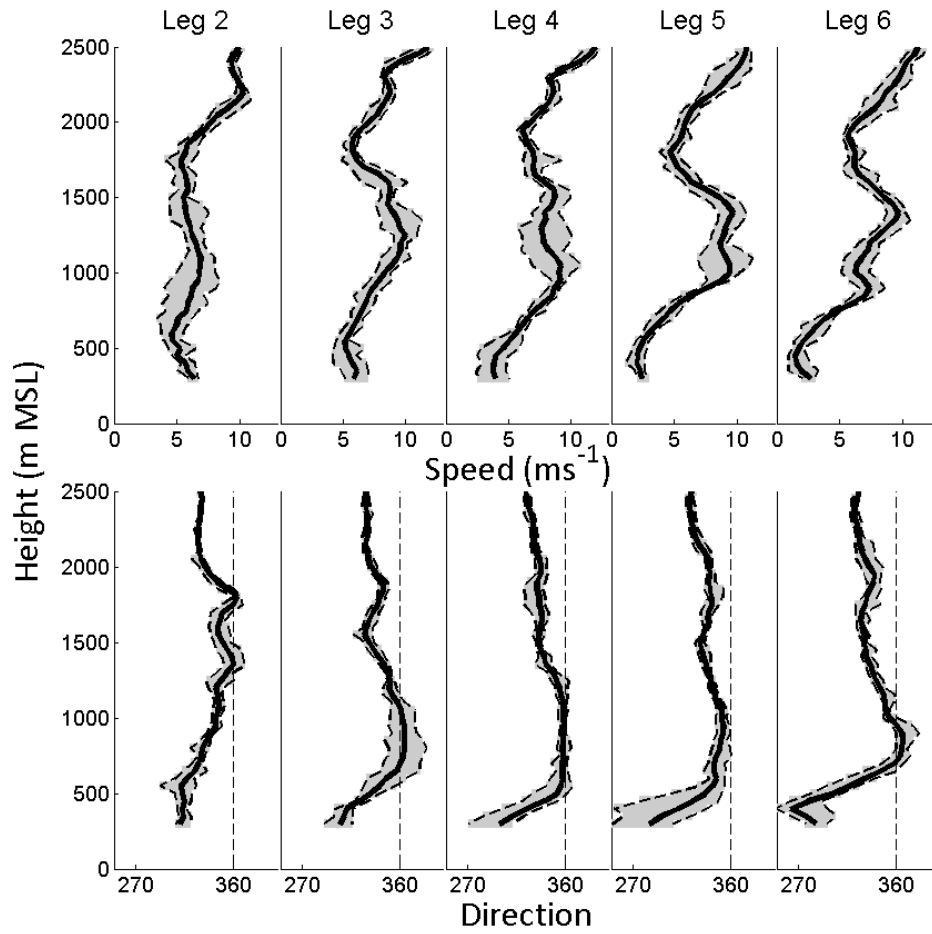


Figure 8: Vertical profiles of TODWL derived wind speed and direction averaged over the flat valley floor section in the Salinas Valley for flight legs 2 to 6 . The grey shading bounded by the dashed lines indicates the range of wind speed and direction among the profiles.

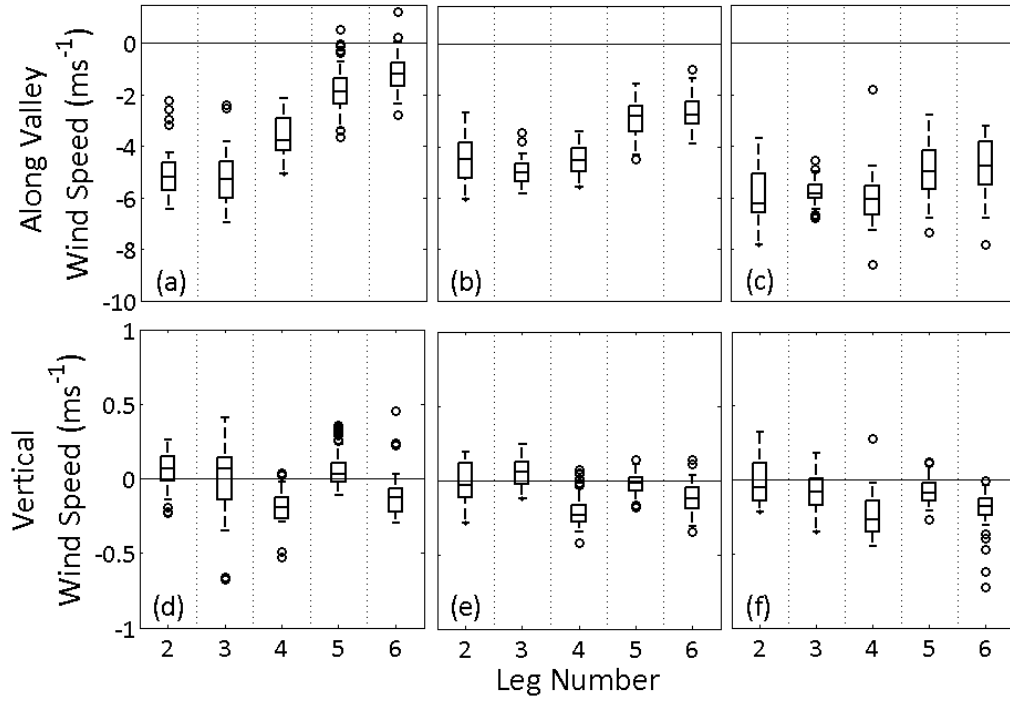


Figure 9: Box and whisker plots of the along-valley and vertical wind speed at heights between 350 and 500m (a,d), 550 and 700 m (b,e) and 750 and 900 m (c,f). Only wind data obtained over the flat Salinas valley floor are used. The number above each box in (b) is the number of data points used for the calculation of the statistics for a particular leg. Negative vertical wind speeds denote sinking motions.

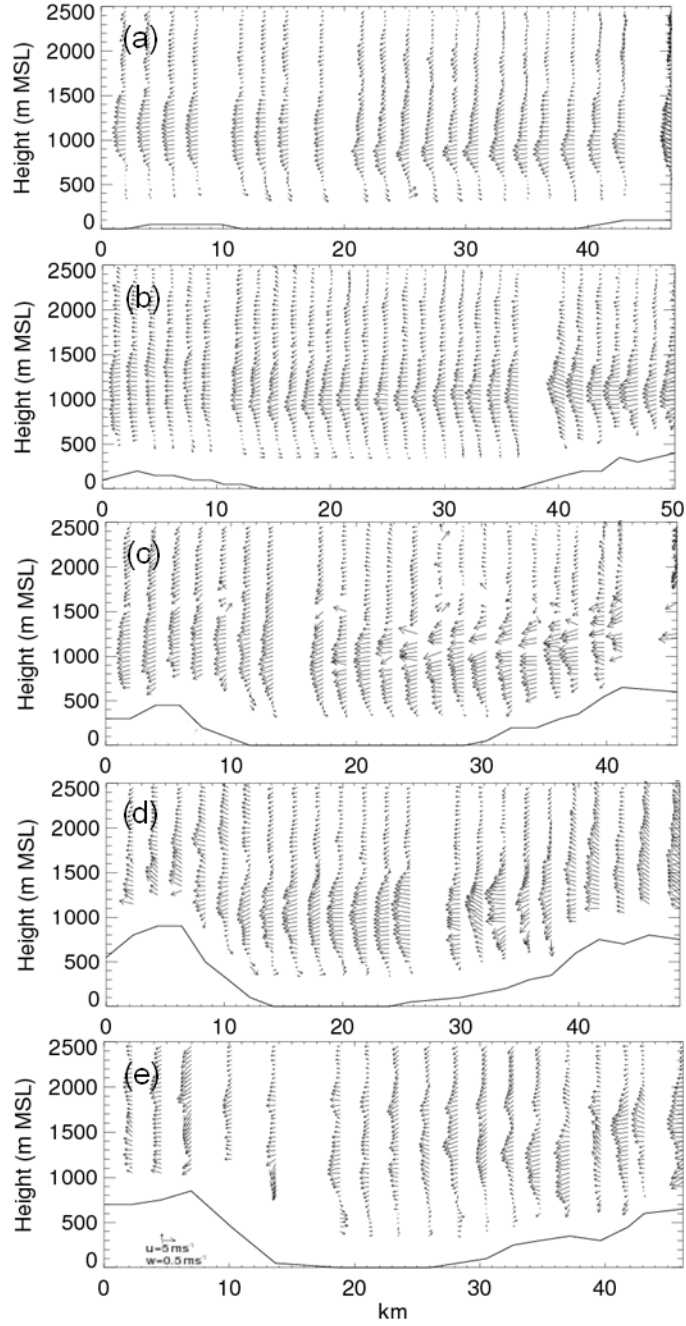


Figure 10: Cross-valley wind component of the wind for (a) flight leg 6, (b) flight leg 5, (c) flight leg 4, (d) flight leg 3, and (e) flight leg 2. The cross-valley direction is defined perpendicular to the dashed line in Fig. 1. Only a subsection of the lidar cross section in Fig.6 is shown.

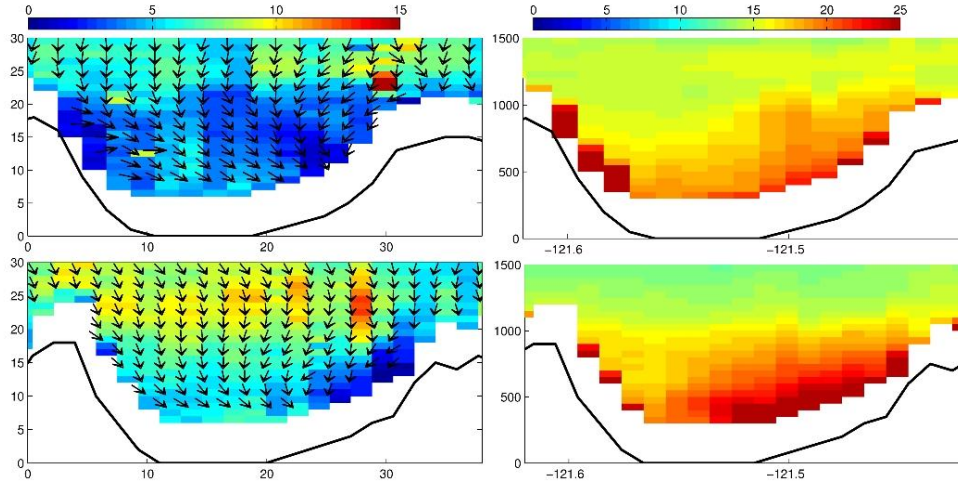


Figure 11: Vertical cross section of the TODWL derived wind and SNR for flight leg 1 (a,b) and flight leg 3 (c,d) at identical locations but flows two hours apart from each other (~1200 LST vs. ~1400 LST).

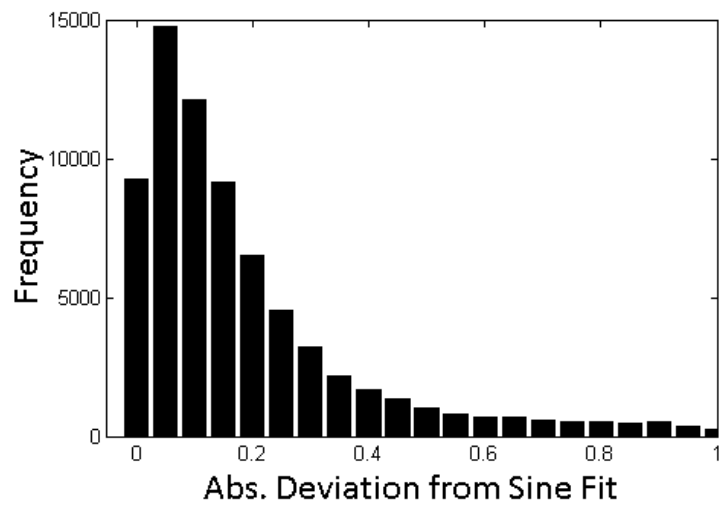


Figure 12: Histogram of all observed radial velocity absolute deviations from the fitted sine wave.

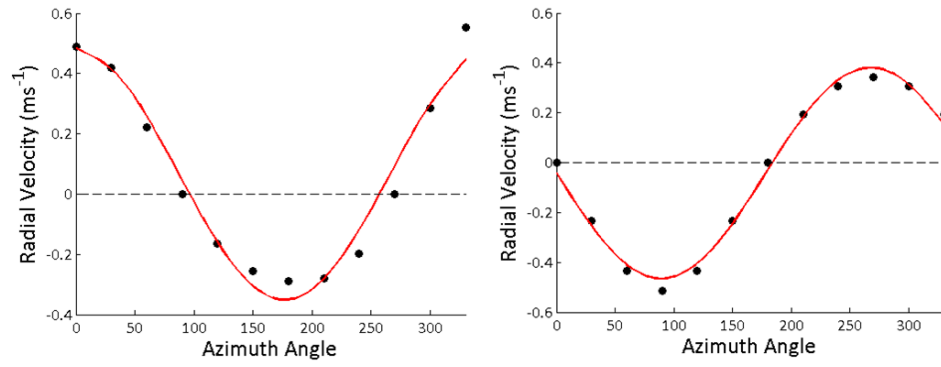


Figure 13: Simulated LOS velocity for a horizontal wind divergence of $2.0 \times 10^{-4} \text{ s}^{-1}$ along the flight path (a) and across the flight path (b). A 20 degree off nadir scan angle is used to convert from u,v,w winds to radial velocity.