

**Terrain-trapped Airflows and Orographic Precipitation
along the Coast of Northern California**

by

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Terrain-trapped Airflows and Orographic Precipitation along the Coast of Northern California

Thesis directed by Dr. David Kingsmill

While several studies have documented kinematic and precipitation structures associated with orographic effects in large-scale mountains (e.g. altitudes > 1.0 km MSL), surveys on small-scale mountains are still relatively rare. Given their lower altitude, these mountain are usually exposed to rain (instead of snow) during the approaching of baroclinic waves and concomitant warm and cold fronts, thus they are prone to flash floods and hydrological disasters during heavy rain episodes.

One factor that influence kinematic and precipitation structures over mountain ranges is associated with airflows induced by the terrain itself, such as during low-level blocking and gap flow episodes. Force balance make these airflows move relatively parallel and in close proximity to the terrain, thus they can be categorized as terrain-trapped airflows (TTAs). TTAs can cause the lifting of incoming synoptic airflows upstream the terrains foothills, initiating and enhancing precipitation well before it would be observed by upslope forcing over the terrain. TTAs and orographic precipitation forcing along the small-scale coastal mountains of northern California have been studied from a 1-dimensional perspective; yet, details of the TTAs 3-dimensional kinematic and precipitation structure, especially in the lowest 500-m MSL, has not being addressed.

In this doctoral thesis I examine physical characteristics and impacts of TTAs on orographic precipitation along the coastal mountains of northern California using a 13-winter season dataset. Selected case studies are employed to document in detail 3-dimensional kinematic and precipitation structures associated with pre-cold-frontal low-level jets (LLJs) and TTAs. The main observational asset is a ground-based X-band dual-polarization scanning Doppler radar located at the coast. Results show that TTAs are normally present during winter time, although with variable duration and offshore extension. In average, they account for 20% and 10% of the long-term rainfall along the coast and over the coastal mountains, respectively. Doppler radar observations depict the lifting

of LLJs offshore forced by TTAs, creating an area of enhanced precipitation about 20 km from the coast. Forcing of the TTA seems to be most commonly associated with gap flows coming from Petaluma Gap and, to a shorter extent, with low-level blocking.

Dedication

I dedicate this work to all of those who have not only illuminated my path with knowledge, but also with life lessons, starting for my mother, Ana. Perhaps you were unaware that many of your teachings would become fundamental for me to reach this achievement. Patricia and Marco are also part of this effort, as well as the rest of my uncles and aunts. Guillermo, Cecilia, and so many other professors were there through primary, secondary, and high school. Without their wisdom and love for teaching I might have not pursued this journey. Finally, I dedicate this work to Luzma, my wife, who joined me in the last part of this journey and gave me her kind support and love. And to Lupe, my daughter, who changed my life.

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List of Abbreviations

AR atmospheric river

BBY Bodega Bay

CFDD contoured frequency by distance diagram

CZD Cazadero

FRS Fort Ross

IVT integrated water vapor transport

IWT column-integrated water vapor

LLJ low-level jet

PPI plane-position indicator

RASS radio-acoustic sounding system

RHI range-height indicator

TTA terrain-trapped airflow

VAD velocity-azimuth display

Chapter 1

Introduction

The interaction between synoptic airflows and mountains has received attention from a theoretical standpoint since at least 1940's (e.g., Queney 1948). Later advance in remote sensing technology, in particular weather Doppler radars, allowed further research from an observational standpoint, addressing questions about kinematic and precipitation structures produced by this interaction (e.g., Browning and Wexler 1968, Browning et al. 1974).

As several theoretical and observational studies show (thorough reviews in Roe 2005, Houze 2012), mountains can modify synoptic-scale precipitation patterns by different mechanisms. One of this mechanism, the upslope flow forcing, entails a weakly stratified or neutral atmosphere and cross-barrier water vapor flux surmounting along the windward side of the mountain, which creates condensation and eventually precipitation. The relationship between synoptic-scale precipitation patterns and mountains (i.e. orographic precipitation, Colle et al. 2013) has been the focus of many studies due to its role in the hydrological cycle (e.g., Trenberth et al. 2003) and because population vulnerability during extreme events (e.g., Neiman et al. 2009).

In the west coast of the United States, winter season precipitation is generally forced by synoptic-scale systems and frequently related with the passage of baroclinic waves and concomitant warm and cold fronts. In particular, the low-level jet (LLJ) embedded in the warm sector of baroclinic waves (ahead of cold fronts) can have significant impact on orographic precipitation along the coast (e.g., Neiman et al. 2002). An additional element added in recent years is an important horizontal flux of water vapor associated with the LLJ, known as Atmospheric River

(e.g., Ralph et al. 2006). Strong horizontal water vapor flux impinging in mountain barriers can produce copious precipitation through the upslope flow mechanism. However, studies also show that this mechanism can be modulated to a large extent by terrain-trapped airflows (TTAs) such as blocked flows (e.g., Peterson et al. 1991, Sinclair et al. 1997).

While orographic precipitation studies addressing kinematic and precipitation structures along the west coast have mainly focused on relatively large-scale barriers (e.g. ridge tops > 1000-m MSL) and lowest analysis levels above 500-m MSL (e.g, Yu and Smull 2000, James and Houze 2005, Houze and Medina 2005, Medina et al. 2007, Yuter et al. 2011), small-scale barriers and analyses below 500-m MSL have received comparatively less attention. Some reasons are the lack of low-level observations from the operational radar network and difficulties to collect data from unperturbed upstream conditions offshore. As a result, in this study I analyze observations from a ground-based X-band dual polarization scanning Doppler radar (X-pol) located near sea level along the coast of norther California to better understand the orographic precipitation forcing associated with TTA regimes in the lowest 500-m MSL and document characteristics of kinematic and precipitation structures.

In Chapter 2, X-pol is employed in a case study to document the interaction between the pre-cold-frontal LLJ and a TTA along the coast of northern California. Documentation of kinematic and precipitation structures allow to describe the TTA influence on orographic precipitation. This study was published in *Monthly Weather Review* (Valenzuela and Kingsmill 2015, Orographic Precipitation Forcing along the Coast of Northern California during a Landfalling Winter Storm).

In Chapter 3, a 13-season wind profiler and surface meteorological observation dataset is employed to characterize TTA regimes along the coast of northern California. In this study, the long-term impact of TTAs on orographic rainfall is quantified and an objective identification of TTAs is developed. This chapter is in preparation to be submitted to *Monthly Weather Review* (Valenzuela and Kingsmill 2017a, Terrain-trapped airflows and orographic rainfall along the coast of northern California. Part I: Kinematic characterization using a wind profiler radar).

In Chapter 4, we employ the objective TTA identification method of Valenzuela and Kingsmill

(2017a) along with X-pol observations to document the kinematic and precipitation structure associated with TTA regimes in 7 storms impacting northern California. Discussion about mean and inter-storm structure variability is provided. This chapter is in preparation to be submitted to *Monthly Weather Review* (Valenzuela and Kingsmill 2017b, Terrain-trapped airflows and orographic rainfall along the coast of northern California. Part II: Structures observed by ground-based scanning Doppler radar).

The overall significance of the present study is thoroughly document and provide context of the importance of TTAs associated with orographic barriers on the order of 500-m MSL altitude. In this way, it is shown that TTAs can effectively divert incoming LLJs upward, forming an enhanced precipitation area along the LLJ/TTA interface well upstream the orographic barrier (Chapter 2). Also, TTA regimes are best identified by a wind direction $< 150^\circ$ in the average 0-500-m MSL layer during at least 2 hours and have a significant contribution to orographic rainfall along the coast, or equivalently, along the mountain foothills (Chapter 3). Finally, although kinematic structures associated with TTA regimes are consistent from storm-to-storm, precipitation structures present large variability, probably associated with variation in the synoptic conditions (Chapter 4).

Chapter 2

Orographic Precipitation Forcing along the Coast of Northern California during a Landfalling Winter Storm

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2.1 Introduction

During the winter season, extratropical cyclones and cold fronts often develop over the northern Pacific Ocean and move toward the West Coast of the United States. A particular airflow embedded within these cyclones, known as the warm conveyor belt, is associated with significant precipitation events along the West Coast mountain ranges. Warm conveyor belts extend over a broad area ahead of surface cold fronts and are associated with poleward heat transport (latent and sensible heat) and isentropic ascent over the surface warm front (Harrold 1973, Schultz 2001, Eckhardt et al. 2004). In addition, they usually present a horizontal wind speed profile with a local maximum at an average altitude of ~ 1 km MSL referred to as the low-level jet (LLJ, Neiman et al. 2002, Ralph et al. 2005). More recently, a narrow subset of the warm conveyor belt coincident with the LLJ core and associated with large horizontal water vapor transport has been documented. This structure is now commonly referred to as an atmospheric river (AR, Zhu and Newell 1994, Ralph et al. 2004; 2005). Winter-season ARs offshore of the U.S. west coast are on average oriented with their long axis from southwest to northeast (Neiman et al. 2008, Wick et al. 2013).

Many west coast mountain ranges can be approximated as nearly two-dimensional barriers. Those in northern California, such as the Sierra Nevada and nearby coastal ranges, have a generally northwest to southeast orientation. Such a topographic characteristic allows the typical AR to impact the mountains with a near perpendicular orientation, favoring the cross-barrier ascent of moist, statically-neutral air (Neiman et al. 2002, Ralph et al. 2005). However, this relatively simple orographic precipitation forcing can become more complicated when a low-level terrain-trapped airflow exists. A terrain-trapped airflow (henceforth TTA) is defined as a relatively narrow air mass consistently flowing in close proximity and approximately parallel to an orographic barrier. In the context of Pacific winter storms making landfall along the west coast, TTAs flow poleward along the western side of coastal mountain ranges.

One mechanism often attributed to TTA formation is low-level blocking (Smith 1979). In this scenario, stably-stratified air parcels decelerating and abutting the windward slope of a mountain are unable to surmount the barrier and instead turn left (right) in the Northern (Southern) Hemisphere as a result of an increasing along-barrier pressure gradient and weakened Coriolis force. Theoretical studies (e.g., Pierrehumbert and Wyman 1985) have shown that an airflow impinging on a two-dimensional barrier can be described by the non-dimensional Froude number (Fr):

$$Fr = \frac{U}{Nh} \quad (2.1)$$

where N represents the Brunt-Visla frequency within the stable air, h the barrier height, and U the cross-barrier wind speed upstream of the stable air and orthogonal to the terrain. Fr values between 0 and 1 indicate favorable conditions for the formation of low level blocking.

TTAs can also be induced by gap flows. Gap flows are usually produced in mountainous regions when a relatively cold continental air mass crosses through gaps (i.e. narrow topographic depressions) as a response to an along-gap pressure gradient associated with approaching synoptic-scale disturbances (e.g., Lackmann and Overland 1989, Neiman et al. 2006, Mayr et al. 2007). Mass et al. (1995; henceforth M95) derived an analytical expression for gap flows that includes

both pressure-gradient forces and frictional effects:

$$u^2(x) = u^2(0) - \frac{PGF}{K}e^{-2Kx} + \frac{PGF}{K} \quad (2.2)$$

Here, $u(0)$ and $u(x)$ represent the airflow at the entrance and at some distance x downstream from the gap entrance (respectively), PGF is the along-gap pressure gradient force ($\rho^{-1}\frac{\delta P}{\delta x}$), and $K = 2.8\frac{C_D}{H}$ is a parameter representing friction through a drag coefficient (C_D) and the average depth of the boundary layer within the gap (H). After exiting the mountain barrier, gap flows are able to turn to the right (left) in the Northern (Southern) Hemisphere with a radius of curvature determined by their wind speed and the Coriolis parameter (e.g., Steenburgh et al. 1998). This effect can be enhanced when an offshore-directed gap-flow joins a large-scale onshore-directed airflow, a phenomenon observed by Loescher et al. (2006) along the coast of Alaska.

A detailed understanding of TTA kinematic structure is essential for clarifying how TTAs impact the intensity and spatial distribution of orographic precipitation, which ultimately can help improve forecasts of storms that have the potential to cause flooding and debris-flows in mid-latitude mountainous regions (Ralph et al. 2006, Neiman et al. 2009). For example, Bousquet and Smull (2003) attributed the errors in a severe precipitation forecast in the Alps to an inadequate representation of a blocking-induced TTA in the model. Likewise, Neiman et al. (2002; 2004), Yu and Bond (2002), Anders et al. (2007), Hughes et al. (2009) have documented the effects of blocking-induced TTAs on the spatial distribution of precipitation.

TTAs produced by gap flows have also been studied (e.g. Overland and Walter 1981, Overland 1984, Lackmann and Overland 1989, Mass et al. 1995, Steenburgh et al. 1998, Colle and Mass 2000) but few of these investigations have focused on the association of gap flows with orographic precipitation. In one of these rare studies, Neiman et al. (2006) examined 915 MHz wind profiler observations along the coast of northern California at Bodega Bay during winter storms from 1997 to 2004. They found that a cold and dry airstream of ~ 500 m depth frequently exited through

the Petaluma Gap and that total rainfall and rain rates increased over Bodega Bay during strong gap-flow cases.

Compared to inland ranges such as the Cascades and Sierra Nevada, coastal ranges in the western U.S. are impacted by landfalling ARs unperturbed by upstream terrain. This relative simplicity has motivated the observational documentation of orographic precipitation mechanisms during winter storms along coastal ranges. For example, (Yu and Smull 2000) examined a case where the pre-cold-frontal air was associated with a weak TTA within ~ 20 km offshore of the Oregon-California coastal mountains and enhanced precipitation along the coast. Also, Neiman et al. (2002) found a statistically significant linear relationship between cross-barrier wind speed in the LLJ at ~ 1 km MSL and rain rate at the surface using two winter seasons of observations along the windward side of northern, central, and southern California coastal ranges. Remarkably, the linear relationship below mountain top was degraded at lower levels when hypothesized blocking-induced TTAs were evident. James and Houze (2005) documented coastal orographic precipitation enhancement during three winter seasons in the Oregon-California border, hypothesizing that offshore enhancement was associated with TTAs.

The aforementioned studies provide clues about the structure of TTAs adjacent to coastal orography and associated impacts on precipitation but were limited in a few important respects. For example, the results of Neiman et al. (2002; 2006) were based on a one-dimensional, vertical-profile perspective. Also, the studies of Yu and Smull (2000) and James and Houze (2005), although based on scanning radars, had incomplete documentation of airflows below the peaks of the coastal orography, which prevents the detailed examination of low-level three-dimensional kinematic structures that are essential for better understanding their influence on orographic precipitation. The present investigation addresses these limitations by observationally documenting a landfalling winter storm along the northern California coast with a combination of scanning and profiling radar data obtained with ground-based instrumentation deployed near sea level. This study is unique because it documents in detail the three-dimensional kinematic structure and orographic precipitation forcing at low-levels (e.g. below 1 km MSL) associated with a small-scale coastal barrier.

Section 2.2 provides information about the observing systems and data processing employed in this study. An overview of the storm is presented in Section 2.3, with synoptic context for the mesoscale observations. Section 2.4 contains a detailed analysis of the kinematic and precipitation structure of the storm, which is separated into two distinct episodes. Theoretical context for the observed airflow structures is discussed in Section 2.5. Finally, section 2.6 presents a summary and conclusions for this study.

2.2 Observing systems and data processing

The data employed in this study were collected along the northern California coast during the 2003-2004 winter season as part of the Hydrometeorology Testbed (HMT, Ralph et al. 2013) operated by the National Oceanic and Atmospheric Administration (NOAA). Locations of key observing systems are shown in Figure 2.1. The primary instrument used in this study is the scanning ground-based NOAA X-band (3.2 cm wavelength) dual polarization Doppler radar called X-POL (Martner et al. 2001, Matrosov et al. 2005) deployed at Fort Ross (FRS).

XPOL executed both slant-horizontal plan-position indicator (PPI) and vertically oriented range-height indicator (RHI) scans (Table 2.1). PPI scans extended to a maximum range of 57 km with 0.23 km gate spacing and were repeated at least once every 6 minutes. At elevation angles $\leq 9.0^\circ$, PPI scans were only directed offshore due to low-level beam obstruction by the coastal mountains. RHI scans extended to a maximum range of 29 km with 0.11 km gate spacing and were executed over the ocean and terrain in a cycle that repeated at least once every 6 minutes.

Individual sweeps were edited to remove artifacts such as ground and sea clutter, range folding (i.e. second trip echoes), and sidelobe echoes, as well as to dealias folded radial velocities. After editing, polar-coordinate scans were interpolated to a Cartesian grid. For PPI scans, the horizontal and vertical grid spacing was 0.5 km and 0.35 km, respectively. For RHI scans, the horizontal grid spacing was 0.1 km and the vertical grid spacing was 0.2 km. A Cressman distance-dependent weighting scheme (Trapp and Doswell 2000) was employed to interpolate values of attenuation-corrected reflectivity (Matrosov et al. 2005) and Doppler radial velocity to each Cartesian grid

point.

Composite vertical cross sections of reflectivity and radial velocity were made by combining north and south RHI scans (e.g., 0° and 180° azimuth). Although the contributing RHI scans were offset by 2-3 minutes, the structure across the interface of the two scans was coherent. The horizontal component of radial velocity in the plane of each composite cross section was calculated toward north (i.e. meridional wind). Elevation angles greater than 30° were excluded to simplify both visualization and interpretation of airflow structures.

A velocity-azimuth-display technique (VAD, Browning and Wexler 1968) was applied to 15° -elevation-angle PPI scans, with one scan available every 12 minutes. The VAD technique allows the retrieval of mean horizontal wind above and centered on the radar by using Fourier analysis of the radial velocity as a function of azimuth. Radial velocities from slant ranges between 0.3 km and 8 km were employed to derive VAD wind profiles from near the surface up to ~ 2 km MSL.

A 915 MHz wind profiling radar (Ecklund et al. 1988, Carter et al. 1995) located at Bodega Bay (BBY, Fig. 2.1) provided hourly profiles of horizontal winds from ~ 0.1 km MSL to ~ 2.3 km MSL with ~ 60 m vertical resolution (low altitude mode). Hourly wind profiler data was processed with the continuity method of (Weber et al. 1993) that checks consistency in the data set over time and height. To complement the wind profiler at BBY, a radio acoustic sounding system (RASS) was also deployed at this site, retrieving hourly profiles of virtual temperature (Clifford et al. 1994). The system retrieved a total of 20 range gates from 141 m MSL to 1268 m MSL at ~ 60 m intervals. Hourly RASS data was processed using the consensus method of Fischler and Bolles (1981) that checks for consistency in time. Virtual potential temperature was then derived using the method of Neiman et al. (1992).

Several other important observing systems contributed to the analysis. Vertical profiles of pressure, temperature, relative humidity, and horizontal wind velocity were retrieved from 7 balloon soundings released at FRS with an average vertical resolution of 9 m. Routine buoy data from the NOAA-National Data Buoy Center provided surface winds offshore. Buoy 46013 (henceforth B13) is located ~ 22 km southwest of BBY and reports hourly averaged wind speed and direction observed

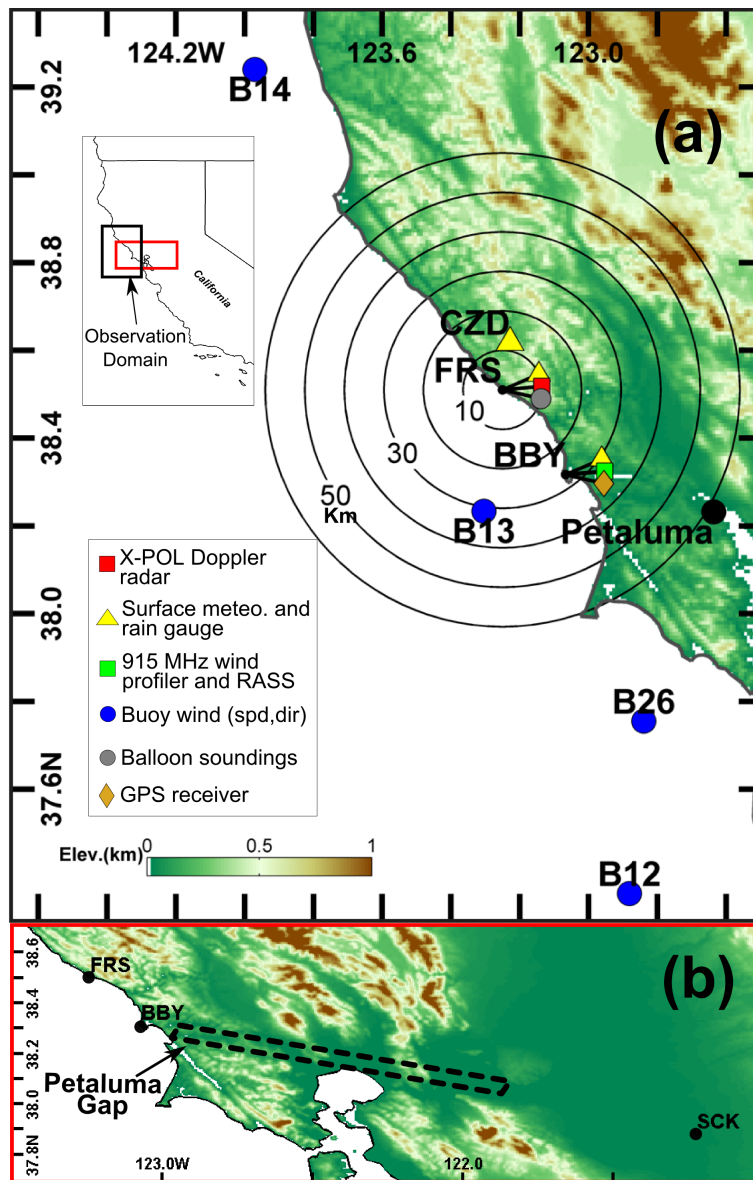


Figure 2.1: (a) Topographic map overlaid with observing systems. Legend of each instrument along with the color scale for terrain elevation (km) is located at the bottom left corner. Range circles centered at Fort Ross (FRS) indicate the X-POL analysis domain. (b) Topography map showing Petaluma Gap terrain and the METAR station at Stockton (SCK). Inset map in (a) provides a reference for both topographic maps relative to the coast of northern California: map (a), black line; map (b), red line.

Table 2.1: PPI and RHI scan strategies performed by X-Pol.

	Rotation angle	Fixed angle
	Azimuth ($^{\circ}$)	Elev ($^{\circ}$)
PPI	150 - 330	0.5, 1.0, 2.0, 3.4, 4.8, 6.2, 7.6, 9.0
	0 - 360	11.0, 13.0, 15.0, 17.0, 19.0
	Elev ($^{\circ}$)	Azimuth ($^{\circ}$)
RHI	0 - 100	0, 15, 30, 45, 60, 75, 90, 105, 120, 135, 150, 180, 210, 240, 270, 300, 330, 345
	0 - 162	6, 147

continuously at 1-minute resolution. In addition, buoys 46014 (B14), 46026 (B26), and 46012 (B12), located ~ 126 km north, ~ 67 km south, and ~ 104 km south from B13 respectively, were included in the analysis (Fig. 2.1a). Hourly averaged column-integrated water vapor (IWT) at BBY was obtained from a ground-based Global Positioning System (GPS) receiver (Wolfe and Gutman 2000), allowing the evolution of water vapor to be monitored while the storm passed over the observing domain. GPS-IWV observations are unaffected by precipitation (Businger et al. 1996). Two-minute resolution surface observations of air temperature, relative humidity, pressure, wind velocity, and precipitation were collected at FRS, BBY, and Cazadero (CZD). These observations were manually checked for outliers and time consistency.

2.3 Event overview

Analyses from the Climate Forecast System Reanalysis (CFSR, Saha et al. 2010) provide synoptic-scale context for the storm (Fig. 2.2). An upper level cyclonic circulation is evident over the Pacific ($\sim 45^\circ\text{N}$, $\sim 145^\circ\text{W}$) at 06 UTC 16 February (Fig. 2.2a) and is associated with developing cold and warm fronts on its southeast flank. The warm front approaches the coast of northern California by 18 UTC 16 February (Fig. 2.2b), making landfall at 00 UTC 17 February and then dissipating (not shown). By 06 UTC 17 February (Fig. 2.2c), the cold front has stalled offshore of northern California as a stationary front. A 500 hPa shortwave trough approaching from the west and a ridge building onshore are also evident at this time. By 18 UTC 17 February (Fig. 2.2d), the shortwave trough aloft begins accelerating toward the California coast. After 18 UTC 17 February, the shortwave trough aloft progresses quickly eastward to California and the cold front consequently moves rapidly southeastward down the California coast (Fig. 2.2e,f).

The characteristics of the stationary front are examined in more detail with hourly averaged winds from buoy observations (Fig. 2.3). There are relatively weak cold-sector winds at B14 during the period from about 22 UTC 16 February to about 06 UTC 17 February, with stronger warm-sector south-southeasterly winds before and after. This suggests that the stationary front sagged southward over B14 during this period and then retreated northward. B13 exhibits a similar trend,

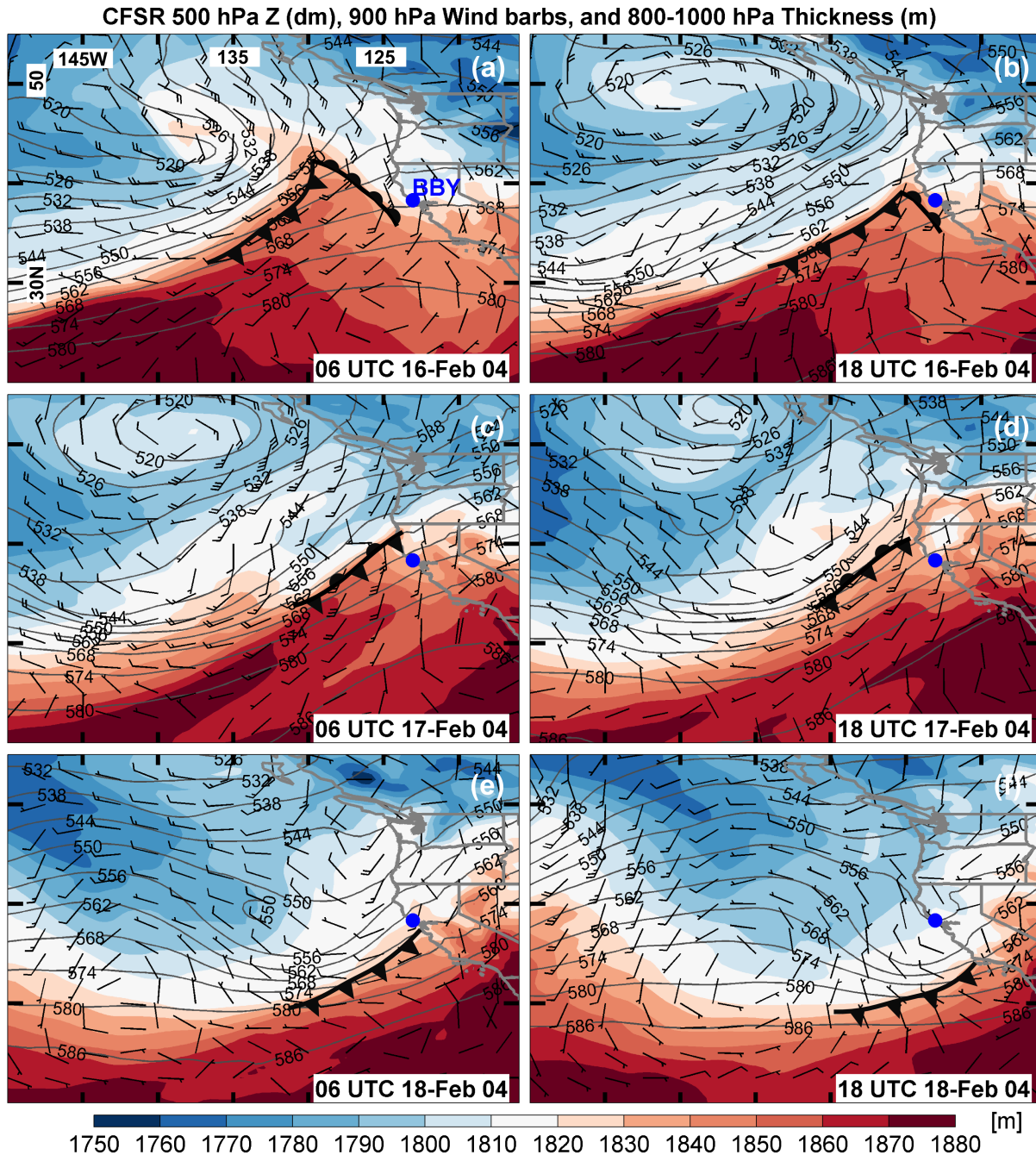


Figure 2.2: Reanalysis-derived 500 hPa geopotential height (Z , contours), 900 hPa wind barbs, and 800-1000 hPa thickness (m, filled contours) from the NCEP-CFSR at (a) 06 UTC, (b) 18 UTC 16 February, 17 February (c,d, respectively), and 18 February (e,f, respectively). Conventional symbols for near-surface cold, warm, and stationary fronts are included. Full barbs are 10 m s⁻¹. Blue dot indicates BBY location.

but delayed by about 4 hours with less of a contrast in wind characteristics and of shorter duration. Further south, B26 and B12 show no effect of the stationary front given that it remained north of these buoys. After 10 UTC 18 February, all the buoys mark a wind-velocity shift from southerly to northwesterly with the southeastward-directed cold-frontal passage (as revealed by the CFSR analyses in Fig. 2.2).

Additional synoptic context is provided by IWV retrievals (Wentz 1997) from the group of three polar-orbiting satellites carrying the Special Sensor Microwave Imager (SSM/I). Figure 2.4 presents composite images of SSM/I-IWV during periods similar to those shown in Fig. 2.2. A long and narrow corridor of IWV greater than 2 cm is evident offshore of California during the entire storm, which is the signature of a landfalling AR (Zhu and Newell 1998, Ralph et al. 2004). Initial AR landfall occurs shortly after 06 UTC 16 February (Fig. 2.4a). Thereafter, from 14 UTC 16 February to 08 UTC 17 February (Fig. 2.4b,c) the AR meanders, probably in association with the stationary front depicted by the CFSR analysis. Finally, from 14 UTC 17 February onwards (Fig. 2.4d,e,f), the AR shows a southward progression in connection with the cold frontal passage indicated by the CFSR at similar times.

Time series of surface rain-rate observations (Fig. 2.5a) highlight relevant meteorological characteristics in northern California during the landfalling storm, which produced total rainfall accumulations of 252, 152, and 125 mm at CZD, FRS, and BBY, respectively. There are two distinct episodes of relatively large surface rain rate: from 06 UTC 16 February to 06 UTC 17 February (episode 1) and from 06 UTC 17 February to 06 UTC 18 February (episode 2). Peak rain rates of $3\text{--}8\text{ mm h}^{-1}$ are evident at all sites during episode 1 from 06 UTC 16 February to 15 UTC 16 February, with slightly higher values at BBY (coast) and slightly lower values at CZD (coastal mountains). Shortly thereafter, rain rates increase dramatically. Peak values of $10\text{--}20\text{ mm h}^{-1}$ are evident between 15 UTC 16 February to 21 UTC 16 February, with the largest rain rates at CZD and smallest at BBY. After the peak, rain rates decrease to less than 5 mm h^{-1} through the end of episode 1. In contrast to episode 1 with its single peak of rain rate, episode 2 is characterized by multiple lower-valued peaks of rain rate. Maximum rain rate at CZD is $\sim 18\text{ mm h}^{-1}$ while at BBY

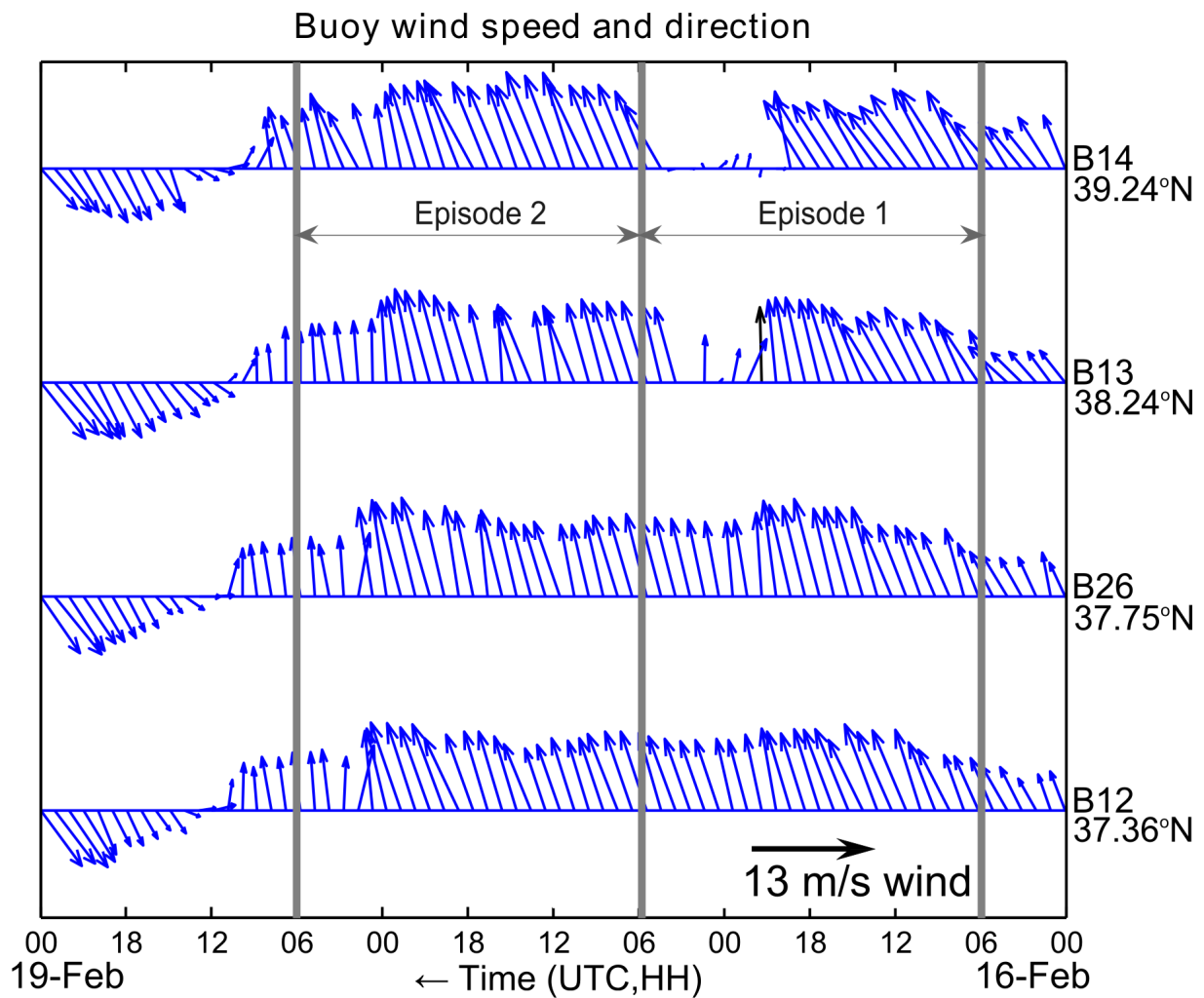


Figure 2.3: Wind vectors from buoys B14 (39.24°N, 123.97°W), B13 (38.24°N, 123.30°W), B26 (37.75°N, 122.84°W) and B12 (37.36°N, 122.88°W) located offshore of the northern California coast and arranged from north to south. Time is from right to left to represent eastward advection of the storm. Black vector wind scale is included. See text for definition of episode 1 and 2.

Composites of SSMIS Integrated Water Vapor

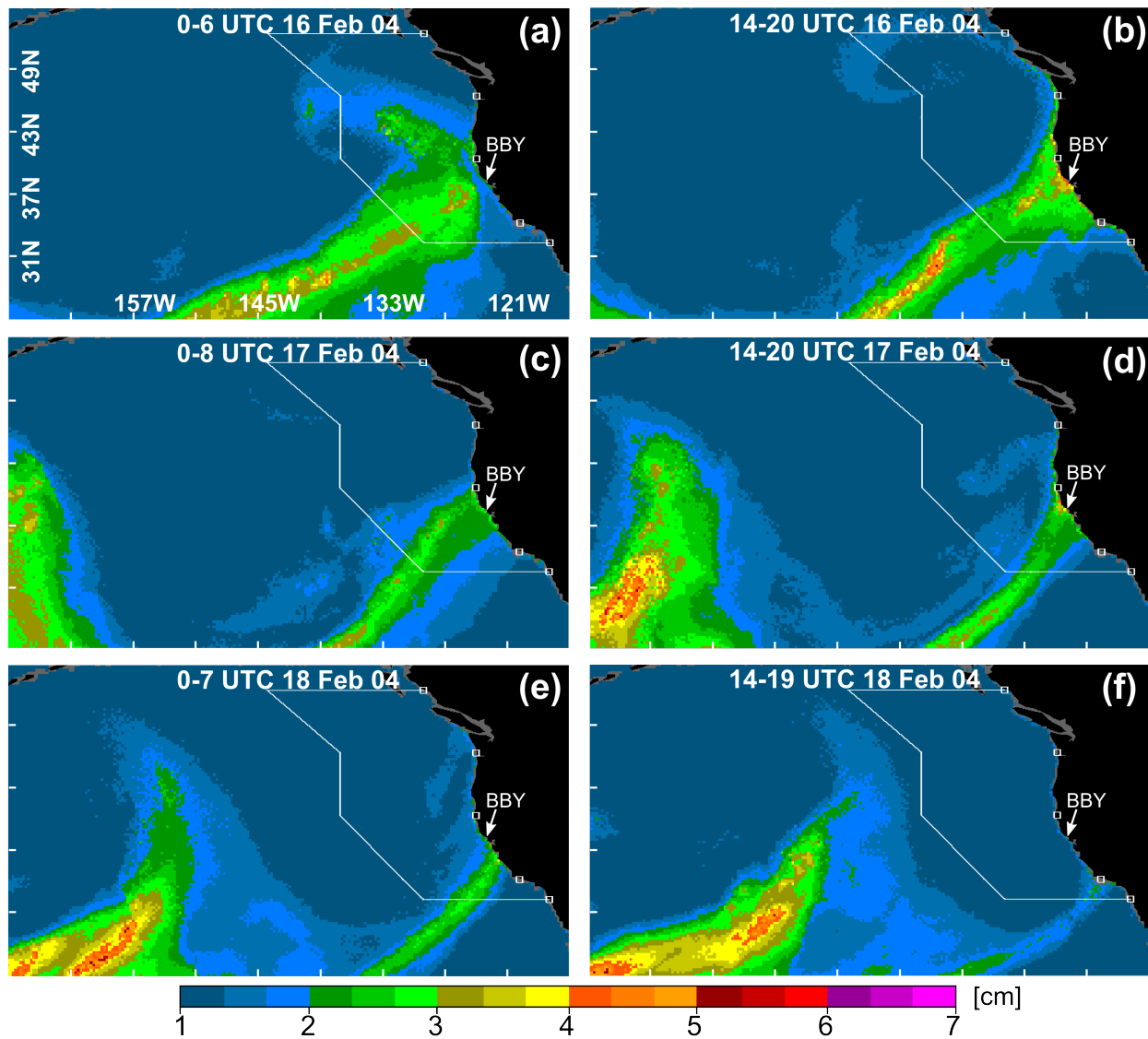


Figure 2.4: IWV composite images observed with the SSM/I instrument over the periods (a) 0-6 UTC 16 February, (b) 14-20 UTC 16 February, (c) 0-8 UTC 17 February, (d) 14-20 UTC 17 February, (e) 0-7 UTC 18 February, and (f) 14-19 UTC 18 February. Color scale for IWV indicated at bottom of panels.

and FRS maximum values are a factor of three smaller ($\sim 6 \text{ mm h}^{-1}$). To further contrast both episodes, Table 2.2 reveals that during episode 1 similar amounts of precipitation fell along the coast and over the mountains (e.g. only 10% to 27% of orographic enhancement); however, episode 2 exhibits a significant difference between precipitation along the coast and the coastal mountains (e.g. 241% to 393% of orographic enhancement).

Table 2.2: Site elevation and total precipitation during episode 1 (0600 UTC 16 Feb 0600 UTC 17 Feb) and episode 2 (0600 UTC 17 Feb 0600 UTC 18 Feb). The CZD ratio is defined as the total precipitation at CZD divided by the total precipitation at each of the three sites.

Station	Elev (m MSL)	Episode 1 (mm)	CZD ratio	Episode 2 (mm)	CZD ratio
CZD	462	124	1.00	133	1.00
FRS	40	113	1.10	39	3.41
BBY	15	98	1.27	27	4.93

Neiman et al. (2009) document that the largest correlation between cross-barrier bulk IWV flux and rain rate over the coastal mountains of northern California is achieved in a $\sim 300\text{-m}$ -deep layer centered at $\sim 1 \text{ km}$ MSL. Thus, the cross-barrier bulk IWV flux at BBY was computed as the product between GPS-IWV and layer-mean ($0.85\text{-}1.15 \text{ km}$ MSL) cross-barrier wind speed (directed toward 50°) observed with the wind profiler (Fig. 2.5b). Although GPS-IWV is an integrated observation of air moisture, balloon soundings (not shown) confirm that most of the water vapor (e.g. mixing ratio $> 6 \text{ g kg}^{-1}$) is on average located below 2.5 km MSL, with maximum values of $\sim 8.5 \text{ g kg}^{-1}$ near the surface. Figure 2.5b illustrates that episode 1 is associated with maximum bulk IWV flux of $80\text{-}90 \text{ cm m s}^{-1}$ over a $\sim 5 \text{ h}$ period coinciding with peak rain rates, whereas episode 2 is characterized by values around 70 cm m s^{-1} over most of the period with less clear association to the rain-rate time series compared with episode 1. The peak in bulk IWV flux at the interface between episodes 1 and 2 (e.g., 06 UTC 17 February) occurs along with rain rates near zero and is associated with relatively large cross-barrier wind speeds ($\sim 35 \text{ m s}^{-1}$) and a local minimum in GPS-IWV ($\sim 2.7 \text{ cm}$). It is worth noting that, despite this single peak, the overall

trend in rain rate seems better correlated with bulk IWV flux compared with GPS-IWV alone, in agreement with the results of Neiman et al. (2009).

2.4 Detailed kinematic and precipitation structure

2.4.1 Episode 1: 06 UTC 16 February to 06 UTC 17 February 2004

Previous studies by Ralph et al. (2004; 2005) and Neiman et al. (2002; 2009) have consistently observed the pre-frontal LLJ at a height of ~ 1 km MSL. As indicated by CFSR analyses (Fig. 2.2), winds at 900 hPa (~ 1 km MSL) near BBY and FRS are characterized by a strong meridional component. Thus, we can infer that the LLJ in this case has a primarily meridional orientation. Wind profiler data at BBY indicates a LLJ centered at ~ 1 km MSL from ~ 10 -14 UTC 16 February (Fig. 2.6a). The LLJ manifests itself as a meridional-component wind speed maximum of $\sim 25 \text{ m s}^{-1}$ from ~ 0.5 -1.5 km MSL. During the same period, a VAD analysis at FRS shows a LLJ of comparable magnitude, thickness and altitude (Fig. 2.6b). After 14 UTC 16 February, the LLJ-center descends at both locations, down to ~ 0.5 km MSL at BBY and down to ~ 0.7 km MSL at FRS. Then, both profiles show a decrease in meridional wind speed after 21 UTC 16 February in association with the cold frontal passage described in section 3. FRS shows the largest decrease in wind speed after 21 UTC 16 February and both profiles return to wind speeds $> 20 \text{ m s}^{-1}$ after 02 UTC 17 February, which is consistent with a cold front sagging southward and retreating northward afterwards (e.g. Fig. 2.2).

Along with the LLJ structure and cold frontal effects, both time-height cross sections show a distinctive vertical gradient of meridional wind speed centered at ~ 0.6 -0.7 km MSL, with values of $\sim 5 \text{ m s}^{-1}$ below and $\sim 25 \text{ m s}^{-1}$ above. At BBY the gradient is observed from 09-15 UTC 16 February while at FRS the gradient is apparent at the same altitude from ~ 09 -18 UTC 16 February. Winds below the gradient are mainly southeasterly at both sites and a jet structure is absent. This suggests that the airflow is directed approximately parallel to the terrain orientation during the periods described above and is therefore consistent with a TTA.

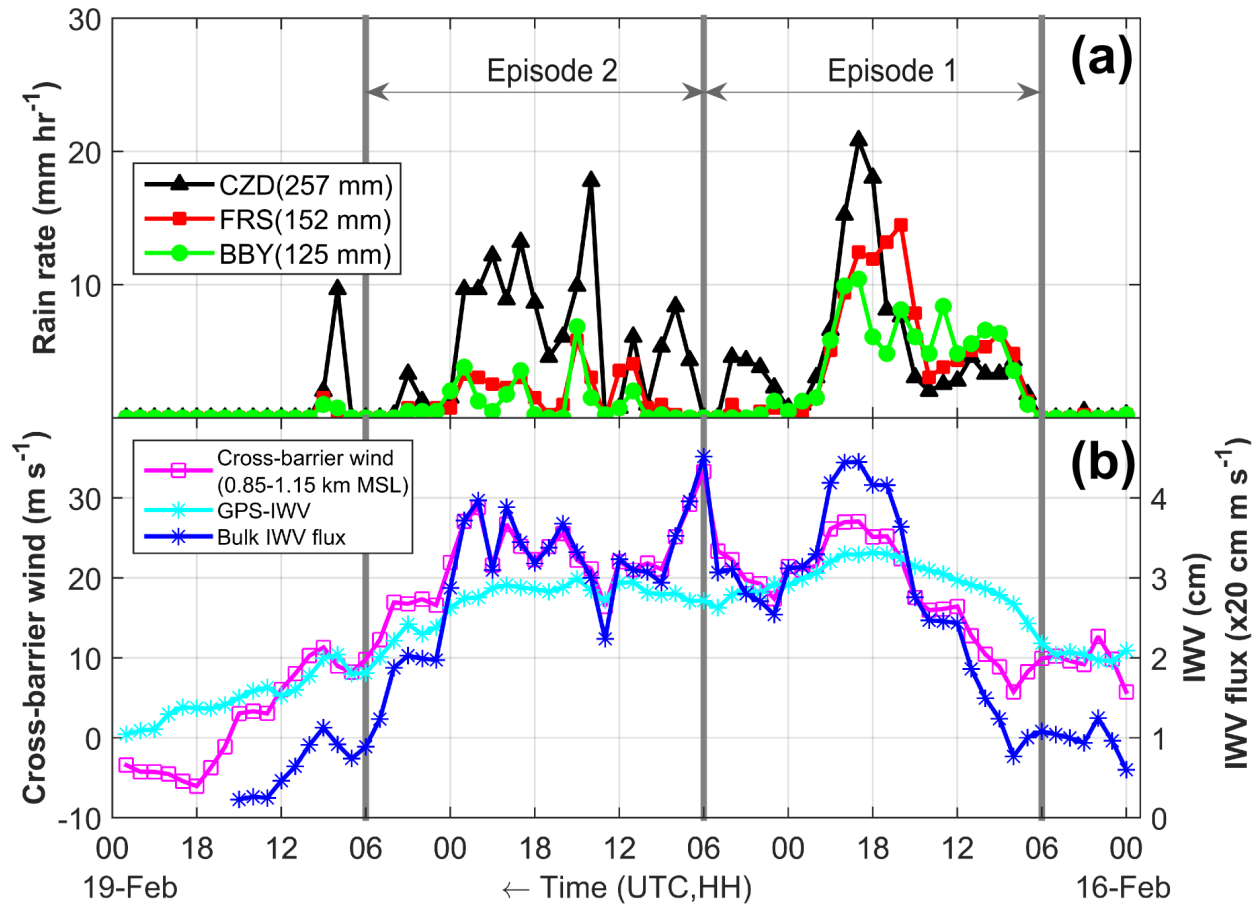


Figure 2.5: (a) Rain rate at CZD (black), FRS (red), and BBY (green). Legend includes total accumulated precipitation at each site over the period from the beginning of episode 1 to the end of episode 2. (b) 0.85-1.15 km MSL layer-mean cross-barrier wind speed (i.e., directed toward 50) from the wind profiler on the left axis and GPS-IWV and bulk IWV flux on the right axis. Bulk IWV flux is derived from the product of the layer-mean cross-barrier wind speed and the GPS-IWV. Time is from right to left to represent eastward advection of the storm. See Fig. 1 for location of CZD, FRS, and BBY.

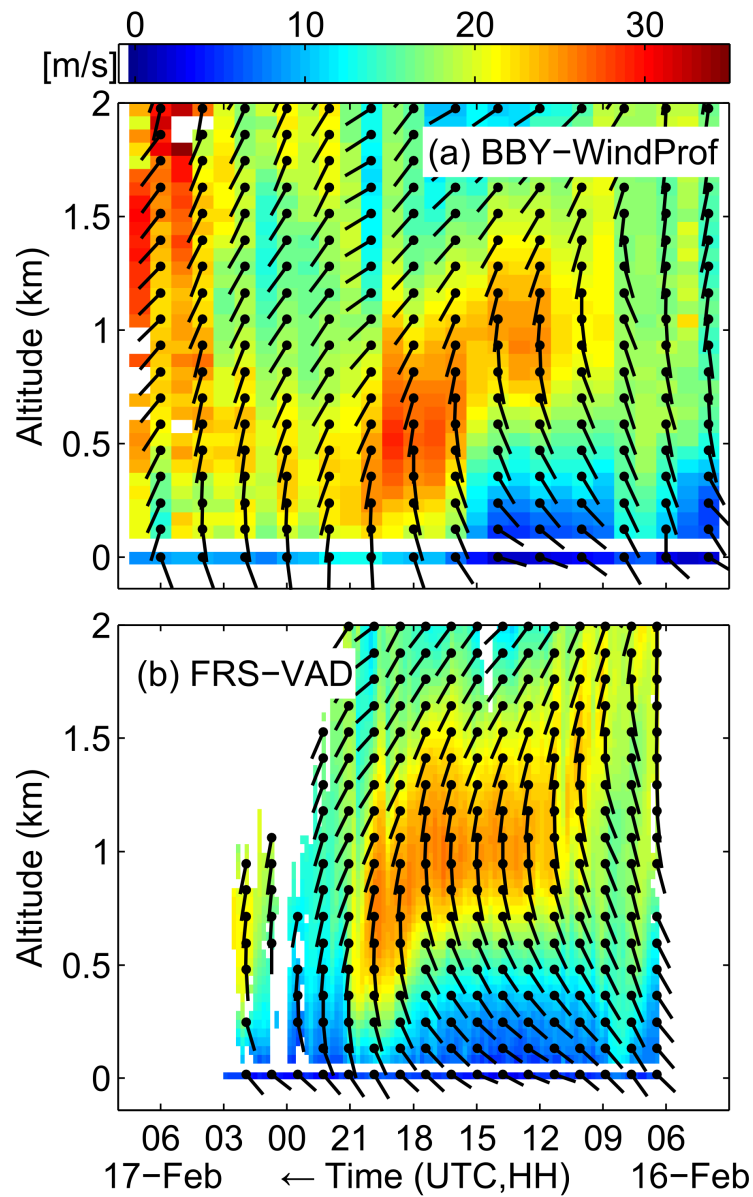


Figure 2.6: Time-height analysis of meridional wind speed (color coded, m s⁻¹) and total wind direction (wind staffs) at (a) BBY from the 915 MHz wind profiler (WindProf) and (b) FRS from an X-POL Velocity-Azimuth Display (VAD) analysis during episode 1. The first level in each panel includes surface observations. Time is from right to left to represent eastward advection of the storm.

The horizontal pattern of Doppler radial velocity associated with the TTA during episode 1 is analyzed with hourly-averaged X-POL 0.5° PPI scans. Figure 2.7a depicts a representative hour (e.g., at 14-15 UTC 16 February) when a sharp curvature in the isoline of zero radial velocity (henceforth called the zero isodop) is evident along the $\sim 225^\circ$ radial at a range of ~ 18 km. This curvature indicates a nearly homogenous southeasterly wind from the coast to ~ 18 km offshore and a significant wind direction shift further offshore, where the zero isodop changes its orientation by about 90° implying south-southeasterly to southerly winds. At ~ 18 km range, the X-POL 0.5° beam is at an altitude of ~ 220 m MSL, which is well below the vertical gradient of wind direction observed in the vertical profiles at BBY and FRS (Fig. 2.6). This suggests that the sharp curvature observed in the zero isodop ~ 18 km offshore is the result of a horizontal gradient in wind direction. Additionally, Wood and Brown (1986) simulated a variety of Doppler patterns using different wind direction and speed profiles. The only case that produced a sharp curvature in the zero-isodop line was the simulation of a horizontal wind direction discontinuity within the radar domain intended to mimic a frontal boundary. Consequently, the sharp curvature in the zero isodop is most likely due to a strong horizontal gradient in wind direction. We assert that this horizontal gradient is associated with the interface between the TTA along the coast and the LLJ further offshore.

The companion average reflectivity (Fig. 2.7b) shows a precipitation pattern dominated by a band of northwest-southeast-oriented enhanced reflectivity from the coast to ~ 20 km offshore. This was a nearly stationary structure that persisted at other hours during episode 1 with both variable extension offshore and reflectivity magnitude. Figure 2.7b also shows 5 thinner reflectivity bands with similar orientation 20-60 km southwest of the radar and two southwest-northeast-oriented reflectivity bands on the western edge of the domain. Individual 6-minute 0.5° PPI scans between 14-15 UTC 16 February (e.g., Fig. 2.7c-d) indicate that the former corresponds to a single northwest-southeast oriented band moving northeastward during the averaging period, whereas the latter is formed by two isolated cells moving toward the northeast and giving the appearance of a southwest-northeast-oriented band.

Vertical context for the converging LLJ and TTA is provided by examination of X-POL RHI

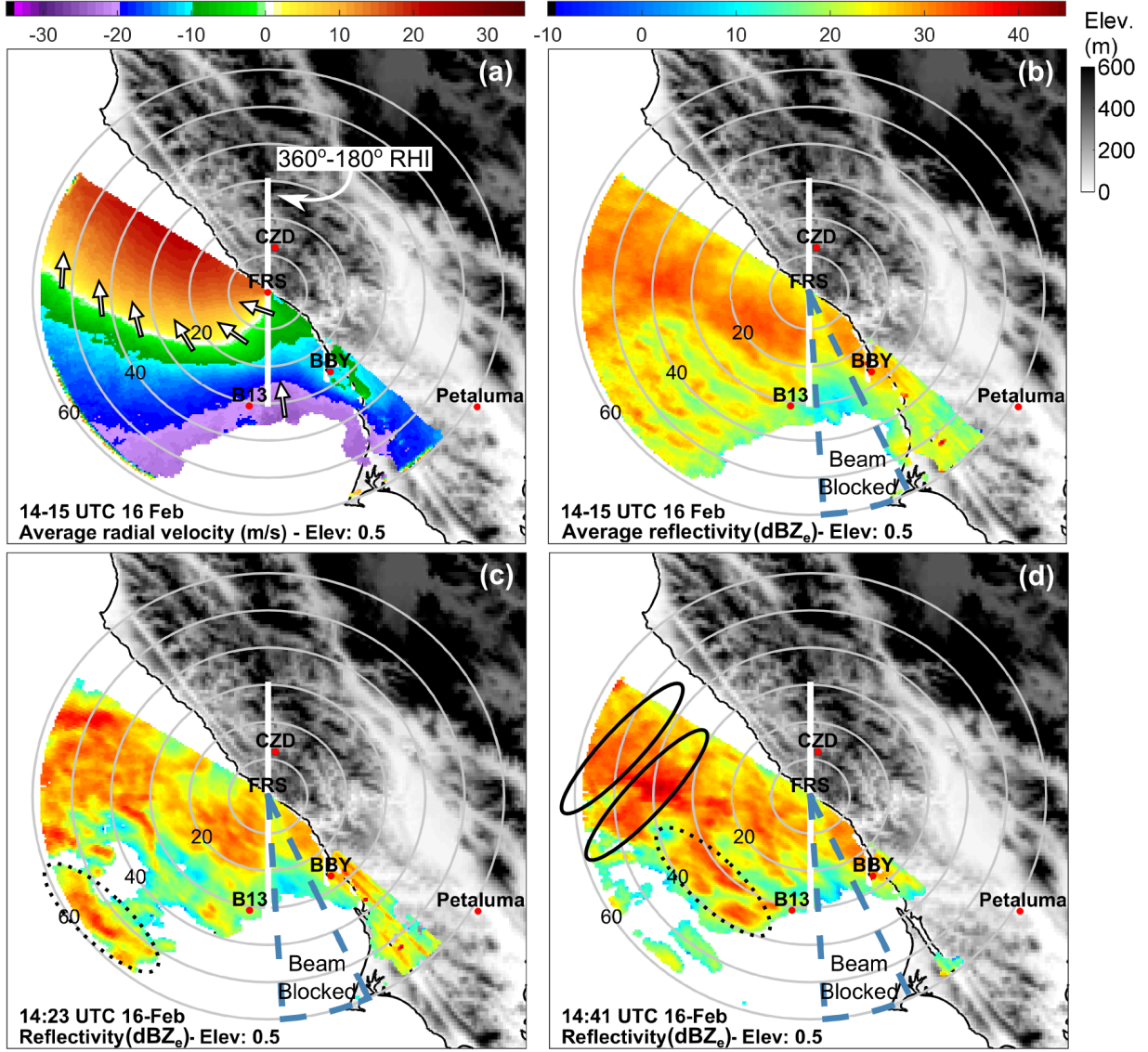


Figure 2.7: (a) Average radial velocity (m s^{-1}) and (b) equivalent reflectivity (dBZe) from plan-position indicator (PPI) scans observed during episode 1 between 14-15 UTC 16 February. (c) and (d) present individual scans of equivalent reflectivity during the same period at 14:23 UTC and 14:41 UTC. Arrows along the isoline of zero radial velocity (zero isodop) in (a) are orthogonal to X-POL beams, thus they aid in visualizing the airflow direction. The isolated arrow along the $\sim 170^\circ$ radial at ~ 30 km range in (a) indicates the direction of maximum approaching radial velocities ($\sim 25 \text{ m s}^{-1}$) associated with the LLJ. The dashed line sector in (b,c,d) indicates an area with decreased reflectivity due to beam blockage by the radar trailer. (c) and (d) highlight transient reflectivity bands northwest-southeast (dotted line) and southwest-northeast (continuous line) oriented, which are visualized in the 14-15 UTC average (b). Panels include the azimuth direction (white line) of range-height indicator (RHI) analyses for episode 1 (see Fig. 2.8).

scans oriented from south to north (e.g. 180° - 360°). The horizontal velocity during 14-15 UTC 16 February (Fig. 2.8a) indicates a strong airflow of ~ 20 - 25 m s^{-1} (meridional component of the LLJ) sloping over a weaker flow of ~ 5 - 10 m s^{-1} (meridional component of the TTA). Horizontal velocity during 17-18 UTC 16 February (Fig. 2.8b) shows a similar structure but with the LLJ sloping over the weaker flow closer to the coast. Assuming that the LLJ/TTA interface near sea level is located 20-25 km offshore from the coast during 14-15 UTC, then the difference between the two hours shows a progression of $\sim 15 \text{ km}$ toward the coastline. As a result, the vertical kinematic structure at these times illustrates how the LLJ slopes upward over the TTA offshore and also how the interface between these two flows gets closer to the coast.

The average vertical structure of reflectivity gives some clues about the precipitation structure associated with the LLJ/TTA interface. During 14-15 UTC 16 February (Fig. 2.8c) there is evidence of a bright band (i.e. melting layer) at $\sim 3 \text{ km}$ MSL that ascends slightly toward the coastal mountains. This upward deviation of the brightband toward the mountains is opposite of that typically observed in association with larger barriers (e.g. Marwitz 1983, Minder and Kingsmill 2012). The physical mechanisms responsible for this behavior are unclear. Above and below the bright band some vertically oriented reflectivity structures resembling fallstreaks are observed. The presence of a bright band and fallstreaks suggests that seeder particles aloft are available to fall into feeder clouds at lower levels and contribute to precipitation development via the seeder-feeder process (e.g. Bergeron 1965, Browning et al. 1974, Rutledge and Hobbs 1983). Below $\sim 1.5 \text{ km}$ MSL and from ~ 2.5 to $\sim 18 \text{ km}$ offshore there is a concentrated area of enhanced precipitation. Interestingly, the enhancement below $\sim 1.5 \text{ km}$ MSL resides downwind from where the LLJ slopes upward (Fig. 2.8a). This suggests that precipitation enhancement is associated with condensate production resulting from lifting of the LLJ by the TTA. Later, during 17-18 UTC 16 February (Fig. 2.8d), the precipitation enhancement zone is closer to the coast and more intense, especially over the windward side of the coastal mountain. Accordingly, an upslope flow effect (i.e. direct orographic ascent) likely has a larger role going forward from this time.

Isochrones of the zero isodop are drawn every 2 hours from 8-21 UTC 16 February to analyze

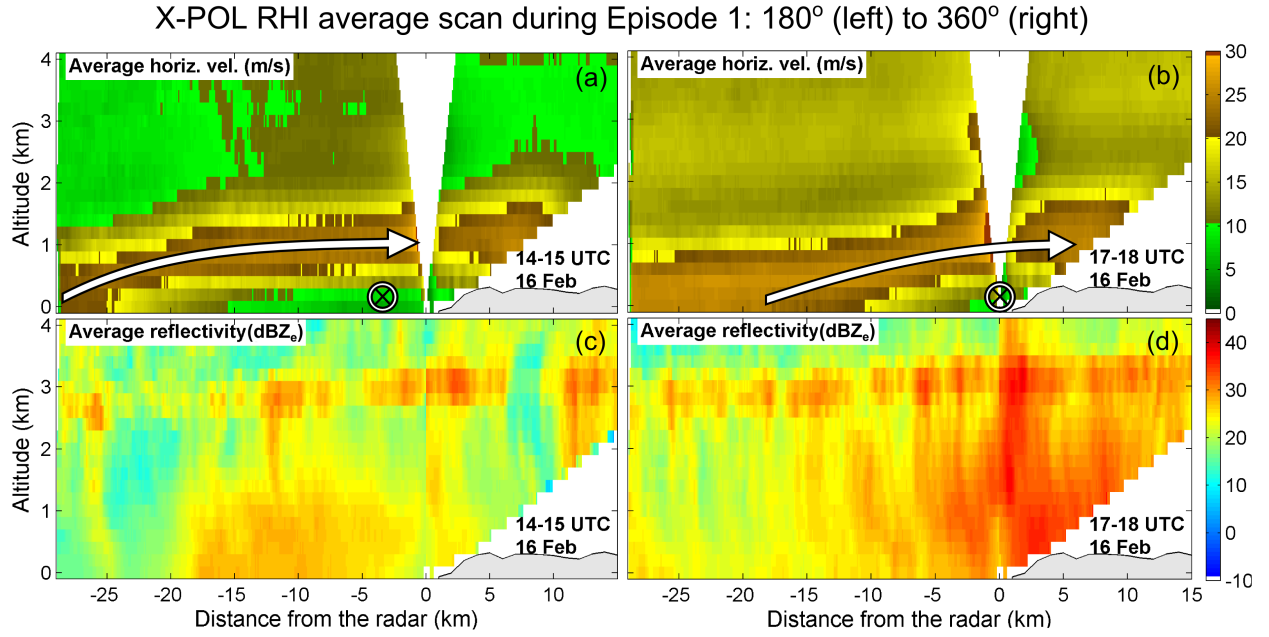


Figure 2.8: (a, b) Average horizontal velocity (m s^{-1} , see color scale) and (c, d) equivalent reflectivity (dBZe, see color scale) from range-height indicator (RHI) scans observed during episode 1 between 14-15 UTC (left panels) and 17-18 UTC (right panels) 16 February in the meridional direction (180°-360° azimuth, indicated in Fig. 2.7). White arrows in horizontal velocity indicate the location of the LLJ, whereas the crossed circle indicates location of the TTA.

the motion of the LLJ/TTA interface over the course of episode 1 (Fig. 2.9). The zero isodop changed its shape significantly during this period. It started as a nearly straight, west-southwest to east-northeast line between 8-9 UTC 16 February. Thereafter, zero-isodop lines attained an increasingly curved appearance with their western extent moving progressively northward and closer to the coast. A short segment of sharp curvature in the zero-isodop lines was first observed during 14-15 UTC 16 February at 18 km range. It is this sharp curvature that is most indicative of the wind discontinuity associated with the LLJ/TTA interface. Later isochrones also exhibit short segments of sharp curvature, but they occur closer to the coastline. This implies that the LLJ/TTA interface moves toward the shore over time. By 20-21 UTC 16 February, the zero-isodop line is so close to the coastline and its length so truncated that the LLJ/TTA interface is difficult to see in the X-POL PPI data.

If the TTA is causing a horizontal wind discontinuity detectable in the Doppler velocity pattern, then variations in the Doppler pattern from one time to another might provide information about the wind-discontinuity motion. Variations in the Doppler pattern were quantified by computing hourly standard deviations of radial velocity from single PPI scans with 0.5° elevation angle. The hourly standard deviation of Doppler radial velocity (StdDev) between 14-15 UTC 16 February (Fig. 2.10a) is in general at or below $\sim 1 \text{ m s}^{-1}$ except for a northwest-southeast oriented band of large variability with values around 2 m s^{-1} or greater. Along with the information provided by the hourly averaged radial velocity, this standard-deviation band (henceforth called StdDev band) identifies the location and progression of the LLJ/TTA interface between 14-15 UTC 16 February. In other words, the interface was northwest-southeast oriented and moved a distance equivalent to the StdDev band width (i.e., from the $\sim 25 \text{ km}$ to $\sim 18 \text{ km}$ range ring) toward the northeast. Figure 2.10b shows isochrones of the LLJ/TTA interface from 13-18 UTC 16 February, indicating that the interface propagated $\sim 22 \text{ km}$ toward the coast over 5 hours, which represents an average speed of $\sim 1.2 \text{ m s}^{-1}$. This means that the interface approaches the coast at a slower speed than the background south-southeasterly airflow of $\sim 12\text{-}15 \text{ m s}^{-1}$ at the surface (indicated by B13) and the south-southeasterly airflow of $\sim 20\text{-}25 \text{ m s}^{-1}$ at $\sim 250 \text{ m}$ MSL (indicated by the PPI and RHI

Isochrones (UTC) of zero isodop in Episode 1

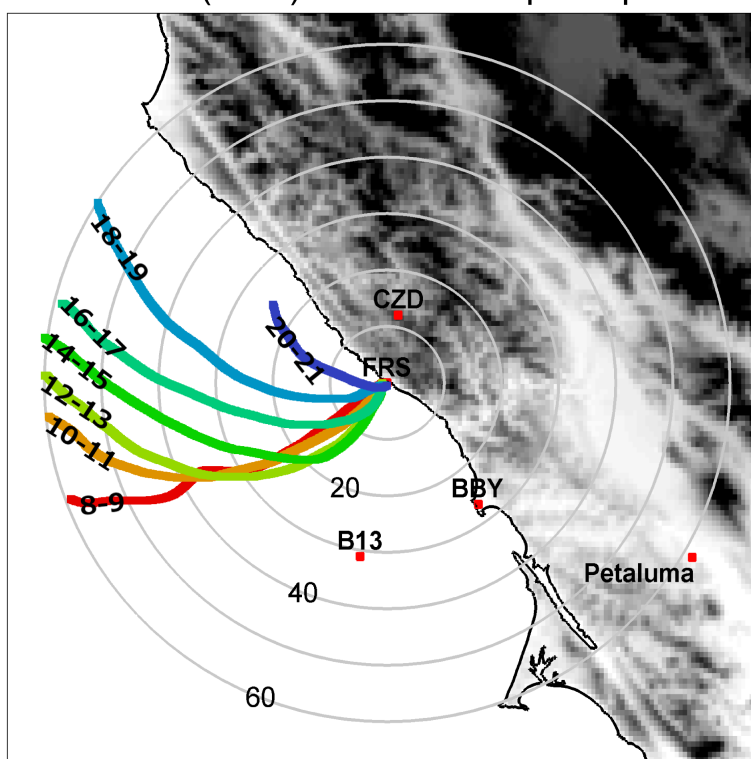


Figure 2.9: Isochrones (in UTC on 16 February 2004) of the zero isodop observed in the average-PPI scans during episode 1. Colors represent the progression of time, from early (red) to late (blue) during the episode.

scans). In addition, the isochrones point out that the LLJ/TTA interface has a skewed orientation relative to the mean coastline orientation during the course of episode 1, with positions that are further offshore toward the north.

The precipitation response to the LLJ/TTA interface motion is investigated by using a contoured frequency by distance diagram (CFDD) of reflectivity for each hour of episode 1. Unlike the contoured frequency by altitude diagram (CFAD, Yuter and Houze 1995), the CFDD provides histograms of a variable as a function of the horizontal distance from a reference point toward some arbitrary direction. In this case, reflectivity histograms are computed as a function of the distance from FRS offshore toward 210° azimuth (i.e. roughly perpendicular to the LLJ/TTA interface) using hourly averages of PPI reflectivity (Fig. 2.11). As a result, changes in reflectivity relative to the LLJ/TTA interface can be detected.

Between 13-14 UTC 16 February (Fig. 2.12a) an increase in the mode of the reflectivity distribution is evident from ~ 50 km offshore (~ 20 dBZe) toward the coast (~ 28 dBZe), with a local maximum (~ 30 dBZe) at ~ 15 km offshore. The reflectivity distribution over the period 14-15 UTC (Fig. 2.12b) is characterized by more variability from ~ 20 km offshore compared with the previous hour, while the local maximum (~ 32 dBZe) migrates to within ~ 11 km of the coast. Between 15-16 UTC (Fig. 2.12c), the reflectivity distribution exhibits little variation from ~ 20 km offshore to further offshore (~ 27 - 29 dBZe). The local reflectivity maximum keeps approaching the coast (~ 8 km) and has a larger magnitude (~ 33 - 35 dBZe). Increased variability of the reflectivity distribution is evident within 30 km of the coast during the period between 16-17 UTC (Fig. 2.12d). However, the dominant mode of the reflectivity distribution has a distinct maximum of ~ 36 dBZe at ~ 5 km offshore. By 17-18 UTC (Fig. 2.12e), the reflectivity distribution is approximately independent of the distance offshore (~ 27 - 29 dBZe) except within 5 km of the coast where modal values increase to ~ 34 dBZe.

The systematic progression of maximum reflectivity toward the coast documented in Fig. 2.12 represents local precipitation enhancement due to the ascent of the incoming LLJ over the TTA. Cloud particles likely reach larger sizes just downwind of the LLJ/TTA interface due to the

StdDev of radial velocity and isochrones (UTC)
of the interface motion in Episode 1

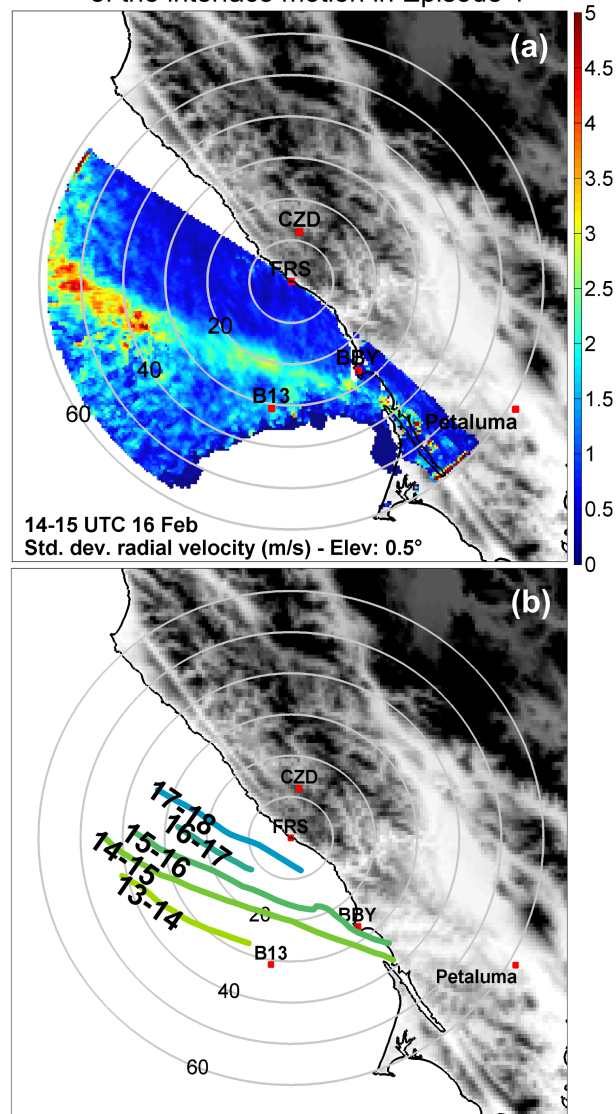


Figure 2.10: (a) Standard deviation of the average PPI radial velocity (m s^{-1}) between 14-15 UTC 16 February showing a band (StdDev band) of large variation that highlights the LLJ/TTA interface motion during the analyzed hour. (b) Isochrones (in UTC on 16 February 2004) of the StdDev band observed during other hours in episode 1. Each isochrone is drawn along the edge of the StdDev band that is closest to the coast for each hour when the band was visually detected.

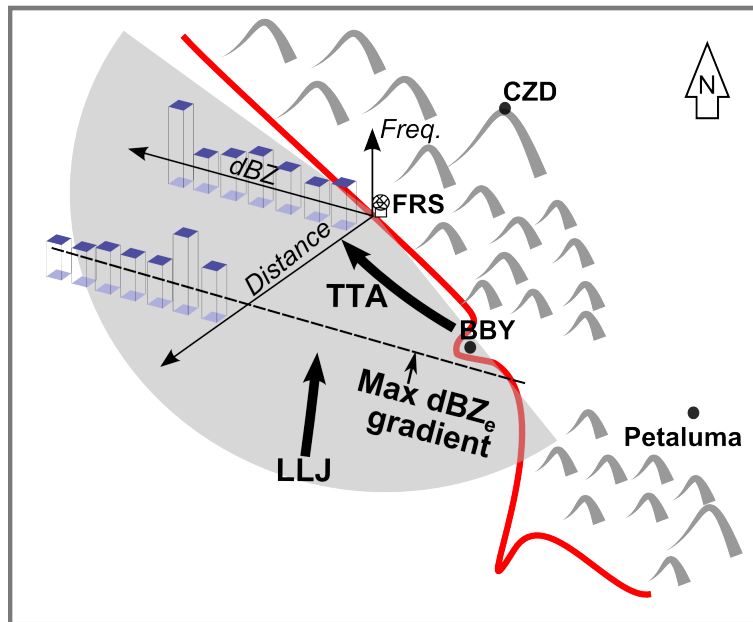


Figure 2.11: Schematic illustration of the contoured frequency by distance diagram (CFDD) employed in this study. The gray-shaded area indicates the spatial domain of X-POL PPI scans and the red line indicates the coastline. The orientation of the distance axis (reference azimuth) is chosen such that it is orthogonal to the maximum horizontal reflectivity gradient and the LLJ/TTA interface at 14-15 UTC 16 February (Fig. 2.7). Columns schematically represent the histograms that are built as a function of distance showing a hypothetical distribution: high frequency of large dBZ_e values near the coast and high frequency of low dBZ_e values offshore. LLJ indicates location and orientation of the low-level jet while TTA indicates location and orientation of terrain-trapped airflow.

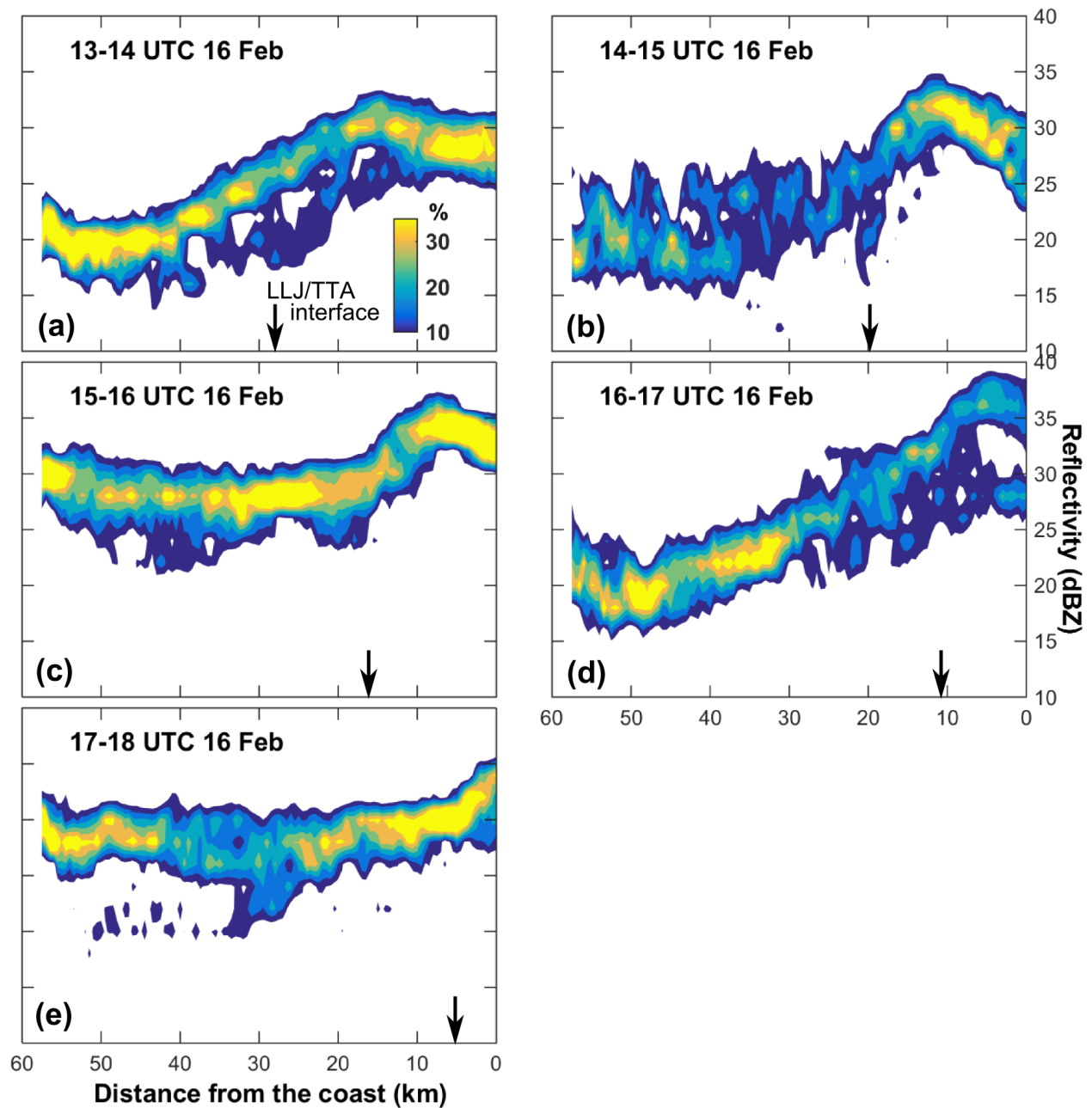


Figure 2.12: Contoured Frequency by Distance Diagrams (CFDD) of X-POL reflectivity (dBZe) for hourly-averaging periods associated with the LLJ/TTA interface progression during episode 1: (a) 13-14 UTC 16 February, (b) 14-15 UTC 16 February, (c) 15-16 UTC 16 February, (d) 16-17 UTC 16 February, and (e) 17-18 UTC 16 February. Black arrow indicates location of the LLJ/TTA interface (Fig. 2.10b). See Fig. 2.11 and the text for geometry and methodology information. Color scale for frequency is indicated in panel (a).

vertical ascent of the LLJ over the TTA and subsequent condensation. Then, once the LLJ stops its ascent and coalesced particles have fallen out, precipitation decays near the coast. Finally, close to the coastal mountains, direct orographic ascent likely contributes to the reflectivity maxima.

Thermodynamic characteristics associated with the TTA during episode 1 are documented with a time-height section of virtual potential temperature (θ_v) derived from the RASS at BBY (Fig. 2.13a). Results show that episode 1 is characterized by relatively cold temperatures between 08-16 UTC 16 February, with ~ 283 - 285 K at the surface and ~ 283 - 287 K in the lowest 500 m MSL above the surface in connection with east-southeasterly to southeasterly winds likely originating over the land. After 16 UTC 16 February, temperatures increase to ~ 285 - 287 K at the surface and ~ 287 - 291 K in the lowest 500 m MSL above the surface in association with south-southeasterly to south-southwesterly winds likely originating over the ocean. Similarly, TTA characteristics at FRS in the lowest 500 m above the surface are examined using balloon-sounding observations at 1758 UTC 16 February (Fig. 2.13b). Equivalent potential temperature (θ_e) is nearly constant (307 K), indicative of moist-neutral conditions. Relative humidity (not shown) exhibits saturated conditions (i.e. $> 90\%$) from just above the surface to mid-levels of the troposphere. Static stability is characterized with moist Brunt-Väisälä frequency (N_m , Durran and Klemp 1982), which indicates a mean value of $N = 4.3 \times 10^{-3} \text{ s}^{-1}$. Winds are southeasterly, with a slightly negative mean cross-barrier component of -1.0 m s^{-1} . While further thermodynamic analysis is addressed in section 2.5, the air within the TTA in episode 1 is characterized by a ~ 500 m depth, relatively weak stratification, and a slightly downslope flow.

2.4.2 Episode 2: 06 UTC 17 February to 06 UTC 18 February 2004

Sporadic LLJ structures are observed at BBY during episode 2 (Fig. 2.14a), with maximum meridional wind speeds of $\sim 30 \text{ m s}^{-1}$ centered at altitudes ranging from 0.6-1.2 km MSL. Wind direction in all these structures is between southerly to south-southwesterly and the distinctive vertical gradient of meridional wind speed observed during episode 1 at ~ 0.6 - 0.7 km is not apparent. After 23 UTC 17 February there is a marked shift to much weaker winds. Above the surface, winds

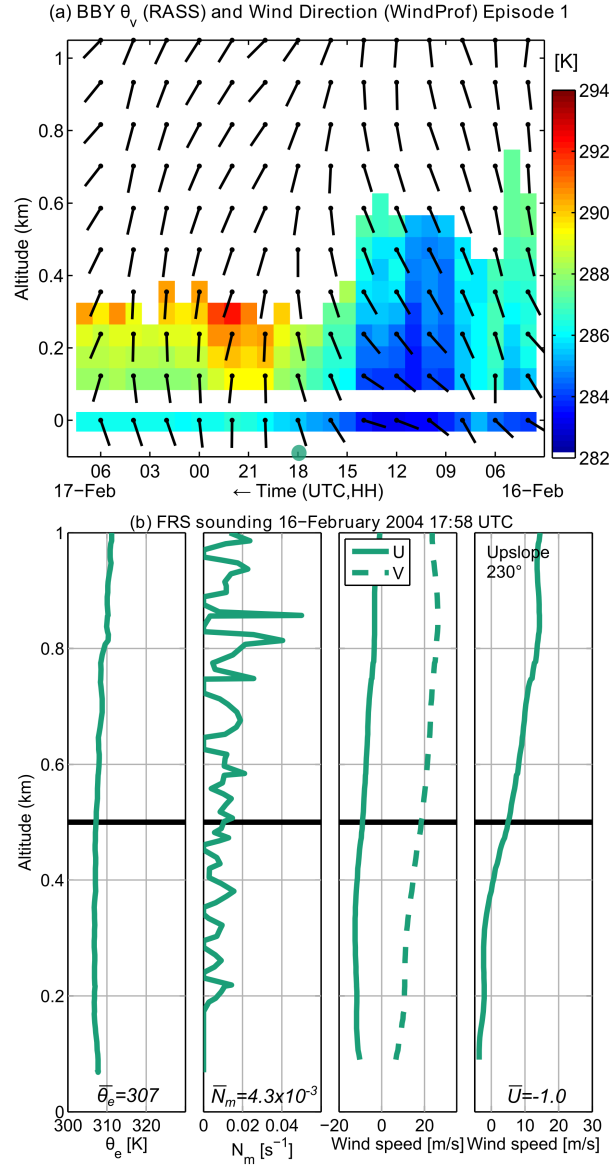


Figure 2.13: (a) Time-height analysis of virtual potential temperature (θ_v) at BBY from RASS overlaid with total wind direction (wind staffs) from the 915 MHz wind profiler at the same location. The light green circle indicates time of balloon sounding release at FRS. Time is from right to left to represent eastward advection of the storm. (b) Balloon sounding profiles at FRS: equivalent potential temperature (θ_e), moist Brunt-Visla frequency (N_m), zonal (U) and meridional (V) wind components, and cross-barrier wind from 230. Black horizontal lines indicate the 500 m altitude and mean-layer values are included.

are mainly southwesterly and meridional speeds decrease from $\sim 30 \text{ m s}^{-1}$ to $\sim 5 \text{ m s}^{-1}$. At the surface, winds are mainly south-southeasterly with meridional speeds of $\sim 10 \text{ m s}^{-1}$ that decrease to $\sim 5 \text{ m s}^{-1}$. This marked shift is most likely associated with a cold frontal passage described in section 2.3 (e.g. Fig. 2.2e). There were no high-elevation-angle X-POL PPI scans executed during episode 2 to derive VAD wind profiles at FRS. As an alternative, balloon-sounding observations were employed to produce a time-height analysis of horizontal winds (Fig. 2.14b). This analysis suggests a comparable behavior at FRS relative to BBY above 0.5 km MSL, with meridional winds of $\sim 30 \text{ m s}^{-1}$ that significantly weaken after 23 UTC 17 February. However, larger differences are evident below 0.5 km MSL, where FRS has weaker meridional wind speeds and slightly larger easterly-component wind directions compared to BBY. The character of this low-level flow structure is further explored with PPI and RHI scans.

Average Doppler radial velocity during 19-20 UTC 17 February (Fig. 2.15a) shows a zero-isodop line indicative of a smooth south-southeasterly to south-southwesterly wind-direction transition. Zero-isodop lines from PPI averages at other hours over the course of episode 2 (not shown) but before the cold frontal passage (section 2.3, Fig. 2.2e) reveal little variation relative to the 19-20 UTC 17 February average. This wind-direction transition occurs out to a range of almost 60 km, which corresponds to a beam height of $\sim 700 \text{ m}$ MSL for a 0.5° PPI. The vertical gradient of wind direction at BBY (Fig. 2.14a) indicates a smooth south-southeasterly to south-southwesterly transition below 700 m MSL from 08 to 20 UTC 17 February, in agreement with inferences from the PPI averages. Therefore, unlike the sharply curved zero isodop observed in episode 1 associated with a strong horizontal gradient in wind direction due to the TTA, the smooth and broadly curved zero isodop observed in episode 2 is more likely produced by the background vertical gradient rather than the horizontal gradient in wind direction. As a consequence, the kinematic structure inferred from the X-POL radial velocity suggests the absence of a TTA during episode 2.

The companion average reflectivity during 19-20 UTC 17 February (Fig. 2.15b) indicates background values of $\sim 18 \text{ dBZe}$ with embedded cells of $\sim 35 \text{ dBZe}$. This pattern suggests scattered convective precipitation at this hour, which is representative of the pattern observed over extended

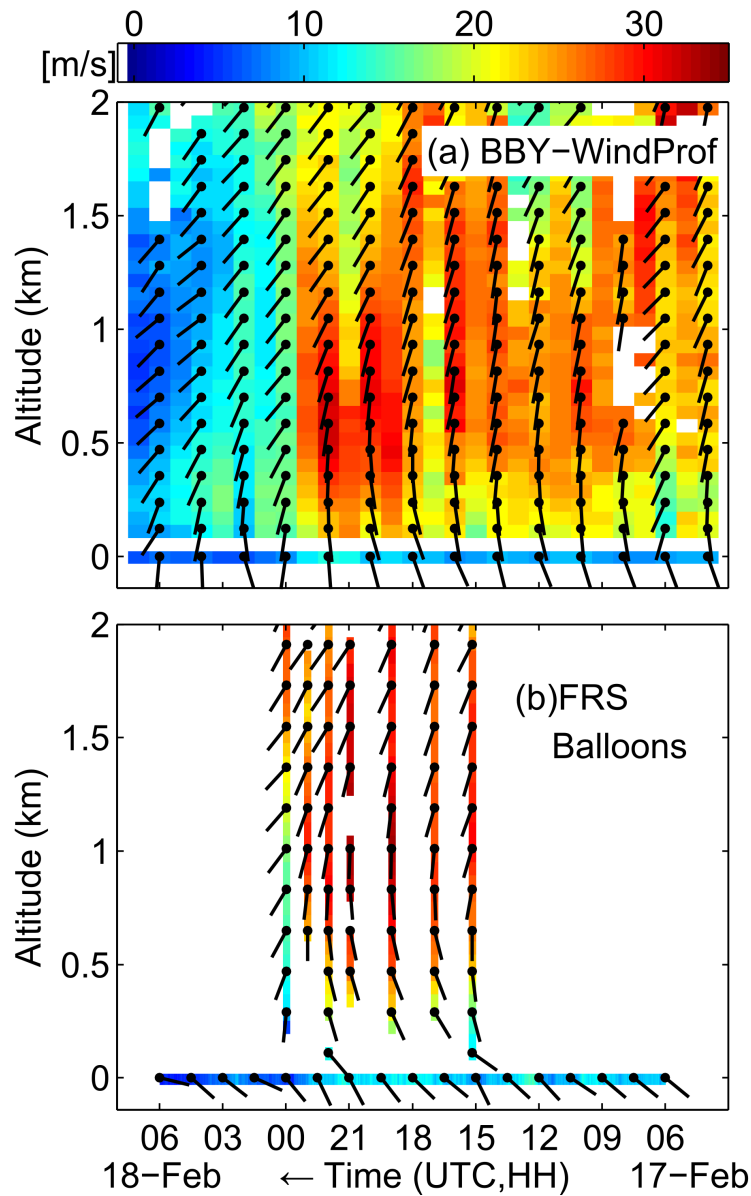


Figure 2.14: Time-height analysis of meridional wind speed (color coded, m s^{-1}) and total wind direction (wind staffs) at (a) BBY from the 915 MHz wind profiler (WindProf) and (b) FRS from balloon soundings during episode 2 with same conventions as Fig. 2.6. Balloon sounding profiles are centered at launch time and each have a 20-minute window to improve visibility.

periods of episode 2. However, the reflectivity enhancement observed offshore during episode 1 (e.g. Fig. 2.7) is absent at this hour and during the course of episode 2.

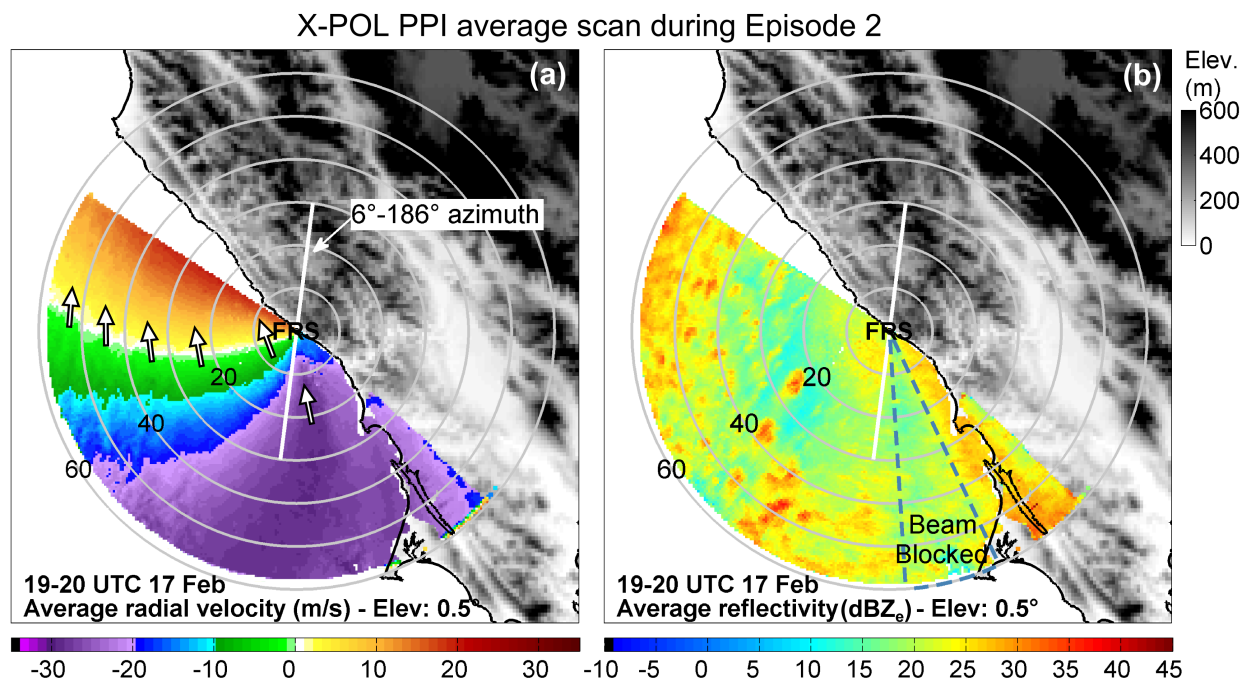


Figure 2.15: Same as Fig. 2.7 but during episode 2 between 19-20 UTC 17 February.

The vertical structure of horizontal winds and reflectivity is analyzed using hourly averages of RHI scans directed toward 6° azimuth (i.e. nearly north). These scans were executed with a maximum elevation angle of $\sim 162^\circ$ (i.e. $\sim 18^\circ$ elevation above the horizon), which allowed a view toward 186° azimuth (i.e. nearly south). The horizontal velocity between 19-20 UTC 17 February (Fig. 2.16a) shows the meridional component of the LLJ with a magnitude of $\sim 30 \text{ m s}^{-1}$ located over a weaker airflow of $\sim 10 \text{ m s}^{-1}$. This kinematic structure varies only slightly during the course of episode 2. Although there is limited RHI context for offshore low-level airflow, we hypothesize that an upward-sloping LLJ is present immediately in advance of the coastal terrain as a result of direct orographic forcing (dashed line Fig. 2.16a), which is in stark contrast to the structure observed during episode 1. Given the absence of a TTA kinematic structure during episode 2, the weaker airflow residing below the LLJ and above the windward slope (also observed with FRS balloon soundings, Fig. 2.14b) is possibly due to frictional effects of the terrain on the lower portion of the LLJ.

Companion average RHI-scan reflectivity between 19-20 UTC 17 February (Fig. 2.16b) exhibits echoes of $\sim 18 \text{ dBZe}$ below 2.5 km MSL, with maximum values of $\sim 25\text{--}28 \text{ dBZe}$ located above the windward slope and onshore over the coastal mountains. These results suggest that the absence of a TTA in episode 2 leads to the cross-barrier flow sloping upward directly over the orographic barrier, so precipitation over the coastal mountains tends to be larger relative to the coast as documented by the precipitation ratios in Table 2.2 and the hourly rain rates in Fig. 2.5.

Further contrast between the two episodes is documented with the thermodynamic characteristics of episode 2. Virtual potential temperatures at BBY (Fig. 2.17a) vary between $\sim 288\text{--}290 \text{ K}$ above the surface and $\sim 286 \text{ K}$ at the surface. Cooling of about 1-2 K after 00 UTC 18 February is most likely associated with a cold-frontal passage (e.g. Fig 2.2d,e). Hence, episode 2 is characterized by less variable and generally higher-valued BBY temperatures below 500 m compared to episode 1. A series of six balloon soundings released at FRS during the period from 1510-2259 UTC 17 February (Fig. 2.17b) indicates generally saturated conditions (i.e. $\text{RH} > 90\%$, not shown) with $\theta_e = 309 \text{ K}$, $N_m = 4.3 \times 10^{-3} \text{ s}^{-1}$, mean wind direction out of the south-southeast and a mean cross-

barrier wind of 7.8 m s^{-1} in the layer below 500 m MSL. Interestingly, the mean static stability in the lowest 500 m above FRS observed during episode 2 in the absence of a TTA is almost the same as that observed during episode 1 with a noticeable TTA. This result is addressed in the following section.

2.5 Forcing of terrain-trapped airflow

As mentioned earlier, the TTA described in this study is associated with a relatively narrow air mass flowing poleward in close proximity and approximately parallel to the west side of the coastal orography. The TTA observed during episode 1 was well documented from a kinematic and thermodynamic perspective, but its dynamic forcing was not addressed. In this section, three hypotheses for forcing of the episode 1 TTA are examined: cold pool, low-level blocking, and gap flow.

A cold pool is typically formed in association with the downdrafts from precipitating convective cells (e.g. Engerer et al. 2008). In this process, raindrops evaporate within a subsaturated environment beneath cloud base, cooling the downdraft air and spreading out horizontally at the surface, forming a cold pool and potentially a density current structure (e.g. Markowski and Richardson 2010). Although some transient and small cellular-convection structures were observed during episode 1 (Fig. 2.7b-d), they crossed the observation domain in less than an hour, which makes it unlikely that they could produce a consolidated cold pool along the coast capable of forming the observed kinematic structure associated with the TTA. Even in the absence of convective cells, a cold pool formed by simple evaporative cooling seems unlikely since surface relative humidity values at BBY, FRS and CZD and sounding profiles at FRS (not shown) are all between 85-100% from 07 UTC 16-February onwards, making it doubtful that sub-saturated conditions could lead to cold pool formation. Therefore, a precipitation-generated cold pool along the coast is rejected as a hypothesis for explaining the TTA kinematic structure.

Low-level blocking on the windward side of mountain barriers is typically associated with stably-stratified cross-barrier flows characterized by Fr (Eq. 2.1) between 0 and 1 (Smith 1979).

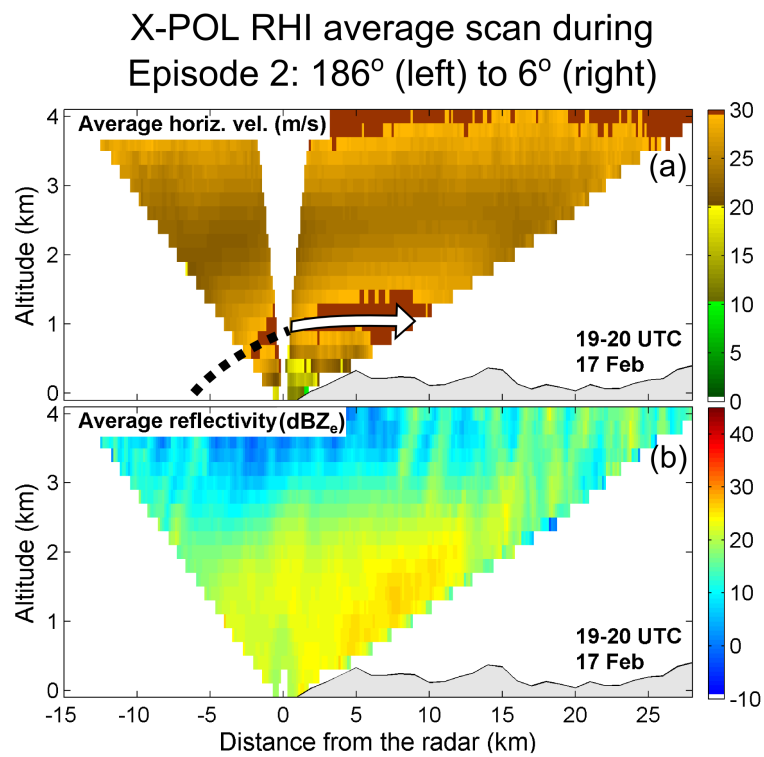


Figure 2.16: Same as Fig. 2.8 except during episode 2 between 19-20 UTC 17 February and with azimuth orientation of 186°-6°. Dashed line indicates hypothesized LLJ location offshore.

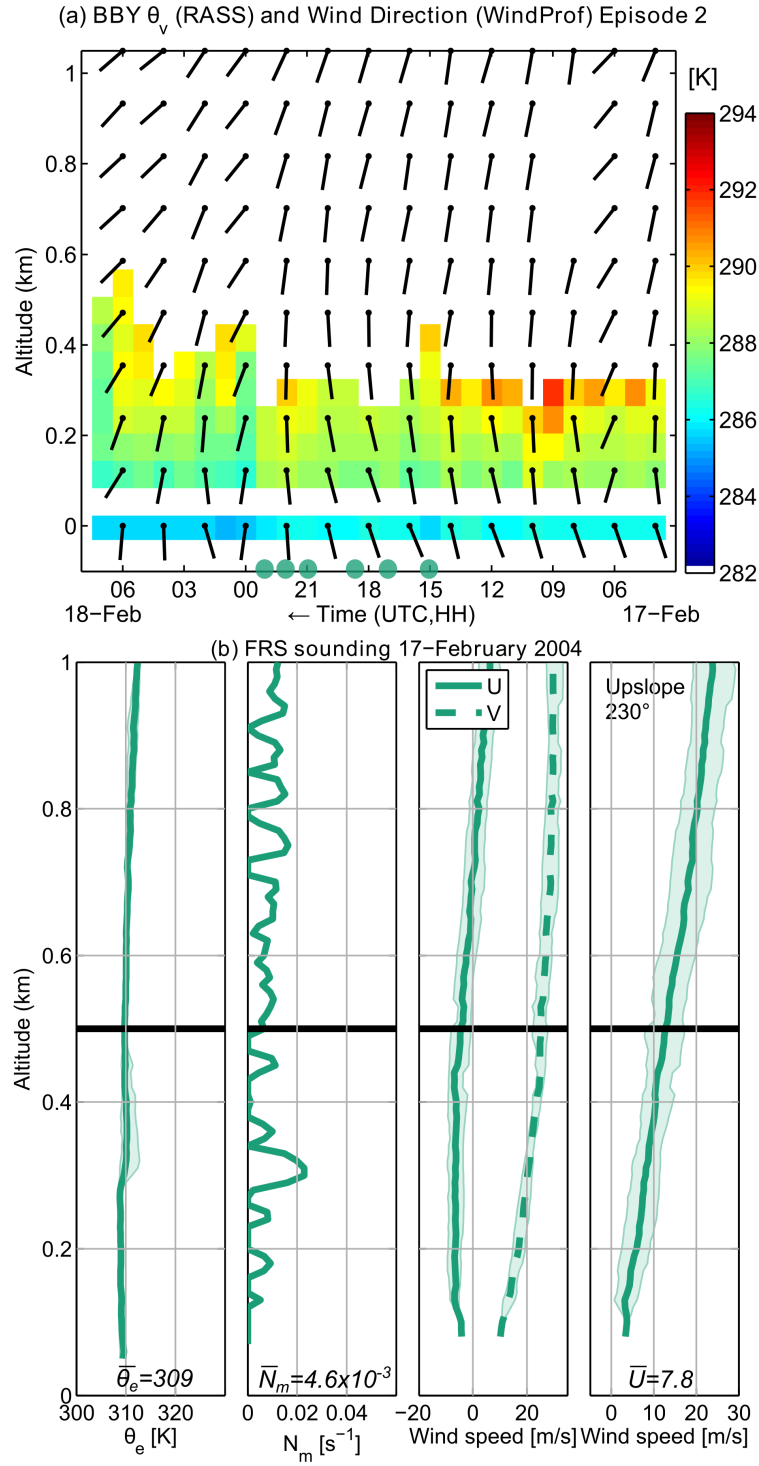


Figure 2.17: Same as Fig. 2.13 but for episode 2. Light green circles in (a) indicate launch time of single soundings. Light green lines in (b) indicate mean values considering six balloon soundings released at 1510 UTC, 1658 UTC, 1859 UTC, 2058 UTC, 2159 UTC, and 2259 UTC 17 February; shaded areas represent minimum and maximum ranges considering all these soundings.

Calculation of Fr for episode 1 employs an interpretation of Eq. 2.1 similar to that used by Hughes et al. (2009) and Kingsmill et al. (2013). A barrier height (h) of 500 m MSL is assumed, which represents an average altitude of nearby coastal-mountain peaks within a radius of 40 km from FRS. N is calculated within the perturbed air associated with the TTA and obtained from the lowest 500 m of the 1758 UTC 16 February balloon sounding at FRS (Fig. 2.13b). Given the saturated conditions (i.e. RH > 90%) observed in this layer (not shown), N_m is the most appropriate theoretical reference with a value of $N_m = 4.3 \times 10^{-3} \text{s}^{-1}$. Average cross-barrier wind speed (U , from 230°) in the lowest 500 m is derived upstream of the perturbed airflow associated with the TTA using X-POL RHI and PPI observations at ~ 1758 UTC 16 February. Winds in this layer are estimated at a range of ~ 30 km along the 180° radial from FRS. This location provides an estimate of wind conditions upwind (i.e., to the south) of the LLJ/TTA interface (see Fig. 2.10b), which is presumably also the offshore extent of stable air associated with the TTA. The 180° RHI scan (not shown) allows derivation of the TTA-layer mean meridional flow upstream of the TTA. A nearly coincident X-POL PPI scan at 0.5° elevation angle (not shown) provides an estimate of low-level wind direction to facilitate calculation of U from the RHI scan. This approach yields U equal to 20 m s^{-1} . Combining the estimates of h , N_m and U produces Fr equal to 9.3, indicating that the TTA is unlikely forced by low-level blocking. The absence of low-level blocking may explain why similar stabilities were observed in the TTA layer during episode 1 and 2 despite the dramatic differences in kinematic structure.

Gap flows are produced in mountainous regions where a relatively cold air mass crosses through gaps in the terrain as a response to pressure gradients associated with approaching synoptic-scale disturbances. Neiman et al. (2006) examined several gap-flow cases along the coast of northern California and found that a relatively cold airflow of ~ 500 m depth frequently exited the Central Valley through the Petaluma Gap (Fig. 2.1b) and crossed over BBY. With this as context, the joint behavior between along-gap pressure difference and the zonal wind at BBY during episode 1 of the present case study is now examined. Hourly pressure difference between BBY and a METAR station located over Californias Central Valley at Stockton (SCK) provides a direct measure of

the along-gap pressure gradient (Fig. 2.1b). Zonal winds were derived using surface and 0-500 m layer-mean (TTA layer) observations at BBY from the 915 MHz wind profiler. These data were temporally segregated before and after 16 UTC 16 February when temperatures below 500 m MSL at BBY transitioned from relatively cold to relatively warm (Fig. 2.13a). This time is also coincident with the LLJ/TTA interface passage over BBY (Fig. 2.10). The theoretical relationship between gap flow, pressure gradient, and friction is also explored by using the M95 equation (Eq. 2.1). The original drag coefficient ($C_D = 7.5 \times 10^{-3}$) was varied by 50% to evaluate the parameter space associated with frictional forces. Wind speed at the gap entrance ($u(0)$) was assumed to be zero since its effect is negligible at the gap exit (100 km downwind). An average air density of 1.24 kg m^{-3} derived from observations at BBY and SCK was employed.

Figure 2.18 reveals that temporal separation of the observations produces two different behaviors. The group before 16 UTC (relatively cold near-surface temperatures) tends to follow the force balance given by the M95 equation. Surface outliers could be due to underestimation of the drag coefficient (i.e. the actual C_D is larger than $1.5 \times C_D$ in M95). Also, since BBY is located west of the Petaluma Gap and therefore exposed to airflows other than the Petaluma Gap flow, the 0-500m outliers could be associated with transient mesoscale features ahead of the approaching warm front (see Section 2.3). In contrast, the group after 16 UTC (relatively warmer temperatures) significantly departs from the M95 equation, suggesting that these observations are not associated with a gap flow. Given that low-level winds over BBY up to 16 UTC 16 February are identified as part of the TTA, this analysis favors the hypothesis that a gap flow forces the TTA. As a result, airflow exiting the Petaluma Gap should have a right-turning curvature. This horizontal airflow pattern may explain the skewed orientation of the LLJ/TTA interface documented in Fig. 2.10, where positions are further offshore toward the north.

2.6 Summary and conclusions

This study has documented orographic precipitation forcing along the coastal mountains of northern California during the landfall of a significant winter storm over the period 16-18 February

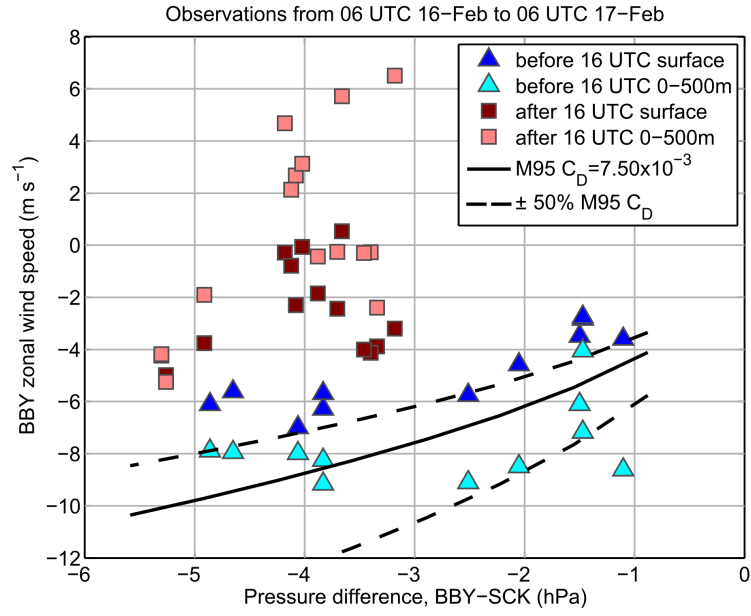


Figure 2.18: Scatter plot of hourly surface pressure difference between Bodega Bay (BBY) and Stockton (SCK) vs hourly zonal winds at BBY at surface level (dark blue, dark red) and averaged in the 0-500 m layer (light blue, light red). Observations were segregated between observations before (tringles, inclusive) and after (squares) 16 UTC 16 February based on θ_v observations (Fig. 2.13). Black lines represent the theoretical relationship between pressure gradient, surface friction, and gap flow winds provided by Mass et al. 1995 (M95, solid black line) with a 50% variation in the drag coefficient (C_D , dashed black line).

2004. Observations from scanning and profiling Doppler radars, balloon soundings, a radio acoustic sounding system (RASS), buoys, a GPS receiver, and surface meteorology sensors were part of the instrumentation employed. The scanning Doppler radar (X-POL) located near sea level on the coast was the main asset to document the kinematic and precipitation structure offshore and along the coast since it provided horizontal and vertical context near the surface.

The landfalling winter storm was associated with a synoptic-scale cyclonic circulation over the Pacific moving eastward. Warm and cold frontal passages, low-level jet (LLJ) structures, and the landfall of an atmospheric river (AR) were also evident. The event was divided into two 24-hour-duration episodes (episode 1 starting at 06 UTC 16 February and episode 2 starting at 06 UTC 17 February) based on different surface rain-rate characteristics along the coast.

LLJ structures were characteristics of both episodes; however, a terrain-trapped airflow (TTA) structure was observed only during episode 1. A schematic that summarizes the kinematic and precipitation structure associated with the LLJ and TTA interaction is presented in Figure 2.19. The TTA structure was observed for ~ 5 hours and up to ~ 25 km offshore of the coastline with a depth of ~ 0.5 km MSL. As episode 1 progressed, the LLJ/TTA interface moved closer to shore with a skewed orientation relative to the mean coastline orientation such that positions are further offshore toward the north. Meanwhile, the incoming LLJ sloped from an average altitude of ~ 0.25 km MSL offshore to ~ 1 km MSL over the coast. FRS presented a weakly stratified atmosphere associated with the TTA; hence, the LLJ was probably displaced upward through isentropic lift because horizontally moving air parcels would tend to conserve their potential temperature (e.g. Saucier 1989, Neiman et al. 2002).

A precipitation enhancement zone was observed offshore in association with convergence from the LLJ/TTA interface. The precipitation enhancement zone moved toward the coast as the LLJ/TTA interface moved coastward. Convergence produced by synoptic flows (e.g. LLJ) and TTAs has been previously documented, such as in Neiman et al. (2004), Yu et al. (2007), Yu and Hsieh (2009). However, while the former studies document the origin of the TTA as low-level blocking, the present study documents a TTA formed by an offshore-directed gap flow.

During episode 2, the kinematic structure along the coast and offshore was fundamentally different. The LLJ/TTA interface was absent, so that the LLJ sloped upward directly over the windward side of the coastal mountains. Consequently, orographic precipitation enhancement over the coastal mountain was likely achieved through an upslope flow effect (i.e. forced ascent by the orography).

The contrast between episode 1 and 2 in terms of LLJ lifting (offshore and onshore, respectively) and precipitation enhancement response is consistent with the observed mountain (CZD)-to-coastal (FRS and BBY) precipitation ratios (average of 1.2 in episode 1 and 4.2 in episode 2; Table 2.2), and with previous studies (0.71 blocked and 5.6 unblocked; Neiman et al. 2002). Hourly rain rates (Fig. 2.5a) are also in agreement with offshore and onshore lifting, showing consistently larger rain rates over FRS and BBY for several hours during episode 1 and larger rain rates over CZD during almost the entire course of episode 2.

Forcing of the TTA was investigated by analyzing cold pool, low-level blocking and gap flow hypotheses. This analysis favored the hypothesis that the TTA is forced by a gap flow exiting the Petaluma Gap that encounters the onshore airstream associated with the LLJ. TTA formation in this manner is consistent with previous studies performed along the Gulf of Alaska (Loescher et al. 2006). Since offshore-directed gap-flows in the Northern Hemisphere have a right-turning curvature, this forcing mechanism could explain the observed skewed orientation of the LLJ/TTA interface where positions are further offshore toward the north.

Although thermodynamic characteristics associated with the TTA along the coast were documented, profiling RASS and balloon soundings only provided observations at single points (BBY and FRS, respectively) and were sparse in time in the case of balloon soundings. As a result, it is unclear if mixing was occurring at the LLJ/TTA interface and if that could have affected the lifting of the LLJ. In addition, it is interesting to contrast these results with Neiman et al. (2006), whose gap flow inventory does not account for the case presented here (perhaps corresponding to a weaker-than-normal gap flow case) and who found that a linear relationship between pressure difference and zonal wind best represented a 5-day gap-flow case. Since in our case study a linear

relationship is less clear, one remaining question is the relative importance of frictional forces in determining gap flow magnitude. Another remaining question is the representativeness of the kinematic and precipitation structures observed in this case relative to other winter storm cases within the same domain. This question will be addressed in future research by the authors.

Chapter 3

Terrain-trapped airflows and orographic rainfall along the coast of northern California. Part I: Kinematic characterization using a wind profiling radar

This chapter is in preparation to be submitted to:

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3.1 Introduction

A number of studies have shown that poleward horizontal water vapor transport from the tropics to midlatitudes is mostly achieved through a narrow corridor of less than ~ 1000 km width and larger than ~ 2000 km length (e.g. Zhu and Newell 1994, Ralph et al. 2004; 2005, Bao et al. 2006). This form of water vapor transport, known as an atmospheric river (AR), is commonly found in the warm sector of extratropical cyclones and collocated with the pre-frontal low-level jet (LLJ, e.g. Ralph et al. 2005). The west coast of the United States often experiences the landfall of these extratropical cyclones and concomitant ARs during wintertime. These storms can produce copious amounts of precipitation with a spatial distribution that is modulated by mountains (e.g., Colle et al. 2008, Hughes et al. 2009). Modulation of the background synoptic precipitation pattern due to the presence of mountains is known as orographic precipitation (e.g. Colle et al. 2013).

West coast terrain, particularly along northern California, is characterized by mountain peaks of 500-1000-m MSL, a ridge orientation of roughly 140° relative to the north, and steep slopes around 45° right along the coastal line. When moist, statically-neutral airflow impacts these slopes,

air is generally lifted and cooled, which produces condensate and, after a finite time, precipitation-sized hydrometeors. This upslope flow mechanism is a relatively simple conceptual model that can explain a significant fraction of orographic rainfall over midlatitudes. However, mountains can produce their own mesoscale circulation and modify the spatial precipitation pattern normally seen during the approach of extratropical cyclones and upslope flow events. One example is the presence of a terrain-trapped airflow (TTA) on the windward side of orographic barriers. According to Valenzuela and Kingsmill (2015; henceforth VK15), a TTA is defined as relatively narrow air mass consistently flowing in close proximity and approximately parallel to an orographic barrier. Thus, in the west coast of the United States TTAs flow poleward along the western side of coastal mountain ranges.

TTA impacts on orographic precipitation have been studied in association with several large-scale mountain ranges (e.g. altitudes above ~ 1000 m MSL) such as the European Alps (e.g., Medina et al. 2005), Southern Alps of New Zealand (e.g., Sinclair et al. 1997), Taiwan (e.g., Yu and Hsieh 2009), Colorado’s Rocky Mountains (e.g., Peterson et al. 1991), and California’s Sierra Nevada (e.g. Kingsmill et al. 2013). In contrast, TTA impacts associated with small-scale mountain ranges (e.g. altitudes below ~ 1000 m MSL) have received much less attention. One of these relatively rare studies (Neiman et al. 2002) used wind profiler and rain gauge data along the coastal mountains of California to describe the relationship between upslope wind speed and orographic precipitation. By studying six west coast winter storms, they found that the presence of a TTA decreased the average mountain to coast precipitation ratio from 5.06 to 1.26 and hypothesized that this occurred due to upstream lifting of the pre-frontal LLJ offshore by the TTA. VK15 was another study that addressed this problem by examining ground-based scanning Doppler radar observations along and offshore of the northern California coastal mountains during a significant landfalling winter storm to detail the three-dimensional kinematic and precipitation structure associated with a TTA. The structures presented in their study showed that the TTA was responsible for upstream lifting of a LLJ. Furthermore, VK15 documented a precipitation enhancement zone roughly 30 km offshore and near parallel to the coast associated with TTA-lifting of the LLJ. Both of these observations

support the hypothesis offered by Neiman et al. (2002).

Although VK15 and Neiman et al. (2002) provide some insights about kinematic characteristics and effects of TTAs on orographic rainfall using case studies, there is still a lack of an objective method to identify TTA events. For example, given the terrain characteristics a wind direction of 140° would work as a first guess for identifying the TTA wind regime but no detailed evaluation of this threshold has been made. As result, in this study we develop an objective identification of TTAs using a long-term 13-season dataset. This study is unique because it provides an objective identification method and documents physical effects of TTAs on orographic rainfall using a statistically significant (13-season) dataset. While this study is applied to the coast of northern California, the method is still applicable to any midlatitude mountain range.

Section 3.2 describes the observing systems and data processing techniques employed in the analysis. A statistical characterization of low-level winds is described in section 3.3, while physical characteristics and orographic rainfall impacts of TTAs is provided in section 3.4. Finally, section 3.5 presents the summary and conclusions of this study. In Part II of the present study (Valenzuela and Kingsmill 2017b), we apply the objective TTA identification method to document kinematic and precipitation structures in 7 case studies observed with a ground-based X-band scanning Doppler radar, performing an inter-storm and regime (TTA, NO-TTA) comparison, and including a theoretical discussion of TTA forcing mechanisms.

3.2 Observing systems

Observations employed in this study were collected along the northern California coast as part of the California Land-Falling Jets (CALJET), Pacific Land-Falling Jets (PACJET) and Hydrometeorology Testbed experiments (Ralph et al. 2013) operated by the National Oceanic and Atmospheric Administrations (NOAA) Earth System Research Laboratory (ESRL). Locations of key observing systems are shown in Figure 3.1. Dates and number of hours included from each season are presented in Table 3.1.

The main observational asset is a 915-MHz wind-profiling radar (Ecklund et al. 1988) located

at Bodega Bay (BBY), which provided hourly vertical profiles of horizontal winds. Each profile was processed with the continuity method of Weber et al. (1993) that checks consistency in the dataset over time and height. Since wind-profiler gate spacing and first gate altitude vary slightly between seasons, each profile was interpolated to a common 40-gate grid having 92-m vertical spacing with the first gate at 160-m MSL. In addition, two-minute resolution measurements of surface winds and rain accumulation were available at BBY and Cazadero (CZD). After quality control, hourly mean winds (speed and direction) and hourly rain accumulations were derived from the native resolution observations at both sites. Hours with missing surface (BBY and CZD) or wind profiler (BBY) data were discarded, ensuring the presence of simultaneous observations.

Table 3.1: Winter seasons included and their start and end dates and time (UTC). Number of hours when rain gauge (CZD and BBY) and surface and wind profiler observations (BBY) were available simultaneously with non-missing observations is included. Last column includes a subset of this group when rain ≥ 0.25 mm was observed at CZD.

Winter season	Start	End	Hours	Hours CZD rain ≥ 0.25 mm
1998	0000 UTC 01 Jan 1998	2300 UTC 31 Mar 1998	1839	604
2001	0000 UTC 12 Jan 2001	2000 UTC 08 Mar 2001	1153	276
2002	0200 UTC 11 Jan 2002	2300 UTC 06 Apr 2002	2062	201
2003	0000 UTC 09 Dec 2002	2300 UTC 09 Apr 2003	2784	458
2004	2200 UTC 13 Dec 2003	2300 UTC 21 Mar 2004	2378	391
2005	1600 UTC 12 Nov 2004	2300 UTC 01 Apr 2005	3330	487
2006	0000 UTC 15 Nov 2005	2300 UTC 25 Apr 2006	3145	726
2007	0000 UTC 01 Dec 2006	2300 UTC 30 Apr 2007	3624	428
2008	0000 UTC 21 Nov 2007	2300 UTC 09 Apr 2008	3384	420
2009	0000 UTC 08 Nov 2008	2300 UTC 11 May 2009	4440	533
2010	0000 UTC 06 Nov 2009	2300 UTC 11 May 2010	4233	594
2011	0000 UTC 02 Nov 2010	2300 UTC 31 Mar 2011	3312	533
2012	0000 UTC 16 Nov 2011	2300 UTC 01 May 2012	4032	430
Total			39716	6081

3.3 Statistical characteristics of low-level winds

The overall joint distribution of wind speed and direction at single vertical levels up to ~ 3000 -m MSL is shown in Figure 3.2. At surface there is a bimodal distribution with one of the modes

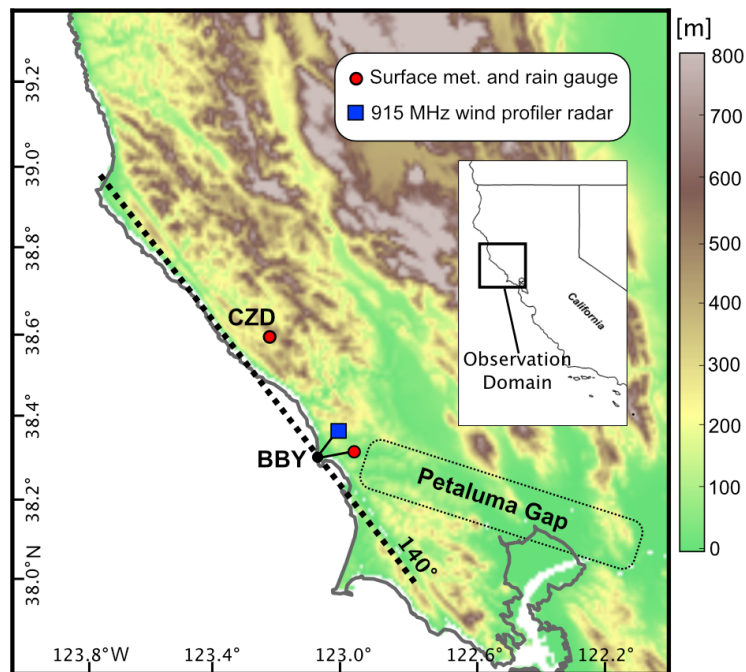


Figure 3.1: Topographic map overlaid with observing systems. Legend identifying each instrument along with the color scale for terrain elevation (m) are included. Dashed line (azimuth 140°) is the reference for the mean orientation of the coastal range (e.g., Neiman et al. 2002).

associated with predominant northwesterly flow and maximum wind speeds of 15 m s^{-1} . The other mode is associated with easterly flows and maximum wind speeds of 9 m s^{-1} . At 160-m and 344-m MSL the wind distribution becomes unimodal, with predominant northwesterly winds and maximum speeds around 21 m s^{-1} . From 528-m MSL, the dominant mode still is associated with northwesterly winds but the distribution start spreading progressively such that at 2001-m MSL it becomes multi-modal. Above 2001-m MSL, modal winds are associated with westerly winds. Considering modal wind directions, Figure 3.2 indicates a relatively small vertical wind shear in horizontal winds.

Since our study is focused on orographic rainfall, we examine how the distribution of winds are different when it is raining over the coastal mountains by selecting hours when rain rate at $\text{CZD} \geq 0.25 \text{ mm}$ (gauge resolution). Figure 3.3 shows that during rainy hours winds are generally stronger than the overall dataset. Surface distribution changes its first mode to southeasterly winds, keeping the second mode associated to easterly flows. Between 160-m and 344-m MSL the wind distribution becomes unimodal, with predominant southeasterly winds and maximum speeds around 21 m s^{-1} . From 528-m MSL upward, wind distributions are more widespread, have multimodal characteristics, and present a wind shift toward southerly and southwesterly winds in the dominant modal wind direction. In addition, increasing frequency of winds above 21 m s^{-1} is observed from 1541-m MSL upward.

Details about the mean wind speed and direction profile is shown in Figure 3.4. Standard deviation for speed and direction is estimated by using sample and angular variance, respectively (Weber 1991). The overall dataset shows monotonically increasing wind speed from $\sim 4 \text{ m s}^{-1}$ at surface to $\sim 14 \text{ m s}^{-1}$ at 3000-m MSL, with speeds increasing rapidly in the lowest ~ 250 -m MSL (Fig. 3.4a). Wind speed standard deviation is nearly $\pm 2.5 \text{ m s}^{-1}$ near surface and $\pm 5 \text{ m s}^{-1}$ from ~ 250 -m MSL upward. Corresponding mean wind direction profile indicates a shift from 300° at surface to 240° at 3000-m MSL (Fig. 3.4b). The angular standard deviation indicates a variation near surface of roughly $\pm 100^\circ$ decreasing toward upper levels to roughly $\pm 60^\circ$. Frequency of non-missing gates indicates a loss of good gates with altitude, especially after 1000-m MSL (Fig. 3.4c).

Wind roses at BBY for winter-season hours

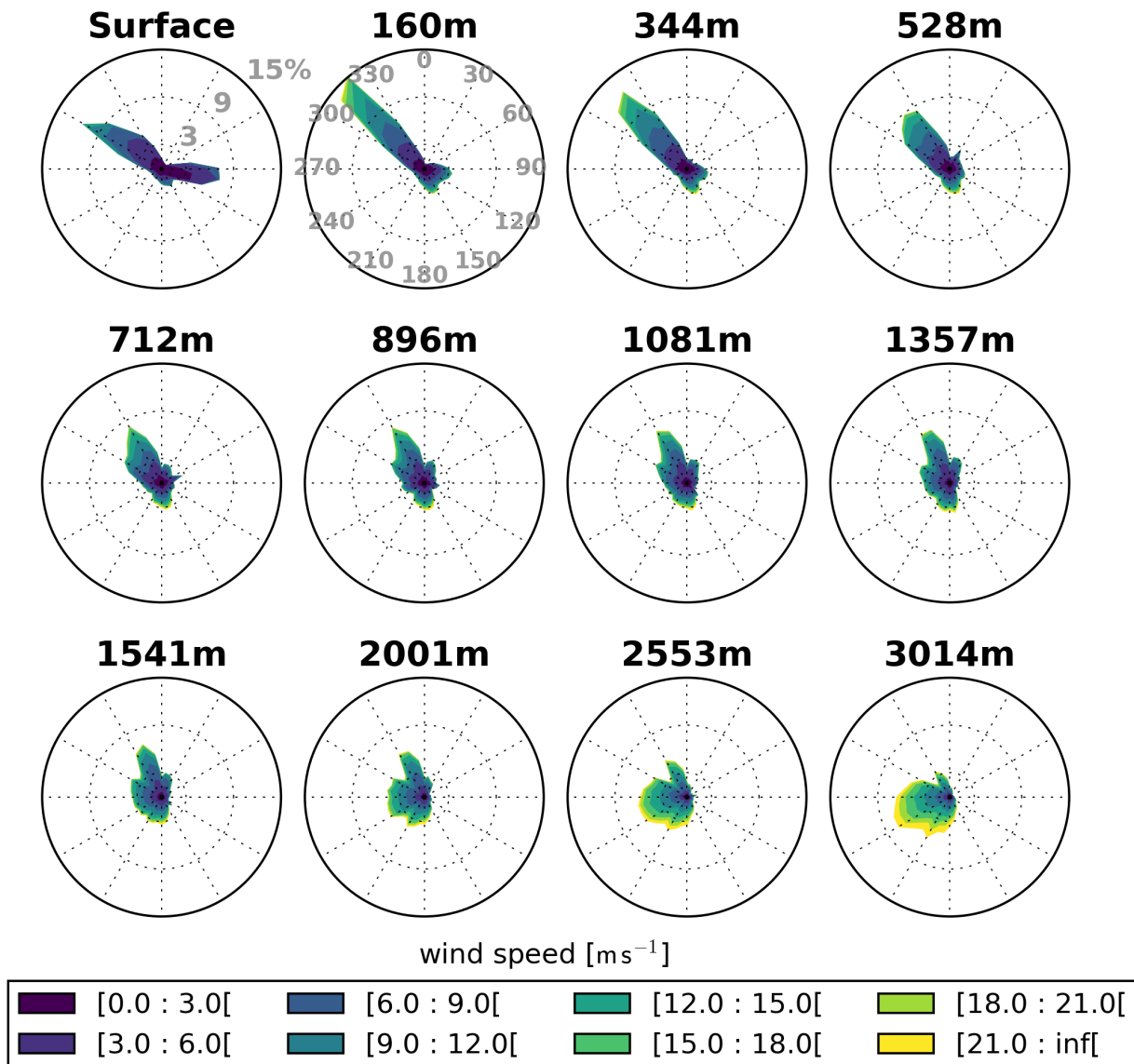


Figure 3.2: Wind roses of observations at Bodega Bay (BBY) including all 13-season dataset from surface and 915 MHz wind profile observations up to ~ 3000 -m MSL. Wind direction and speed bins are 10° and 3 m s^{-1} , respectively.

Wind roses at BBY for czd-rain hours

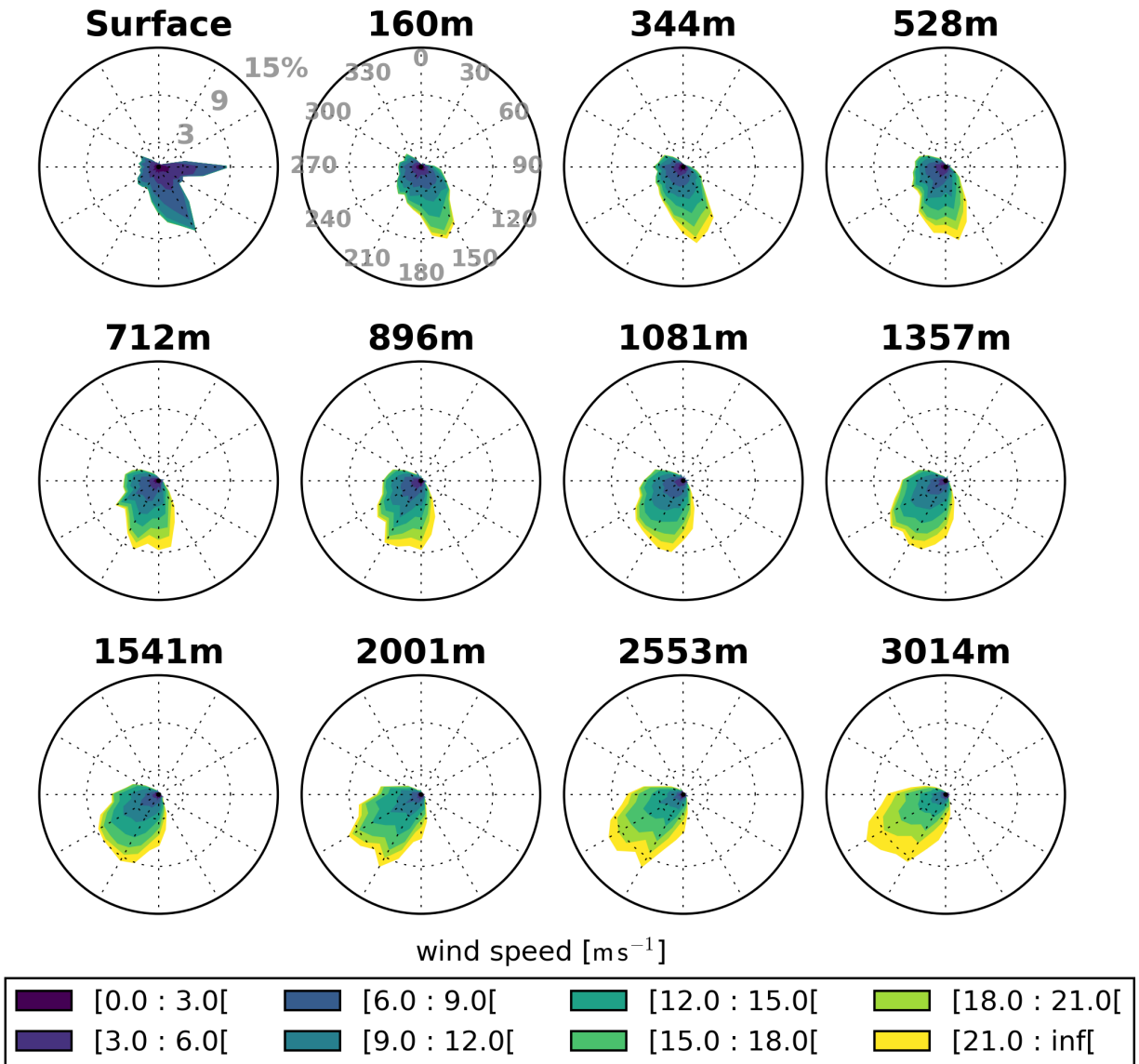


Figure 3.3: Same as Figure 3.2 but for hours when rainfall ≥ 0.25 mm is observed at CZD.

When selecting rainy hours, the mean wind speed profile present winds of $\sim 6 \text{ m s}^{-1}$ at the surface increasing to $\sim 17 \text{ m s}^{-1}$ at 3000-m MSL (Fig. 3.4d). The standard deviation shows nearly $\pm 2.5 \text{ m s}^{-1}$ near surface and $\pm 5 \text{ m s}^{-1}$ in the rest of the profile except for a slightly larger variation between 250-m and 1250-m MSL. Corresponding mean wind direction profile shows winds shifting from $\sim 170^\circ$ at surface to $\sim 230^\circ$ at 3000-m MSL (Fig. 3.4e). The angular standard deviation indicates a variation near surface of roughly $\pm 60^\circ$ decreasing toward upper levels to roughly $\pm 30^\circ$. Frequency of non-missing gates indicates that the rainy subset has less loss of good gates with altitude, with more than 90% of good gates below 2000-m MSL (Fig. 3.4f).

Further details of the mean wind component profile is presented in Figure 3.5. The overall mean profile of U-component exhibits only positive mean velocities along the profile, with values around 1 m s^{-1} from surface to 900-m MSL and monotonically increasing velocities up to 8 m s^{-1} at 3000-m MSL (Fig. 3.5a). Profile dispersion (one standard deviation) holds almost constant with altitude, slightly increasing after 2500-m MSL. The V-component shows a weak negative velocity below ~ 250 -m MSL, mean velocities of $\sim 0 \text{ m s}^{-1}$ from 250-m to 1500-m MSL, and monotonically increasing velocities up to 4 m s^{-1} at 3000-m MSL (Fig. 3.5b). Profile dispersion increases from surface to 3000-m MSL and exhibits a larger magnitude relative to the U-component. Larger dispersion in the meridional component might be linked with the passage of synoptic systems and cyclonic circulation.

When selecting only rainy hours, the U-component shifts its mean velocity near the ground to roughly -1 m s^{-1} , monotonically increasing to 11 m s^{-1} from ~ 250 -m MSL upward (Fig. 3.5d). Profile dispersion is nearly constant with altitude and, unlike the overall U-component dispersion, indicates the presence of an easterly jet structure below 500-m MSL with maximum absolute velocity of 7 m s^{-1} considering one standard deviation. The V-component is significantly different relative to the overall dataset, with a dominance of southerly winds along the entire profile, most likely associated with pre-frontal environment of baroclinic wave passages (Fig. 3.5e). Velocity increases rapidly from $\sim 3 \text{ m s}^{-1}$ to 8 m s^{-1} in the lowest 500-m, from where start exhibiting a small fluctuation around 8 m s^{-1} up to 2500-m MSL and monotonically increase afterwards reach-

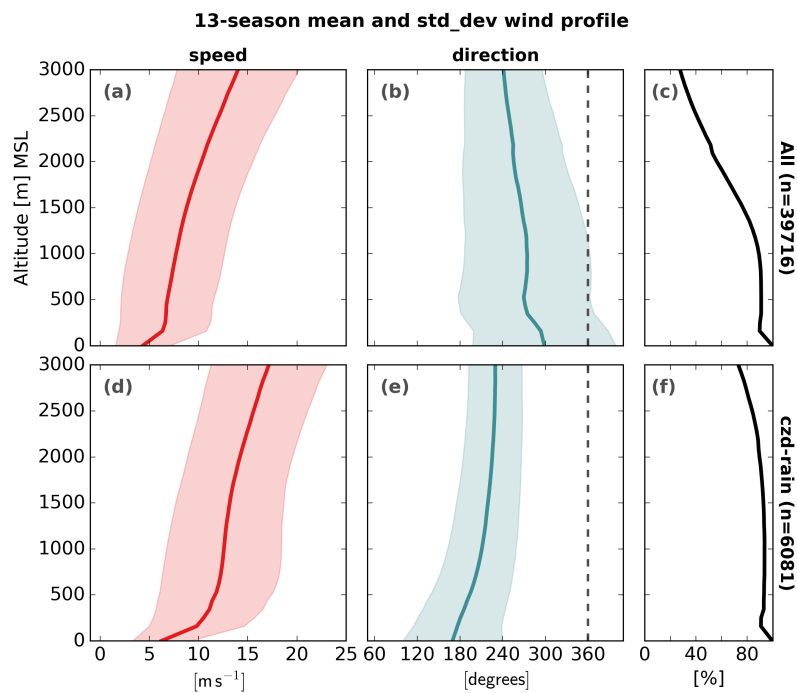


Figure 3.4: 13-season profile of mean wind speed and direction (bold lines) and standard deviation (shaded area) for entire dataset (a,b) and the subset of hours when rainfall ≥ 0.25 mm is observed at CZD (c,d). Count of good gates at each vertical level is included for the entire dataset (c) and the subset of rainy hours (f)

13-season mean and std_dev wind component profile

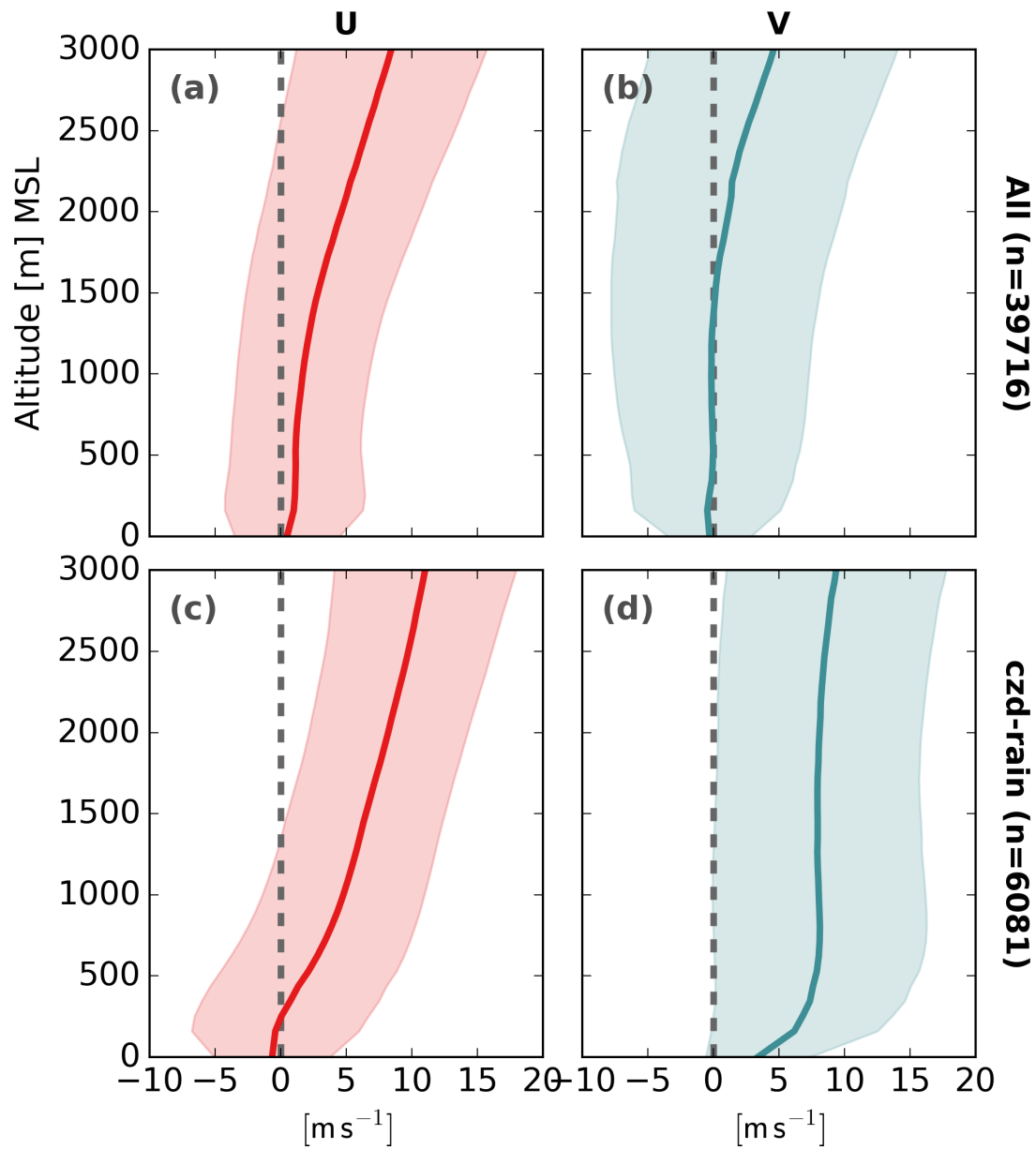


Figure 3.5: 13-season mean profile of U (zonal) and V (meridional) wind components (bold lines) for the entire dataset (a,b) and the subset of hours when rainfall ≥ 0.25 mm is observed at CZD (d,e). Shaded areas indicate one standard deviation.

ing 9 m s^{-1} at 3000-m MSL. Profile dispersion is relatively small near the surface and increases rapidly until ~ 500 -m MSL, from where it holds relatively constant with altitude.

One striking difference between the overall and rainy subset is the rapid change of velocities with altitude in the bottom part of the U and V profiles, especially near or below 500-m MSL. We examine these changes with profiles of vertical wind shear. In the overall dataset, U and V component shear profile (Figure 3.6a,b) indicate mostly positive shear with maximum values of $5 \times 10^{-3} \text{ s}^{-1}$. The resulting vector magnitude indicates nearly equal contribution from both components. During rainy hours, the U-component shear features a maximum of $\sim 9 \times 10^{-3} \text{ s}^{-1}$ at 500-m MSL, with values around $2 \times 10^{-3} \text{ s}^{-1}$ along the rest of the profile. The V-component features a rapidly decreasing shear from $\sim 17 \times 10^{-3} \text{ s}^{-1}$ near surface to $\sim 1 \times 10^{-3} \text{ s}^{-1}$ at 500-m MSL. Afterwards, the shear profile oscillates around $1 \times 10^{-3} \text{ s}^{-1}$. Consequently, the vector magnitude indicates a relatively large wind shear near surface decreasing rapidly from $17 \times 10^{-3} \text{ s}^{-1}$ to $8 \times 10^{-3} \text{ s}^{-1}$ at 500-m MSL and oscillating around $4 \times 10^{-3} \text{ s}^{-1}$ after 1000-m MSL. The strong vertical shear below 1000-m MSL is probably linked with enhanced surface friction along the coastal mountains.

These results suggest that winds in the rainy environment have characteristics that are distinctly different from those in the overall dataset. For example, the rainy subset presents in general larger velocities along the profile with a relatively small variation in wind directions and predominant southerly airflows, probably related with a predominant meridional orientation of baroclinic wave passages. In addition, there are some characteristics that appear linked with the lowest 500-m MSL, which is approximately the same depth as the nearby coastal terrain: rapidly increasing values of V-component velocities, maximum vertical shear, and easterly jet structure.

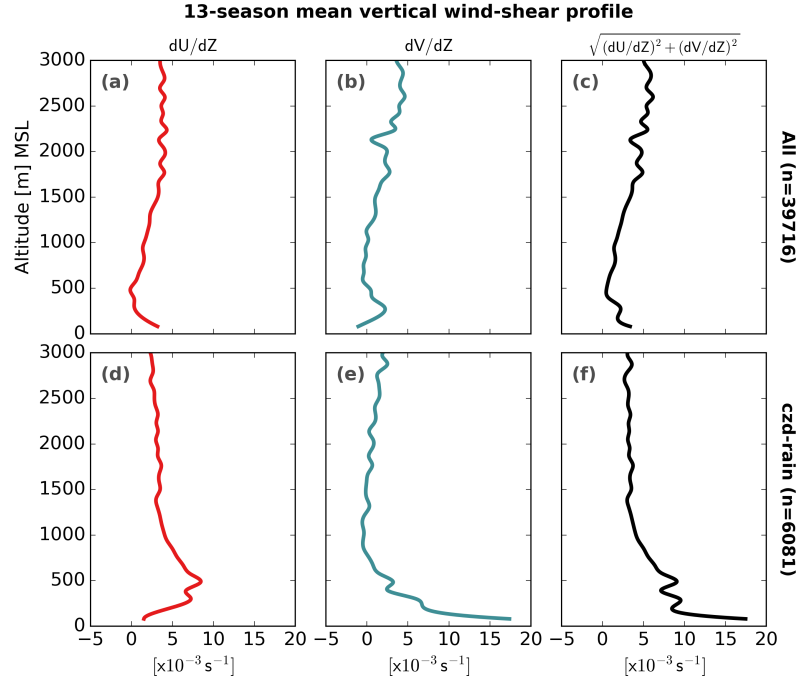


Figure 3.6: 13-season mean vertical shear profile of U and V components (a,b) and vector magnitude (c) for the entire dataset. Similar profiles for the subset of hours when rainfall ≥ 0.25 mm is observed at CZD is showed in d,e,f.

3.4 Terrain-trapped airflows

3.4.1 Objective identification

The relationship between upslope flow and orographic rainfall in the coastal mountains of northern California was studied in Neiman et al. (2002). By observing a detailed topographic map, they determined that the coastal ridge have an average orientation of 320° - 140° (roughly northwest to southeast) and therefore winds from 230° were considered as cross-barrier (i.e. upslope) winds. The 230° wind direction has also been used in other studies to compute upslope component of incoming winds and bulk integrated water vapor flux (Neiman et al. 2005; 2009, Kingsmill et al. 2016). As a result, a wind direction of 140° would be indicative of TTA winds. In addition to the terrain orientation, examination of a detailed topographic map also indicates that the average altitude of this ridge is approximately 500-m nearby BBY and Cazadero (CZD). This altitude is also associated with gap flows observed along Petaluma Gap, nearby BBY (e.g., Neiman et al. 2006).

Considering the TTA definition and aforementioned terrain characteristics, we select hourly TTA wind profiles when the mean wind direction in the lowest 500-m MSL ($\overline{WDIR}_{500}$) is less than 140° . Given that wind directions less than 90° are uncommon (Fig. 3.4d), we can safely use an upper-limit wind direction to discriminate wind regimes associated with TTAs and allow fluctuation in the terrain-parallel airflow direction. The NO-TTA subset is then composed by those profiles with $\overline{WDIR}_{500} \geq 140^\circ$. Zonal (U) and meridional (V) wind components are derived for the TTA and NO-TTA subset. Since the focus is on orographic rainfall, we apply this analysis to hours when rain rate at CZD is $\geq 0.25\text{mm}$.

The mean U-component profile in the TTA regime shows easterly winds below $\sim 1400\text{-m}$ MSL and westerly winds above this altitude (Fig. 3.7a). In addition, there is a jet structure of easterly winds with maximum absolute velocity of $\sim 8 \text{ m s}^{-1}$ residing below 500-m MSL and centered at $\sim 250\text{-m}$ MSL. The mean V-component profile indicates a rapid increase of southerly winds from ~ 0 to 10 m s^{-1} between surface and 1000-m MSL and a small fluctuation around 10 m s^{-1} above (Fig.

3.7b). Contrasting profiles of NO-TTA regime result in a mean U-component profile monotonically increasing and with westerly winds (Fig. 3.7c), while the mean V-component profile present a more modest increase of southerly winds with altitude, fluctuating around $\sim 7 \text{ m s}^{-1}$ between $\sim 250\text{-}3000\text{-m}$ MSL (Fig. 3.7d).

To estimate an average time-scale for the TTA regime, we select hours when $\overline{WDIR}_{500} < 140^\circ$ and count the total number of hours in each winter season using the rainy hour subset. Similarly, to estimate the number of TTA events per season we computed the number of continuous hours, where a single hour is counted as one event. Figure 3.8a shows that the number of TTA hours is as low 25 in 2002 and as large as 144 in 2005. These hours are distributed in 13 and 54 events, respectively (Fig. 3.8b). The number of hours per events in each winter season shows that long-lasting TTA events occurred in 2003 with roughly 3.2 hours per event, whereas short-lasting events occurred during 2002, with approximately 1.9 hours per event (Fig. 3.8c). The average duration of TTA events during rainy conditions is 2.5 hours.

We apply the objective identification to a storm observed during 16 February 2004, which correspond to episode 1 in VK15. In this storm, a TTA was documented between 0900-1600 UTC, with a LLJ-TTA interface crossing BBY between 1500-1600 UTC. Figure 3.9 indicates that by using $\overline{WDIR}_{500} < 140^\circ$ during at least 1 hour ($\text{nh} \geq 1$), a TTA regime between 1100-1500 UTC is identified. This period is shorter than the TTA period identified in VK15 and misses the time when LLJ-TTA crosses BBY.

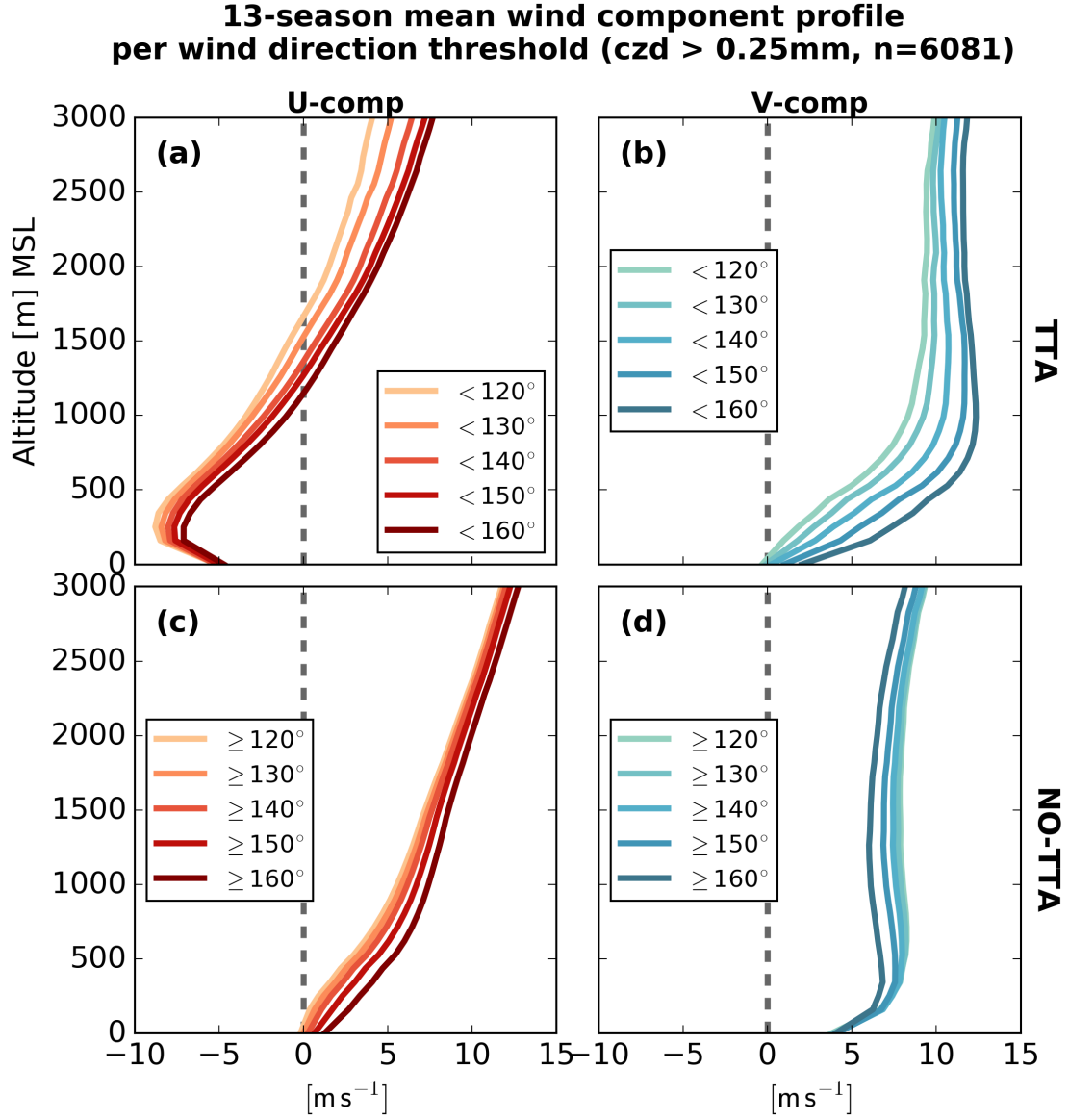


Figure 3.7: 13-season mean profile of U (zonal) and V (meridional) wind components for different thresholds of TTA (a,b) and NO-TTA (c,d) regime during hours when rainfall ≥ 0.25 mm is observed at CZD.

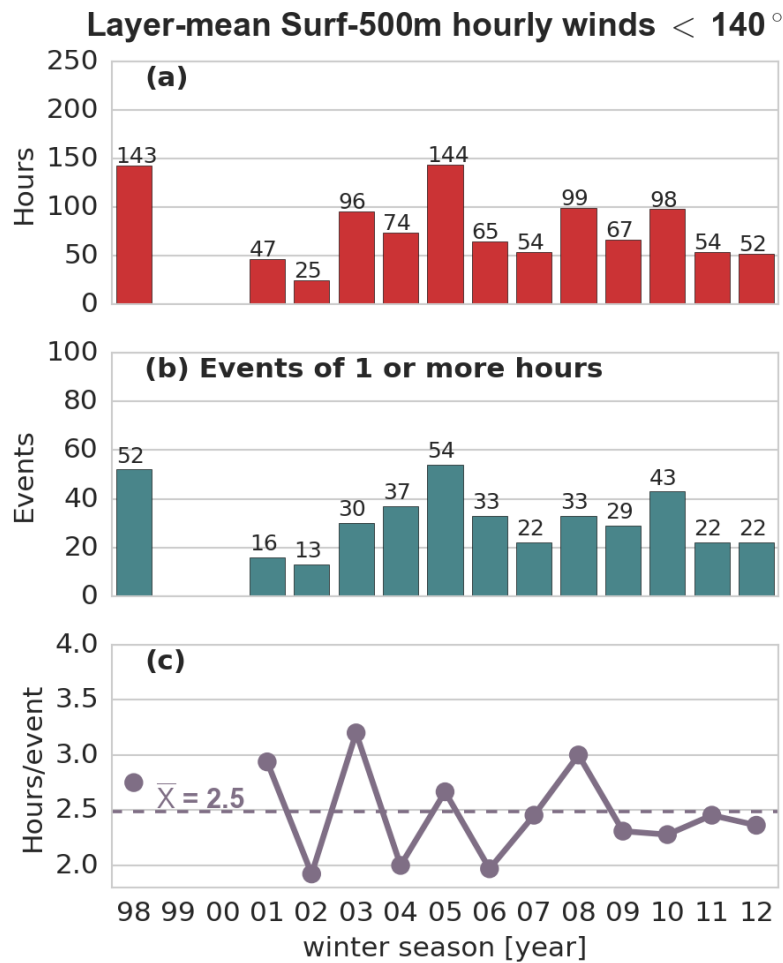


Figure 3.8: Seasonal statistics of TTAs: (a) total number of hours of observed TTAs in each season, (b) number of TTA events in each season, and (c) number of hours per TTA event. Statistics are from the subset of hours when rainfall ≥ 0.25 mm is observed at CZD. Dashed line in (c) shows the average duration of TTA events.

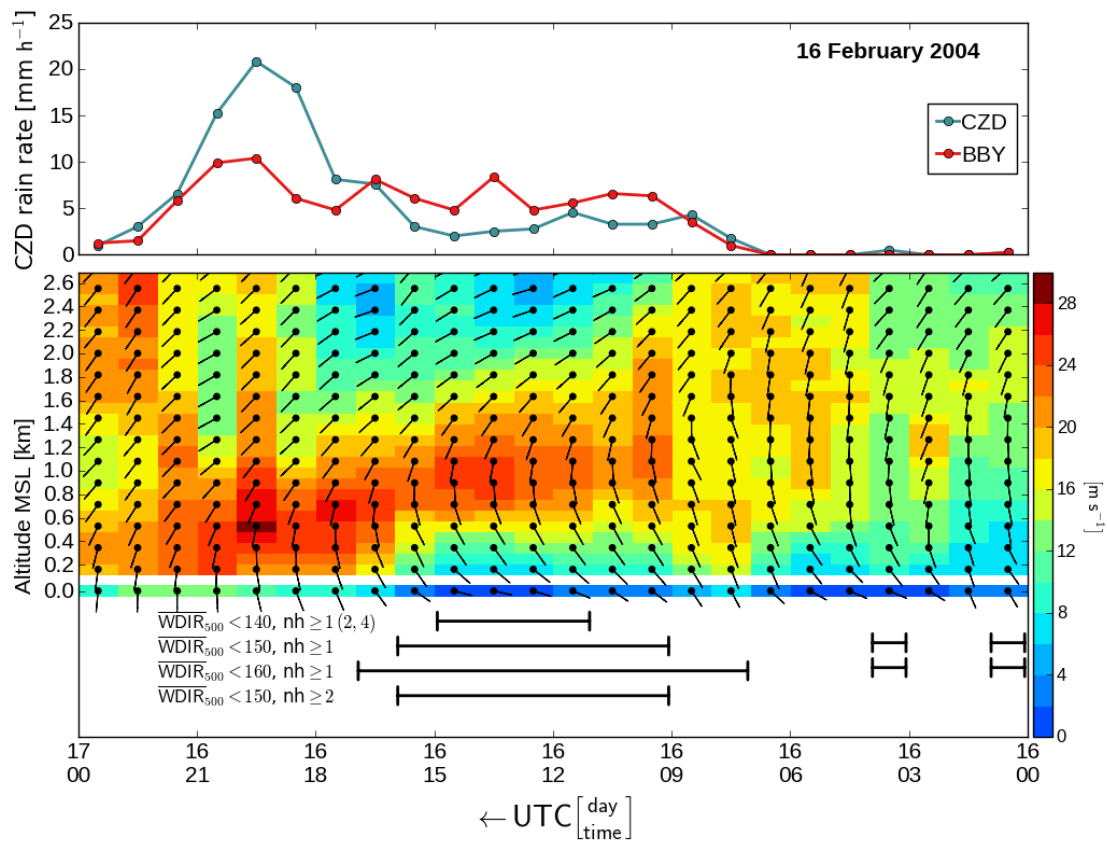


Figure 3.9: Rain rate at CZD and BBY (top) and time-height section of hourly surface and wind profile observations at BBY (bottom) during 16 February 2004 storm. Color coded is the wind meridional component (V-component). Wind staffs indicate total wind direction. Annotations in the wind profile time-high section indicate TTA objective identification using different criteria.

3.4.2 Relationship to rainfall

Contribution of the TTA regime ($\overline{WDIR}_{500} < 140^\circ$) to orographic rainfall is presented Table 3.2. Total 13-season rainfall is consistent with an upslope forcing mechanism producing larger amounts of rainfall at CZD (14589 mm) than BBY (5068 mm), resulting in a CZD/BBY ratio of 2.9, a slightly larger value than the average winter season ratio of ~ 2.2 found by Neiman et al. (2002), White et al. (2003), Neiman et al. (2009). The larger ratio in our analysis is observed in the rainy subset (i.e. a ratio of ~ 2.2 is obtained when using the overall dataset). When comparing TTA partitioned values, CZD (1425 mm) and BBY (1018 mm) present closer amounts of rainfall and therefore a small orographic enhancement (CZD/BBY ratio of 1.4). On the other hand, NO-TTA rainfall at CZD (13164 mm) and BBY (4050 mm) shows an orographic enhancement of 3.2. TTA and NO-TTA rainfall ratio differences are in agreement with VK15 results and previous studies about TTA effects on precipitation (e.g., Neiman et al. 2002); namely, the presence of TTAs reduce the precipitation gradient between foothills and mountain top.

TTA rainfall represents a relative large fraction of the coastal rainfall (BBY, 20%) compared with mountain rainfall (CZD, 10%). This result is consistent with VK15 observations of a TTA precipitation enhancement initiated upwind (offshore) and moving progressively toward the shore. Consequently, these results show that coastal (or foothills) locations are prone to receive relatively more rainfall during a TTA regime compared with mountain sites. Rain rates present slightly larger values at CZD (1.4 mm hr^{-1}) compared with BBY (1.0 mm hr^{-1}) during TTA regime; however, a significant difference is observed during NO-TTA regime (2.6 and 0.8 mm hr^{-1} , respectively).

TTA (NO-TTA) rainfall partition is consistent with documented decrease (increase) of orographic rainfall enhancement during TTA (NO-TTA) periods (e.g., VK15, Neiman et al. 2002); yet, the TTA period identified in the 16 February 2004 case misses a portion of time when the TTA structure was documented in VK15. As a result, we examine in more detail the relationship between $\overline{WDIR}_{500}$ and orographic rainfall to identify any shortcoming associated with the initial guess of a TTA regime associated with 140° .

Table 3.2: Rainfall partition for TTA regime (layer-mean 0-500-m MSL wind direction $<140^\circ$) during the subset of hours when rain ≥ 0.25 mm was observed at CZD. Included are total number of hours, rainfall accumulation (mm), relative contribution to total rainfall (%), and rate rate (mm hr⁻¹) at CZD and BBY, as well as the mountain to coast rainfall ratio. NO-TTA regime correspond to hours outside the TTA regime.

		TTA	NO-TTA	Total
Total time	[hr]	1018	5063	6081
CZD	[mm]	1425	13164	14589
	[%]	10	90	100
	[mm h ⁻¹]	1.4	2.6	2.4
BBY	[mm]	1018	4050	5068
	[%]	20	80	100
	[mm h ⁻¹]	1.0	0.8	0.8
CZD/BBY ratio		1.4	3.2	2.9

We grouped hourly rain rates at CZD and BBY by selecting hours when corresponding layer-mean airflow emanated from wind direction bins of 10° width and centered from 90° to 270° (total of 19 wind direction bins). The accumulated rainfall divided by the number of hours in each bin produced the rain rate associated to the bin at CZD and BBY. Confidence interval for each rain rate was estimated with a bootstrap technique (e.g., Wilks 2011). For a given wind direction bin, this technique produces different values of accumulated rainfall due to a random resampling with replacement. After dividing by the number of hours (which holds the same for a given bin) the 2.5% and 97.5% percentiles are computed out of 5000 resampled rain ratios (when values converge), returning the 95% confidence interval of the rain rate. In addition to estimating the rain rate per wind direction bin, the ratio between CZD and BBY rainfall (e.g. precipitation enhancement), was computed for each wind direction bin. Rainfall ratios equal or closer to one indicate no or low rainfall enhancement, while values significantly larger than one indicate large enhancement.

Figure 3.10a indicates that rain rates at CZD are maximized for airflows from $180^\circ \pm 5^\circ$, while they decrease for any other direction. Although more subtle, rain rates at BBY present a maximum at $140^\circ \pm 5^\circ$. Figure 3.10b suggests that there is a low rainfall enhancement of roughly 1.5 for wind directions less than 150° ; however, the enhancement is almost twice for wind directions pass this direction.

To explore the mean relationship between wind direction and rainfall, two statistical models are fitted to the observed rain rates and rainfall ratios. A three-parameter Gaussian model, given by:

$$f(x | \mathbf{p}) = \frac{A}{\sigma\sqrt{2\pi}} e^{-\frac{(x-\mu)^2}{2\sigma^2}} \quad (3.1)$$

where $\mathbf{p} \equiv [A, \mu, \sigma]$ is the parameter vector, A is the amplitude, μ is the mean, and σ is the standard deviation of the model (e.g., Larson et al. 2001), was selected for rain rates due to the bell-shaped relationship with wind direction. In addition, its parameter μ can be physically interpreted as the wind direction that maximizes the rain rate.

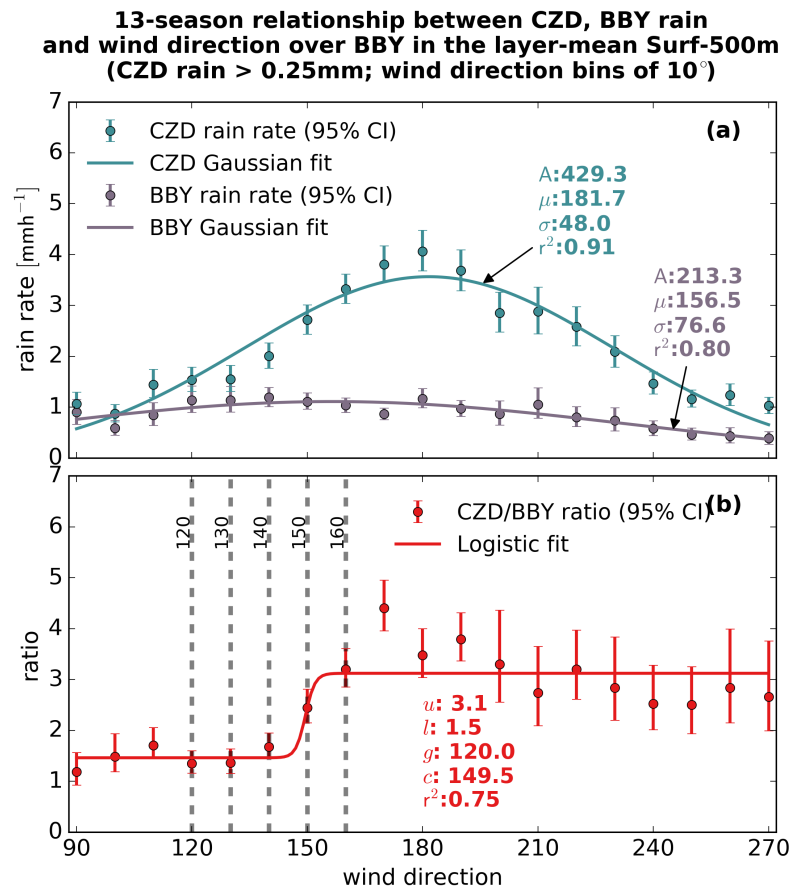


Figure 3.10: (a) Relationship between rain rate (CZD, BBY) and wind direction observed at BBY. Observations are grouped in wind direction bins of 10°. (b) Same as (a) but for CZD/BBY rainfall ratio. Vertical solid lines in each point correspond to a 95% confidence interval estimated with bootstrap. Vertical dashed lines in (b) correspond to wind directions included in the sensitivity analysis of Figure 3.7. Statistical model fitted lines (Gaussian and Logistic) and their parameters are included.

A four-parameter logistic model, given by:

$$f(x | \mathbf{p}) = u + \frac{l - u}{1 + (\frac{x}{c})^g} \quad (3.2)$$

where $\mathbf{p} \equiv [u, l, g, c]$ is the parameter vector, u and l the upper and lower asymptote, respectively, g is the growth rate, and c the center of the model (e.g., Gottschalk and Dunn 2005), was selected for rainfall ratios due the sigmoidal relationship with wind direction. In addition, 3 of the logistic model parameters have physical interpretations: the upper (u) and lower (l) asymptote represent the average rainfall enhancement in the upper and lower regime, whereas the center (c) represents the wind direction where the regime changes. Parameters of each model were obtained using a least squared method.

The Gaussian fit for CZD rain rate indicates that values are maximized for a wind direction of 181.7° , whereas BBY rain rate is maximized for wind directions of 156.5° . The logistic fit of rainfall ratios indicates that the lower and upper regime have average ratios of 1.5 and 3.1, respectively. The change in rainfall ratio regime occurs at 149.5° . All these models show a reasonable agreement with observations, with coefficient of determination (r^2) of 0.91, 0.80, and 0.75, respectively. Parameters and coefficients of determination have no significant change when using bins of 30° width (not shown).

Since the most prominent change in rainfall ratio is associated with $\overline{WDIR}_{500}$ of approximately 150° , it is reasonable to assume that this wind direction (instead of 140°) separates TTA and NO-TTA regimes more accurately, with average rainfall ratios of 1.5 and 3.1, respectively.

3.4.3 Sensitivity analysis

Since the relationship between wind direction and rainfall indicates that a threshold of 150° more precisely divide TTA and NO-TTA regimes, we re-examine the mean wind-component profile using this threshold. Additionally, we employ 120° , 130° , 160° to provide a more complete sensitivity test around the initial threshold of 140° . TTA threshold variations in the U-component have no significant difference in structure below 1000-m MSL when using values up to 150° (Fig.

3.7a). Beyond 150° , the magnitude of the easterly flow decreases slightly faster. Above 1000-m MSL, lower-valued thresholds show a bias toward easterly directions. Threshold variations in the V-component indicates that the larger the threshold, the stronger the southerly winds and the more rapid the change in velocity with altitude (Fig. 3.7b). NO-TTA regimes show that the U-component keeps a westerly direction regardless the threshold, with more significant difference in velocity for a threshold of 160° (Fig. 3.7c). Similarly, the V-component shows the more significant difference for a threshold of 160° with a weaker southerly flow (Fig. 3.7d).

We complement the sensitivity analysis by computing the TTA and NO-TTA rainfall partition for different thresholds of wind direction, time, and layer-mean depth. Table 3.3 shows that the initial wind direction threshold of 140° produces nearly the same rainfall ratios regardless the time constrain; yet, the longer the time threshold, the smaller the total number of hours of TTA regime. Considering different wind direction thresholds, rainfall ratios increase for larger wind direction angles. This is consistent with results showed in Figure 3.10b, where there is an increase in the rainfall ratio from 120° to 160° . Also, by using larger wind direction thresholds there is an increase in the total number of TTA hours; thus, more NO-TTA winds are classified as TTA. When considering a layer-mean depth in the lowest 1000-m MSL and initial wind direction threshold of 140° , the rainfall ratio is reasonable for a TTA regime but the total number of TTA hours decrease significantly.

We employ sensitivity parameters of Table 3.3 in the 16 February 2014 storm and include resulting TTA periods in Figure 3.9 for parameters that found a TTA regime. We observe that for threshold of 140° , $nh \geq 4$ produce the same result as $nh \geq 1$ since the TTA period for wind direction threshold $< 140^\circ$ identified 4 hours. Wind direction threshold of 150° produce a TTA period more consistent with VK15 results but it also includes hours at 0000 and 0300 UTC. Although these hours fit the criteria imposed, rain rate at CZD and BBY indicates that only a minimal amount of rainfall was observed during these times so TTA effects over orographic rainfall can be neglected. Wind direction threshold of 160° identifies a TTA period longer than 0900-1600 UTC (VK15), including times of the NO-TTA regime before 0900 and after 1600 UTC. As with 150° , there are

isolated hours that can be neglected.

Table 3.3: Sensitivity analysis of thresholds used to identify TTA regimes. Results include variations in rain rate (mm hr^{-1}) at CZD and BBY, mountain to coast rainfall ratio, and total number of hours (hr) for TTA and NO-TTA regimes.

Layer [m]	Hours $\geq$	Wind direction < [deg]	TTA				NO-TTA			
			CZD [mm h ⁻¹]	BBY [mm h ⁻¹]	Rainfall ratio	Hours	CZD [mm h ⁻¹]	BBY [mm h ⁻¹]	Rainfall ratio	Hours
0-500	1	140	1.4	1.0	1.4	1018	2.6	0.8	3.2	5063
0-500	2	140	1.5	1.1	1.4	783	2.5	0.8	3.1	5298
0-500	4	140	1.6	1.2	1.3	579	2.5	0.8	3.1	5502
0-500	8	140	1.6	1.2	1.3	301	2.4	0.8	3.0	5780
0-500	1	120	1.2	0.8	1.5	571	2.5	0.9	2.8	5510
0-500	1	130	1.3	0.9	1.4	759	2.5	0.8	3.1	5322
0-500	1	150	1.7	1.0	1.7	1399	2.6	0.8	3.2	4682
0-500	1	160	2.0	1.0	2.0	1934	2.5	0.8	3.1	4147
0-1000	1	140	1.2	0.8	1.5	502	2.5	0.9	2.8	5579
0-500	2	150	1.8	1.2	1.5	1118	2.5	0.8	3.1	4963

3.5 Summary and conclusions

This study has developed an objective identification of terrain-trapped airflows (TTAs) along the coast of northern California and documented their physical effects on orographic rainfall with a long-term, 13-season dataset. Observations from a 915 MHz wind profiling radar and surface meteorology were the main asset employed in this study.

The overall 13-winter-season dataset indicates large frequency of northwesterly flows in the lowest 1000-m MSL; however, when selecting a subset of rainy hours (rain at CZD > 0.25 mm), winds shift to a predominant southerly component. In addition, the rainy subset shows stronger wind speeds, likely associated with the passage of baroclinic waves over the observation domain, and a particular behavior in the lowest 500-m MSL: a rapidly increasing meridional velocity with altitude, maximum vertical wind shear of horizontal winds, and an easterly jet structure.

Terrain characteristics indicates that terrain-parallel winds are likely associated with wind directions of 140° within the average terrain altitude of 500-m MSL. As a result, these parameters are considered as a first guess for identifying the TTA regime. Indeed, selecting hourly profiles with $\overline{WDIR}_{500} < 140^\circ$ highlights the easterly jet structure identified in the rainy subset, as well as it shows stronger southerly velocities in the meridional profile. By counting the total number of hours and the number of continuous hours when the TTA regime was observed, we determined that the average TTA time-scale is 2.5 hours.

The TTA (NO-TTA) regime in the rainy subset results in a mountain/coast rainfall ratio of 1.4 (3.2). It also produces a relatively larger contribution to rainfall along the coast (20%) than over the mountain (10%), which is consistent with VK15 results. However, by using $\overline{WDIR}_{500} < 140^\circ$ some inconsistencies are evident relative to the TTA documented in VK15. A more detailed analysis of the relationship between wind direction and orographic rainfall reveals that a threshold of 150° divides more precisely two regimes of orographic enhancement, one with average rainfall ratio of 1.5 (TTA regime) and the other with 3.1 (NO-TTA regime).

A sensitivity analysis using different thresholds for wind direction, time, and layer-mean

depth indicates that using $\overline{WDIR}_{500} < 150^\circ$ produce results more consistent with VK15 but some negligible hours are included. As a result, and considering the 13-season average duration of 2.5 hours of TTA events, we conclude that a $\overline{WDIR}_{500} < 150^\circ$ for at least 2 hours is the best joint criteria to identify TTA events, allowing to filter out isolated hours associated with negligible amounts of rainfall (Fig 3.9).

In part 2 of this study we incorporate the aforementioned TTA identification criteria to isolate TTA periods in 7 case studies and produced composited kinematic and precipitation structures using a ground-based scanning Doppler radar located at a coastal site north of BBY.

Chapter 4

Terrain-trapped airflows and orographic rainfall along the coast of northern California. Part II: Horizontal and vertical structures observed by a scanning Doppler radar

This chapter is in preparation to be submitted to:

Valenzuela, R. A., and D. E. Kingsmill, 2017b: Terrain-trapped airflows and orographic rainfall along the coast of northern California. Part II: Horizontal and vertical structures observed by a scanning Doppler radar. *Monthly Weather Review*, **To be submitted**

4.1 Introduction

Extratropical cyclones and concomitant atmospheric rivers (ARs, e.g., Ralph et al. 2004) hit the coast of northern California every year during the winter season. Normally, the moist airflow associated with ARs ascends over the coastal mountains, producing condensation and in several cases precipitation due to the mechanical lift of ARs by steep orography. However, precipitation produced by this upslope flow mechanism is occasionally modulated due to the presence of terrain-trapped airflows (TTAs) along the coast.

One forcing mechanism responsible of TTAs is linked to blocking of low-level airflow (e.g., Smith 1979, Chen and Smith 1987, McCauley and Sturman 1999, Bousquet and Smull 2003, Kingsmill et al. 2013). Blocking occurs when stably-stratified air parcels approaching a mountain are unable to surmount the barrier and instead turn left (right) in the Northern (Southern) Hemisphere in response to an increased along-barrier pressure gradient and weakened Coriolis force. Theoretical studies suggest that the non-dimensional Froude number (e.g., Pierrehumbert and Wyman 1985) can describe low-level blocking conditions upstream of mountain barriers. An-

other forcing mechanism is through gap flows. In this case, a cold pool of continental air in close proximity to the lowest point along a mountain ridge (i.e., a gap) tends to be channeled in the presence of an along-gap pressure gradient (e.g. Lackmann and Overland 1989, Mass et al. 1995, Neiman et al. 2006, Mayr et al. 2007). Thus, the gap flow depth is equal to or lower than the gap depth. Once past the mountain gap, the airflow is able to turn left (right) in the Northern (Southern) Hemisphere with a radius of curvature that is a function of the gap flow wind speed and Coriolis parameter (e.g. Steenburgh et al. 1998). An analytical expression was proposed by Mass et al. (1995; henceforth M95) to describe gap flow forcing.

Only a few studies have addressed the documentation of structures and effects of TTAs along the coast of northern California. For instance, Neiman et al. (2002) showed that the presence of TTAs weakened the linear relationship between moist upslope flow and orographic precipitation. Valenzuela and Kingsmill (2015; henceforth VK15) used a ground-based X-band Doppler radar to document the detailed kinematic structure of a TTA interacting with incoming low-level jet in a single case study. Although the aforementioned studies provide insights about TTA structures and effects on orographic precipitation, they were limited in a few important respects. For example, Neiman et al. (2002) was based on only one winter season, which invites questions regarding sample size and the representativeness of that one season. Similarly, VK15 documented a coastal-mountain TTA with unprecedented detail but the study was based on only a single storm. Accordingly, there is uncertainty about the generality of their results.

In Part I of this study (Valenzuela and Kingsmill 2017a; henceforth VK17) we developed an objective identification of TTA regimes using a 13-winter-season dataset of hourly wind profile observations. In this second part, we employ VK17 results to identify TTA regimes during the landfalling of 7 winter storms along the coast of northern California and thus address concerns about the generality in the results of previous studies. We employ a ground-based scanning Doppler radar to analyze the detailed kinematic and precipitation structures of these storms. This study is unique because characterizes the mean properties and variability of three-dimensional kinematic and precipitation structures associated with coastal TTAs. Section 4.2 describes the observing

systems and data processing techniques employed in the analysis. An overview of each storm is presented in section 4.3, while detailed kinematic and precipitation structures associated with TTAs in each storm are described in section 4.4. Finally, section 4.5 presents the summary and conclusions of this study.

4.2 Observing systems and data processing

Observations employed in this study were collected along the northern California coast as part of the Hydrometeorology Testbed experiments (Ralph et al. 2013) operated by the National Oceanic and Atmospheric Administrations (NOAA) Earth System Research Laboratory (ESRL). Locations of key observing systems are shown in Figure 4.1 and observation periods are defined in Table 4.1.

The main asset is a ground-based scanning X-band (3.2-cm wavelength) dual-polarization Doppler radar (X-Pol, Martner et al. 2001, Matrosov et al. 2005) located at Fort Ross (FRS), California. X-Pol executed both slant-horizontal plan-position indicator (PPI) and vertically oriented range-height indicator (RHI) scans (Table 4.1). PPI scans extended to a maximum range of 57 km with 0.23-km gate spacing and were repeated at least once every 6 min. The analysis employed PPI scans with an elevation angle of 0.5° to best resolve low-level structures. These scans were only directed offshore due to low-level beam obstruction from the coastal mountains. RHI scans extended to a variable maximum range between 28 and 38 km with 0.11-km gate spacing and were executed over the ocean and terrain in a cycle that repeated at least once every 12 min. These scans started by viewing one horizon then rotated upward toward and past zenith, but did not extend all the way down to the opposite horizon. The analysis employed RHI scans that were directed in the meridional plane.

Each radar sweep was quality controlled by removing artifacts such as ground and sea clutter, range folding (i.e. second trip echo), sidelobe echoes, and by dealiasing folded Doppler radial velocities. After the quality control process, polar-coordinate sweeps were interpolated into a Cartesian grid. For PPI scans, the horizontal and vertical grid spacings were 0.5 and 0.35 km, respectively.

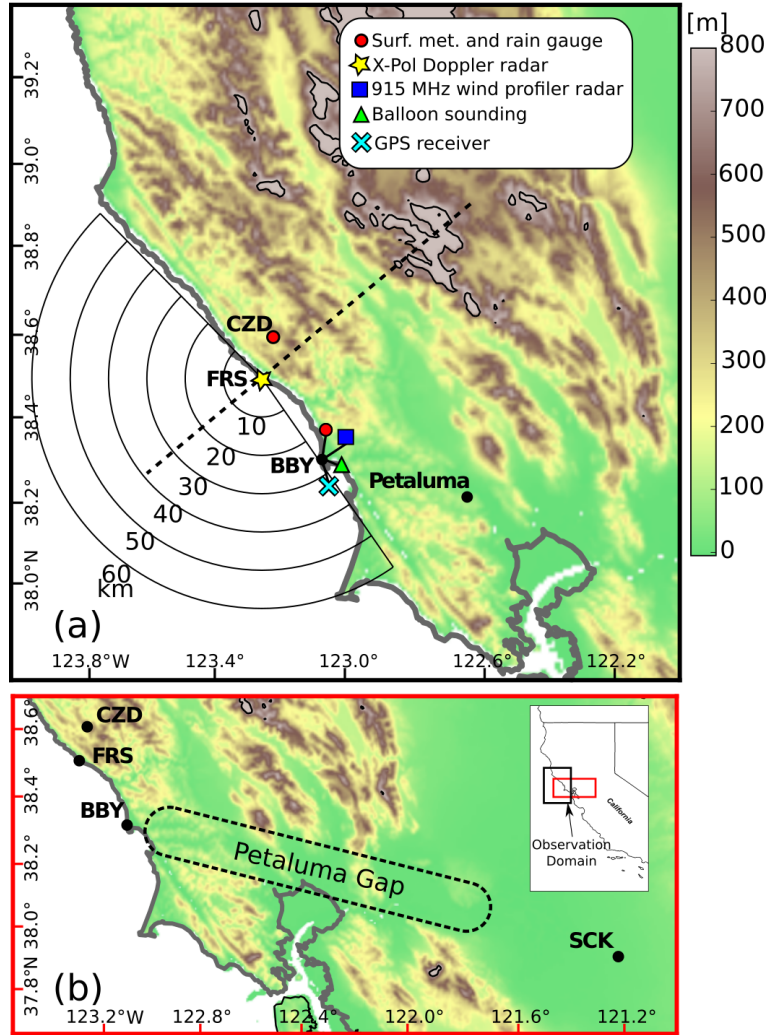


Figure 4.1: (a) Topographic map overlaid with observing systems. Legend identifying each instrument along with the color scale for terrain elevation (m) are included. Semi circles centered at Fort Ross (FRS) indicate the X-Pol analysis domain. Dashed line (azimuth 50°-230°) is the reference for elevation profile in Figure 4.15. Black contour line over the terrain indicates altitude of 800 m. (b) Topographic map showing Petaluma Gap terrain and the METAR station at Stockton (SCK). Inset map in (b) provides a reference for both topographic maps relative to the coast of Northern California: map (a), black line; map (b), red line.

Table 4.1: Start and end dates and time (UTC) of observations in the 7-storm dataset and X-Pol scans employed. Angular limits of azimuth and elevation sectors are indicated.

Storm	Surface and wprof		X-Pol		PPI			RHI		
	Start	End	Start	End	Fixed elevation	Azimuth sector	Sweeps	Fixed azimuth	Elevation sector	Sweeps
12-14Jan03	0000	0000	0500	1600						
	12 JAN	15 JAN	12 JAN	14 JAN	0.5	134-320	570	180	0-155	283
	2003	2003	2003	2003						
21-23Jan03	0000	0000	0500	0900						
	21 JAN	24 JAN	21 JAN	23 JAN	0.5	134-320	372	180	0-155	180
	2003	2003	2003	2003						
15-16Feb03 *	0000	0000	2000	1500						
	15 FEB	17 FEB	15 FEB	16 FEB	0.5	134-320	190	180	0-155	96
	2003	2003	2003	2003						
09Jan04	0000	0000	1500	0000						
	09 JAN	10 JAN	09 JAN	10 JAN	0.5	110-320	77	0 180	10-101 0-101	40 40
	2004	2004	2004	2004						
02Feb04	0000	0000	0900	2000						
	02 FEB	03 FEB	02 FEB	02 FEB	0.5	110-320	111	6	10-162	110
	2004	2004	2004	2004						
16-18Feb04	0000	0000	0700	1800						
	16 FEB	19 FEB	16 FEB	18 FEB	0.5	110-320	602	0 180	10-101 0-101	105 105
	2004	2004	2004	2004						
25Feb04	0000	0000	0600	1900						
	25 FEB	26 FEB	25 FEB	25 FEB	0.5	110-320	128	0 180	10-101 0-101	64 64
	2004	2004	2004	2004						

* wprof ends at 0400 UTC 16 FEB 2003

For RHI scans, the horizontal and vertical grid spacings were 0.1 and 0.2 km, respectively. A Cressman distance-dependent weighting scheme (Trapp and Doswell 2000) was employed to interpolate values of attenuation-corrected reflectivity (Matrosov et al. 2005) and Doppler radial velocity to each Cartesian grid point.

Merged vertical cross sections of reflectivity and radial velocity were made by combining north and south RHI scans (e.g. 0° and 180° azimuth) for 9 January 2004, 16-18 February 2004, and 25 February 2004 storms (Table 4.1). Although the contributing RHI scans were offset by 2–3 min, the structure across the merged interface of the two scans was coherent. The horizontal component of radial velocity in the plane of each cross section was calculated toward north (i.e. meridional wind). Elevation angles between 65° and 115° were excluded from merged RHI to simplify both visualization and interpretation of airflow structures.

A 915-MHz wind-profiling radar (Ecklund et al. 1988) located at Bodega Bay (BBY) provided hourly profiles of horizontal winds from ~ 0.1 to ~ 4.0 km MSL with ~ 100 -m vertical resolution (high-altitude mode). Each profile was processed with the continuity method of Weber et al. (1993) that checks consistency in the dataset over time and height. Additional information for 2004 storms is provided by hourly averaged column-integrated water vapor (IWV) observations collected at BBY from a ground-based global positioning system (GPS) receiver (Wolfe and Gutman 2000), allowing the evolution of water vapor to be monitored while storms passed over the observing domain. GPS-IWV observations are unaffected by precipitation (Businger et al. 1996).

4.3 Storms overview

Surface rainfall traces from BBY and CZD (Fig. 4.2) indicate that most of the storms have peak rain rates of ~ 10 - 15 mm h $^{-1}$. However, the 16-18 February 2004 storm (Fig. 4.2f) is characterized by two rain-rate peaks of ~ 15 - 20 mm h $^{-1}$. Total accumulated rainfall is consistently larger over the mountains (CZD), with mountain to coast rainfall ratios between 1.9 and 5.7.

Synoptic context for each storm is provided by analyses from the Climate Forecast System Reanalysis (CFSR, Saha et al. 2010) less than 6 h before significant precipitation was observed

along the coast and over the coastal mountains (Fig. 4.3). Common features in all the analyses are surface pressure depressions of $\sim 980\text{--}990$ hPa centered well offshore of the Washington-Oregon coast (around $\sim 45^\circ\text{N}$, $\sim 140^\circ\text{W}$) and integrated water vapor transport (IVT) of at least $250 \text{ kg m}^{-1}\text{s}^{-1}$ associated with either existing or dissipated ARs. They differ however in the equivalent-potential-temperature gradients of approaching baroclinic waves. The 16-18 February 2004 storm is characterized by the strongest gradient (Fig. 4.3f) while the weakest gradient is associated with the 2 February 2004 storm (Fig. 4.3e). Comparable equivalent-potential-temperature gradients are evident for the remaining storms. There are also differences in AR characteristics. While the 15-16 February 2003 storm (Fig. 4.3c) is relatively weak in terms of IVT and the 2 February 2004 storm (Fig. 4.3e) contains the remnants of a dissipated AR, the storm on 16-18 February 2004 (Fig. 4.3f) is associated with a strong and clearly defined AR. Orientation of ARs for the storms on 21-23 January 2003 and 9 January 2004 (Fig. 4.3b,d) is approximately meridional. In contrast, the storms on 12-14 January 2003, 15-16 February 2003, 16-18 February 2004, and 25 February 2004 (Fig. 4.3a,c,f,g) have ARs with nearly southwest-northeast orientations. However, in these latter cases, the IVT pattern takes on a more meridional orientation closer to the coast, suggesting that the low-level meridional component of flow increases as the ARs approach the continent.

Coastal airflow characteristics of each storm are now examined with data from the 915-MHz wind profiler at BBY. As a means to highlight LLJ structures, the meridional component of flow is analyzed based on aforementioned inferences drawn from CFSR data. Meridional LLJ structures are evident during the 2004 storms on 2 February, 16-18 February, and 25 February (Fig. 4.4e,f,g), with magnitudes between $8\text{--}32 \text{ m s}^{-1}$ centered between 0.8 and 1.2 km MSL. LLJ structures during the 9 February 2004 storm are not evident (Fig. 4.4d). The remaining 2003 storms are characterized by more modest LLJ magnitudes but centered at similar altitudes. Orographic forcing associated with these LLJs is distilled by examination of upslope wind speed (i.e., component from 230°) averaged over the 0.85-1.15 km layer and its product with GPS-IWV called bulk upslope IWV flux (Fig. 4.5), an approach employed by Neiman et al. (2009) and Kingsmill et al. (2016). All of the storms meet or exceed upslope wind speed and, when available, bulk upslope IWV flux thresholds

(12.5 m s^{-1} and 25 cm m s^{-1} , respectively) for AR conditions during at least a portion of their lifetimes. Additionally, it is evident that the maxima in orographic forcing are nearly coincident with CZD rain-rate maxima (Fig. 4.2).

4.4 TTA kinematic and precipitation structures

4.4.1 X-Pol analysis approach

Fig. 4.4 shows the time coverage of X-Pol observations and the TTA periods for each of the 7 storms determined with the objective identification of VK17: layer-mean 0-500-m MSL wind direction $< 150^\circ$ during at least 2 hours. Two of the storms (15-16 February 2003 and 25 February 2004) did not experience any TTA conditions. NO-TTA periods are defined by times with X-Pol observations (Table 4.1) outside of TTA periods. X-Pol sweeps were partitioned into TTA and NO-TTA periods and combined to investigate composite structures. We first examine the 7-storm composite as a whole and then explore composites for individual storms to evaluate inter-storm variability.

Composited kinematic structures were determined by time-averaging Doppler velocity, similar to the VK15 approach. Composited precipitation structures were determined by deriving the frequency of attenuated-corrected reflectivity exceeding a given threshold. Yuter et al. (2011) employed a similar approach to examine precipitation structures near Portland, Oregon with operational radar observations. They used this methodology to compute precipitation frequency that was relatively insensitive to the presence or absence of a radar bright band. Yuter et al. (2011) utilized exceedance thresholds of 13 dBZ and 25 dBZ, which correspond to rain rates of ~ 0.2 and 1.3 mm h^{-1} , respectively. This worked well for their relatively large 117-storm dataset. However, there are just 7 storms in our dataset, which increases the impact of reflectivity-magnitude variability from storm to storm. As a means to reduce these impacts, exceedance thresholds were based on the median value of reflectivity for each grouping of X-Pol data analyzed. The median depends only on the cumulative frequency distribution of reflectivity in each group and, among

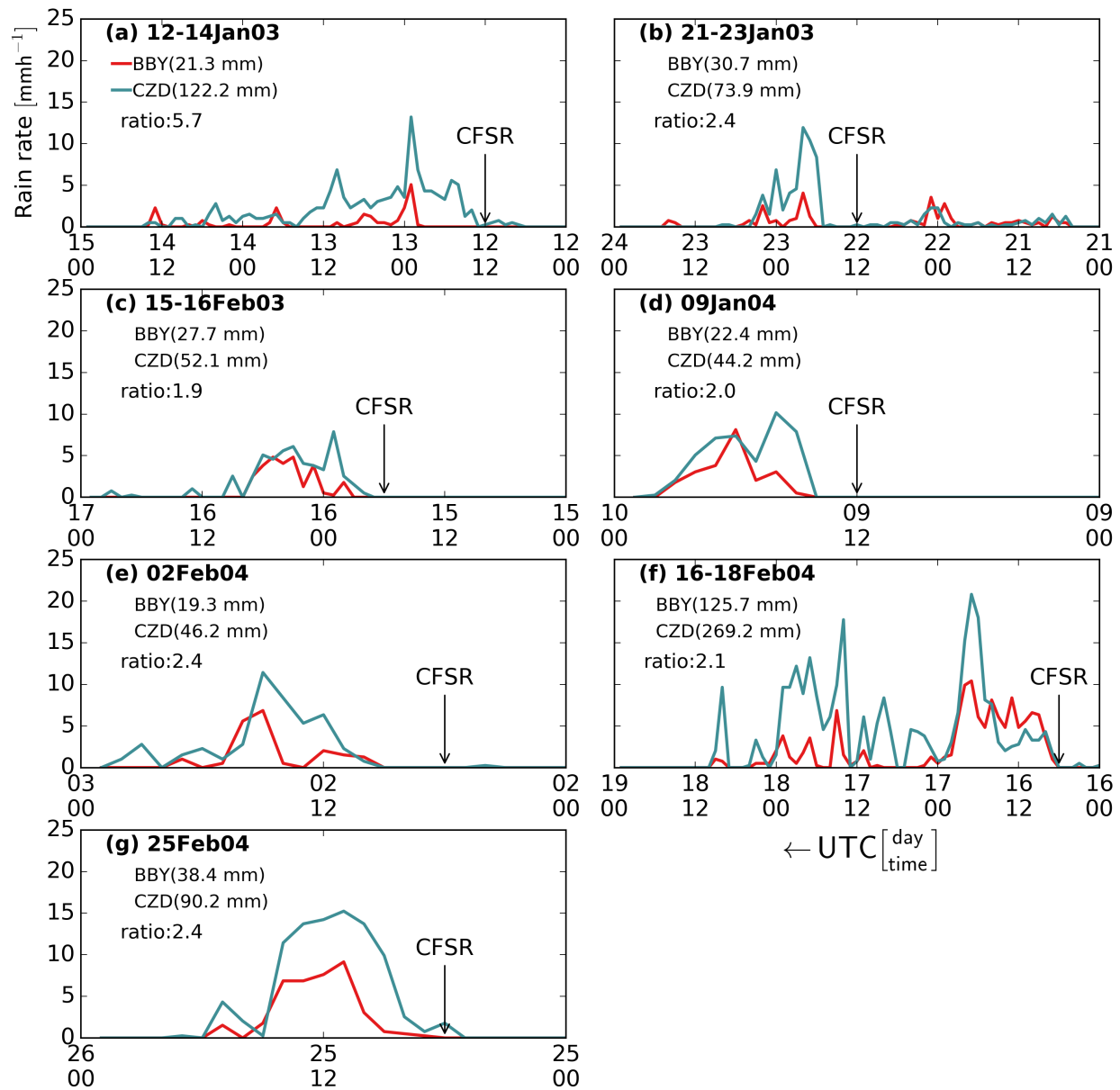


Figure 4.2: Time series of surface rainfall at Bodega Bay (BBY) and Cazadero (CZD). Legend includes total rainfall at each site (t) and mountain to coast (CZD/BBY) rainfall ratio. CFSR annotation indicates analysis time in Figure 4.3.

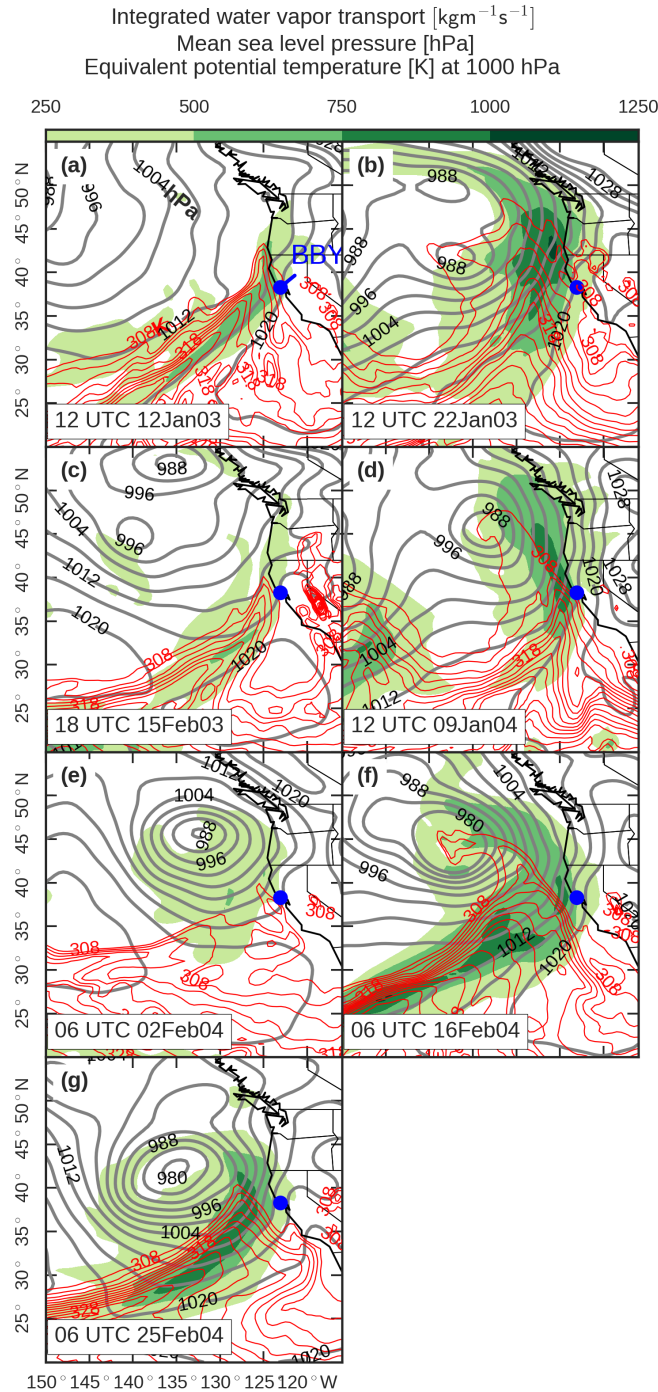


Figure 4.3: Reanalysis-derived integrated water vapor transport (IVT, $\text{kg m}^{-1} \text{s}^{-1}$, filled contours), mean sea level pressure (hPa, gray contours), and equivalent potential temperature (θ_e) at 1000 hPa (K, red contours) from the CFSR during 2003 storms at (a) 1200 UTC 12 January, (b) 1200 UTC 22 January, (c) 1800 UTC 15 February, and 2004 storms at (d) 1200 UTC 09 January, (e) 0600 UTC 02 February, (f) 0600 UTC 16 February, and (g) 0600 UTC 25 February. θ_e contours included between 308 and 338 K every 2 K.

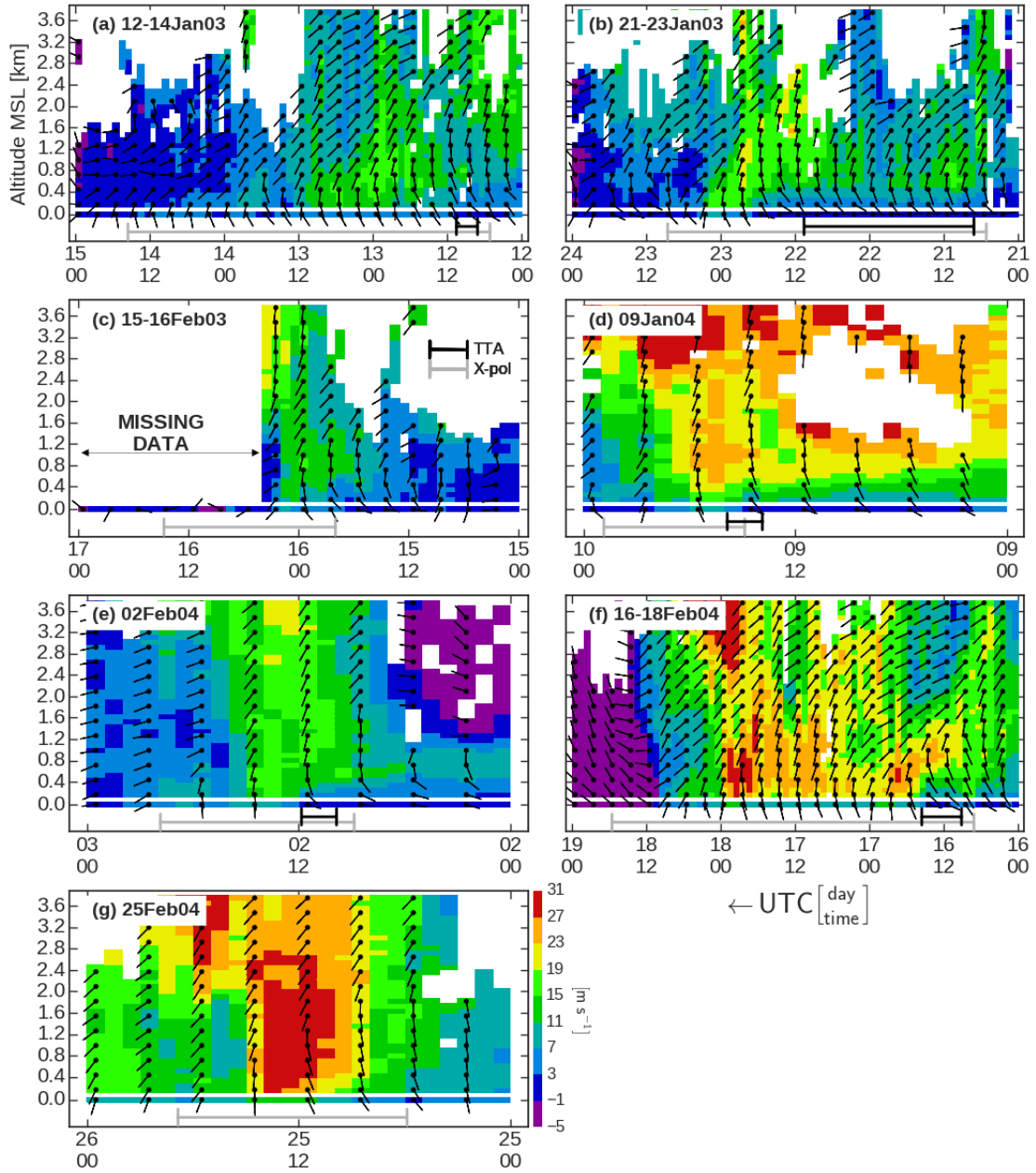


Figure 4.4: Time-height analyses of meridional wind speed (color coded) and total wind direction (wind staff) observed at Bodega Bay (BBY) from the 915-MHz wind profiling radar for each of the storms indicated in Table 4.1. Some of the analyses have extended periods relative to Table 4.1 to provide better context of the storm. Each panel includes time interval of TTA periods (black line) and X-Pol observations at Fort Ross (FRS, gray line) at the bottom. First level in each analysis includes surface observations. Time is from right to left to represent eastward advection of storms.

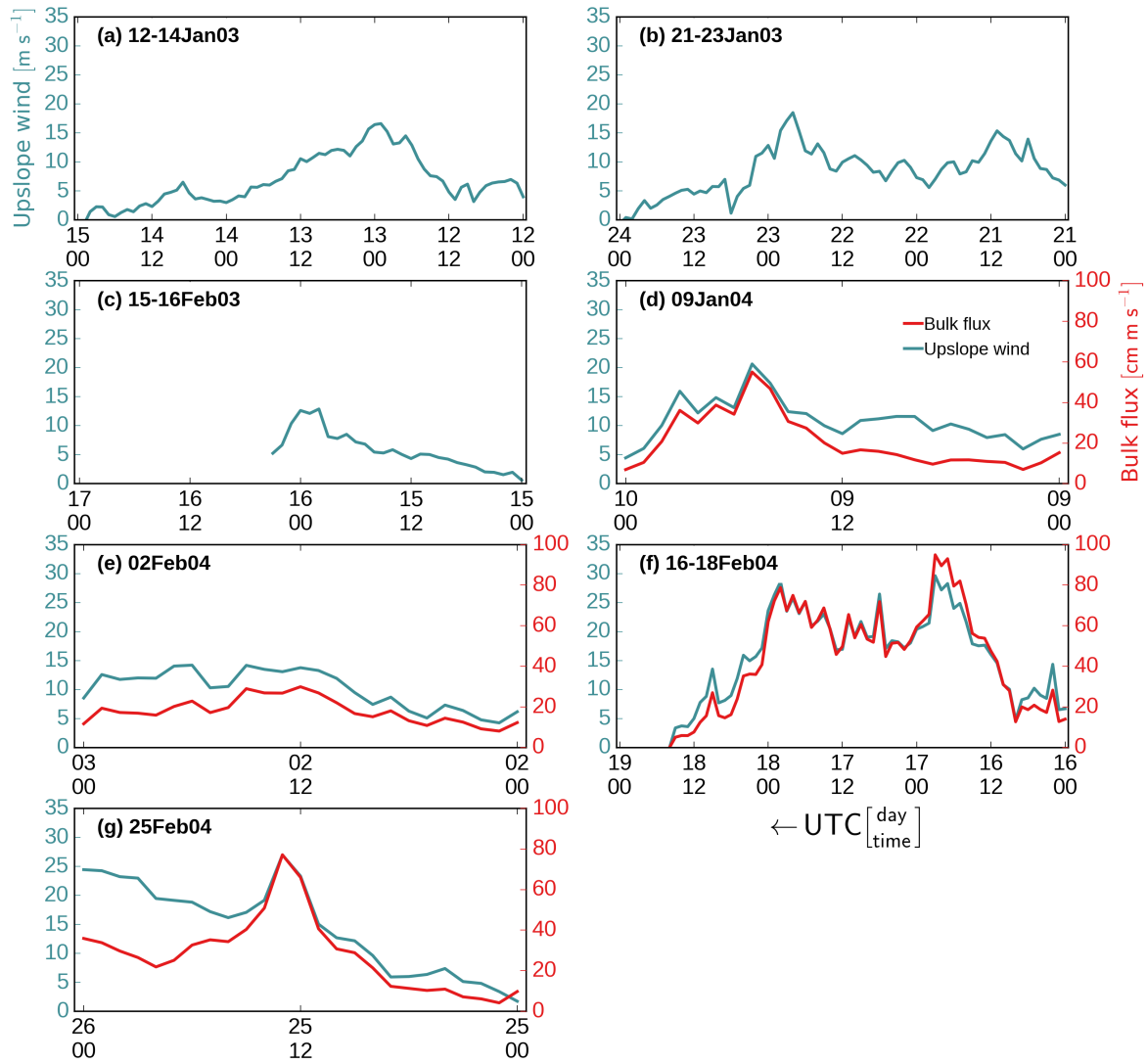


Figure 4.5: Similar to Figure 4.2 but for time series of upslope wind (from 230° , ms^{-1}) and bulk water vapor flux (cm m s^{-1}) at Bodega Bay (BBY) and Cazadero (CZD). Storms during 2003 only have upslope wind available.

central tendency metrics, has the advantage of being less sensitive to variations in extreme values. Exceedance frequency for each radar grid point of each X-Pol group was computed by summing radar grid points greater than or equal to the median of the corresponding cumulative frequency distribution, dividing by the total number of radar grid points in the group and multiplying by 100 to obtain a percentage.

In the PPI group, TTA and NO-TTA median values are the same for the 7-storm composite (Fig. 4.6a), but the individual-storm composites have TTA median values that are sometimes higher and sometimes lower than NO-TTA median values. These variations might be linked with variations in the LLJ magnitude (e.g., Fig. 4.4), such that stronger (weaker) LLJs would produce stronger (weaker) uplift in the LLJ-TTA interface (e.g., VK15) and thus faster (slower) condensation rates, favoring (hindering) microphysical growth through collision-coalescence. In the RHI group, there is a slightly larger median value in the NO-TTA category for the 7-storm composite (Fig. 4.7a). Similar to the PPI group, individual-storm RHI composites have TTA median values that are sometimes higher and sometimes lower than NO-TTA medians. However, medians in the RHI group present lower values relative to the PPI group produced by larger frequency of weaker radar reflectivity from the top part of the vertical domain.

The X-Pol RHI observation domain (i.e. maximum range and elevation angle) varied between storms (Table 4.1), which produced spatial discontinuities in the vertical TTA and NO-TTA 7-storm composite structures. To mitigate this problem and improve interpretation we use the smallest maximum range of the various RHI scans. In addition, RHIs from the 2 February 2004 storm were removed from the vertical 7-storm composite analysis because they did not extend to low enough elevation angles offshore, preventing a clear view and analysis of the TTA and LLJ interaction. The removal of 2 February 2004 RHI scans from the 7-storm composite analysis did not fundamentally influence the resulting kinematic and precipitation structures other than to eliminate spatial discontinuities.

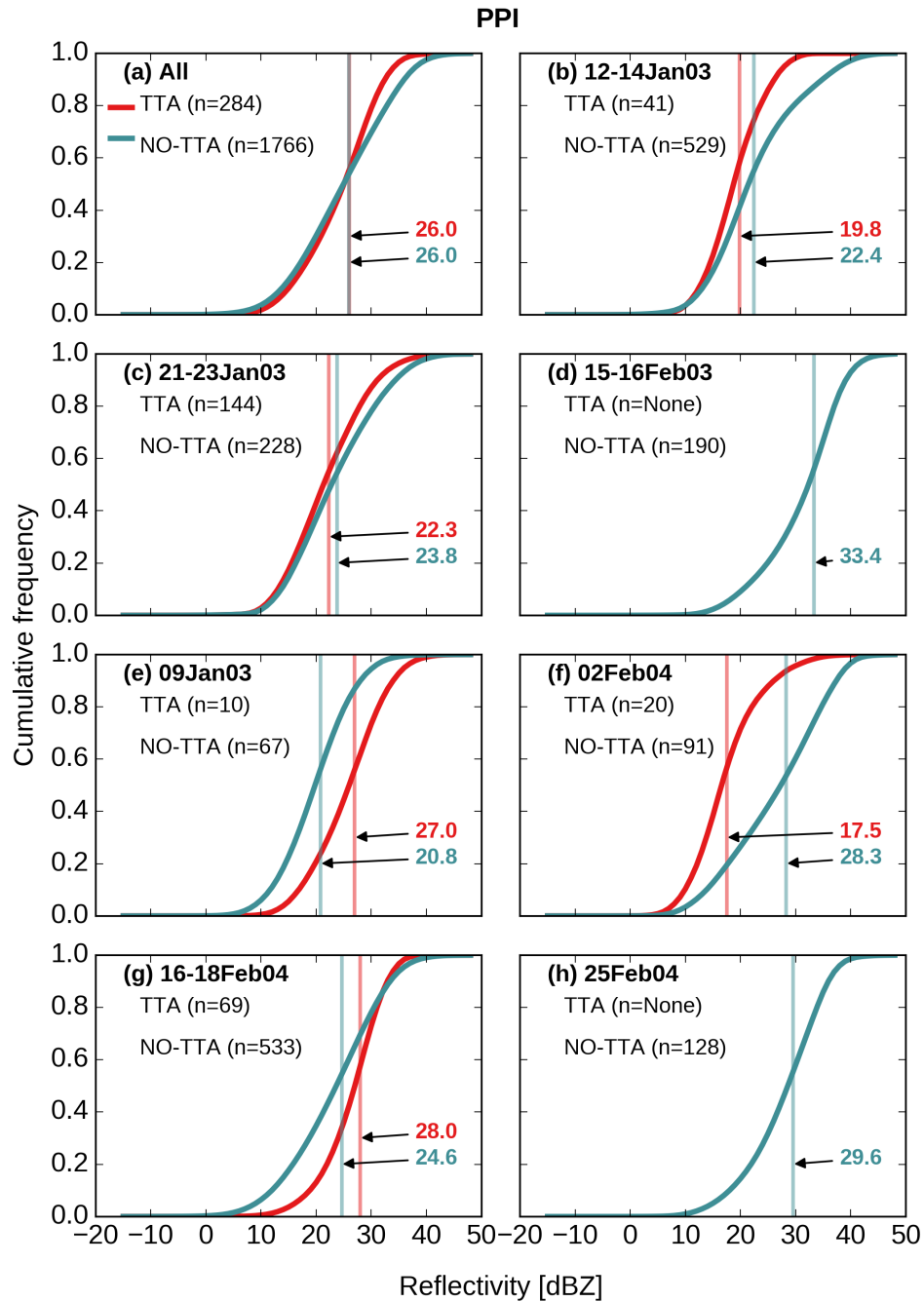


Figure 4.6: Cumulative frequency distribution (CFD) of attenuated-corrected reflectivity for plan-position indicator (PPI) scans in each of the 7 storms and separated in TTA and NO-TTA groups. Vertical line indicates median value.

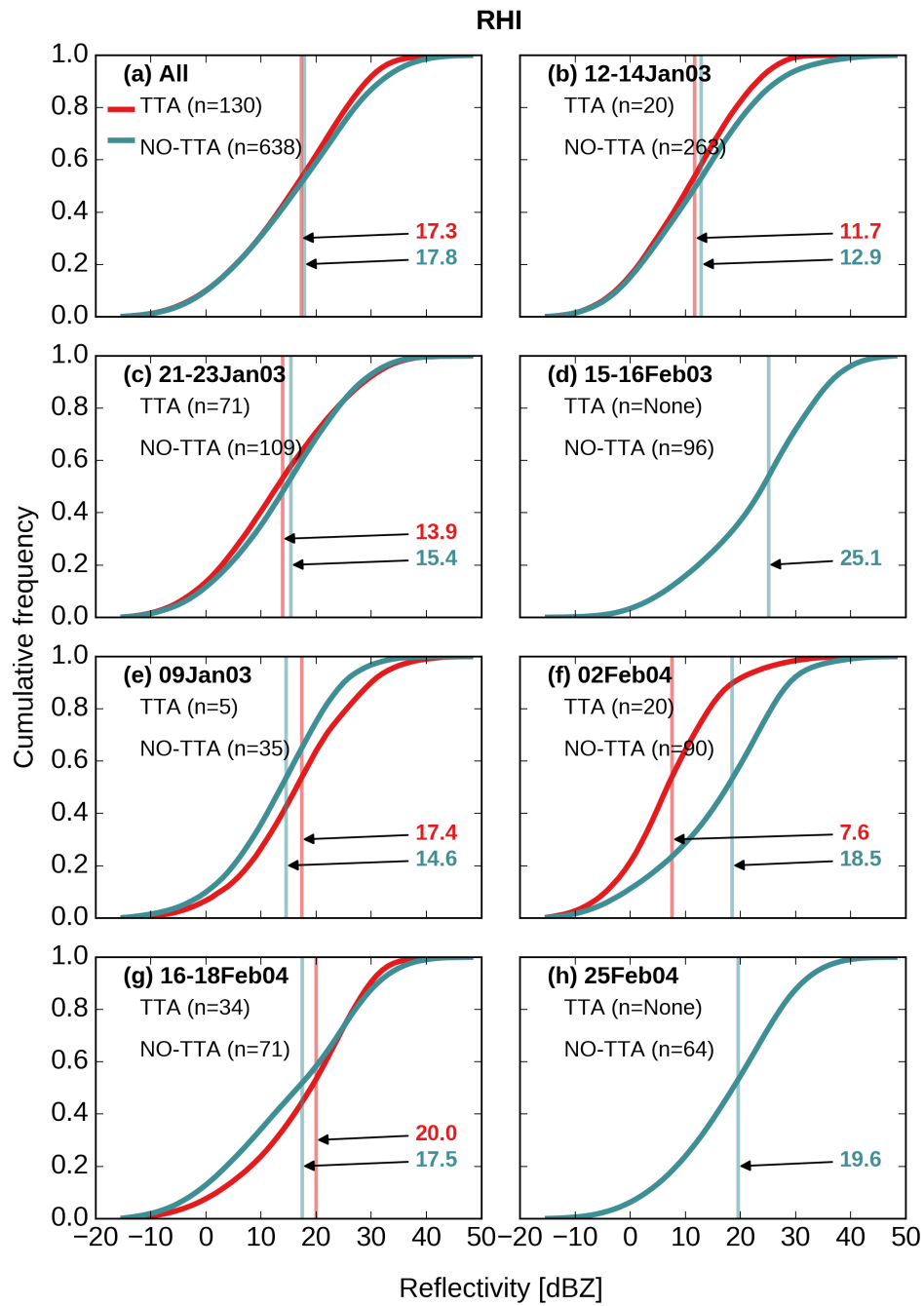


Figure 4.7: Same as Figure 4.6 but for RHI scans.

4.4.2 Storm-composite analysis

Figure 4.8 presents TTA (a,c,e,g) and NO-TTA (b,d,f,h) composited structures. TTA composite includes a total of 414 sweeps (284 PPI and 130 RHI) while NO-TTA composite comprise a total of 2404 sweeps (1766 PPI and 638 RHI). PPI sweeps were obstructed by the X-Pol radar trailer over an azimuth sector of $\sim 130^\circ$ - 175° . Impacts from this obstruction were most evident in the analysis of precipitation frequency (Fig. 4.8e,f).

In terms of kinematic structure, the horizontal TTA composite shows a sharper curvature of the isoline of zero radial velocity (henceforth called zero isodop, Fig 4.8a) compared to the NO-TTA composite (Fig. 4.8b). The airflow associated with the TTA composite is characterized by inferred southeasterly winds from the coast out to ~ 20 km offshore, southerly winds beyond ~ 40 km and a transition zone from ~ 20 to 40 km. In contrast, the airflow associated with the NO-TTA composite is characterized by inferred southeasterly winds out to ~ 10 km offshore, southerly winds beyond ~ 30 km, and a transition zone from ~ 10 km to ~ 30 km. There are also variations in the location and strength of LLJ structures. The TTA composite (Fig 4.8a) shows maximum inbound radial velocities of $\sim -18 \text{ m s}^{-1}$ located south of the radar at ranges beyond ~ 35 km that are associated with LLJ structures upstream of TTAs. This differs from the NO-TTA composite (Fig 4.8b) where maximum inbound radial velocities are smaller ($\sim -14 \text{ m s}^{-1}$) and located closer to the radar at ranges beyond ~ 20 km.

The vertical TTA composite (Fig. 4.8c) depicts a meridional LLJ exceeding 20 m s^{-1} riding up and over a weaker airflow of less than 10 m s^{-1} extending from the coast to ~ 15 km offshore and less than 0.5 km depth at the coast, corresponding to the TTA. This vertical structure is comparable to the TTA vertical structure described by VK15. The mean altitude of the LLJ shifts from ~ 0.5 km MSL at ~ 30 km range to ~ 1.0 km MSL over the radar. Conversely, the vertical NO-TTA composite (Fig. 4.8d) has a meridional flow structure that does not clearly indicate the presence of a LLJ, a result that requires some explanation. The NO-TTA composite is associated with a relatively large fraction of X-Pol observation time (Fig. 4.4). Therefore, any LLJ structures

that are present during NO-TTA conditions are likely smoothed out by the longer sampling period, including times after the baroclinic wave passage. This may also explain the weaker LLJ signature observed for NO-TTA conditions in the PPI composite (Fig 4.8b). More details about this result are provided in the inter-storm analysis.

Analysis of the horizontal TTA precipitation structure indicates a relatively large frequency of echoes exceeding the median reflectivity of 26.0 dBZ at $\sim 5 - 25$ km offshore from the coast (Fig. 4.8e). In contrast, the NO-TTA composite shows relatively large median-reflectivity exceedance frequencies that are in closer proximity to the coast and only extend offshore ~ 15 km (Fig. 4.8f). The vertical TTA precipitation structure reveals a relatively large frequency of echoes exceeding the median reflectivity of 17.3 dBZe below ~ 0.5 km MSL and extending from the coast to almost 20 km offshore (Fig. 4.8g). This differs substantially from the vertical NO-TTA composite (Fig. 4.8h) where the largest median-reflectivity (17.8 dBZe) exceedance frequencies are confined to within ~ 5 km of the coast and extend slightly above 1.0 km MSL.

Results from the 7-storm composite analysis show that TTA and NO-TTA periods are associated with two distinct kinematic and precipitation structures. These structures strongly suggest that TTAs are responsible for elevating the mean LLJ altitude offshore and enhancing precipitation upstream of the coastal mountains out to a range of 20 km from the coast. Conversely, NO-TTA periods are associated with a more homogeneous kinematic structure and precipitation enhancement closer to the coastal mountains.

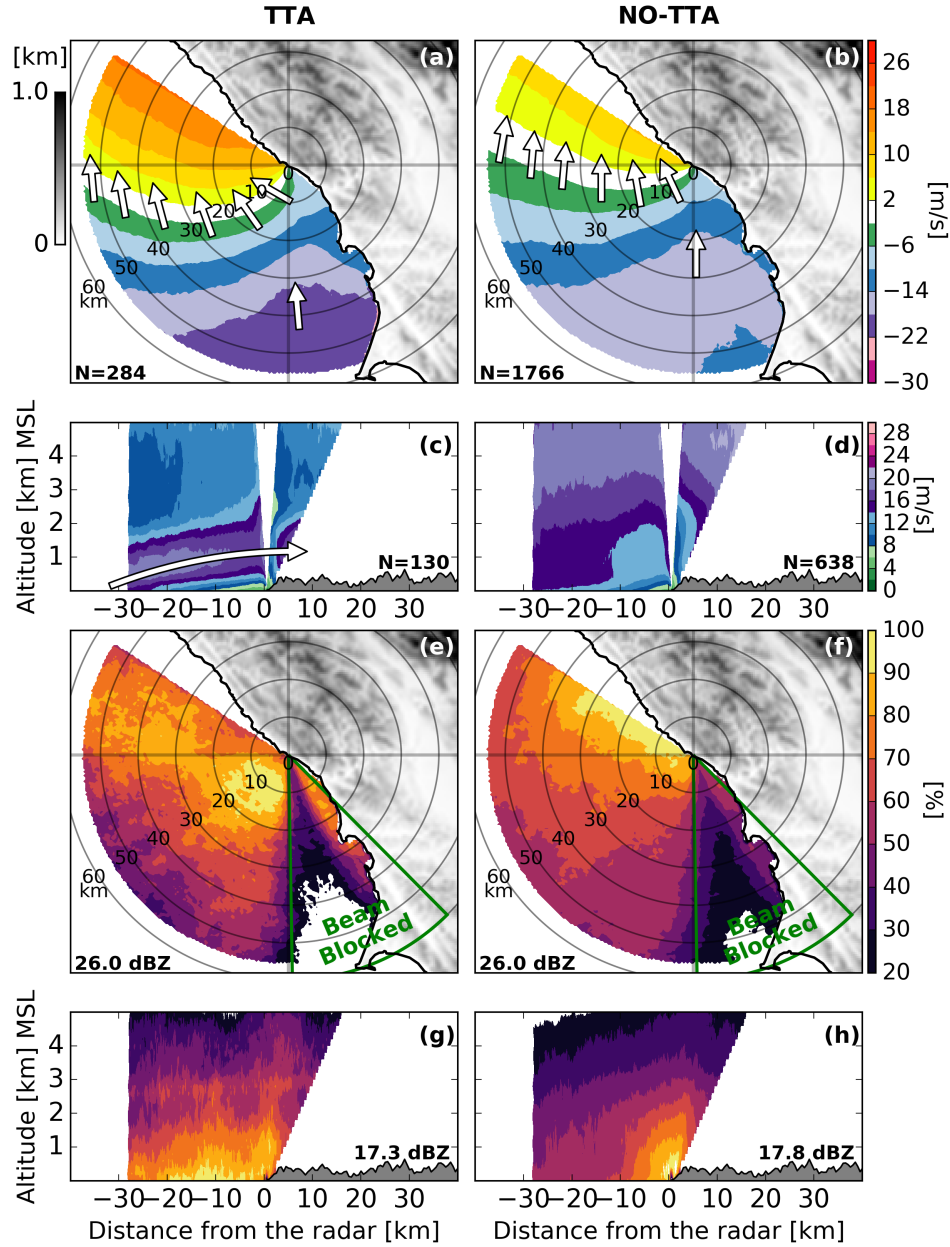


Figure 4.8: Time composite analysis of X-Pol observations for TTA (left panels) and NO-TTA (right panels) periods for 7 storms: average horizontal (a, b) and vertical (c, d) kinematic structures derived from Doppler radial velocity; normalized frequency in exceeding threshold of horizontal (e, f) and vertical (g, h) attenuation-corrected reflectivity. Vertical sections are meridionally oriented (i.e. azimuth 0° - 180°). Beam blockage by the radar trailer is indicated (e, f). Vertical sections (c, d, g, h) exclude 2 February 2004 storm (see text for details). Arrows were added over the zero isodop (white contour in a,b) to facilitate wind field interpretation. A single arrow south of the radar indicates maximum radial velocity associated with low-level jets.

4.4.3 Inter-storm variability analysis

The 7-storm-composite analysis depicts average TTA and NO-TTA kinematic and precipitation structures but does not provide context about inter-storm variability. In this section, details about the kinematic and precipitation structures for individual storms that comprise the composite are examined. As mentioned before, TTA conditions were not observed for the 15-16 February 2003 and 25 February 2004 storms.

The horizontal kinematic structures of TTA periods observed during 12-14 January 2003, 21-23 January 2003, 9 January 2004, 2 February 2004, and 16-18 February 2004 storms (Figure 4.9a,c,e,g,i) exhibit a relatively consistent pattern in the shape of the zero isodop: inferred southeasterly winds are observed from the coast out to a distance of ~ 20 km, followed by a transition zone of southeasterly to southerly winds from ~ 20 km to ~ 30 km, and then southerly to south-southwesterly winds beyond ~ 30 km. During 12-14 January 2003 (Fig. 4.9a) the zero isodop transition is somewhat smoother, likely produced by a synoptic pattern with larger south-southeasterly component flow and thus a smaller difference relative to the southeasterly winds associated with the TTA. Maximum inbound radial velocities south of the radar vary from ~ -10 to -18 m s $^{-1}$ for all of these storms except 9 January 2004 and 16-18 February 2004 (Fig. 4.9e,j) when they fluctuate between ~ -22 and -30 m s $^{-1}$.

Median-reflectivity exceedance frequencies during TTA conditions depict horizontal precipitation structures with offshore enhancement that are approximately parallel to the coast during 21-23 January 2003, 2 February 2004, and 16-18 February 2004 (Fig. 4.9d,h,j). For the 21-23 January 2003 storm, this enhancement is ~ 20 to 30 km offshore, whereas for the 2 February 2004 and 16-18 February 2004 storms (Fig. 4.9f,h) it is ~ 10 -20 km offshore. The 12-14 January 2003 storm (Fig. 4.9b) has a distinctly different horizontal precipitation structure that is the result of isolated and scattered echoes. Finally, the 9 January 2004 storm has only one hour of X-Pol observations during TTA conditions ($n=10$, Fig. 4.9f), which makes precipitation-enhancement patterns difficult to detect due to a relatively small sample size.

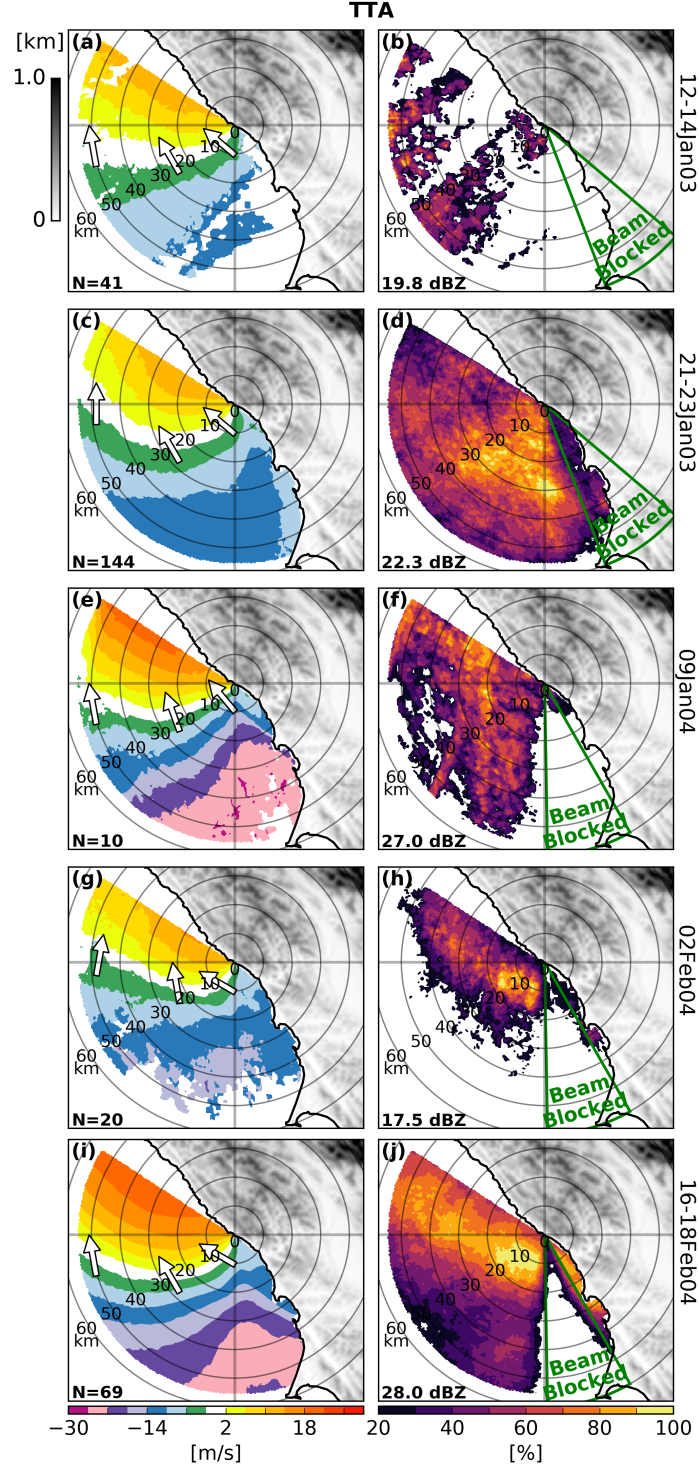


Figure 4.9: Horizontal kinematic (left panels) and precipitation (right panels) structures for individual storms during TTA periods: 12-14 January 2003 (a,b), 21-23 January 2003 (c, d), 9 January 2004 (e, f), 2 February 2004 (g, h), and 16-18 February 2004 (i, j). Number of sweeps (N) in each analysis, reflectivity threshold, and rings indicating range from X-Pol are included. See Figure 4.4 for details of each TTA period. Beam blockage by the radar trailer is indicated.

Vertical kinematic structures for TTA periods (Fig. 4.10a,c,e,g,i) are generally characterized by a meridional LLJ riding up and over a weaker airflow that corresponds to the TTA. However, there are some non-trivial storm-to-storm differences. X-Pol observations during the 12-14 January 2003 and 9 January 2004 storms (Fig. 4.10a,e) capture only a fraction of the TTA period when the meridional LLJ is relatively weak or absent (Fig. 4.4a,d). Nevertheless, it is evident that meridional winds of ~ 12 and 26 m s^{-1} , respectively, ride up and over a weaker airflow of ~ 10 and 12 m s^{-1} , respectively. The 21-23 January 2003 storm (Fig. 4.10c) shows a clear LLJ structure with meridional winds of $\sim 14 \text{ m s}^{-1}$ riding up and over a $\sim 10 \text{ m s}^{-1}$ airflow associated with the TTA. X-Pol observations during the 2 February 2004 storm (Fig. 4.10g) also capture only a fraction of the TTA period when the LLJ is weak or absent (Fig. 4.4e). In addition, a lack of low-level observations offshore prevents a clear depiction of the LLJ and TTA. The 16-18 February 2004 storm (Fig. 4.10i) is characterized by the strongest and most clearly defined LLJ with meridional winds of $\sim 24 \text{ m s}^{-1}$ being lifted above an $\sim 12 \text{ m s}^{-1}$ airflow associated with the TTA from ~ 0.5 km MSL offshore to ~ 1.0 MSL over the coastal mountains.

Corresponding median-reflectivity exceedance frequencies for TTA periods show vertical precipitation structures that are mostly consistent with offshore enhancement. The 21-23 January 2003 storm (Fig. 4.10d) indicates enhancement in a shallow layer from the surface up to $\sim 0.5 - 1.0$ km MSL and at a distance of $\sim 10-20$ km from the coast. Offshore enhancement is also clearly observed with the 16-18 February 2004 storm (Fig. 4.10j) but it occurs in a slightly deeper layer from the surface up to $\sim 1.0 - 1.5$ km MSL and has a wider horizontal extent from the coast to ~ 20 km offshore. Like its horizontal counterpart, the 12-14 January 2003 storm (Fig. 4.10b) exhibits a very different structure characterized by more isolated and cellular structure of median-reflectivity exceedance frequency. Interpretation of results for the 9 January 2004 storm is hampered by a small number of vertical scans ($n=5$, Fig. 4.10f), which makes it difficult to identify clear precipitation-enhancement patterns at low levels. Although the 2 February 2004 storm (Fig. 4.10h) shows enhancement at low levels over the coastal mountains, it is not possible to determine offshore enhancement due to radar scanning limitations for this case.

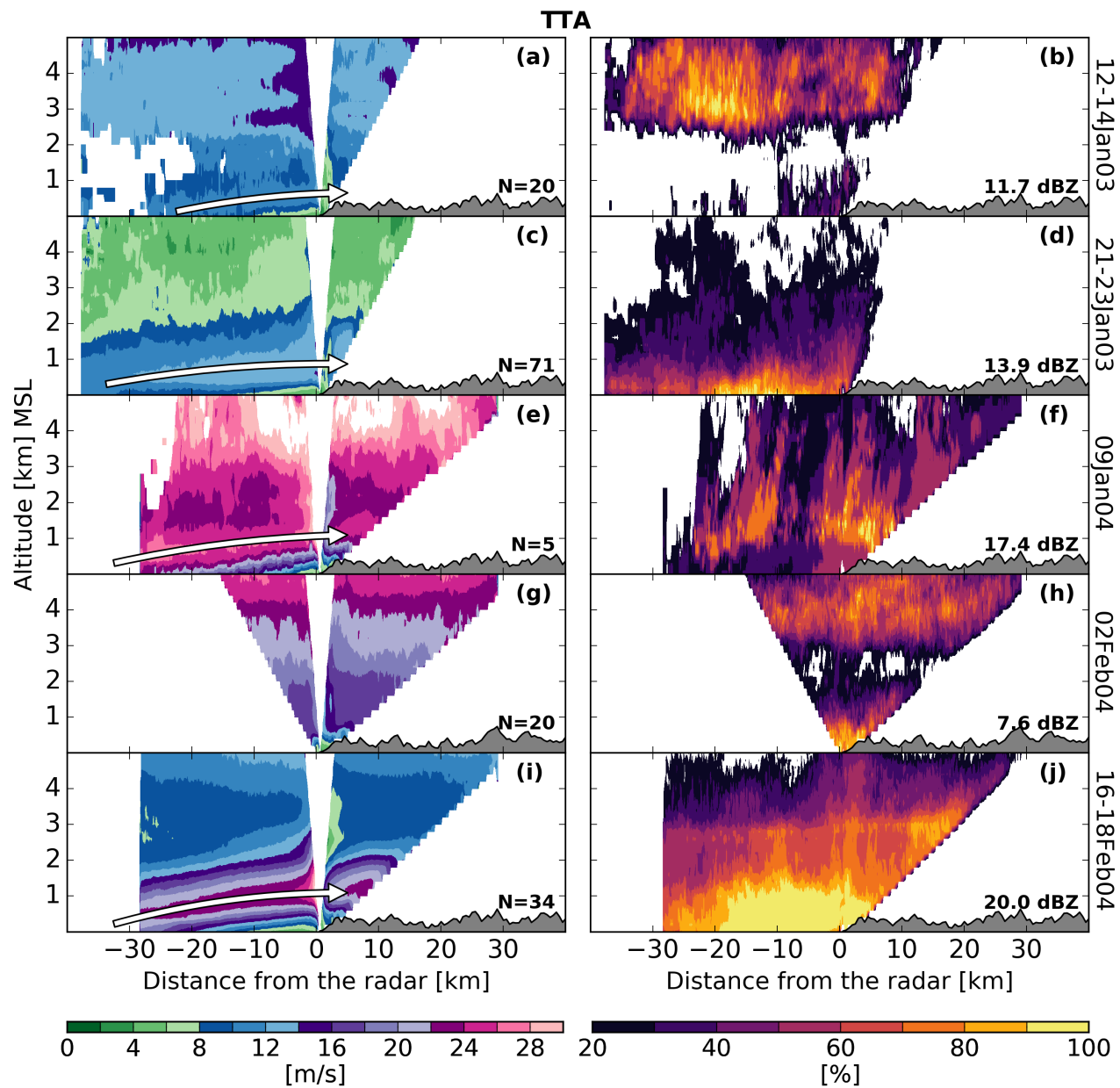


Figure 4.10: As in Fig. 4.9 but for vertical kinematic (left panels) and precipitation (right panels) structures. Arrow indicates location of low-level jet.

To strike a contrast with the TTA kinematic and precipitation structures just discussed, attention is now shifted to the inter-storm variability of NO-TTA periods. The zero isodop associated with the 12-14 January 2003 storm (Fig. 4.11a) exhibits a curvature at ~ 10 km range and then is roughly straight at distances further offshore. The 15-16 February 2003 (Fig. 4.11c) storm depicts a zero isodop with curvature at ~ 15 km range and extending mostly over the northwest quadrant of X-Pol, with inferred southwesterly winds beyond 20 km range. The 25 February 2004 storm (Fig. 4.11e) shows a zero isodop extending almost in a straight line over the third quadrant of X-Pol, indicating southeasterly winds with nearly absence of vertical wind direction shear. Maximum inbound radial velocities south of the radar for these storms are -18, -14, and -30 m s^{-1} , respectively.

Median-reflectivity exceedance frequencies during NO-TTA conditions on 12-14 January 2003 (Fig. 4.11b) portray a pattern with precipitation enhancement concentrated within ~ 10 km of the coast. The 15-16 February 2003 case (Fig. 4.11d) also shows precipitation enhancement concentrated within ~ 10 km of the coast but embedded in a more homogeneous distribution of precipitation. The 25 February 2004 storm (Fig. 4.11f) presents a homogeneous distribution of precipitation with no particular concentration of precipitation enhancement.

Kinematic and precipitation structures in NO-TTA conditions observed during 12-14 January 2003 storm (Fig. 4.11a,b) are similar to 21-23 January 2003, 9 January 2004, and 16-18 February 2004 storms, except for a maximum inbound radial velocity of -18, -22, and -26 m s^{-1} (respectively), while the 15-16 February 2003 pattern (Fig. 4.11c,d) is similar to the 2 February 2004 storm (not shown). Homogeneous structures in the NO-TTA median-reflectivity exceedance frequencies pattern during 15-16 February 2003 (Fig. 4.11d), 2 February 2004 (not shown), and 25 February 2004 (Fig. 4.11f) suggest a relatively even spatial distribution of precipitation. Let us recall that the median-reflectivity exceedance frequencies are computed over time for each grid point. Therefore, a spatially homogeneous distribution means that the majority of the grid points exceeded the median value most of the time. This pattern could be explained by widespread stratiform precipitation over the domain with small change over time.

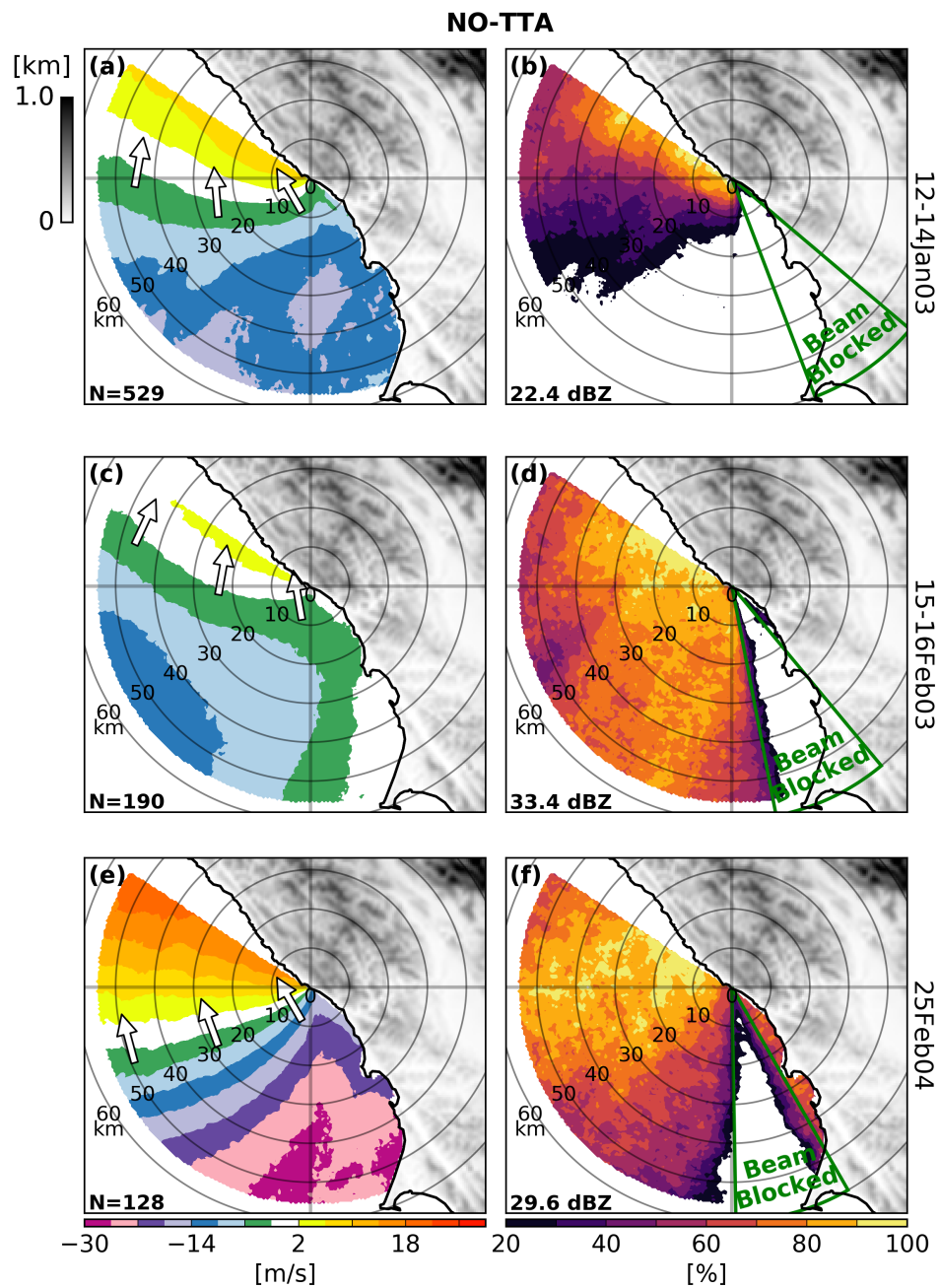


Figure 4.11: As in Fig. 4.9 but for representative horizontal kinematic (left panels) and precipitation (right panels) structures of individual storms during NO-TTA periods: 12-14 January 2003 (a,b), 15-16 February 2003 (c, d), and 25 February 2004 (e, f).

The common characteristic of vertical kinematic structures for NO-TTA periods of all storms is the apparent absence of a LLJ. Figure 4.12a,c,e illustrates this pattern for the 12-14 January 2003, 15-16 February 2003, and 25 February 2004 storms. Despite the fact that LLJs are evident in individual vertical scans during portions of NO-TTA periods (not shown), the compositing process smooths them out due to the contribution of vertical scans from extended periods of time when LLJs are effectively absent. Another characteristic of NO-TTA periods is the low-valued meridional winds within 10 km of the coast and below 500 m MSL. At first glance, this structure seems similar to what is observed during TTA conditions, just with a shorter offshore extension. By definition, this NO-TTA grouping of X-pol data is based on application of the TTA objective identification at BBY (VK17). However, the observed structure is confined so close to the radar that it may be an effect near FRS but not BBY, which would be consistent with the NO-TTA designation. Given its apparently small scale, one possible explanation for this structure might be a smaller scale TTA unresolved by wind profiler observations at BBY. Another explanation might be meridional wind deceleration and convergence produced by differential sea-land friction (e.g., Doyle 1997, Colle et al. 2008).

Vertical precipitation structures for NO-TTA periods show the 12-14 January 2003 storm (Fig. 4.12b) with maximum values of median-reflectivity exceedance frequency concentrated within the first 5 km from the coast and in the lowest ~ 1.2 km MSL. In contrast, the 15-16 February 2003 storm (Fig. 4.12d) has enhanced frequencies spread over at least 30 km from the coast and below ~ 2.0 km MSL (similar for 21-23 January 2003 and 16-18 February 2004 storms, not shown). The 25 February 2004 storm (Fig. 4.12f) is characterized by a relatively large spread over the domain but enhanced frequencies are concentrated over the coastal mountains and below ~ 3.0 km MSL (same for 9 January 2004 storm, not shown). However, this pattern is influenced by a radar beam blockage artifact along the 180° radial that is best visualized in the corresponding PPI analysis (Fig. 4.11f). Although vertical X-Pol observations during the 2 February 2004 storm are limited (e.g. Fig. 4.10h) enhanced frequencies are observed toward the coastal mountains (not shown). As in the horizontal NO-TTA precipitation analysis, vertical patterns with relatively homogeneous

spread over the domain (e.g., 21-23 January 2003, 15-16 February 2003, 9 January 2004, 16-18 February 2004, 25 February 2004) could be explained by widespread stratiform precipitation with small change over time.

4.4.4 Theoretical context

The previous section documented kinematic and precipitation structures of TTAs, but their forcing mechanisms were not addressed. Two hypotheses that have been used to explain observed TTAs along midlatitude mountain barriers during winter time are gap flow and low-level blocking. In this section we discuss some details about these theories and attempt to apply them to the TTA periods of storms examined in this study (Fig. 4.4).

In a multi-winter analysis of gap flow episodes, Neiman et al. (2006) determined that the Petaluma Gap, whose exit is located at BBY (Fig. 4.1), has a significant influence on the local distribution of rainfall. Additionally, VK15 studied the 16-18 February 2004 storm in detail and concluded that a TTA likely formed from a gap flow and created an enhanced precipitation zone offshore. The following discussion employs gap-flow theory as a theoretical framework to examine the forcing of each TTA observed. Hourly along-gap pressure difference is determined between BBY and a surface station located at Stockton (SCK, Fig. 4.1b). Zonal wind speed is derived at BBY from surface and layer-mean wind profiler observations between 0.1-0.5 km MSL (e.g., Neiman et al. 2006). These observations are examined using the analytical expression proposed by M95:

$$u^2(x) = u^2(0) - \frac{PGF}{K}e^{-2Kx} + \frac{PGF}{K} \quad (4.1)$$

where $u(0)$ and $u(x)$ represent the airflow at the entrance and at some distance x downstream the gap entrance (respectively), PGF is the along-gap pressure gradient force $(-(\rho^{-1})\delta P/\delta x)$, and $K = 2.8C_D/H$ is a parameter representing friction through a dimensionless drag coefficient ($C_D = 7.5 \times 10^{-3}$) and the average depth of the boundary layer within the gap (H). An average value of $\rho=1.24 \text{ kg m}^{-3}$ and a constant value H equal to 0.5 km MSL are assumed, which are in close agreement with Neiman et al. (2006).

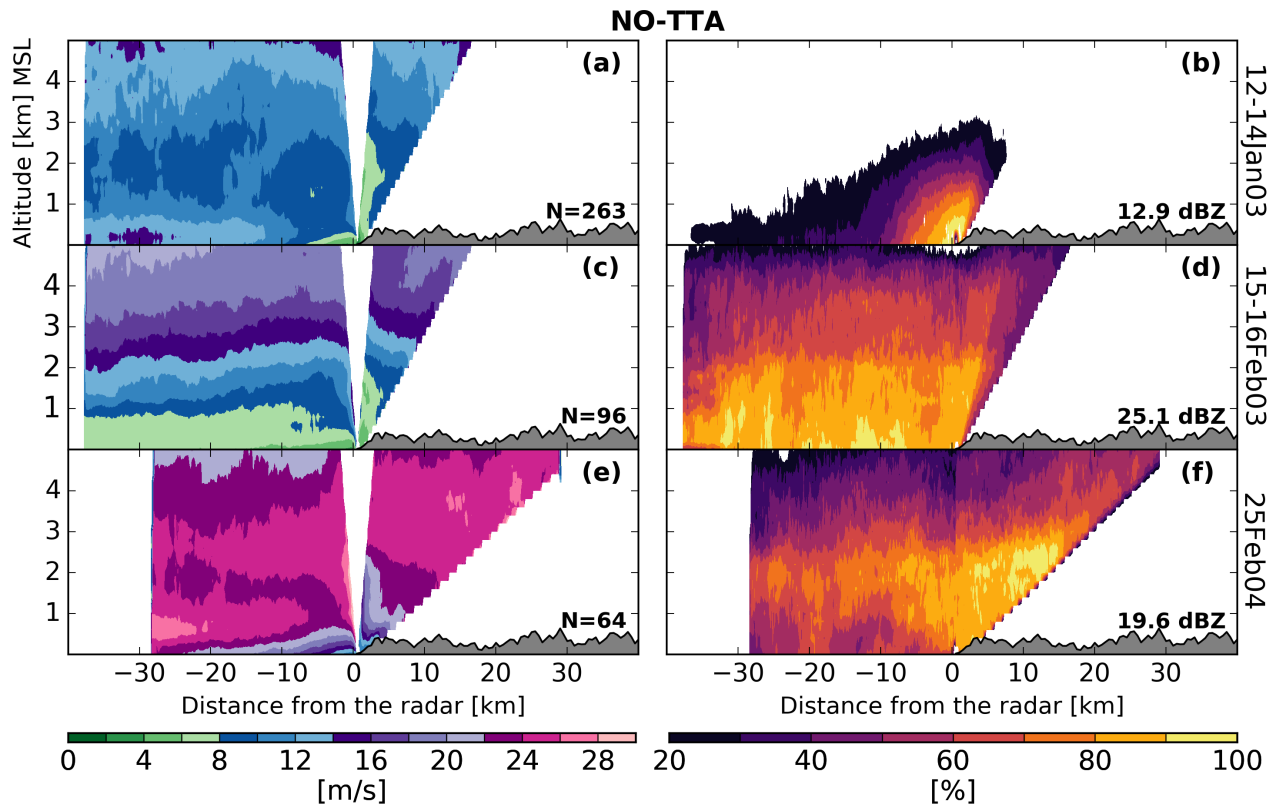


Figure 4.12: As in Fig. 4.11 but for vertical kinematic (left panels) and precipitation (right panels) structures.

The gap flow analysis of TTA periods for the 12-14 January 2003 and 21-23 January 2003 storms (Fig. 4.13a,b) show no evidence of gap flow since all points are outside of the theoretical envelope for gap-flow conditions (i.e., $\pm 50\%$ variation of C_D). In contrast, data points for the 9 January, 2 February, and 16-18 February 2004 storms are closer to or within the gap-flow theoretical envelope, which suggests that gap flows might be forcing the TTAs associated with those storms. With the uncertainties in specifying parameters such as C_D and H of Eq. 4.1, it is reasonable to expect some observed points to lie slightly outside of the theoretical envelope, even if the TTA is actually forced by a gap flow.

An alternative theory for explaining the existence of a TTA is low-level blocking. Theoretically, low-level blocking on the windward side of mountain barriers occurs when the magnitude of the upslope wind component U (e.g. kinetic energy) is lower than the buoyancy frequency N within a layer represented by the mountain height h (e.g. potential energy), which is equivalent to a Froude number (Fr) between 0 and 1 (e.g. Smith 1979, Pierrehumbert and Wyman 1985) where:

$$Fr = \frac{U}{Nh} \quad (4.2)$$

The value of N is usually represented by the Brunt-Väisälä frequency. If conditions are moist (i.e., relative humidity equal to or larger than 90%), then the moist N (N_m) is employed (Durran and Klemp 1982) to account for the latent heat of condensation. N values are computed from balloon soundings launched at BBY. Since observations of U along the coast would be biased by the presence of a TTA, we estimated U offshore using X-Pol PPI and RHI scans at a range of 40 km from the coast as in VK15.

Only the 21-23 January 2003 and 16-18 February 2004 storms had simultaneous X-Pol and balloon-sounding observations during the TTA period and therefore they are employed in the low-level blocking analysis (Figure 4.14). N is derived for two layers: the lowest 0.5 km MSL and 1.0 km MSL. The median values in each layer are used for the analysis¹. It is evident that most points are in unblocked space assuming $h = 0.5$ km but results are mixed for $h = 1.0$ km. Also,

¹ results hold the same when using mean values

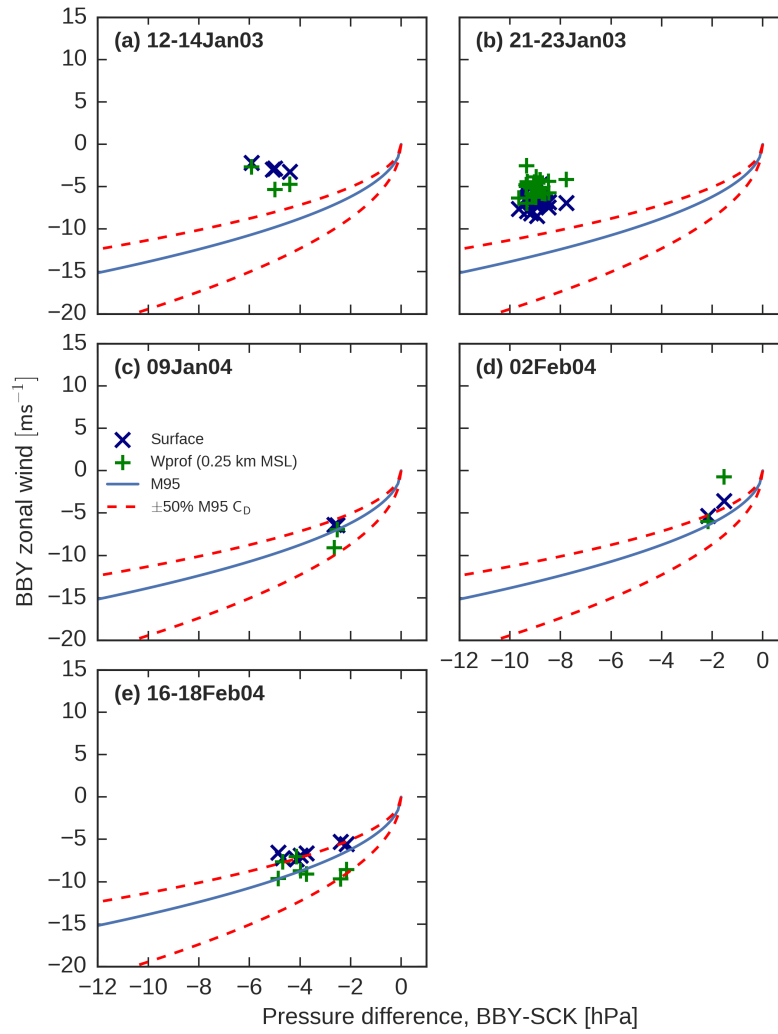


Figure 4.13: Gap flow analysis using (Mass et al. 1995; M95) analytical expression for storms with TTA. Blue line uses same drag coefficient (C_D) as in M95. Red dashed lines include a variation of plus (top) and minus (bottom) 50% of M95 C_D . A constant boundary layer altitude of 500 m is assumed.

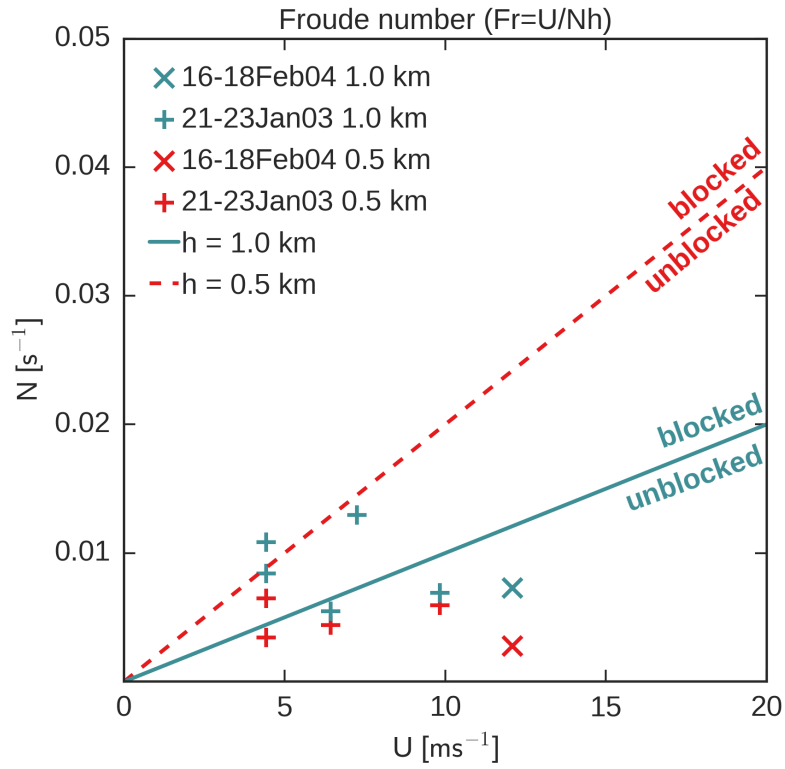


Figure 4.14: Non-dimensional Froude number analysis. Dotted and continuous line indicate $Fr = 1$ for a mountain altitude (h) of 0.5 km and 1.0 km, respectively. Annotation for blocked and unblocked areas relative to each altitude is included.

most points for the 21-23 January 2003 storm are in blocked space whereas the only point for the 16-18 February 2004 storm is in unblocked space.

The Rossby radius of deformation ($L_r = Nh/f$) represents the theoretical upstream extent of low-level blocking for a given barrier. Assuming $h = 1.0$ km, $N = 0.01$ s⁻¹, and $f = 1 \times 10^{-4}$ s⁻¹, then $L_r = 100$ km. Mountain heights of ~ 1.0 km are located ~ 50 km inland from the coast. Then, a TTA with offshore extension of ~ 40 km would be consistent with $L_r = 100$ km for a low-level blocked airflow induced by the inland mountains (Figure 4.15). As a consequence, Fr analysis suggests that at least some part of the TTA observed during 21-23 January 2003 might be forced by inland orography (~ 1.0 km altitude) rather than coastal orography (~ 0.5 km altitude) through low-level blocking. This result is consistent with Loescher et al. (2006), whose findings show a TTA influenced by terrain features as much as 100 km inland.

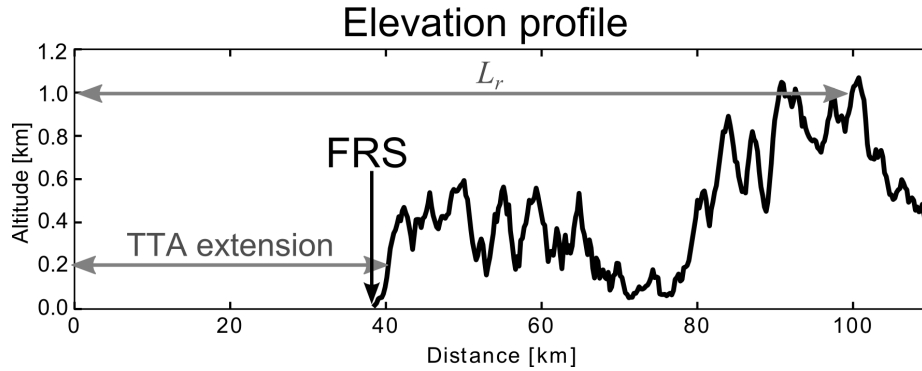


Figure 4.15: Elevation profile along the line indicated in Fig. 4.1, with azimuth 50°-230° (i.e. upslope component). Horizontal arrows indicate extension of the theoretical Rossby radius of deformation (L_r) and approximate TTA extension observed during storm 2. Vertical arrow indicates location of X-Pol radar at Fort Ross (FRS).

4.5 Summary and conclusions

This study has documented TTA kinematic and precipitation structures associated with 7 storm. Observations from scanning (X-Pol) and profiling Doppler radars were the main assets employed. Balloon soundings, surface meteorological sensors, and a GPS receiver were also part of the study.

Using X-Pol observations in a seven-storm composited analysis reveal that the average TTA kinematic structure is characterized by a significant horizontal wind direction gradient with predominant southeasterly winds along the coast and mainly from the south to southwest at a range of ~ 50 km from the coast. Average vertical kinematic structure indicates low-level jet (LLJ) of $\sim 20 \text{ m s}^{-1}$ surmounting a weaker airflow of $\sim 10 \text{ m s}^{-1}$ that corresponds to the TTA. LLJ center is displaced upward by the TTA from ~ 0.5 (offshore) to ~ 1.0 km MSL (along the coast). Horizontal and vertical precipitation structures indicate an enhanced precipitation zone offshore and roughly parallel to the coast. The precipitation enhancement zone is horizontally centered at a range of ~ 15 km from the coast and has a vertical extension of ~ 0.5 km MSL. Composited NO-TTA structures show predominant southerly component with modest horizontal directional shear, weak or absent LLJ structure, and precipitation enhancement within the first 10 km from the coast and below 1.0 km MSL.

Inter-storm variability analysis reveals relatively small variation in the TTA kinematic structure, although LLJ magnitude does present differences of $\sim 10 \text{ m s}^{-1}$ between storms, likely associated with synoptic forcing. Precipitation structures present the largest inter-storm variations during TTA periods. The precipitation structure observed during 16-18 February 2004 storm consisting in a enhanced precipitation zone offshore (~ 30 km range) and near parallel to the coast (also documented in VK15), is similar to 21-23 January 2003 and 2 February 2004 storm, with the former presenting the enhanced precipitation zone centered at ~ 25 km.

To better understand the forcing mechanisms of TTAs we examined TTA periods using the analytical expression for gap flows proposed by M95 and the non-dimensional Froude number (Fr).

Results indicate that a gap flow forcing is likely operating in the 2 February 2004 and 16-18 February 2004 TTAs. On the other hand, low-level blocking seems to be operating in the 21-23 January 2003 TTA, induced by ~ 1.0 km MSL terrain located further inland. Computation of the Rossby radius of deformation is consistent with this estimation. Although 12-14 January 2003 TTA seems to be forced by other than gap flow, there are no observations allowing to test low-level blocking, so its forcing mechanism is inconclusive.

It is worth noting that low-level blocking and gap flow are two forcing mechanisms associated with TTAs but they are not mutually exclusive. Given certain atmospheric conditions, one might be contributing more to the TTA formation and maintenance than the other. For example, in the absence of a nearby mountain gap or the presence of either weak along-gap pressure gradient or weak inland cold pool, low-level blocking might better explain TTA formation. Similarly, an initially neutral atmosphere upstream a mountain range could become increasingly stable forced by cold air brought through a gap flow, in which case a gap flow could better explain the TTA formation.

Although the inter-storm variability analysis indicates a consistent horizontal kinematic structure of sharp curvature in the zero isodop for TTA periods, precipitation structures presented significant differences during these periods. One explanation could be associated with the magnitude of upslope IVT. For example, assuming the presence of the same TTA structure, differences in the upslope component of IVT could make more or less precipitation to form and thus precipitation structures to differ.

One interesting results is a TTA (21-23 January 2003) forced by low-level blocking of mountains located further inland instead of the mountains immediately next to the coast. Since ground-based radars struggle in resolving the three-dimensional wind field in a highly complex terrain, using an airborne Doppler radar would be a better approach for a future study.

Chapter 5

Summary of major findings

Several major findings are part of this study. In Chapter 2, the use of a X-band dual polarization scanning Doppler radar (X-pol) during a winter storm observed on 16-18 February 2004 allowed to document a terrain-trapped airflow (TTA) structure along the coast of northern California. The TTA structure was observed for ~ 5 hours and up to ~ 25 km offshore of the coastline with a depth of ~ 0.5 km MSL. A pre-cold-frontal low-level jet (LLJ) impacting the coastal mountains and the TTA formed an interface offshore that moved closer to shore with a skewed orientation relative to the mean coastline orientation (i.e. interface position was further offshore toward the north). The incoming LLJ sloped from an average altitude of ~ 0.25 km MSL offshore to ~ 1 km MSL over the coast in response to the TTA. In addition, a precipitation enhancement zone was observed offshore in association with convergence from the LLJ/TTA interface. The precipitation enhancement zone moved toward the coast as the LLJ/TTA interface moved coastward.

In Chapter 3, a 13-winter-season dataset composed of hourly 915 MHz wind profiler radar observations at the coast and rain gauge observations at the coast and over the coastal mountains allowed to examine in detail the airflow associated with the small-scale mountains of northern California, the physical characteristics of TTA regimes, and TTA effects on precipitation. Considering local terrain characteristics, a wind direction $< 140^\circ$ in the average 0-500-m MSL layer was used as a first guess to identify TTAs. Considering rainy hours (CZD rain rate > 0.25 mm), the average 13-season TTA duration is 2.5 hours. The TTA (NO-TTA) regime results in a mountain/coast rainfall ratio of 1.4 (3.2) and produces a relatively larger contribution to rainfall along the coast

(20%) than over the mountain (10%). However, when used in a case study (16 February 2004), the 140° threshold produce some inconsistencies relative to the TTA identified in Chapter 2. A more detailed analysis of the relationship between wind direction and orographic rainfall reveals that a threshold of 150° more accurately divides two regimes of orographic enhancement, one with average rainfall ratio of 1.5 (TTA regime) and the other with 3.1 (NO-TTA regime). Based on a sensitivity analysis using different thresholds for wind direction, time, and layer-mean depth, it is concluded that TTA regimes are best identified by a wind direction $< 150^\circ$ in the average 0-500-m MSL layer during at least 2 hours.

In Chapter 4, using the objective TTA identification derived in Chapter 3 (layer-mean 0-500-m MSL wind direction $< 150^\circ$ during at least 2 hours) along with observations collected with X-pol during 7 storms impacting northern California allowed to study the mean kinematic and precipitation structures of TTA regimes, examine the inter-storm consistency of these structures, and compare structures with NO-TTA regimes. Kinematic structures associated with TTA regimes are consistent from storm-to-storm, presenting a horizontal structure with sharp curvature in the zero isodop (line of zero radial velocity) around 30 km from the coast, indicating strong direction shear. Vertical structures indicate a LLJ surmounting the TTA. Corresponding precipitation structures present a horizontally oriented enhanced precipitation area extending near parallel to the coast and centered around 20 km from the coast, whereas vertical structures show precipitation enhancement up to 20 km from the coast and below 1.0-km MSL. Precipitation structures however present a significant inter-storm variability, probably associated with variation in the synoptic conditions and the magnitude of the horizontal water vapor flux. They could also be affected by a relatively small sample size in specific storms. In comparison, NO-TTA kinematic structures present horizontal zero isodop suggesting modest horizontal directional shear and weak LLJ structure that are smoothed out during the relatively longer NO-TTA period. Precipitation enhancement is observed within the first 10 km from the coast and below 1.0 km MSL.

Finally, forcing mechanism of TTAs documented in Chapter 2 and 4 suggest that in most cases the terrain-trapped airflow is formed by a gap flow exiting through a narrow depression in the

coastal mountains, known as Petaluma Gap. In these cases, a pool of cold air formed in California's Central Valley accelerates through Petaluma Gap when an along-gap pressure gradient is elicited by a coastal-approaching cyclone. Yet, there is evidence suggesting that at least one TTA was formed in response to low-level blocking of 1000-m MSL terrain, even though the nearby terrain has altitudes of about 500-m MSL. Terrain of 1000-m MSL can be found about 60-km inland from the coast. This result is consistent with results of Loescher et al. (2006), where TTAs are the response not only to the most proximate terrain but also to terrain covering a broader area. Using an airborne radar would be a better approach to examine this hypothesis in a future study.

Bibliography

- Anders, A. M., G. H. Roe, D. R. Durran, and J. R. Minder, 2007: Small-Scale spatial gradients in climatological precipitation on the Olympic Peninsula. *Journal of Hydrometeorology*, **8** (5), 1068–1081, doi: 10.1175/JHM610.1.
- Bao, J. W., S. A. Michelson, P. J. Neiman, F. M. Ralph, and J. M. Wilczak, 2006: Interpretation of enhanced integrated water vapor bands associated with extratropical cyclones: Their formation and connection to tropical moisture. *Monthly Weather Review*, **134** (4), 1063–1080.
- Bergeron, T., 1965: On the low-level redistribution of atmospheric water caused by orography. *Suppl. Proc. Int. Conf. Cloud Phys.*
- Bousquet, O., and B. F. Smull, 2003: Observations and impacts of upstream blocking during a widespread orographic precipitation event. *Quarterly Journal of the Royal Meteorological Society*, **129** (588), 391–409, doi: 10.1256/qj.02.49.
- Browning, K. A., F. F. Hill, and C. W. Pardoe, 1974: Structure and mechanism of precipitation and the effect of orography in a wintertime warm sector. *Quarterly Journal of the Royal Meteorological Society*, **100** (425), 309–330, doi: 10.1002/qj.49710042505.
- Browning, K. A., and R. Wexler, 1968: The determination of kinematic properties of a wind field using doppler radar. *Journal of Applied Meteorology*, **7**, 105–113.
- Businger, S., and Coauthors, 1996: The promise of GPS in atmospheric monitoring. *Bulletin of the American Meteorological Society*, **77** (1), 5–18, doi: 10.1175/1520-0477(1996)077<0005:TPOGIA>2.0.CO;2.
- Carter, D. A., K. S. Gage, W. L. Ecklund, W. M. Angevine, P. E. Johnston, A. C. Riddle, J. Wilson, and C. R. Williams, 1995: Developments in UHF lower tropospheric wind profiling at NOAA's aeronomy laboratory. *Radio Science*, **30** (4), 977–1001, doi: 10.1029/95RS00649.
- Chen, W.-D., and R. B. Smith, 1987: Blocking and deflection of air by the Alps. *Monthly Weather Review*, **115**, 2578–2597.
- Clifford, S. F., J. C. Kaimal, R. J. Lataitis, and R. G. Strauch, 1994: Ground-based remote profiling in atmospheric studies: an overview. *Proceedings of the IEEE*, **82** (3), 313–355, doi: 10.1109/5.272138.
- Colle, B. A., Y. Lin, S. Medina, and B. F. Smull, 2008: Orographic modification of convection and flow kinematics by the Oregon coast range and Cascades during IMPROVE-2. *Monthly Weather Review*, **136** (10), 3894–3916, doi: 10.1175/2008MWR2369.1.

- Colle, B. A., and C. F. Mass, 2000: High-resolution observations and numerical simulations of easterly gap flow through the strait of Juan de Fuca on 910 December 1995. *Monthly Weather Review*, **128** (7), 2398–2422, doi: 10.1175/1520-0493(2000)128<2398:HROANS>2.0.CO;2.
- Colle, B. A., R. B. Smith, and D. A. Wesley, 2013: Theory, observations, and predictions of orographic precipitation. *Mountain Weather Research and Forecasting*, F. K. Chow, S. F. J. D. Wekker, and B. J. Snyder, Eds., Springer Atmospheric Sciences, Springer Netherlands, 291–344, doi: 10.1007/978-94-007-4098-3_6.
- Doyle, J. D., 1997: The influence of mesoscale orography on a coastal jet and rainband. *Monthly Weather Review*, **125** (7), 1465–1488.
- Durran, D. R., and J. B. Klemp, 1982: On the effects of moisture on the Brunt-Väisälä frequency. *Journal of the Atmospheric Sciences*, **39** (10), 2152–2158, doi: 10.1175/1520-0469(1982)039<2152:OTEOMO>2.0.CO;2.
- Eckhardt, S., A. Stohl, H. Wernli, P. James, C. Forster, and N. Spichtinger, 2004: A 15-year climatology of warm conveyor belts. *Journal of climate*, **17** (1), 218–237.
- Ecklund, W. L., D. A. Carter, and B. B. Balsley, 1988: A UHF wind profiler for the boundary layer: Brief description and initial results. *Journal of Atmospheric and Oceanic Technology*, **5** (3), 432–441, doi: 10.1175/1520-0426(1988)005<0432:AUWPFT>2.0.CO;2.
- Engerer, N. A., D. J. Stensrud, and M. C. Coniglio, 2008: Surface characteristics of observed cold pools. *Monthly Weather Review*, **136** (12), 4839–4849, doi: 10.1175/2008MWR2528.1.
- Fischler, M. A., and R. C. Bolles, 1981: Random sample consensus: A paradigm for model fitting with applications to image analysis and automated cartography. *Commun. ACM*, **24** (6), 381–395, doi: 10.1145/358669.358692.
- Gottschalk, P. G., and J. R. Dunn, 2005: The five-parameter logistic: A characterization and comparison with the four-parameter logistic. *Analytical Biochemistry*, **343** (1), 54–65.
- Harrold, T. W., 1973: Mechanisms influencing the distribution of precipitation within baroclinic disturbances. *Quarterly Journal of the Royal Meteorological Society*, **99** (420), 232–251.
- Houze, R. A., 2012: Orographic effects on precipitating clouds. *Reviews of Geophysics*, **50** (1), doi: 10.1029/2011RG000365.
- Houze, R. A., and S. Medina, 2005: Turbulence as a mechanism for orographic precipitation enhancement. *Journal of the Atmospheric Sciences*, **62** (10), 3599–3623, doi: 10.1175/JAS3555.1.
- Hughes, M., A. Hall, and R. G. Fovell, 2009: Blocking in areas of complex topography, and its influence on rainfall distribution. *Journal of the Atmospheric Sciences*, **66** (2), 508–518, doi: 10.1175/2008JAS2689.1.
- James, C. N., and R. A. Houze, 2005: Modification of precipitation by coastal orography in storms crossing northern California. *Monthly Weather Review*, **133** (11), 3110–3131.
- Kingsmill, D. E., P. J. Neiman, B. J. Moore, M. Hughes, S. E. Yuter, and F. M. Ralph, 2013: Kinematic and thermodynamic structures of Sierra barrier jets and overrunning atmospheric rivers during a landfalling winter storm in northern California. *Monthly Weather Review*, **141** (6), 2015–2036, doi: 10.1175/MWR-D-12-00277.1.

- Kingsmill, D. E., P. O. G. Persson, S. Haimov, and M. D. Shupe, 2016: Mountain waves and orographic precipitation in a northern Colorado winter storm. *Quarterly Journal of the Royal Meteorological Society*, **142** (695), 836–853, doi: 10.1002/qj.2685.
- Lackmann, G. M., and J. E. Overland, 1989: Atmospheric structure and momentum balance during a gap-wind event in Shelikof Strait, Alaska. *Monthly Weather Review*, **117** (8), 1817–1833, doi: 10.1175/1520-0493(1989)117<1817:ASAMBD>2.0.CO;2.
- Larson, V. E., R. Wood, P. R. Field, J.-C. Golaz, T. H. V. Haar, and W. R. Cotton, 2001: Small-Scale and Mesoscale Variability of Scalars in Cloudy Boundary Layers: One-Dimensional Probability Density Functions. *Journal of the Atmospheric Sciences*, **58** (14), 1978–1994.
- Loescher, K. A., G. S. Young, B. A. Colle, and N. S. Winstead, 2006: Climatology of barrier jets along the Alaskan coast. Part I: Spatial and temporal distributions. *Monthly Weather Review*, **134** (2), 437–453.
- Markowski, P. M., and Y. P. Richardson, 2010: *Mesoscale Meteorology in Midlatitudes*. John Wiley & Sons, 424+ pp.
- Martner, B. E., K. A. Clark, S. Y. Matrosov, W. C. Campbell, and J. S. Gibson, 2001: NOAA/ETLs polarization-upgraded x-band hydro radar. *30th International Conf. on Radar Meteorology*.
- Marwitz, J. D., 1983: The kinematics of orographic airflow during sierra storms. *Journal of the Atmospheric Sciences*, **40** (5), 1218–1227, doi: 10.1175/1520-0469(1983)040<1218:TKOOAD>2.0.CO;2.
- Mass, C. F., S. Businger, M. D. Albright, and Z. A. Tucker, 1995: A windstorm in the lee of a gap in a coastal mountain barrier. *Monthly Weather Review*, **123** (2), 315–331, doi: 10.1175/1520-0493(1995)123<0315:AWITLO>2.0.CO;2.
- Matrosov, S. Y., D. E. Kingsmill, B. E. Martner, and F. M. Ralph, 2005: The utility of X-band polarimetric radar for quantitative estimates of rainfall parameters. *Journal of hydrometeorology*, **6** (3), 248–262.
- Mayr, G. J., and Coauthors, 2007: Gap flows: Results from the mesoscale alpine programme. *Quarterly Journal of the Royal Meteorological Society*, **133** (625), 881–896, doi: 10.1002/qj.66.
- McCauley, M. P., and A. P. Sturman, 1999: A study of orographic blocking and barrier wind development upstream of the Southern Alps, New Zealand. *Meteorology and Atmospheric Physics*, **70** (3), 121–131, doi: 10.1007/s007030050029.
- Medina, S., B. F. Smull, R. A. Houze, and M. Steiner, 2005: Cross-barrier flow during orographic precipitation events: Results from MAP and IMPROVE. *Journal of the Atmospheric Sciences*, **62** (10), 3580–3598, doi: 10.1175/JAS3554.1.
- Medina, S., E. Sukovich, and R. A. Houze, 2007: Vertical structures of precipitation in cyclones crossing the Oregon Cascades. *Monthly Weather Review*, **135** (10), 3565–3586, doi: 10.1175/MWR3470.1.
- Minder, J. R., and D. E. Kingsmill, 2012: Mesoscale variations of the atmospheric snow line over the northern sierra nevada: Multiyear statistics, case study, and mechanisms. *Journal of the Atmospheric Sciences*, **70** (3), 916–938, doi: 10.1175/JAS-D-12-0194.1.

- Neiman, P. J., F. Martin Ralph, P. O. G. Persson, A. B. White, D. P. Jorgensen, and D. E. Kingsmill, 2004: Modification of fronts and precipitation by coastal blocking during an intense landfalling winter storm in southern California: Observations during CALJET. *Monthly Weather Review*, **132** (1), 242–273.
- Neiman, P. J., P. T. May, and M. A. Shapiro, 1992: Radio acoustic sounding system (RASS) and wind profiler observations of lower- and midtropospheric weather systems. *Monthly Weather Review*, **120** (10), 2298–2313, doi: 10.1175/1520-0493(1992)120<2298:RASSAW>2.0.CO;2.
- Neiman, P. J., F. M. Ralph, A. B. White, D. E. Kingsmill, and P. O. G. Persson, 2002: The statistical relationship between upslope flow and rainfall in California’s coastal mountains: Observations during CALJET. *Monthly Weather Review*, **130** (6), 1468–1492.
- Neiman, P. J., F. M. Ralph, A. B. White, D. D. Parrish, J. S. Holloway, and D. L. Bartels, 2006: A multiwinter analysis of channeled flow through a prominent gap along the northern California coast during CALJET and PACJET. *Monthly Weather Review*, **134** (7), 1815–1841.
- Neiman, P. J., F. M. Ralph, G. A. Wick, J. D. Lundquist, and M. D. Dettinger, 2008: Meteorological characteristics and overland precipitation impacts of atmospheric rivers affecting the west coast of North America based on eight years of SSM/I satellite observations. *Journal of Hydrometeorology*, **9** (1), 22–47, doi: 10.1175/2007JHM855.1.
- Neiman, P. J., A. B. White, F. M. Ralph, D. J. Gottas, and S. I. Gutman, 2009: A water vapour flux tool for precipitation forecasting. *Proceedings of the Institution of Civil Engineers - Water Management*, **162** (2), 83–94, doi: 10.1680/wama.2009.162.2.83.
- Neiman, P. J., G. A. Wick, F. M. Ralph, B. E. Martner, A. B. White, and D. E. Kingsmill, 2005: Wintertime nonbrightband rain in California and Oregon during CALJET and PACJET: Geographic, interannual, and synoptic variability. *Monthly Weather Review*, **133** (5), 1199–1223.
- Overland, J., and B. A. Walter, 1981: Gap winds in the strait of Juan de Fuca. *Monthly Weather Review*, **109**, 2221–2233.
- Overland, J. E., 1984: Scale analysis of marine winds in straits and along mountainous coasts. *Monthly Weather Review*, **112** (12), 2530–2534, doi: 10.1175/1520-0493(1984)112<2530:SAOMWI>2.0.CO;2.
- Peterson, T. C., L. O. Grant, W. R. Cotton, and D. C. Rogers, 1991: The effect of decoupled low-level flow on winter orographic clouds and precipitation in the Yampa River Valley. *Journal of Applied Meteorology*, **30**, 368–386.
- Pierrehumbert, R. T., and B. Wyman, 1985: Upstream effects of mesoscale mountains. *Journal of the Atmospheric Sciences*, **42** (10), 977–1003.
- Queney, P., 1948: The problem of air flow over mountains: A summary of theoretical studies. *Bull. Am. Meteorol. Soc.*, **29**, 16–26.
- Ralph, F. M., P. J. Neiman, and R. Rotunno, 2005: Dropsonde observations in low-level jets over the northeastern Pacific Ocean from CALJET-1998 and PACJET-2001: Mean vertical-profile and atmospheric-river characteristics. *Monthly Weather Review*, **133** (4), 889–910.

- Ralph, F. M., P. J. Neiman, and G. A. Wick, 2004: Satellite and CALJET aircraft observations of atmospheric rivers over the eastern North Pacific Ocean during the winter of 1997/98. *Monthly Weather Review*, **132** (7), 1721–1745.
- Ralph, F. M., P. J. Neiman, G. A. Wick, S. I. Gutman, M. D. Dettinger, D. R. Cayan, and A. B. White, 2006: Flooding on California’s Russian River: Role of atmospheric rivers. *Geophysical Research Letters*, **33** (13), doi: 10.1029/2006GL026689.
- Ralph, F. M., and Coauthors, 2013: The emergence of weather-related test beds linking research and forecasting operations. *Bulletin of the American Meteorological Society*, **94** (8), 1187–1211, doi: 10.1175/BAMS-D-12-00080.1.
- Roe, G. H., 2005: Orographic precipitation. *Annual Review of Earth and Planetary Sciences*, **33** (1), 645–671, doi: 10.1146/annurev.earth.33.092203.122541.
- Rutledge, S. A., and P. Hobbs, 1983: The mesoscale and microscale structure and organization of clouds and precipitation in midlatitude cyclones. VIII: A model for the seeder-feeder process in warm-frontal rainbands. *Journal of the Atmospheric Sciences*, **40** (5), 1185–1206, doi: 10.1175/1520-0469(1983)040<1185:TMAMSA>2.0.CO;2.
- Saha, S., and Coauthors, 2010: The NCEP climate forecast system reanalysis. *Bulletin of the American Meteorological Society*, **91** (8), 1015–1057, doi: 10.1175/2010BAMS3001.1.
- Saucier, W. J., 1989: *Principles of Meteorological Analysis*. Dover Publications.
- Schultz, D. M., 2001: Reexamining the cold conveyor belt. *Monthly Weather Review*, **129** (9), 2205–2225.
- Sinclair, M. R., D. S. Wratt, R. D. Henderson, and W. R. Gray, 1997: Factors affecting the distribution and spillover of precipitation in the Southern Alps of New Zealand-A case study. *Journal of Applied Meteorology*, **36** (5), 428–442.
- Smith, R., 1979: *The influence of mountains on the atmosphere*, Vol. 21. Academic Press, New York; San Francisco; London.
- Steenburgh, W. J., D. M. Schultz, and B. A. Colle, 1998: The structure and evolution of gap outflow over the Gulf of Tehuantepec, Mexico. *Monthly Weather Review*, **126** (10), 2673–2691.
- Trapp, R. J., and C. A. Doswell, 2000: Radar data objective analysis. *Journal of Atmospheric and Oceanic Technology*, **17** (2), 105–120.
- Trenberth, K. E., A. Dai, R. M. Rasmussen, and D. B. Parsons, 2003: The changing character of precipitation. *Bulletin of the American Meteorological Society*, **84** (9), 1205–1217, doi: 10.1175/BAMS-84-9-1205.
- Valenzuela, R. A., and D. E. Kingsmill, 2015: Orographic precipitation forcing along the coast of Northern California during a landfalling winter storm. *Monthly Weather Review*, **143** (9), 3570–3590, doi: 10.1175/MWR-D-14-00365.1.
- Valenzuela, R. A., and D. E. Kingsmill, 2017a: Terrain-trapped airflows and orographic rainfall along the coast of northern California. Part I: Kinematic characterization using a wind profiling radar. *Monthly Weather Review*, **To be submitted**.

- Valenzuela, R. A., and D. E. Kingsmill, 2017b: Terrain-trapped airflows and orographic rainfall along the coast of northern California. Part II: Horizontal and vertical structures observed by a scanning Doppler radar. *Monthly Weather Review*, **To be submitted**.
- Weber, B. L., D. B. Wuertz, D. C. Welsh, and R. McPeck, 1993: Quality controls for profiler measurements of winds and RASS temperatures. *Journal of Atmospheric and Oceanic Technology*, **10** (4), doi: 10.1175/1520-0426(1993)010<0452:QCFPMO>2.0.CO;2.
- Weber, R., 1991: Estimator for the standard deviation of wind direction based on moments of the Cartesian components. *Journal of Applied Meteorology*, **30** (9), 1341–1353, doi: 10.1175/1520-0450(1991)030.
- Wentz, F. J., 1997: A well-calibrated ocean algorithm for special sensor microwave / imager. *Journal of Geophysical Research: Oceans*, **102**, 8703–8718, doi: 10.1029/96JC01751.
- White, A. B., P. J. Neiman, F. M. Ralph, D. E. Kingsmill, and P. O. G. Persson, 2003: Coastal orographic rainfall processes observed by radar during the California Land-Falling Jets Experiment. *Journal of Hydrometeorology*, **4** (2), 264–282.
- Wick, G. A., P. J. Neiman, and F. M. Ralph, 2013: Description and validation of an automated objective technique for identification and characterization of the integrated water vapor signature of atmospheric rivers. *IEEE Transactions on Geoscience and Remote Sensing*, **51** (4), 2166–2176, doi: 10.1109/TGRS.2012.2211024.
- Wilks, 2011: *Statistical Methods in the Atmospheric Sciences, Volume 100, Third Edition*. 3rd ed., Academic Press, Amsterdam ; Boston.
- Wolfe, D. E., and S. I. Gutman, 2000: Developing an operational, surface-based, GPS, water vapor observing system for NOAA: Network design and results. *Journal of Atmospheric and Oceanic Technology*, **17** (4), 426–440, doi: 10.1175/1520-0426(2000)017<0426:DAOSBG>2.0.CO;2.
- Wood, V. T., and R. A. Brown, 1986: Single doppler velocity signature interpretation of nondivergent environmental winds. *Journal of Atmospheric and Oceanic Technology*, **3** (1), 114–128.
- Yu, C.-K., and N. A. Bond, 2002: Airborne doppler observations of a cold front in the vicinity of vancouver island. *Monthly Weather Review*, **130** (11), 2692–2708, doi: 10.1175/1520-0493(2002)130<2692:ADOOAC>2.0.CO;2.
- Yu, C.-K., and Y.-H. Hsieh, 2009: Formation of the convective lines off the mountainous coast of southeastern taiwan: A case study of 3 january 2004. *Monthly Weather Review*, **137** (9), 3072–3091, doi: 10.1175/2009MWR2867.1.
- Yu, C.-K., D. P. Jorgensen, and F. Roux, 2007: Multiple precipitation mechanisms over mountains observed by airborne doppler radar during MAP IOP5. *Monthly Weather Review*, **135** (3), 955–984, doi: 10.1175/MWR3318.1.
- Yu, C.-K., and B. F. Smull, 2000: Airborne doppler observations of a landfalling cold front upstream of steep coastal orography. *Monthly Weather Review*, **128** (6), 1577–1603, doi: 10.1175/1520-0493(2000)128<1577:ADOOAL>2.0.CO;2.

- Yuter, S. E., and R. A. Houze, 1995: Three-dimensional kinematic and microphysical evolution of florida cumulonimbus. part II: Frequency distributions of vertical velocity, reflectivity, and differential reflectivity. *Monthly Weather Review*, **123** (7), 1941–1963, doi: 10.1175/1520-0493(1995)123<1941:TDKAME>2.0.CO;2.
- Yuter, S. E., D. A. Stark, J. A. Crouch, M. J. Payne, and B. A. Colle, 2011: The impact of varying environmental conditions on the spatial and temporal patterns of orographic precipitation over the Pacific northwest near Portland, Oregon. *Journal of Hydrometeorology*, **12** (3), 329–351, doi: 10.1175/2010JHM1239.1.
- Zhu, Y., and R. E. Newell, 1994: Atmospheric rivers and bombs. *Geophysical Research Letters*, **21** (18), 1999–2002.
- Zhu, Y., and R. E. Newell, 1998: A proposed algorithm for moisture fluxes from atmospheric rivers. *Monthly Weather Review*, **126** (3), 725–735.