

**The Evening Transition of the Atmospheric
Boundary Layer: Inertial Oscillations and
Boundary-Layer Dynamics**

by

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The Evening Transition of the Atmospheric Boundary Layer: Inertial Oscillations and
Boundary-Layer Dynamics

Thesis directed by Prof. William Blumen

In this study, I use observations and a first-order closure model to study inertial oscillations in the fair-weather atmospheric boundary layer over land. I attempt to determine their origin and confirm the hypothesis developed by Blackadar, that inertial oscillations commence at the evening transition of the boundary due to decoupling of winds aloft from those beneath the surface inversion.

I use two months of data from 915 MHz boundary-layer wind profilers associated with the CASES field experiments. Using the data and a new technique for extracting inertial oscillations from boundary-layer winds, I document quantitatively that inertial oscillations are generated at the evening transition as well as at other times in the diurnal cycle. I verify that inertial oscillations are present in the wind speed maximum of the nocturnal low-level jet.

I run a one-dimensional first-order closure numerical model to model the dynamics of one night. The model reproduces observed inertial oscillations. The amplitudes of the observed inertial oscillations strongly correlate with the magnitude of the ageostrophic wind at model initiation. Residual mixing aloft causes the amplitudes of the inertial oscillations to deviate slightly from the profile dictated by the initial ageostrophic wind. Using a simple analytical model, I explore the limit in which turbulent stress divergence is small compared to the ageostrophic wind to confirm that whenever the turbulent stress divergence becomes small compared to the ageostrophic term in the momentum equations, inertial oscillations commence with amplitudes equal to the magnitude of the ageostrophic component of the wind at that time.

Dedication

To my parents, Charles and Linda Lundquist.

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Contents

Chapter

1	Introduction	2
1.1	The atmospheric boundary layer	2
1.2	Overview of the CASES field programs	6
1.3	Overview of thesis	7
2	Datasets and Instrumentation	11
2.1	The CASES-97 and CASES-99 datasets	11
2.2	The suitability of new instrument platforms for boundary-layer research	14
2.2.1	GPS sondes	14
2.2.2	915 MHz boundary-layer wind profilers	22
3	Identifying inertial oscillations in boundary-layer data	31
3.1	Spectral analysis of wind profiler data	33
3.2	A new technique for identifying inertial oscillations	35
3.3	Observed inertial oscillations	48
3.3.1	Inertial oscillation commencement times	53
3.3.2	Discerning the time of initiation of an inertial oscillation	56
3.3.3	Comparison between observations and the Blackadar scenario	63
3.4	Interaction of inertial oscillations with the LLJ	69
3.5	Summary	74

4	Climatology of the Low-Level Jet	77
4.1	Background and Importance of the LLJ	78
4.2	LLJs observed in CASES-97 and CASES-99	81
4.2.1	Diurnal variability of LLJs	84
4.2.2	Variation of LLJ height with time of day	85
4.2.3	Wind Speed and Wind Direction at jet max	91
4.3	Correlation of LLJs and inertial oscillations	91
4.4	Summary of LLJ Climatology	94
5	Inertial oscillations in a first-order 1D numerical model	100
5.1	Model Equations	101
5.2	Parameterization of longwave radiative cooling	105
5.3	Parameterizations of k_m , the turbulent exchange coefficient	108
5.4	Model Simulations	115
5.5	Analytical verification of the importance of the initial ageostrophic wind	137
6	Conclusions and Suggestions for Future Research	144
	Bibliography	150
	Appendix	
A	Supplementary Spectral Analysis Figures	156

Figures

Figure

1.1	Map of the Walnut River Watershed	8
2.1	Map of CASES-97 instrumentation	13
2.2	Map of CASES-99 instrumentation	15
2.3	Intercomparison of GLASS and HRDL wind data	17
2.4	Lowest usable wind data points for CASES-99 soundings	19
2.5	Heights of the 10th lowest data point for CASES-99 soundings	20
2.6	Number of the 302 CASES-99 soundings with at least one usable wind data point within each 10m height bin.	21
2.7	Intercomparison of GLASS and profiler winds, 11 Oct 1999 0300 UTC, Beaumont	27
2.8	Intercomparison of GLASS and profiler winds, 11 Oct 1999 0300 UTC, Oxford	28
2.9	Backscatter from Oxford profiler, 11 October 1999	29
3.1	Power Spectra, Oxford profiler, CASES-97	36
3.2	Power Spectra, Oxford profiler, CASES-99	37
3.3	Power Spectra, Whitewater profiler, CASES-97	38
3.4	Power Spectra, Whitewater profiler, CASES-99	39
3.5	Power Spectra, El Dorado profiler, CASES-99	40

3.6	Schematic for finding the steady-state components of wind time series. .	44
3.7	Hodograph of a successful inertial oscillation fit	47
3.8	PDF for a successful inertial oscillation fit	49
3.9	Characteristics of a successful inertial oscillations fit	50
3.10	Comparison of best-fit inertial oscillations using two different techniques	51
3.11	Blackadar's schematic of wind profile with inertial oscillations	54
3.12	Update of Blackadar's schematic for a varying geostrophic wind	55
3.13	Commencement times of inertial oscillations, CASES-97 data	57
3.14	Commencement times of inertial oscillations, CASES-99 data	58
3.15	CASES-97: Times of alignment between inertial oscillations and mean wind	61
3.16	CASES-99: Times of alignment between inertial oscillations and mean wind	62
3.17	Comparison of 4-5 May 1997 to Blackadar's hypothesis	65
3.18	Comparison of 10-11 May 1997 to Blackadar's hypothesis	66
3.19	Comparison of 10-11 October 1999 to Blackadar's hypothesis	67
3.20	Comparison of 11-12 October 1999 to Blackadar's hypothesis	68
3.21	Comparison of 22-23 October 1999 to Blackadar's hypothesis	70
3.22	Wind speed and potential temperature profiles from 10-11 October 1999, GLASS data	71
3.23	Wind speed profiles from 10-11 October 1999, El Dorado profiler data .	72
3.24	Characteristics of inertial oscillations starting at 0000 UTC 11 October 1999	75
4.1	Diurnal frequencies of CASES-97 LLJs	86
4.2	Diurnal frequencies of CASES-99 LLJs	87
4.3	CASES-97: Diurnal variation in heights of wind speed maximums	89

4.4	CASES-99: Diurnal variation in heights of wind speed maximums	90
4.5	CASES-97: Wind speed and direction at the height of the wind speed maximum	92
4.6	CASES-99: Wind speed and direction at the height of the wind speed maximum	93
4.7	CASES-97: Diurnal variation of LLJs with inertial oscillations at the wind speed maximum	95
4.8	CASES-99: Diurnal variation of LLJs with inertial oscillations at the wind speed maximum	96
4.9	CASES-97: Diurnal variation of LLJs without inertial oscillations at the wind speed maximum	97
4.10	CASES-99: Diurnal variation of LLJs without inertial oscillations at the wind speed maximum	98
5.1	Cooling rate as calculated by the MDTERP narrow band model for a CASES-99 sounding, 5 Oct 1999, 0310 UTC	107
5.2	Contours of potential temperature and longwave cooling rates from 5-6 October 1999	110
5.3	Profiles of k_m resulting from the inverse modeling attempt	112
5.4	Surface radiative temperature and 2 m potential temperature, 2300 UTC 11 October - 1500 UTC 12 October 1999	117
5.5	Potential temperature profiles, 0700 UTC 12 October 1999	118
5.6	Model results and radiosonde observations for model run initialized with the 2300 UTC El Dorado sounding of 11 October 1999	120
5.7	Model results and radiosonde observations for model run initialized with the 2309 UTC Beaumont sounding of 11 October 1999	121

5.8	Model results and radiosonde observations for model run initialized with the 2314 UTC Smileyburg sounding of 11 October 1999	122
5.9	Profiler and sonde winds for 2300 UTC 11 October 1999 at El Dorado .	123
5.10	El Dorado profiler and Beaumont sonde winds for 2309 UTC 11 October 1999	124
5.11	El Dorado profiler and Smileyburg sonde winds for 2314 UTC 11 October 1999	125
5.12	Observed IO amplitudes and results of model run initiated with the 2300 UTC El Dorado sounding of 11 October 1999	126
5.13	Observed inertial oscillation amplitudes and results of model run initiated with the 2309 UTC Beaumont sounding of 11 October 1999	127
5.14	Observed inertial oscillation amplitudes and results of model run initiated with the 2314 UTC Smileyburg sounding of 11 October 1999	128
5.15	Turbulent and ageostrophic terms for model run initiated with the 2300 UTC El Dorado sounding of 11 October 1999	130
5.16	Turbulent and ageostrophic terms for model run initiated with the 2309 UTC Beaumont sounding of 11 October 1999	131
5.17	Turbulent and ageostrophic terms for model run initiated with the 2314 Smileyburg sounding of 11 October 1999	132
5.18	Initial ageostrophic profile and amplitudes of IOs generated in model run initialized with the 2300 UTC El Dorado sounding of 11 October 1999 .	134
5.19	Initial ageostrophic profile and amplitudes of IOs generated in model run initialized with the 2309 UTC Beaumont sounding of 11 October 1999 .	135
5.20	Initial ageostrophic profile and amplitudes of IOs generated in model run initialized with the 2314 UTC Smileyburg sounding of 11 October 1999	136

5.21	Model results and radiosonde observations for model run initialized with the 2309 Beaumont sounding of 11 October 1999, k_m parameterized with (5.4)	138
5.22	Initial ageostrophic profile and amplitudes of IOs generated in model run initialized with the 2309 UTC Beaumont sounding of 11 October 1999, k_m parameterized with (5.4)	139
A.1	Power spectra, Beaumont profiler, CASES-97	157
A.2	Power spectra, Beaumont profiler, CASES-99	158
A.3	Power spectra, Smileyburg, CASES-99	159

Tables

Table

2.1	Specifications for CASES boundary-layer profilers	12
3.1	CASES-97: nights with IOs generated near the time of the evening transition	59
3.2	CASES-99: nights with IOs generated near the time of the evening transition	59
4.1	Classification Categories for LLJ structures, as defined by Bonner (1968) and expanded by Whiteman et al. (1997)	83
4.2	LLJ Frequencies (%) for the Bonner (1968) and Whiteman et al. (1997) studies	83
4.3	% LLJ Classification of all good profiles	83
4.4	% LLJ Classification of all good profiles	83
4.5	Correlation between data outages and llj frequency	85
4.6	Correlation between data outages and llj frequency	85
4.7	Frequency of IO occurrence during LLJ events at the height of the wind speed maximum, using CASES-97 (with respect to number of LLJ events)	94
4.8	Frequency of IO occurrence during LLJ events at the height of the wind speed maximum, using CASES-99 (with respect to number of LLJ events)	94

5.1 Atmospheric mixing ratios [g/kg] of chemical species used in the MDTERP radiation model	106
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TO THE SUN-SET BREEZE

Ah, whispering, something again, unseen,
Where late this heated day thou enterest at my window, door,
(Thou hast, O Nature! elements! utterance to my heart beyond the
rest - and this is of them,)
So sweet thy primitive taste to breathe within - thy soothing fingers on
my face and hands,
I feel the sky, the prairie vast - I feel the mightly northern lakes,
I feel the ocean and the forest - somehow I feel the globe itself
swift-swimming in space;
Minister to speak to me, here and now, what word has never told, and
cannot tell,
Art thou not universal concrete's distillation? Law's, all Astronomy's
last refinement?
Hast thou no soul? Can I not know, identify thee?

Walt Whitman, 1891

Chapter 1

Introduction

Knowledge may have its purposes,
but guessing is always
more fun than knowing.

W.H. Auden, “Archaeology”, 1973

1.1 The atmospheric boundary layer

For most humans, every breath that we take comes from a region of the atmosphere known as the atmospheric boundary layer. This small sliver of atmosphere tucked between the Earth’s surface and the free atmosphere is the only portion of the atmosphere that directly feels the effect of the Earth’s surface. As the Sun’s heating of the surface varies over the course of a day, so does the boundary layer’s response to that surface forcing. Although the vertical extent of the fair-weather atmospheric boundary layer over land varies from approximately 100 m at night or over a cool surface to as high as 6 km over hot desert environments, the dynamics of the entire 100 km of atmosphere are strongly influenced by the boundary layer. Over half of the atmosphere’s kinetic energy sink is in the boundary layer (Garratt, 1992), and all water vapor entering the atmosphere from the surface must pass through the boundary layer as it evaporates. For numerical weather prediction models or global climate models to succeed in describing the behavior of the Earth’s atmosphere, these models must incorporate accurate parameterizations of the boundary layer.

Heat, momentum, and moisture are exchanged between the free atmosphere and the Earth's surface by turbulent processes within the boundary layer. During the day, when solar heating of the surface generates sweeping convective thermals with length scales as large as the depth of the boundary layer, the boundary layer over land is dominated by convective turbulence. The boundary layer constantly mixes its own turbulent frictionally-constrained air, full of moisture and pollutants from the surface, with the less turbulent air of the free atmosphere above through entrainment. When dominated by convection, the boundary layer teems with turbulent eddies of all sizes from the huge thermals to tiny fluctuations with length scales on the order of a few millimeters. The dynamics of turbulence under this quasi-stationary regime, when the time rate of change of turbulent quantities is small compared to the turbulent production and loss terms, are reasonably well-understood (Nieuwstadt and Brost, 1986). But as the sun's influence wanes in the late afternoon, finally disappearing at sunset, the energetic turbulence of the boundary layer dissipates. The rapidly cooling surface cools the atmosphere immediately above it. A surface-based inversion develops and the boundary layer becomes stably stratified. As convection ceases near the surface, winds also calm (Mahrt, 1981).

This evening transition of the boundary layer from a convective to a stably-stratified regime is a magical time both close to the surface and aloft. Near the surface, the temperature inversion and low wind speeds make the early evening an optimal time for the transmission of low-frequency sound: the range of a 15 Hz signal is 200 km under these conditions, while the daytime range of such a signal is 30 km. In the African savannah, elephants exploit this enhanced range for communication with distant herds (Larom et al., 1997). Meanwhile, aloft, remnants of convective turbulent plumes decay and turbulence ceases to be isotropic. Equilibrium between turbulent production and loss no longer exists, and the internal dynamics of turbulence must be considered to describe the time evolution of the boundary layer. In their large-eddy simulation study

of an idealized evening transition (surface heat flux becomes zero abruptly instead of gradually), Nieuwstadt and Brost (1986) find that large eddies decay first, from the bottom up, while entrainment in the upper extent of the boundary layer can continue unmitigated even while fluxes have decreased by an order of magnitude in most of the boundary layer. In a later study that allows surface flux to drop off gradually, Sorbjan (1997) confirms that entrainment aloft can continue even while a surface inversion has developed. Many open questions regarding the decay of convective turbulence remain, including the effects of surface heterogeneity and clouds on the decay process.

While the details of the decay of convective turbulence are still being investigated, a forty-year-old one-dimensional model of the effects of that decay has proven quite useful. In 1957, Blackadar proposed that the decoupling of winds above the surface inversion from winds closer to the surface induces accelerations aloft. This decoupling occurs rapidly after sunset. The ageostrophic component of the wind, which was previously constrained by the turbulent stresses, can now accelerate. Because of the Earth's rotation, the ageostrophic wind vector also rotates in the form of inertial oscillations, horizontal circles in the wind field with a period equal to $2\pi/f$, where f is the Coriolis parameter. For example, at a latitude of 45° , an inertial oscillation would complete an entire circle in 16.9 hours. At each level aloft, these inertial oscillations eventually align with the mean wind, forming a low-level jet (LLJ) structure in the wind profile. The wind speed maximum, according to Blackadar, will be located at the height above the inversion where the ageostrophic wind was largest during the day. Thorpe and Guymer (1977) use a two-layer slab model, with smaller turbulent stresses in the upper layer and larger turbulent stresses in the lower layer, to verify that the layers above the surface inversion inertially oscillate to form a wind speed maximum above the surface inversion. This model has proven useful to many researchers who seek to explain observed nocturnal wind profiles: the paper of Thorpe and Guymer (1977) has been cited 49 times between 1977 and 1999 according to the Meteorological and Geophysical Data

Abstracts database (www.mganet.org).

An enhanced understanding of the evening transition of the boundary layer would be useful, even beyond the contributions to accurate global climate modeling. Predictions for the hazards of fog development (Derbyshire, 1999) and pollution dispersion (McNider et al., 1988; Singh et al., 1997; Banta et al., 1998; Pielke and Uliasz, 1998) depend on a thorough understanding of the dynamics of the stable boundary layer. McNider et al. (1988) have shown that horizontal dispersion of pollutants in the planetary boundary layer is controlled by shear that develops overnight, due to mesoscale diurnal-inertial oscillations. Banta et al. (1998) also show that inertial oscillations magnify the effect of an urban ozone plume on local air quality. When mesoscale winds are otherwise stagnant, nocturnal inertial oscillations can spread an urban ozone plume through the surrounding rural areas in all directions, instead of simply advecting it away from the source. On the next day, instead of mixing the new growing urban plume with “clean” rural air, the new plume is mixed in with the previous day’s “dirty” air. In a case where this cycle persists for three days, they find ozone levels reaching 140 ppb, where 100 ppb is considered “high”, and the Environmental Protection Agency requires that a region’s ozone levels not exceed 125 ppb on more than three days a year. Another motivation for understanding the development of inertial oscillations is that of airport management: the runway spacing for departing and arriving aircraft depends on the dispersion or drift of aircraft-induced wake vortices (Kaplan et al., 2000). A low-level jet, possibly caused by inertial oscillations, alters atmospheric dispersion rates and leads to a less predictable residence time for wake vortices. With poor knowledge of vortex dispersion, airports must allow for larger runway spacing in a time when most airports are already operating at their capacity limits. Thus, improved predictions of inertial oscillations and LLJs would be an asset in numerous fields.

Despite the need for a better understanding of inertial oscillations and the LLJ, few studies have used actual observations to examine the development of inertial oscil-

lations, aside from verifying a veering in time in a hodograph of winds. This lack of observational verification of the generation of inertial oscillations (most studies simply assume the evening transition proceeds as described by Blackadar) is partially due to a dearth of observations with adequate time and height resolution. The recent development of radar boundary-layer wind profilers has provided unprecedented temporal and vertical resolution of boundary-layer winds. This improved resolution is adequate for a quantitative study of inertial oscillations and their generation mechanisms. The Cooperative Atmosphere-Surface Exchange Study (CASES) field programs use multiple boundary-layer profilers and a wide array of other instruments for probing the atmospheric boundary layer. Therefore, the data collected there are especially well-suited for determining where and how often inertial oscillations actually occur in the fair-weather atmospheric boundary layer over land. I use the CASES data to understand the influence of inertial oscillations on the low-level jet and to examine how often the one-dimensional Blackadar mechanism actually applies to the boundary layer.

1.2 Overview of the CASES field programs

The Cooperative Atmosphere-Surface Exchange Study (CASES) site was designed to “provide a long-term facility for scientists to study the mesoscale processes of meteorology, hydrology, climate, chemistry, ecology and their complex linkages, and to serve as a focal point to provide field experience for students of the natural sciences” (Pflaum, 1995). In the Walnut River Watershed of southeastern Kansas, a foundation of observational platforms (wind profilers, stream gauges, and surface flux stations) managed by Argonne National Laboratory is periodically reinforced by more intensive concentrations of instruments for more focused studies. The CASES site is conveniently located within the Department of Energy Atmospheric Radiation Measurement (ARM) Southern Great Plains (SGP) Clouds and Radiation Testbed (CART) site. Measurements from the ARM-CART instrumentation supplement CASES observations (see Figure

1.1). The first of the CASES field projects took place (LeMone et al., 2000) from 21 April to 17 June 1997. One component of CASES-97 entailed studying the role that surface processes play in the diurnal evolution of the atmospheric boundary layer. This facet of CASES-97 involved frequent profiling of the boundary layer with radiosondes, aircraft, and boundary-layer profilers in addition to direct observation of surface fluxes at twelve sites throughout the Walnut River Watershed. The second component of CASES-97 involved the testing of radar-based precipitation algorithms. The observations from CASES-97 and their incorporation into this study are discussed in more detail in Chapter 2.

CASES-99 took place in October of 1999. This field campaign focused on surface-atmosphere exchange processes during the evening transition of the boundary layer, the duration of the stable boundary layer, and the morning transition of the boundary layer. CASES-99 sought to identify the sources of the wide range of phenomena that inhabit the nocturnal boundary layer. These phenomena include internal gravity waves, Kelvin-Helmholtz billows, inertial oscillations, shallow drainage flows and intermittent turbulent bursts (Poulos et al., 2001). One of the stated goals of CASES-99 was to assess the role of the evening transition period in the initiation of inertial oscillations and the enhancement of low-level jets 100 to 300 m above the surface.

1.3 Overview of thesis

This dissertation uses data from both CASES-97 and CASES-99 to fulfill the CASES-99 goal of exploring the generation of inertial oscillations by the evening transition. This research is divided into three components: observations of inertial oscillations, observations of the LLJ, and one-dimensional modeling of the generation of inertial oscillations. Because each component of this study is relatively independent of the other components, more specific and detailed discussions of the literature are included at the beginning of Chapters 3, 4, and 5. Before commencing the discussion of the observa-

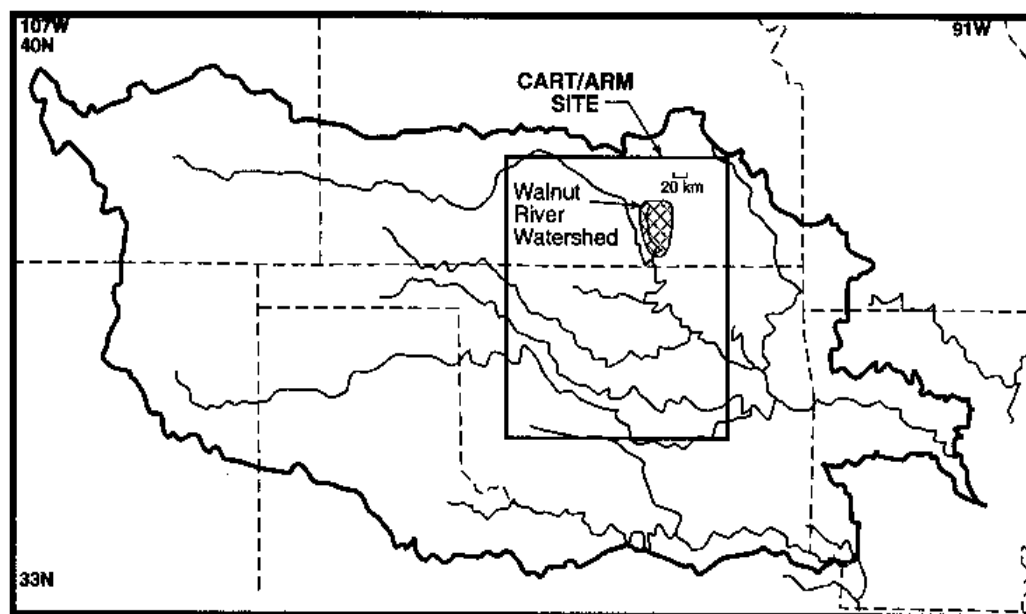


Figure 1.1: Map of the Walnut River Watershed

tions, I discuss, in Chapter 2, the instrumentation used in CASES-97 and CASES-99 to collect the data. This is the first time that the instrumentation used for collecting wind speeds and directions from rawinsondes in CASES-99 has been used for lower-boundary-layer research, and so I document the weaknesses of this platform and discuss its suitability for this type of research. I also discuss the strengths and limitations of 915 MHz boundary-layer wind profiler data, and how those pitfalls are addressed in this study.

In Chapter 3, I use those data in a study of inertial oscillations in the atmospheric boundary layer. After summarizing the literature on the importance of inertial oscillations to boundary-layer dynamics, I use spectral analysis to document qualitatively the contribution of motions at the diurnal-inertial scale to the power spectra of winds in the boundary layer. I then develop a technique to identify constant-amplitude inertial oscillations in hourly wind profiler data. I document the improvement over a technique I have used in previous investigations. The boundary-layer wind profiler data from CASES-97 and CASES-99 show that inertial oscillations occur frequently, and are, in fact, present at all points in the diurnal cycle rather than just after the evening transition of the boundary layer. To test Blackadar’s hypothesis, that inertial oscillations are generated at the evening transition due to decoupling, I focus on the nights that show inertial oscillations developing at the evening transition. I discuss the characteristics of the inertial oscillations on each night, compared with what one would expect if the Blackadar hypothesis encompassed the relevant dynamics on each night. I then conduct a case study of one night where inertial oscillations contribute to the formation of a low-level jet structure in the winds.

To consider low-level jets more broadly, I use the boundary-layer wind profiler data in a climatology of the low-level jet (Chapter 4). I apply the established climatological techniques for identifying low-level jets to boundary-layer wind profiler data. I confirm the observations of previous researchers with respect to the diurnal variability

of the jet. I combine the observations of inertial oscillations in Chapter 2 with the LLJ data to establish a correlation between inertial oscillations generated at the evening transition and low-level jets, in support of the theory proposed by Blackadar (1957).

To further explore the generation of inertial oscillations at the evening transition, whether or not they lead to LLJs, I develop a simple numerical model of the evening transition and the stable boundary layer (Chapter 5). This one-dimensional model uses first-order turbulence closure, and parameterizes turbulence as a function of the local Richardson number. For the night of CASES-99 that exhibits inertial oscillations following the Blackadar mechanism (the only night with dynamics simple enough to simulate with a one-dimensional model), the model succeeds in generating inertial oscillations with the same variation in height as those observed. I show that the variation of the eddy viscosity with height and time does not significantly influence the development of inertial oscillations on this night. Rather, the ageostrophic component of the wind at the evening transition directly predicts the amplitudes of the observed inertial oscillations. This result, that the short-time-scale variations in the decay of convective turbulence are not critical to the development of inertial oscillations, is verified with an analytical model.

The final chapter is a summary of the results and a discussion of the conclusions drawn from this investigation. Although the one-dimensional Blackadar mechanism is shown to explain the inertial oscillations observed on one night in the CASES field experiment, most other nights show the influence of other three-dimensional mechanisms, including slope flow (Holton, 1967). Additionally, the numerous cases of inertial oscillations indicate that other mechanisms are responsible for their generation at other times in the diurnal cycle.

Chapter 2

Datasets and Instrumentation

Science, like Art, is fun, a playing with truths.

W.H. Auden, “Unpredictable but Providential”,
1972

The observational component of this study (primarily Chapters 3 and 4) relies on winds measured with boundary-layer wind profilers and radiosondes during the CASES-97 and CASES-99 field experiments. The modeling component of this study (Chapter 5) utilizes these wind data, thermodynamic data from the radiosondes, and time series of surface temperature from the same field programs. This chapter discusses the CASES field experiments, the data sources, and the limitations of the dataset.

2.1 The CASES-97 and CASES-99 datasets

The first of the CASES field efforts, CASES-97, took place from 21 April to 17 June 1997. CASES-97 had two components: the first, which lasted from 21 April to 22 May 1997, focused on the diurnal evolution of the atmospheric boundary layer and the role of surface processes in that evolution. The second component involved the testing of radar precipitation algorithms.

Although the dataset collected for the first component of CASES-97 was designed to study the surface-induced heterogeneity of the atmospheric boundary layer, the extensive profiling of winds and thermodynamic variables performed for CASES-97

Table 2.1: CASES boundary-layer profiler information

Name	Lat.	Long.	Elev.	Lowest Gate (AGL)	Δz	Operation Dates (used in this study)	Inertial Period
BEA-97	37.627	- 96.538	478 m	87 m	57 m	2-31 May 1997	19.6 h
OXF-97	37.273	-97.095	360 m	310 m	75 m	2-31 May 1997	19.8 h
WHI-97	37.850	-97.187	420 m	82 m	56 m	2-31 May 1997	19.5 h
BEA-99	37.627	- 96.538	478 m	146 m	56 m	3-29 Oct 1999	19.6 h
OXF-99	37.273	-97.095	360 m	137 m	55 m	3-29 Oct 1999	19.8 h
WHI-99	37.841	-97.186	416 m	137 m	55 m	3-29 Oct 1999	19.5 h
ELD-99	37.78	-96.82	561 m	137 m	56 m	3-29 Oct 1999	19.5 h
SMI-99	37.52	-96.86	545 m	137 m	56 m	3-29 Oct 1999	19.6 h

Intensive Observing Periods (IOPs) makes this dataset suitable for inclusion in a study of the dynamics of the evening transition of the boundary layer. As Figure 2.1 shows, three 915 MHz boundary layer profilers, located at Beaumont, Oxford, and Whitewater, form a triangle of approximately 60 km to a side within the Walnut River Watershed. These profilers return hourly profiles of wind speed and direction from approximately 150 m to 2000 m. The characteristics of the data provided by the three boundary-layer wind profilers during CASES-97 are summarized in Table 2.1. Hourly consensus winds for all three profilers are utilized in this study. An extensive discussion of wind profiler data quality follows later in this chapter. During IOPs, the wind data from the profilers were supplemented with rawinsondes. Research-quality Cross-chain Loran Atmospheric Sounding System (CLASS) rawinsondes were launched, nominally every 90 minutes, from the three profiler sites. These soundings record temperature, pressure, relative humidity, wind speed, and wind direction every 10 seconds. Details on the other CASES-97 instrumentation are available in LeMone et al. (2000).

During the CASES-99 field program, the 915 MHz radar boundary-layer wind profilers at Beaumont, Whitewater, and Oxford were supplemented with two NCAR boundary-layer profilers, located at Smileyburg and El Dorado (see Figure 2.2).

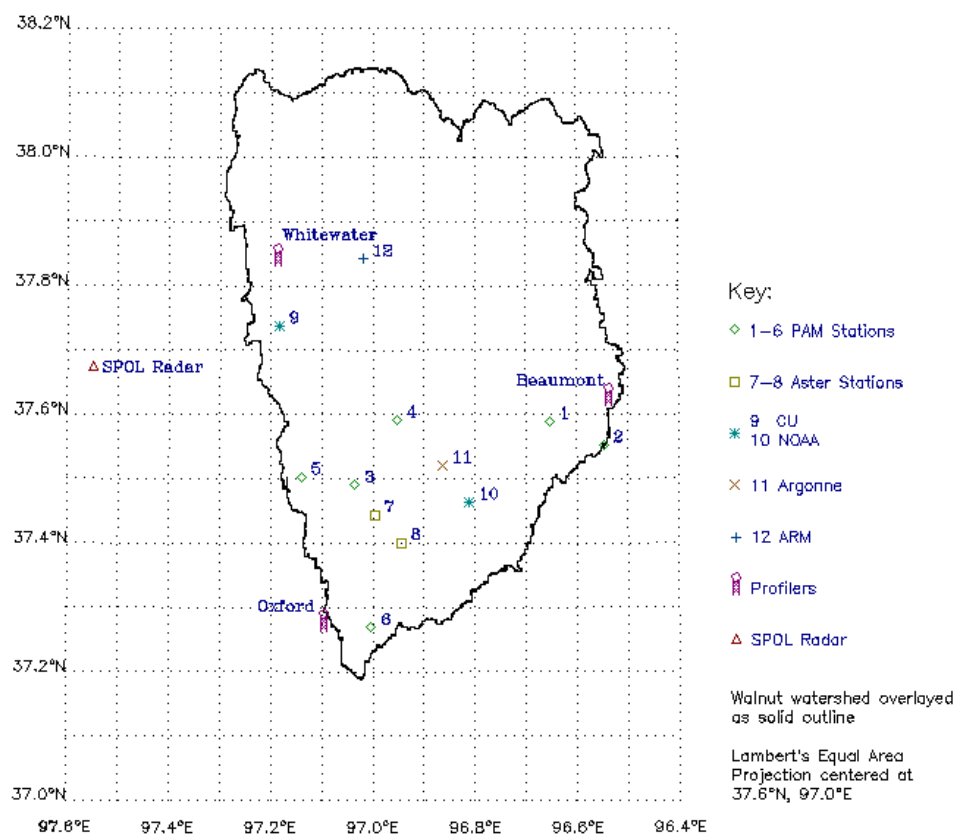


Figure 2.1: This map of the Walnut River Watershed depicts the CASES-97 instrumentation. Numbered sites represent eddy-correlation surface flux stations (the symbol identifies the sponsoring institution). Boundary-layer profilers are located at Beaumont, Oxford, and Whitewater. NCAR sounding stations are co-located with the profilers. This figure is provided by the UCAR Joint Offices for Science Support, at <http://www.joss.ucar.edu/cases/gifs/sites.gif>

Throughout the program, Global Positioning System (GPS) rawinsondes were launched from four sites within the Walnut River Watershed: Beaumont, Smileyburg, El Dorado, and at the central Leon site, co-located with the 55 m tower that anchors the CASES-99 experiment. The wind profiler and radiosonde data are used extensively in this study. For model boundary conditions, five-minute averages of temperature and pressure measured at 2 m are used (Chapter 5), as are 5-minute averages of surface temperature measured with an Everest radiometer at the site of the 55 m tower. More details are available in LeMone et al. (2000) and Poulos et al. (2001).

2.2 The suitability of new instrument platforms for boundary-layer research

The analysis discussed in this thesis relies on data collected with several in situ and remote sensing systems. Data from GPS radiosondes are being applied to boundary-layer research for the first time. In the case of the radar boundary-layer wind profilers, their utility in boundary-layer research is still being explored (Pekour and Coulter, 1999; Cohn and Angevine, 2000). As a result, this section explores the strengths and weaknesses of these instrument platforms for studies of the evening transition of the boundary layer.

2.2.1 GPS sondes

The radiosonde, the instrument package that measures pressure, ambient temperature, and relative humidity that is often attached to a weather balloon, was first used by the U.S. Weather Bureau in 1936¹. Radiosondes provide measurements of slanted or oblique profiles of the atmosphere, and are thus useful, though expensive when launched frequently. In the CASES-97 field program, radiosondes were supplemented with standard LORAN navigation equipment, allowing measurements of wind

¹ <http://www.meteor.wisc.edu/hopkins/wx-inst/wxi-raob.htm>

Plan view of CASES-99

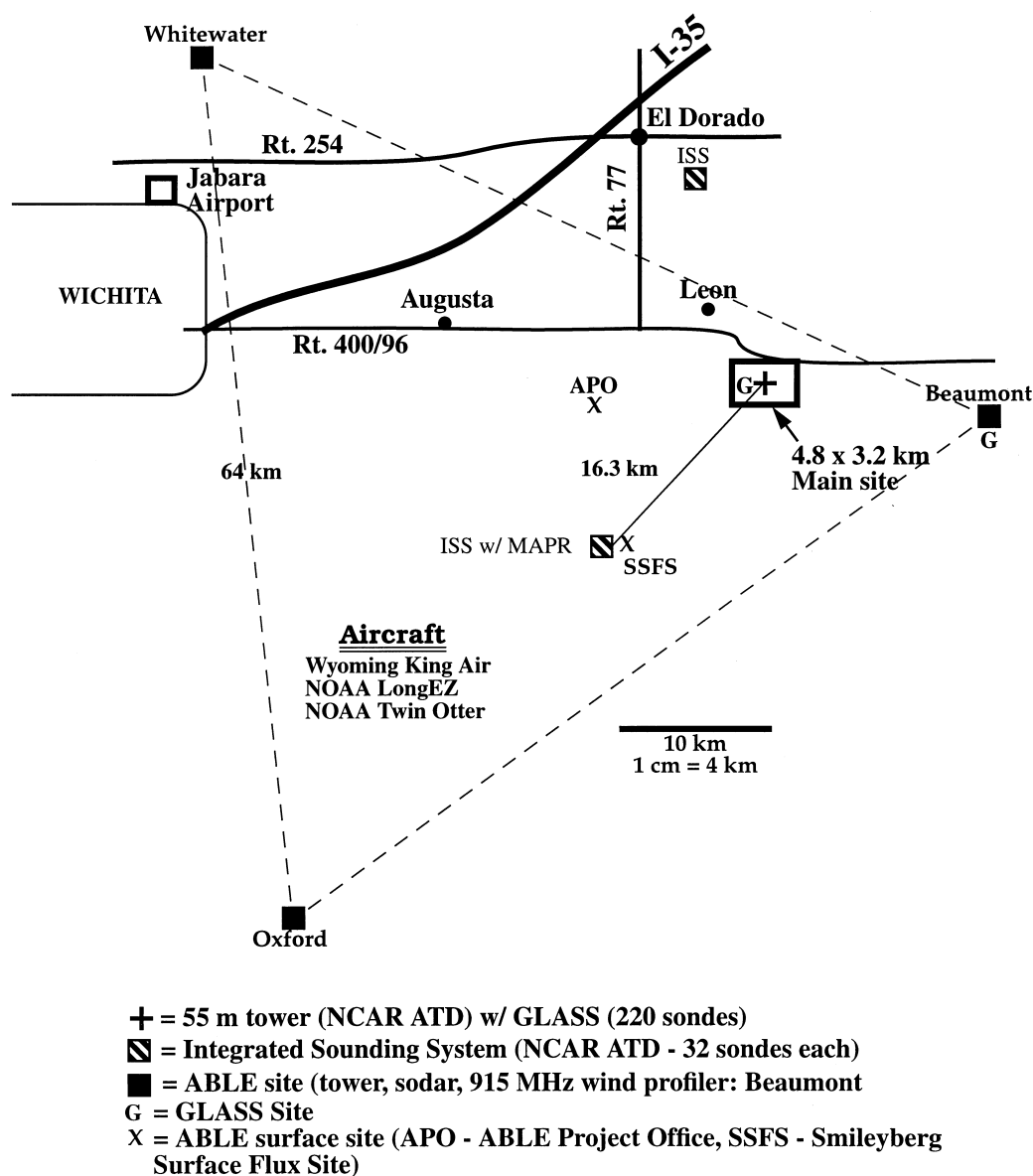


Figure 2.2: Plan view of the CASES-99 experiment. The three Argonne Boundary Layer Facility boundary-layer profilers at Beaumont, Oxford, and Whitewater are joined by two National Center for Atmospheric Research Integrated Sounding System boundary-layer profilers at El Dorado and Smileyburg (Parsons et al., 1994). Rawinsondes are launched four times a day (and more often during IOPs) from the main site, Beaumont, El Dorado, and Smileyburg. Figure courtesy of Greg Poulos, Colorado Research Associates.

speed and direction from the drift of the sonde². Though the sondes were only launched every 90 minutes, they provide useful profiles.

In 1999, the Atmospheric Technology Division of the National Center for Atmospheric Research, the group in charge of sounding operations for CASES-99, changed their radiosonde navigation system from one based on finding LORAN stations to one based on finding Global Positioning System (GPS) satellites. When using GPS instead of LORAN-C navigation data for wind finding, the sonde must maintain contact with at least 3 satellites during the flight to ensure accurate location-finding, and therefore accurate winds. (Wind speed and wind direction are measured by tracking the elevation and azimuth of the radiosonde using Navigational Aid signals from moving satellites in space (Chamberlain 2000, <http://www.atd.ucar.edu/sssf/facilities/class/class.html>). Unfortunately, the Väisälä RS 80-15GH radiosonde, and the accompanying MWG201 GPS Processor Card for processing navigation signals from the sonde, often lose contact with satellites upon release. Without contact with satellites, the wind speed and wind direction of the radiosonde cannot be determined. This loss of wind data occurs during the beginning of the flight, over the lowest 100 - 200 m of the profile, and in many cases prevents any sampling of the nocturnal LLJ structure, especially the data required to discern whether or not the jet is located atop the inversion. Figure 2.3 shows an inter-comparison between winds retrieved from the High-Resolution Doppler Lidar (HRDL) and a typical sounding. Though sonde wind speeds match the HRDL data well from 250 m on up, the height of the wind speed maximum (about 75 m, according to HRDL) cannot be determined from the sonde data. The sonde is also unable to document the shear through the jet layer. For an overview of the entire CASES-99 sounding dataset, including sondes from all four sites, see Figure 2.4, which depicts a summary of the

² According to the United States Coast Guard (<http://www.navcen.uscg.mil/policy/loran/h-book/book-1.txt>), “the name LORAN is an acronym for long-range navigation. It is a radio-navigation system using land-based radio transmitters (operated in the United States by the USCG) and receivers to allow mariners, aviators, and (more recently) those interested in terrestrial navigation to determine their position. Loran-C is the federally provided radio-navigation system for the U.S. Coastal Confluence Zone (CCZ). ... Loran-C is also approved as a supplemental air navigation system.

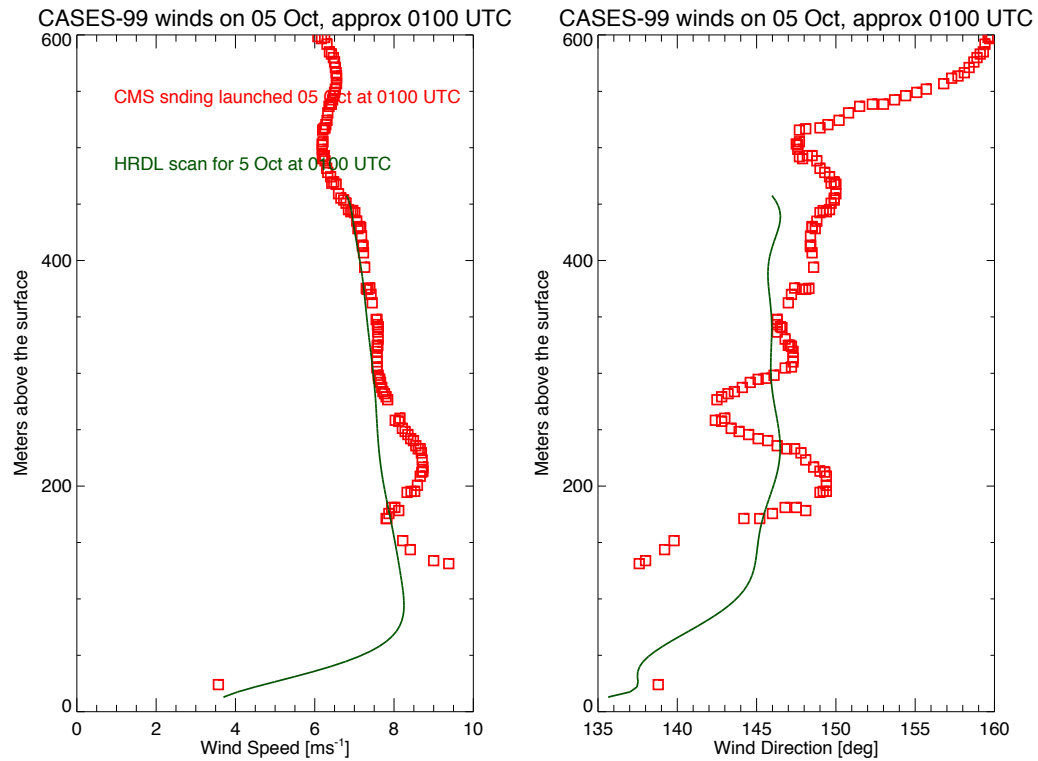


Figure 2.3: Intercomparison of GLASS and lidar winds. GLASS data (squares) are from a sounding launched at 0100 UTC 5 October at the main site, while the High-Resolution Doppler Lidar (HRDL) data (solid line) represents an average of scans from from 0045-0115 UTC 5 October.

lowest height levels available from all sondes. Note that just under half of the soundings have their lowest usable wind point above 100 m. Because contact with GPS satellites is often intermittent, the lowest usable data point may be the only usable wind data point within hundreds of meters. Thus, even soundings with one wind data point below 100 m may be useless.

If sondes rise at 5 ms^{-1} and take data every second, a well-resolved profile would have its 10th lowest point at around 50 - 75 m above the surface. Most soundings fail this test (only 30 of the 302 soundings have 10 data points below 80 m). Figure 2.5 shows the tallies of the height of the 10th lowest wind data point for each sounding. Only half the CASES-99 soundings have 10 acceptable wind data points in the lowest 140 m. Finally, Figure 2.6 shows which heights are resolved well. The abscissa shows the number of the 302 soundings with at least one data point within a given 10 m bin. The GPS data outages of the GLASS soundings inhibit research-quality resolution of the winds in the lower boundary layer.

Radiosondes do have the advantage of being able to provide simultaneous measurements of wind, pressure, ambient temperature, and relative humidity. Studies of instability within the boundary layer often depend on accurate calculations of the Richardson number:

$$\text{Ri} = \frac{g}{\theta_0} \frac{\frac{\partial \theta}{\partial z}}{\left[\left(\frac{\partial u}{\partial z} \right)^2 + \left(\frac{\partial v}{\partial z} \right)^2 \right]} \quad (2.1)$$

where g is gravity, θ_0 is an average potential temperature, $\frac{\partial \theta}{\partial z}$ is the local rate of change of potential temperature with height, and $\left(\frac{\partial u}{\partial z}^2 + \frac{\partial v}{\partial z}^2 \right)^{1/2}$ is the magnitude of the vertical shear. All these quantities can be determined from one sounding. When calculating a bulk version of this gradient Richardson number, using differences instead of differentiating, one must carefully choose the Δz over which to make the differences.

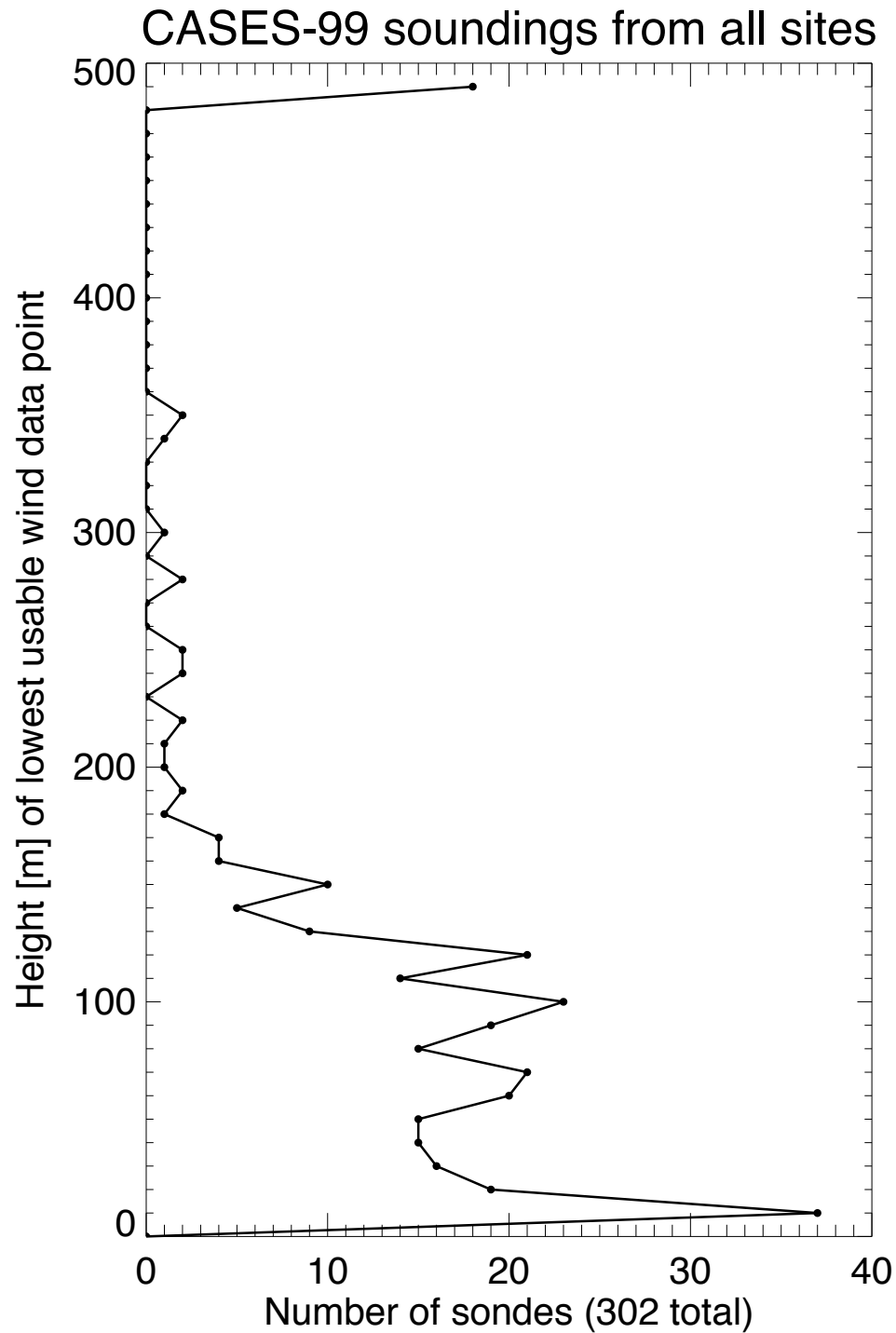


Figure 2.4: The number of soundings out of the 302 available for all of CASES-99 with their lowest usable wind data point at a given height. Note that for 18 soundings, the lowest usable wind datapoint is above 500m. Additionally, having one data point in the lowest 150 m is inadequate for research purposes if it is the only point in the lowest 50 m.

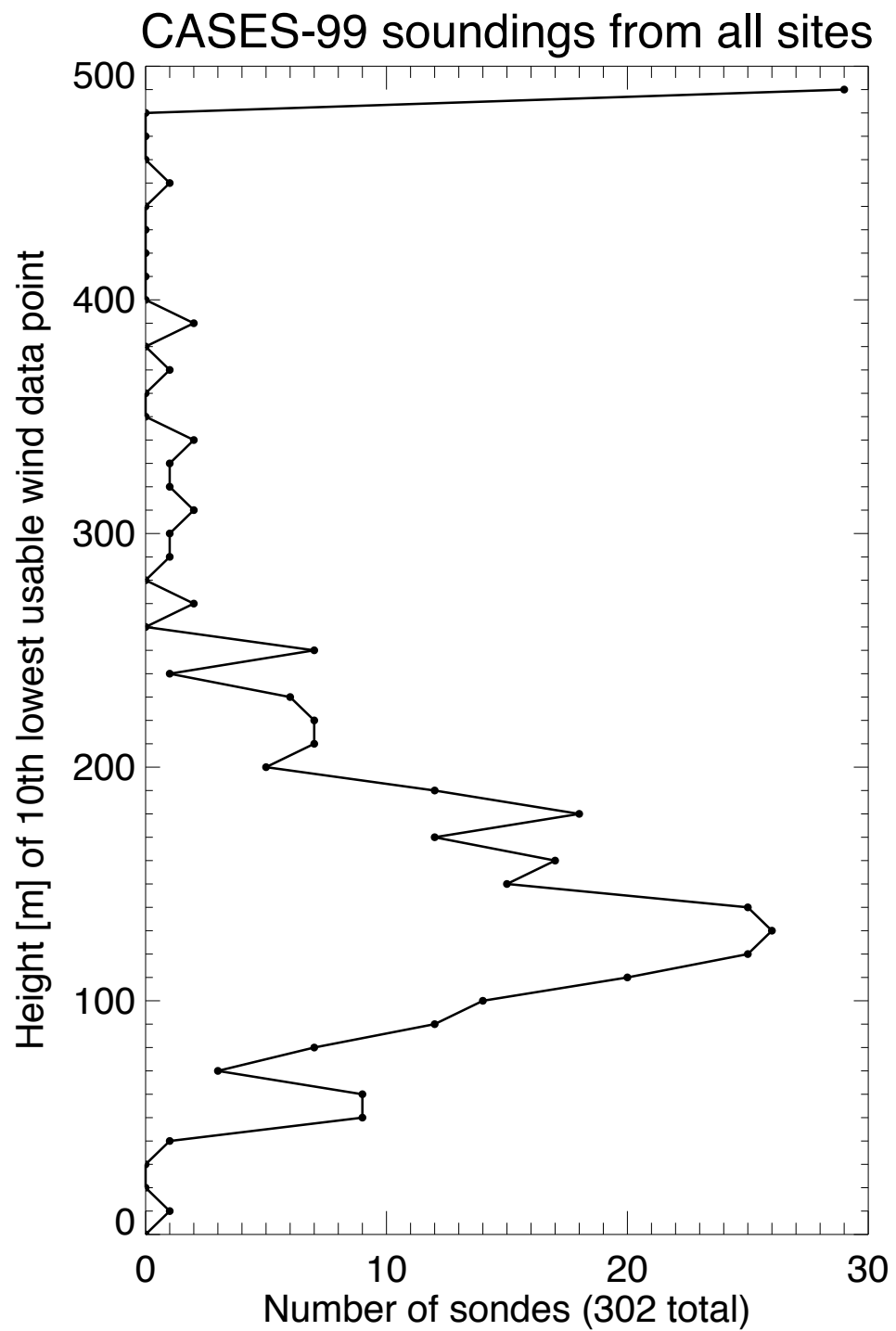


Figure 2.5: Heights of the 10th lowest data point for the 302 CASES-99 soundings. Half of the soundings have their 10th lowest point above 140 m.

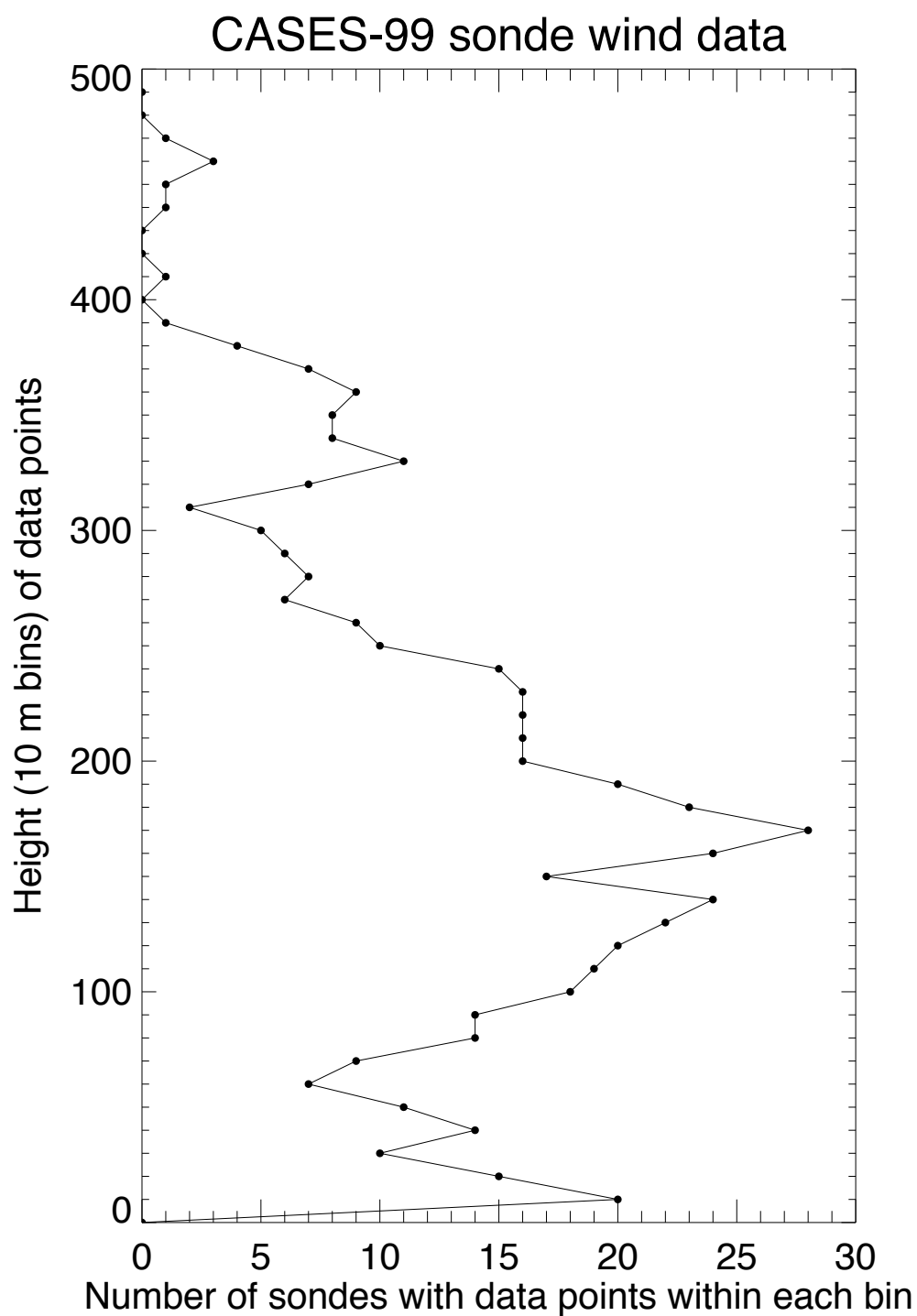


Figure 2.6: Number of the 302 CASES-99 soundings with at least one usable wind data point within each 10m height bin.

2.2.2 915 MHz boundary-layer wind profilers

Radar wind profilers offer the opportunity to improve the time resolution of wind profiles; soundings are only available when someone is available to launch the sondes. As wind profiler scans are made every six minutes and then averaged into hourly (half-hourly during CASES-99) values, they provide adequate time resolution to study the evolution of the LLJ and have in fact been used to observe other phenomena that evolve on even faster time scales (Ralph et al., 2000). Until recently, the paucity of inexpensive and widespread instrumentation was partially responsible for the dearth of observations of inertial oscillations related to the evening transition. The advent (and proliferation) of boundary-layer profilers has enabled long-term studies of the dynamics of the atmospheric boundary layer. For this study, boundary-layer profiler data from the CASES array during the CASES-97 (May 1997) and CASES-99 (October 1999) field campaigns are used. Table 2.1 contains some fundamental information about the data returned from each profiler; throughout this study, I refer to eight different profiler data sets as listed in the table.

Boundary-layer profilers, which send a radar signal at 915 MHz, operate in two modes: in the low-power mode, they scan from approximately 150 m to 2110 m in 60 m range gates. The high power mode usually extends higher, but has a higher initial range gate. The low power mode offers useful vertical resolution, but data below 150 m are generally not available. Previous studies of the Great Plains nocturnal jet have found the mean height of the LLJ wind speed maximum to be above this level. Bonner, using 22 radiosonde sounding stations over the Great Plains, finds an average jet height of 785 m with a standard deviation of 127 m (Bonner, 1968). (It should be noted that Bonner has access to data with very limited vertical resolution: surface, 150, 300, 500, 1000, 1500, 2000, 2500, and 3000 were the levels of his data.) With more modern rawinsondes and finer vertical resolution, Whiteman et al find a mean height of 596 m, with a median

height of 430 m (1997). For the boundary layer profilers to resolve the underside of the jet, the height of the jet max should be no lower than the third range gate, or 270m.

However, this study finds jet structures much lower than these previous studies. Even with the frequent low-level data outages, radiosonde data from the CASES-99 experiment shows that 20 of 25 nights have jets, or inversion wind maxima, below 270 m. The profilers cannot resolve these jets.

In addition to vertical resolution problems, wind profilers are limited in the supplementary data they can provide. Although wind profilers are often co-located with Radio Acoustic Sounding Systems (RASS) that provide profiles of virtual temperature, actual temperature and mixing ratios are unavailable. (And during the CASES-99 program, the RASS was turned off to allow for better temporal resolution in the winds.) When information on atmospheric stability is vital, then, the profilers must be complemented with an instrument which can provide such data.

The wind profiler data provided by Argonne National Laboratory has undergone quality control by that laboratory. Prior to any analysis, remaining outliers among the boundary-layer wind profiler data (caused by “ringing” or instrument noise in the profiler) are removed, and I have interpolated the approximately hourly (and sometimes half-hourly) data onto a regular hourly grid. Quality-control flags identify interpolated data points (actual data points have a flag of 1, while data points interpolated between two data points separated by time t , where t is measured in hours, has a flag of $1/t$. For the studies in Chapters 3 and 4, I set a threshold for the minimum total value of the quality control flags for a time series or a profile to ensure that the studies are based on reliable data.

2.2.2.1 Backscatter problems

The most significant drawback to the use of radar boundary-layer wind profiler data is the prospect of contamination of profiles from radar backscatter due to migrating

birds (Wilczak et al., 1995). The refractive index of the atmosphere is usually a function of moisture and temperature (Stephens, 1994), and variations in moisture and temperature caused by turbulence cause a very weak backscatter of a radar signal. Radar wind profilers measure the Doppler-shifted frequency of the backscatter from a conical volume, defined by one vertical beam and two or four offset beams. The backscattered power of the profiler signal is given by the radar equation

$$P_r = P_t \frac{\eta G^2 \lambda^2 \theta^2 h}{1024 \pi^2 \ln 2 r^2}, \quad (2.2)$$

where η is the radar reflectivity per unit volume, P_t is the transmitted peak power of the radar, G is the antenna gain, λ is the radar wavelength, r is the range from the power source to the source of the backscatter, θ is the beamwidth, and h is the pulse length or twice the range resolution (Cohn and Angevine, 2000). The radar reflectivity per unit volume, η , is a function of the refractive index structure constant C_n^2 :

$$\eta = 0.38 C_n^2 \lambda^{-1/3}, \quad (2.3)$$

where C_n^2 is a function of the radio refractive index n :

$$n = 1 + \left[\frac{776p}{T} \left(1 + \frac{7.73q}{T} \right) \right] \times 10^{-6}, \quad (2.4)$$

where T is temperature (Kelvin), p is pressure (kPa) and q is mixing ratio (g kg^{-1}).

Unfortunately, the beam backscatter sensed by the radar is not only due to refractive index irregularities due to atmospheric variability. When birds, bats, insects, airplanes, precipitation, or other unwanted targets are present in the beam, they will also scatter the radar's signal. (Additionally, ground clutter, sea clutter, precipitation, and clouds present in low-angle sidelobes of a radar can produce signals that can be confused for atmospheric return from the main beam (Merritt, 1995).) The radar is most sensitive to variations on the scale of one-half of the radar wavelength; the wavelength for the 915 MHz profiler is 33 cm. With a typical wingspan of 15 cm, songbirds are ideally suited for contamination of 915 MHz profiler winds. Wilczak et al. (1995)

show that in calm winds, only 1.2×10^{-7} birds m^{-3} are required for continuous contamination of the returned Doppler spectrum for a particular range gate (1.2×10^{-7} birds m^{-3} is roughly equivalent to one bird in a cube measuring 202 m on each side). When winds are 20 ms^{-1} , only 0.54×10^{-7} birds m^{-3} (or one bird in a cube 265 m to a side) will contaminate the signal. If the bird signal is not removed, the radar wind profilers return information on the bird speed and direction instead of the wind speed and direction.

In the CASES region, heavy bird migration at night has been documented to occur from March to June and August through November (Pekour and Coulter, 1999). Birds tend to migrate during the night, beginning at civil twilight (defined as when the sun is 6° below the horizon) and continuing throughout the night (Wilczak et al., 1995). Migrating birds can reach concentrations of over 100 000 individual birds crossing a mile-wide front each hour (Merritt, 1995), and they prefer to fly in favorable wind conditions (northerly in the autumn, southerly in the spring in the Northern Hemisphere). Therefore, winds can be systematically overestimated, especially during southerly (northerly) winds in spring (autumn). Wind direction values are affected as well because birds sometimes fly at an angle to, or even against, the wind in light wind conditions. The fact that birds do not always travel with the prevailing wind makes it very difficult to remove their effects uniformly with a consistency test or QC procedure; instead, in the presence of persistent bird echoes, these methods are liable to (and do) choose the bird signal (Pekour and Coulter, 1999). The average air speed of a bird is 8 to 15 ms^{-1} (Wilczak et al., 1995), so they can cause sizable errors in wind speed and direction observed by 915 MHz profilers.

The bird signal can sometimes be identified as the non-Gaussian portion of the backscatter. A statistical averaging method (SAM) (Merritt, 1995) is in widespread use with boundary-layer profilers. It attempts to eliminate all signals but those from atmospheric scattering and system noise. Even though all profilers used in this study

incorporated some form of the SAM in their spectral processing routine, contamination from migrating birds is still obvious when profiler winds are compared to radiosonde winds over the course of a night with heavy bird migration. In Figure 2.7, profiler data deviate from the sonde data above 1500 m. Although the profiler observes the appropriate wind direction (almost exactly northerly), wind speeds between 1500 and 2000 m are enhanced due to the bird echoes. The other profilers show similar patterns on this night except for Oxford (Figure 2.8), where the bird contamination was so pervasive that profiler winds show little resemblance to sonde winds. Instead, winds at Oxford are more northerly (following preferred bird migration routes in October) and show little variation with height.

Because birds have such a strong radar echo, nights with heavy bird contamination can be identified qualitatively by looking at variations of the signal-to-noise ratio (SNR). Large peaks in SNR can indicate the presence of birds. Figure 2.9 depicts the backscatter from the Oxford profiler on the night discussed above, where the profiler winds show little resemblance to sonde winds. Note the high values (red) of SNR in the regions where the profile from the profiler diverges from the profile from the sonde. Although methods relying on an absolute measure of SNR as a “cut-off” for identifying bird-ridden data have been proposed (Engelbart et al., 1998), these techniques tend to eliminate good data along with the bad (Coulter, personal communication). For this study, all the nights under close consideration during both CASES-97 and CASES-99 have been visually inspected to ensure that bird contamination does not dominate the wind profile returns in the region where inertial oscillations have been identified.

The SNR profiles of the nights that show extensive inertial oscillation (IO) activity during CASES-97 and CASES-99 are available at <http://paradox.colorado.edu/~jkl/profilersnr.html>. Even if some small amount of bird echoes is present, and survives the SAM quality control, it is not expected that bird contamination could lead to false positive identifications of inertial oscillations.

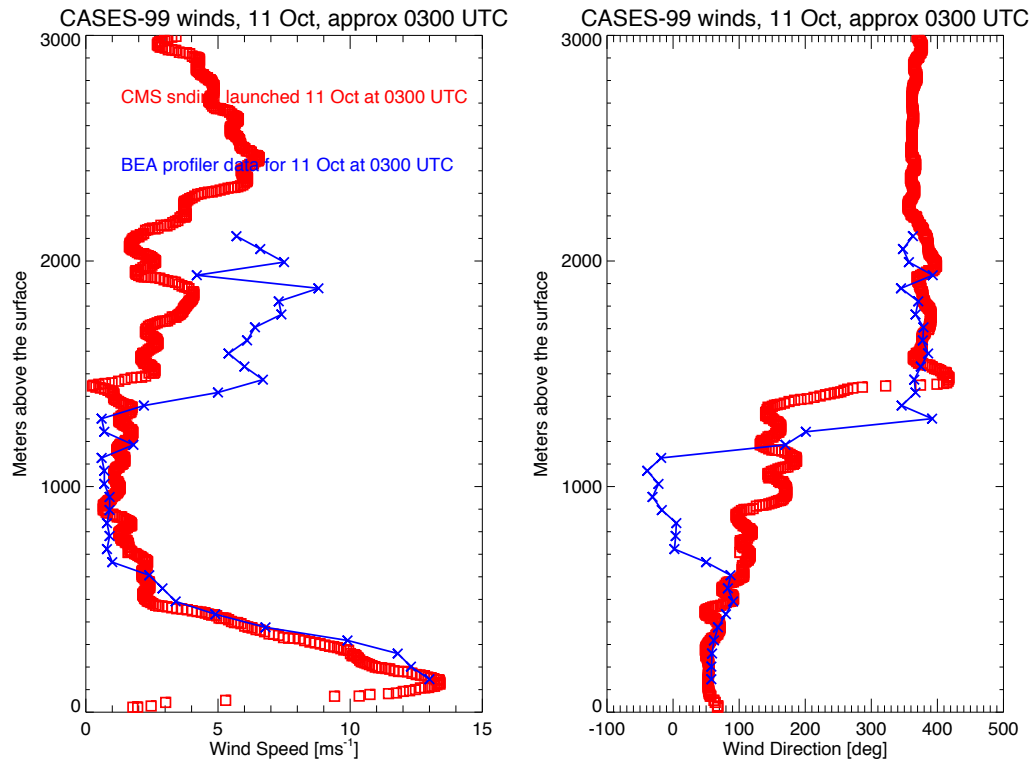


Figure 2.7: Intercomparison of GLASS winds (squares) with winds returned from the boundary-layer profiler at Beaumont (line with x). The wind direction has been modified for continuity following van Doorn et al. (2000). (The variation in wind direction between 700 and 1000 m is not important as wind speeds are near 0 at these levels.)

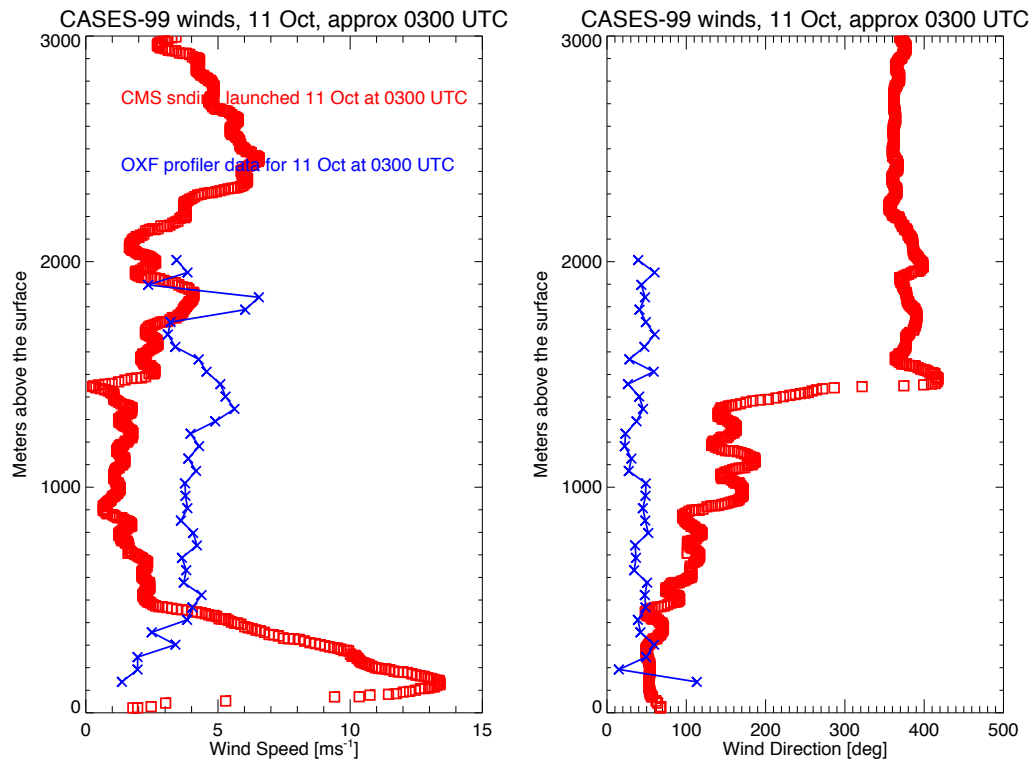


Figure 2.8: Intercomparison of GLASS winds (squares) with winds returned from the boundary-layer profiler at Oxford (line with x). The wind direction has been modified for continuity following van Doorn et al. (2000). The profiler backscatter from this period shows strong evidence of bird contamination.

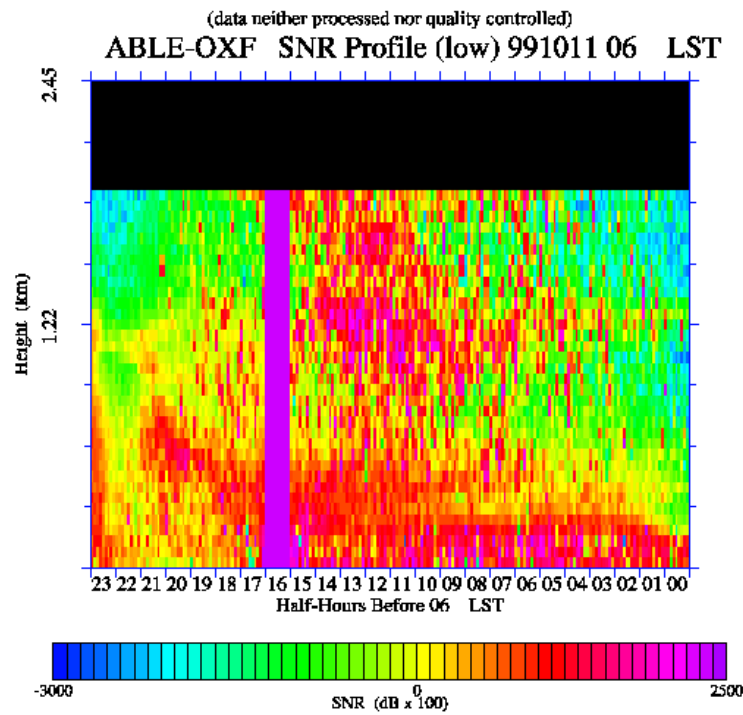


Figure 2.9: Backscatter from the Oxford profiler on the night of 11 October. Red indicates likely bird echoes. Time increases from left to right and is denoted in half-hour increments, with 0000 UTC 11 Oct at the left and 1200 UTC at the right. (LST is 6 hours behind UTC time.) The purple region denotes a data outage. 0300 UTC 11 October corresponds to “18 half-hours before 06 LST”.

As Chapter 3 discusses, the identification of inertial oscillations is dependent on the rate of change of wind speed relative to the ground. Because birds do not rotate their migration route with respect to the ground over the course of a night, their presence would only affect the mean or noise components of the wind. Therefore, the presence of birds in a signal could cause a false negative in the inertial oscillation test, but not a false positive. This possibility is discussed further in the chapter on inertial oscillation identification.

Chapter 3

Identifying inertial oscillations in boundary-layer data

So soft upon the Scene
The Act of Evening fell
We felt how neighborly a Thing
Was the Invisible.

Emily Dickinson # 1278

Blackadar (1957) and Buajitti and Blackadar (1957) highlight the importance of inertial oscillations in the diurnal cycle of the fair weather atmospheric boundary layer over land, proposing that inertial oscillations are responsible for the oft-observed nocturnal jet profile. During the day, the sun heats the surface, triggering strong surface-based convection that increases turbulent stresses throughout the boundary layer. As the sun moves lower in the sky and finally sets, the heat flux at the surface changes sign. The surface cools relative to the air above it, and surface-based convection ceases. A surface-based temperature inversion develops, decoupling layers aloft from layers closer to the surface. According to Blackadar, the winds above this inversion can accelerate due to the decrease in turbulent stresses. Within the reference frame fixed to the Earth, this acceleration will appear to rotate the wind vector relative to the surface. The rotation of the wind vector is known as the inertial oscillation, and has a period of $2\pi/f$, where f is the Coriolis parameter, defined as $2\Omega\sin\phi$, Ω is the rotation rate of the Earth ($7.292 \times 10^{-5} \text{ sec}^{-1}$) and ϕ is latitude. At a latitude of 45° , the inertial oscillation has a period of 16.9 hours. According to Blackadar, the amplitude of this

inertial oscillation is determined by the ageostrophic component of the wind during the day. At levels where the geostrophic wind is strongly constrained by friction or turbulent stresses in the afternoon, the inertial oscillation would have a large amplitude. At other levels, where the winds are almost geostrophic during the day, the nocturnal inertial oscillation would have a small amplitude.

Eventually, the inertially oscillating component of the wind aloft will align with the geostrophic wind, while winds in the layer below the inversion will still be constrained by surface friction (Thorpe and Guymer, 1977). Blackadar proposes that because the amplitudes of inertial oscillations are largest just above the surface inversion, a jet profile in the atmosphere will form, with a maximum just above the inversion. Though other mechanisms have been postulated for the low-level jet (LLJ), including cooling over sloped terrain (Holton, 1967) and synoptic-scale pressure gradients (Uccellini, 1980; Djurić and Ladwig, 1983), a careful modeling study (Zhong et al., 1996) concludes that for one observed jet over the Great Plains of the United States, the effect of the inertial oscillation dominates any effect from baroclinicity due to sloping terrain. Despite the relevance of inertial oscillations to the LLJ, and though inertial oscillations are thought to be ubiquitous in the stable boundary layer, no quantitative climatology documents the presence of inertial oscillations in the stable boundary layer. Additionally, no observation-based verification of their generation during the evening transition has been completed. Quantitative observational studies (as opposed to qualitative observations of veering in a wind hodograph) of inertial oscillations have focused on their occurrences in the vicinity of fronts, rather than their connection to diurnal atmospheric boundary layer processes (Mori, 1990; Ostdiek and Blumen, 1997).

This chapter discusses techniques for identifying inertial oscillations in a time series of winds at one height. After discussing the limitations of applying spectral analysis to wind profiler data, I will discuss a new technique for extracting inertial oscillations from a time series of wind data. I will summarize the results of that technique

as applied to boundary-layer radar wind profiler data from the CASES-97 and CASES-99 experiments.

3.1 Spectral analysis of wind profiler data

Motions at the inertial scale strongly affect boundary layer processes. (Throughout this discussion, all references to “inertial” do not refer to the inertial subrange, but rather to inertial oscillations or other motions with periods on the order of $2\pi/f$, where f is the Coriolis parameter; the inertial period at the central CASES site is 19.5 hours.) Banta et al. (1998) show that nocturnal accelerations like inertial oscillations dominate the evolution of an urban ozone plume. When mesoscale winds are otherwise stagnant, nocturnal inertial oscillations can spread an urban ozone plume through the surrounding rural areas. On the next day, instead of mixing the new growing urban plume with “clean” rural air, the new plume is mixed in with the previous day’s “dirty” air, enhancing the pollution concentration in that region. Other studies (Gupta et al., 1997; Singh et al., 1997, 1993; McNider et al., 1988) verify that inertial oscillations must be considered when predicting plume dispersion in the stable boundary layer.

Gupta et al. (1997) have enlisted spectral analysis to show that motions at the Coriolis frequency contribute significantly to the variance of winds in the atmospheric boundary layer. They analyze power spectra from boundary layer profilers in Tennessee during summer 1995, as part of the 1995 Southern Oxidants Study (1997). In their data, the diurnal-inertial peak is strongest at the lowest observational level (120 m) and weakens with height, whereas the synoptic scale peak is weak at lower levels and strengthens with height. This result, consistent for three of the five profilers, confirms the importance of inertial oscillations to the energy budget of boundary layer winds (Gupta et al., 1997). They conclude that realistic modeling of plume dispersion requires the inclusion of this diurnal-inertial peak in the power spectrum. This peak includes power contributed by inertial oscillations occurring during the observational period.

I apply their method to confirm the significance of the diurnal-inertial peak in the power spectrum of CASES-97 and CASES-99 wind profiler data. The eight gridded and interpolated hourly wind profiler datasets used throughout this thesis (see Table 2.1) briefly are examined here with Gupta’s spectral analysis technique. By using both spring and fall data, I can explore seasonal variation in the significance of the diurnal-inertial peak.

The nominally hourly wind data are first interpolated onto a regular hourly grid. Missing data points are interpolated in time. These data points are data which have been excluded due to bird contamination or “ringing” in the profiler, as discussed in Chapter 2. The gridded anomaly time series (the mean for each height has been removed) at each height is detrended before the Hamming window FFT is applied to the time series. Figures 3.1 - 3.5 for Oxford spring (CASES-97), Oxford fall (CASES-99), Whitewater spring (CASES-97), Whitewater fall (CASES-99), and El Dorado fall (CASES-99) show the non-normalized power spectrum S multiplied by frequency f , plotted against frequency, for every height returned by the profilers. (Figures for Beaumont and Smileyburg are included in the Appendix). A constant has been added to each height to illustrate the vertical variation in the spectra. A dotted vertical line marks the inertial frequency at each station while the diurnal frequency is at the period of 1 day.

The most obvious feature in these spectra is the strength in the mesoscale or synoptic peak at 3-7 days. This feature dominates the meridional (v) spectra, and is much stronger in fall (Figures 3.2, 3.4, 3.5) than in spring (Figures 3.1 and 3.3). In comparison to this large synoptic peak (caused by the passage of strong and distinct pressure and frontal systems in the fall), the diurnal-inertial peak is more subtle. In the zonal wind, a peak between the inertial and the diurnal frequencies is obvious at all profilers in both seasons. This diurnal-inertial peak is usually confined to the lowest 1500 m, implying a source within the boundary layer. In fact, the boundary layer never

grew past 1700 m during either CASES-97 or CASES-99. As McNider et al. (1988) and Banta et al. (1998) have shown, the energy represented by this diurnal-inertial peak can modify smoke plume dispersion within the nocturnal boundary layer. Others (Blumen, 2000; Blumen and Lundquist, 2000) have shown that nonlinear interactions between motions at this diurnal-inertial frequency will modify frontogenesis.

Spectral analysis can only provide suggestive evidence for the presence of inertial oscillations generated at the evening transition; as a technique, it is limited in two ways. First, at mid-latitudes, the inertial period is very close to the diurnal period. At Smileyburg, the central point between the three Argonne boundary-layer profilers, the inertial period is 19.6 hr. Therefore, it is difficult to distinguish inertial oscillations from diurnal oscillations (Gupta et al. refer to an “inertial-diurnal” peak because of this problem). We would expect a strong diurnal peak even at the equator, where no inertial oscillations are possible. A more discriminating method of investigating the inertial influence on this peak would be useful. Second, spectral analysis relies on the assumption that the data contain an entire period of the frequency of interest. However, if Blackadar’s (1957) theory of the generation of inertial oscillations by the evening decrease of turbulent stresses in the stable boundary layer applies, then the growth of the convective boundary layer in mid-morning and the ensuing increase in turbulent stresses would suppress the signal of the inertial oscillation. An inertial oscillation could only be observed during the stable boundary layer regime, which lasts about 15 hours during spring and fall at mid-latitudes. Therefore, we could not expect any spectral analysis technique to prove useful in the study of inertial oscillations in an atmospheric boundary layer with a strong diurnal cycle.

3.2 A new technique for identifying inertial oscillations

Because inertial oscillations in the atmospheric boundary layer are not expected to persist for an entire period, spectral analysis has not proven unambiguously useful

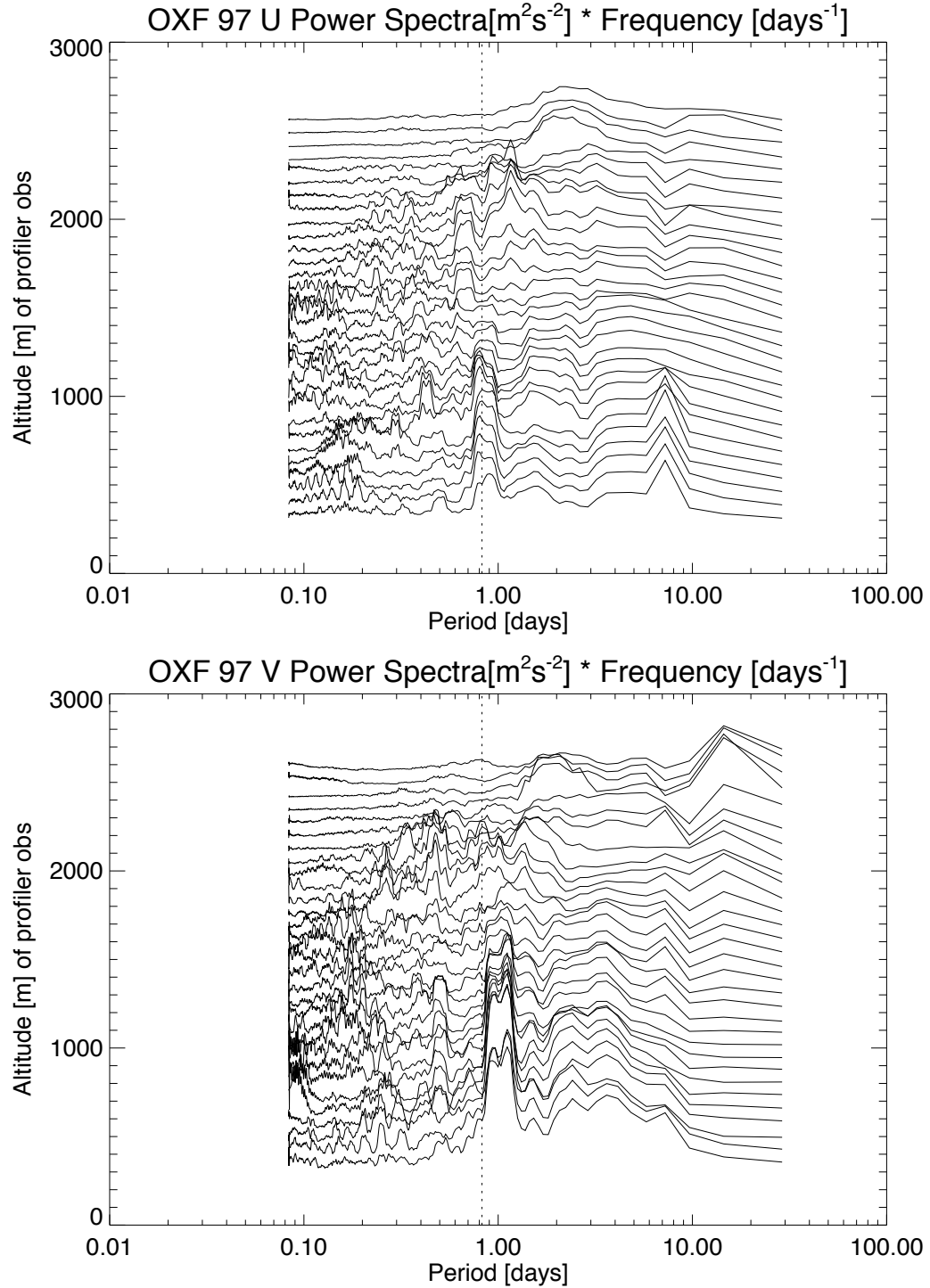


Figure 3.1: Power spectra $[\text{m}^2\text{s}^{-2}]$ times frequency $[\text{days}^{-1}]$ plotted against frequency $[\text{days}^{-1}]$ for each level of data returned from the Oxford wind profiler in May 1997. A constant has been added to the spectrum for each height to illustrate the vertical variation in the spectra. The vertical dotted line marks the inertial period at this station.

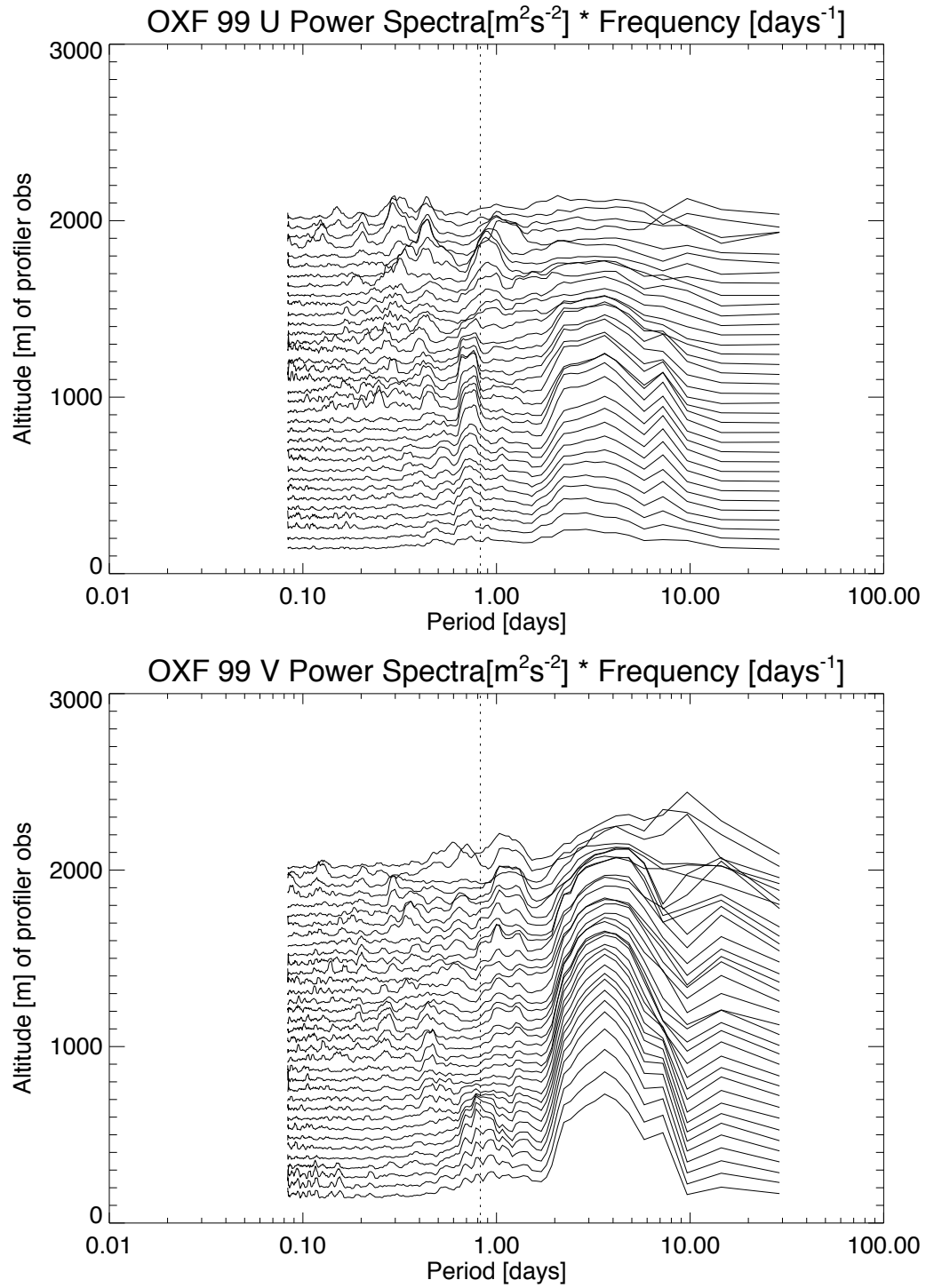


Figure 3.2: As in Figure 3.1 but for Oxford, October 1999.

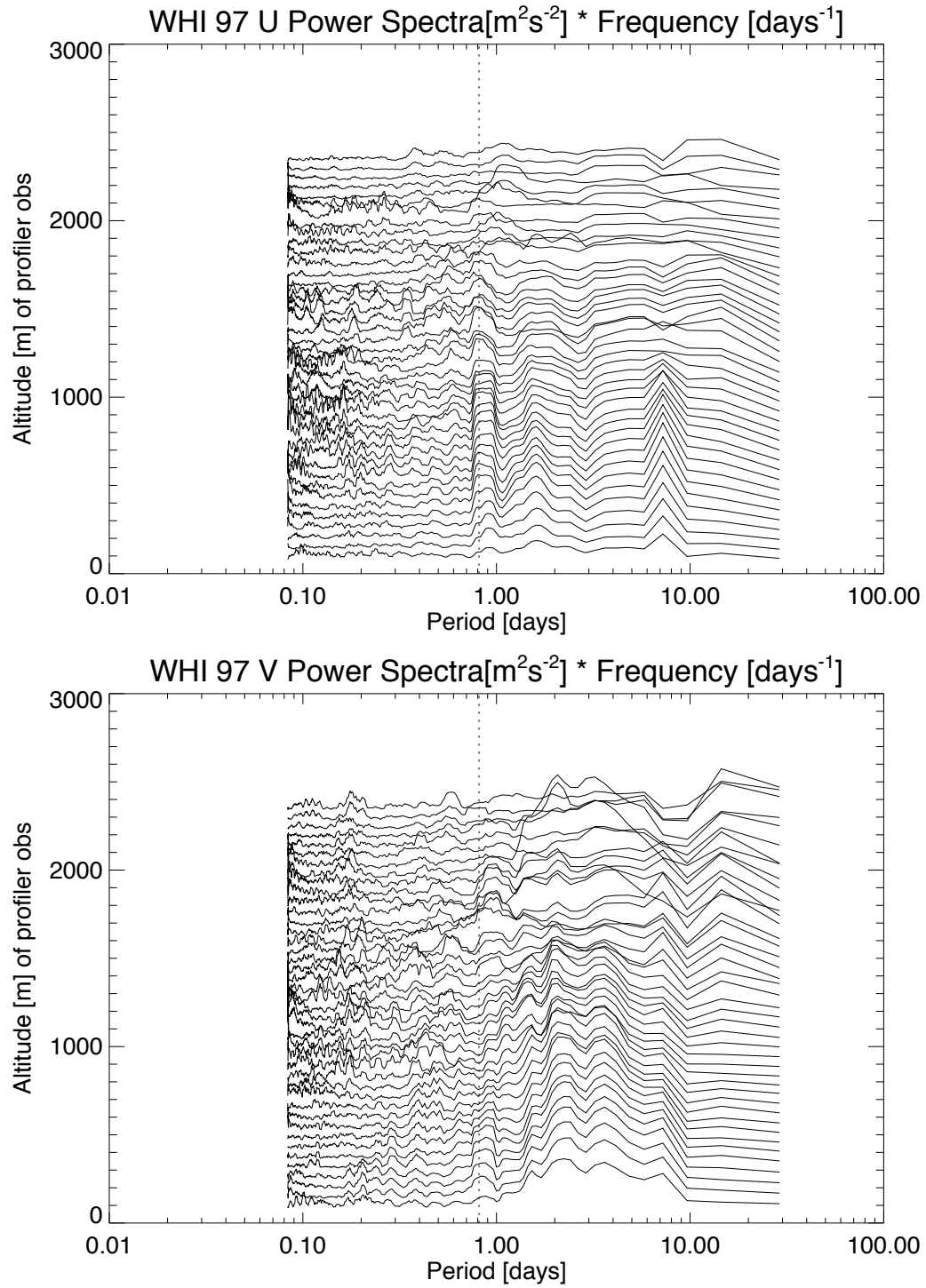


Figure 3.3: As in Figure 3.1 but for Whitewater, May 1997.

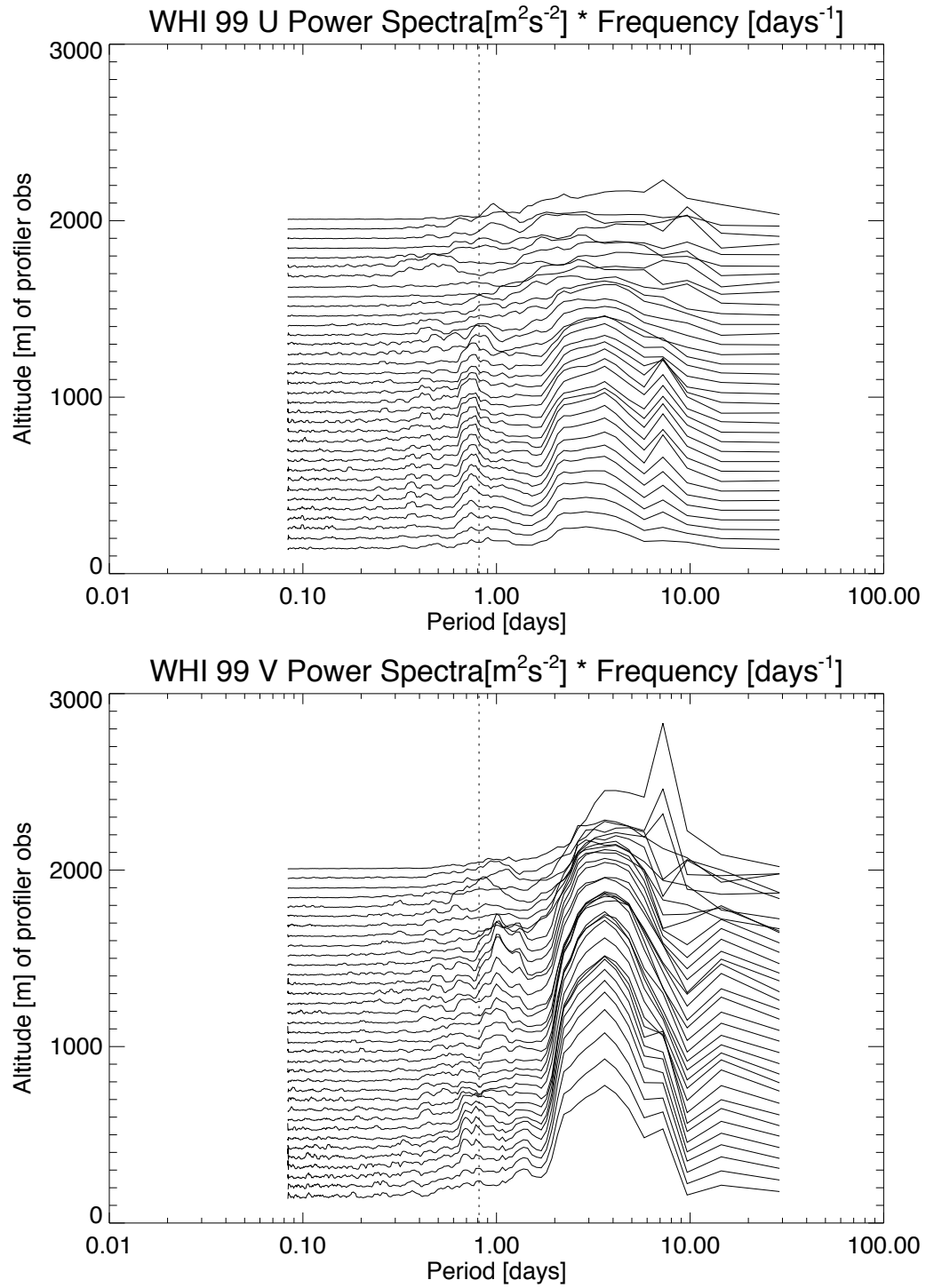


Figure 3.4: As in Figure 3.1 but for Whitewater, October 1999.

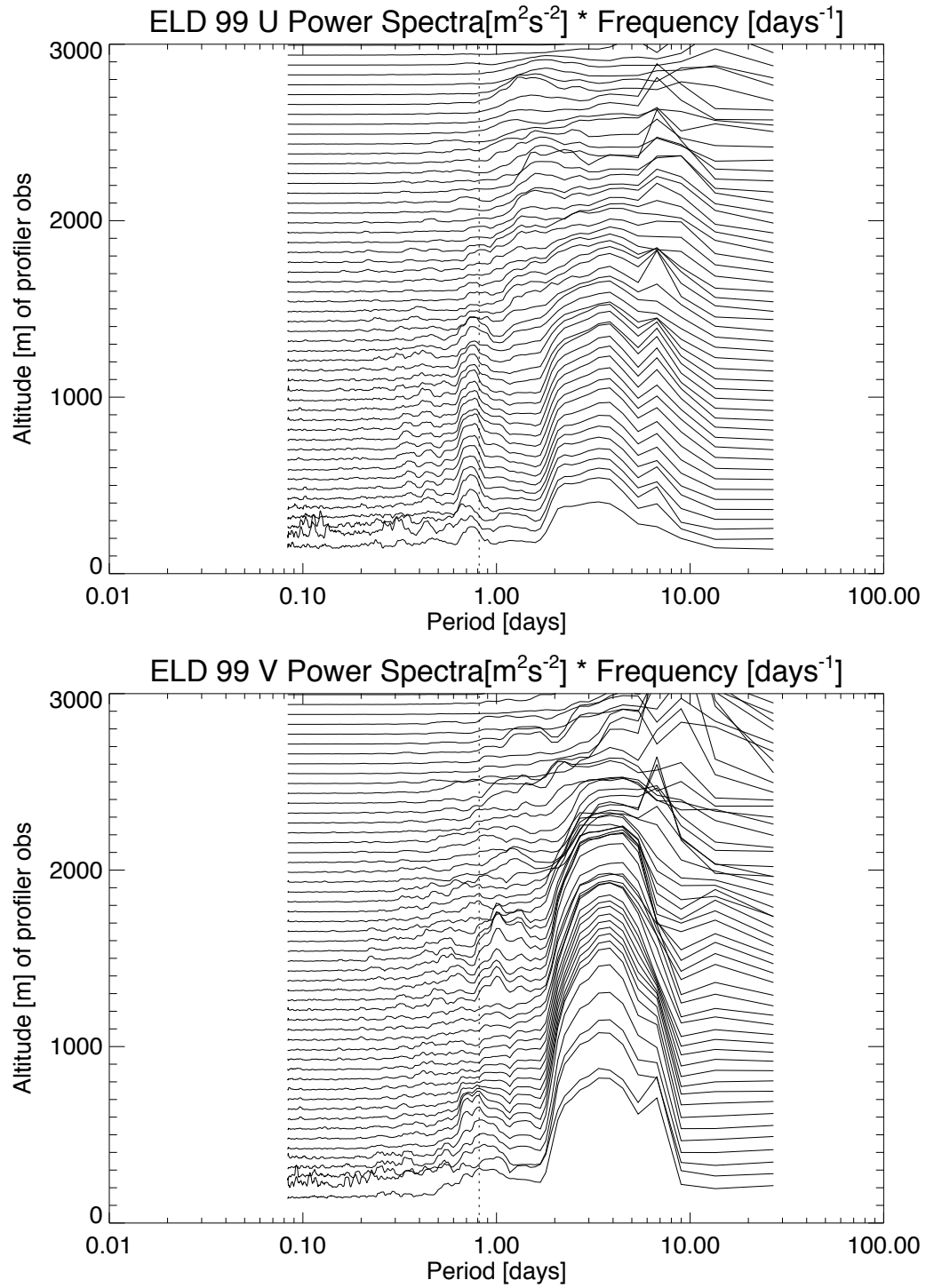


Figure 3.5: As in Figure 3.1 but for El Dorado, October 1999.

in identifying inertial oscillations. Despite the wide acceptance of Blackadar’s theory that inertial oscillations are generated at the evening transition, the actual generation of inertial oscillations has not been systematically documented. In studies of inertial oscillations using time series shorter than the period of the oscillation, a 4-parameter curve-fit technique (4PCF) has been applied to identify inertial oscillations. (Ostdiek and Blumen, 1997; Lundquist and Blumen, 1998b,a, 1999). In this section, I discuss the 4PCF technique and its limitations, and then describe the new technique that I develop for identifying inertial oscillations in a time series of wind profiler data. This new technique is based on considering the rotation of the wind direction directly rather than minimizing the error in an iterative curve fit, as is done with the 4PCF method. I then document the improvement over the old method. I show that this technique is free from the numerical uncertainties of the 4PCF method: the constant-amplitude inertial oscillations identified with this technique more closely approximate observations than those found with the 4PCF technique.

To determine the presence of an inertial oscillation, the 4PCF technique assumes that if an inertial oscillation is present, four parameters within the inertial oscillation function can be modified to fit the data. The zonal and meridional winds at each level are assumed to evolve according to

$$\begin{aligned} u(z_i, t_j) &= u_{ss}(z_i) + A(z_i) \sin(ft_j + \phi(z_i)) \\ v(z_i, t_j) &= v_{ss}(z_i) + A(z_i) \cos(ft_j + \phi(z_i)) \end{aligned} \tag{3.1}$$

where z_i and t_j are the heights and times, respectively, at which observations are taken, $u_{ss}(z_i)$ and $v_{ss}(z_i)$ are the height-dependent “steady-state” winds, and $A(z_i)$ and $\phi(z_i)$ are the height-dependent amplitude and phase, respectively, of the inertial oscillation. Consider a time series u_{ij} and v_{ij} winds at height i and times j , where $j = 0, 1, \dots, J$. To find the best-fit inertial oscillation for this time series of winds using the 4PCF method, values of $u_{ss}(z_i)$, $v_{ss}(z_i)$, $A(z_i)$, and $\phi(z_i)$ are varied to minimize

the combine root-mean-square (RMS) error between the observed winds u_{ij} and v_{ij} and those given in (3.1):

$$\text{RMS} = \sqrt{\frac{1}{2J} \sum_{j=0}^J \left((u_{ij} - u(z_i, t_j))^2 + (v_{ij} - v(z_i, t_j))^2 \right)}. \quad (3.2)$$

An initial guess value for each parameter must be supplied. The RMS error of the fit function is calculated with the initial guesses for each parameter. Each parameter value is then nudged slightly. The RMS error is recalculated, and the parameters are nudged again; when a local minimum in the RMS (3.2) is reached, the “best-fit” parameters are recorded. Iterative techniques such as this one can be unreliable. Their success hinges on close initial guesses for the parameters. Otherwise, the local minimum in RMS that is found through iterations of nudging is not actually the global minimum of RMS (3.2).

To avoid these numerical uncertainties, I develop a technique for finding the best-fit inertial oscillations in a time series of wind data, based on the rotation of the wind direction vector. The identifying characteristic of the constant-amplitude inertial oscillation is the directional rotation of the wind vector. In the Northern (Southern) Hemisphere, this rotation takes place in the clockwise (counter-clockwise) direction. The inertial oscillation is often superimposed on a much stronger mean wind, so that the total zonal (u_{TOT}) and meridional (v_{TOT}) winds at a certain level can be represented by

$$\begin{aligned} u_{TOT} &= u_{IO} + u_{SS}, \\ v_{TOT} &= v_{IO} + v_{SS} \end{aligned} \quad (3.3)$$

where the steady-state components u_{SS} and v_{SS} need not actually be steady, but only evolve slowly compared to the inertially oscillating terms u_{IO} and v_{IO} . Equation (3.3) is analogous to (3.1) above.

Equation (3.3) can be applied to a time series of winds $u(z_i, t)$ and $v(z_i, t)$ observed at height z_i from an initial time $t = 0$ to a final time $t = T$ to find the best-fit inertial

oscillation at that height z_i . (From now on, references to z_i will be dropped and it may be assumed that all variables refer to the same observation level; each height level is treated independently.) First, the steady-state components of the winds are calculated and removed from the time series. If the length of the time series were equal to the inertial period, then the steady-state components would simply be the statistical means of the time series, $\overline{u(t)}$ and $\overline{v(t)}$. However, because we often have less than an entire inertial period (P hours) of data, the steady-state component is calculated as the mean of a collection of pairs of data points:

$$\begin{aligned}
 t = 0 \quad \text{and} \quad t = P/2, \\
 t = 1 \quad \text{and} \quad t = P/2 + 1, \\
 t = 2 \quad \text{and} \quad t = P/2 + 2, \dots \\
 t = n \quad \text{and} \quad t = P/2 + n
 \end{aligned} \tag{3.4}$$

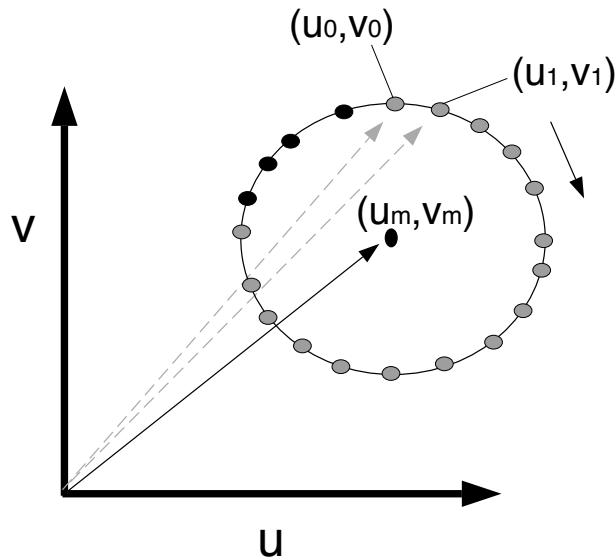
until $P/2 + n \geq T$ where T is the length of the time series. Figure 3.6 provides a graphical explanation of this method. The steady-state components are calculated as the mean of these pairs of data points. The steady-state components are then removed from the raw data to obtain the anomalies or fluctuating components u_{REM} and v_{REM} :

$$u_{REM}(t) = u(t) - u_{SS}, \tag{3.5}$$

$$v_{REM}(t) = v(t) - v_{SS}. \tag{3.6}$$

The remaining wind is converted into wind speed $M(t)$ and direction $\theta(t)$. Because wind direction is a periodic variable, artificial and sudden changes of wind direction from 0° to 360° (and vice versa) are often apparent in the data (van Doorn et al., 2000). To avoid analysis problems introduced by this discontinuity, we follow the suggestion of van Doorn et al. (2000) and calculate a new variable for wind direction, the time step change in wind direction:

$$\Delta\theta(t) = \theta(t) - \theta(t - 1). \tag{3.7}$$



This figure depicts a schematic of an inertial oscillation superimposed on a mean wind, (u_m, v_m) . The observed wind vector (gray dashed lines) rotates around through the inertial oscillation, starting from (u_0, v_0) and turning to the right.

If we are observing a nocturnal inertial oscillation that has a period (here, 20 hours) greater than the period of observation (here, 16 hours), only some fraction of the inertial oscillation is actually observed (the gray points).

Therefore, to calculate the mean wind (u_m, v_m) about which the inertial oscillation rotates, only an average of certain points must be considered: points 0-5 and 10-15.

Including 6-9 would offset the mean value about which the winds were inertially oscillating.

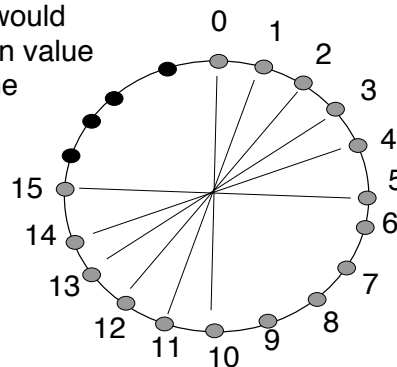


Figure 3.6: Schematic for finding the steady-state components of wind time series.

We recognize that all values of $|\Delta\theta(t)| \geq 330^\circ$ are due to the circular nature of wind direction, and replace them with

$$\Delta\theta_c = \Delta\theta - \frac{\Delta\theta}{|\Delta\theta|} 360^\circ. \quad (3.8)$$

A new time series of wind direction is calculated from the “corrected” $\Delta\theta_c$ so that the time series of wind direction is continuous.

A “best-fit” inertial oscillation is then calculated for the time series of wind speed and wind direction. If the time series $u_{REM}(t)$ and $v_{REM}(t)$ represented a perfect inertial oscillation, $\theta(t)$ would increase at a constant rate given by the Coriolis parameter $f = 2\Omega \sin \phi$. This rate can be converted into degrees, rather than radians, easily:

$$\left(\frac{d\theta}{dt}\right)_{IO} = \frac{360^\circ}{2\pi/f} = \frac{360^\circ \Omega \sin \phi}{\pi}, \quad (3.9)$$

where Ω is the rotation rate of the earth and ϕ is the latitude of the observer. (The period of the best-fit inertial oscillation would be given by $2\pi/f$.) A best-fit initial wind direction $\theta_{IO}(0)$ is calculated using

$$\theta_{IO}(0) = \overline{\theta_{REM}(t) - t \left(\frac{d\theta}{dt}\right)_{IO}}. \quad (3.10)$$

A time series of ideal wind directions is calculated as:

$$\theta_{IO}(t) = \frac{d\theta}{dt}t + \theta_{IO}(0). \quad (3.11)$$

The time series defined by (3.11) is the time series of the wind direction of the inertial oscillation that best fits the data. The amplitude of the best-fit oscillation can be determined from the wind speeds $M(t)$:

$$\text{AMP}_{IO} = \overline{M(t)}. \quad (3.12)$$

Once the best-fit inertial oscillation (described by (3.11) and (3.12)) is calculated from the observations, the quality of the fit is determined by the residual signal remaining after the best-fit inertial oscillation is removed. The measure of the fit is defined as

the ratio of the mean of the residual (the original wind speed minus the mean and the inertial oscillation components) to the amplitude of the oscillation

$$R_{ws} = \frac{\overline{M_{REM-IO}}}{AMP_{IO}}. \quad (3.13)$$

The time series is considered an inertial oscillation if

$$R_{ws} \leq 0.5. \quad (3.14)$$

This criterion (3.14) is based on extensive examinations of best-fit inertial oscillations, as discussed below. Figure 3.7 depicts one time series that fulfills the criterion to be considered an inertial oscillation; R_{ws} is 0.32 for this case. This technique can be applied to time series of varying durations and commencement times.

To prove the suitability of this “ R_{ws} test” criterion, we examine the probability density function (PDF) of the signal remaining after the inertial oscillation is removed to ensure that the remaining signal is actually just noise. Previous research (van Doorn et al., 2000) has shown that the PDF of fluctuations in the time rate of change wind direction can indicate whether or not all meaningful signals have been removed from a time series. If the PDF is Gaussian, then the fluctuations consist of noise. If the PDF is non-Gaussian, then some remaining signal is present. If the PDF has a Gaussian shape but is not centered about zero, then a linear trend is present in the time rate of change of the wind direction. Inspection of the PDFs of the time rate of change of wind direction for those time series that have been identified as inertial oscillations show that, once the mean and the inertial oscillation have been removed, the fluctuation time series is close to Gaussian and therefore is noise. Figure 3.8 show the PDFs of the time rate of change of wind direction for one particular time series. When the mean wind is removed, the mostly-Gaussian signal is still centered around a positive non-zero value, indicating that a positive linear trend in wind direction influences the signal. After the inertial oscillation is removed, the PDF is close to centered around zero, indicating that the inertial oscillation was mostly responsible for the deviation from Gaussian.

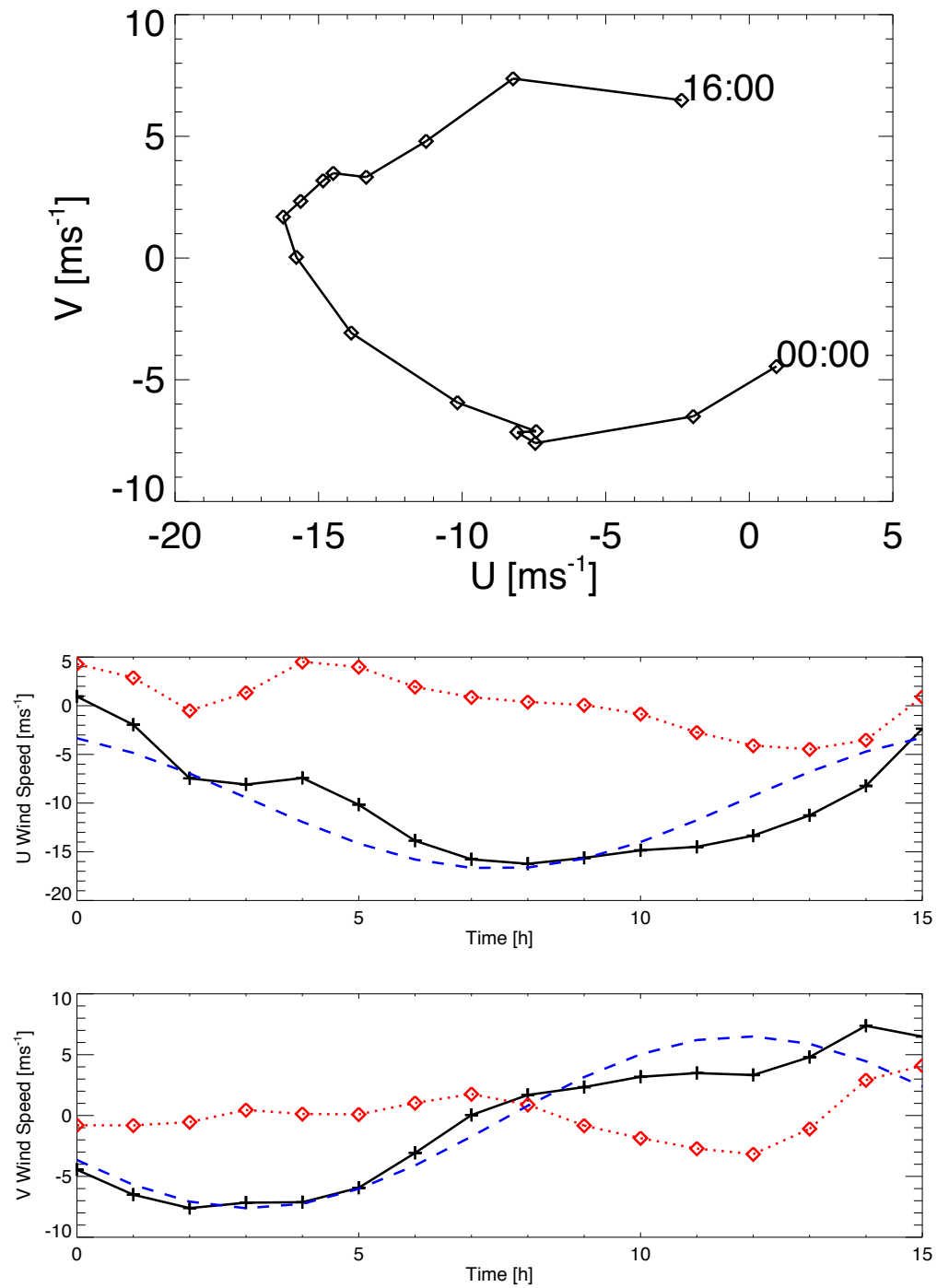


Figure 3.7: The top panel is the hodograph of the winds observed from 0000 UTC to 1600 UTC 11 October at El Dorado at 361 m. The bottom two panels show time series of the zonal and meridional winds. The solid line represents the observed data; the dashed line is the “best-fit” inertial oscillation; the dotted line is the noise remaining after the mean winds and the inertial oscillation have been removed from the time series

To further support the utility of this technique, Figure 3.9 shows characteristics of inertial oscillations identified with this technique on one particular night. Only inertial oscillations between the surface and 600 m pass the R_{ws} test (3.14). Within this region, the basic characteristics of the inertial oscillations (especially initial wind direction and mean wind direction) are well-defined and consistent between the 5 profilers that provide the data. Further aloft, the amplitude of noise is much larger, the mean wind direction varies widely for the best-fit inertial oscillations, and none of the best-fit inertial oscillations pass the R_{ws} test (3.14).

Finally, I note that the inertial oscillations identified with technique are an improvement over those identified with the 4PCF technique, used in previous studies (Ostiek and Blumen, 1997; Lundquist and Blumen, 1998b,a, 1999). Figure 3.10 depicts a time series of data and the inertial oscillations fit to the data with both this technique and the 4PCF technique. The RMS error of this technique, 0.97 ms^{-1} , calculated according to (3.2), shows an improvement over the previous technique, which has an RMS error of 1.11 ms^{-1} .

3.3 Observed inertial oscillations

To explore the generation and persistence of inertial oscillations, the new curve-fit method, described in the previous section, is applied to boundary-layer profiler data from May 1997 (coincident with the CASES-97 field program) and October 1999 (coincident with the CASES-99 field program) to extract best-fit inertial oscillations. This data analysis tests the validity of Blackadar's hypotheses, that 1) inertial oscillations are generated in the late afternoon by the decay of convective turbulence aloft, and 2) constant-amplitude inertial oscillations persist throughout the lifetime of the stable boundary layer until the growth of the convective boundary layer the next morning.

The R_{ws} test (3.14) is applied to the best-fit inertial oscillations from a time series of wind profiler data at each height z_i returned by the profiler, for

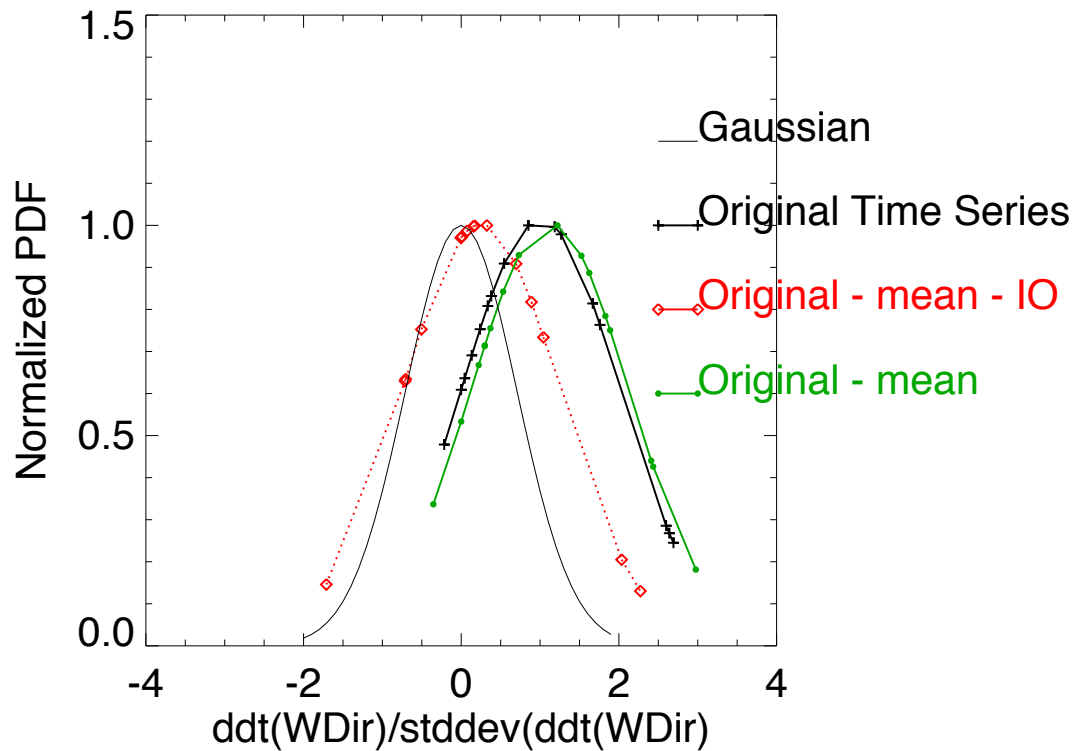


Figure 3.8: The probability density function (PDF) for the time rate of change of wind direction for the 137 m El Dorado profiler from 0000 UTC 11 October to 1600 UTC 11 October. A perfectly Gaussian curve centered about zero indicates that only noise remains in the signal. The time rate of change of wind direction PDFs for both the original time series (solid line and + symbol) and the original time series with the mean removed (solid line and small circles) are both offset from zero, indicating that a constant positive linear trend in time rate of change of wind direction is present. Once the inertial oscillation has been removed (dotted line with diamonds), the curve is almost Gaussian and centered about zero, indicating that the fluctuations are mostly noise and that no other linear trend is present.

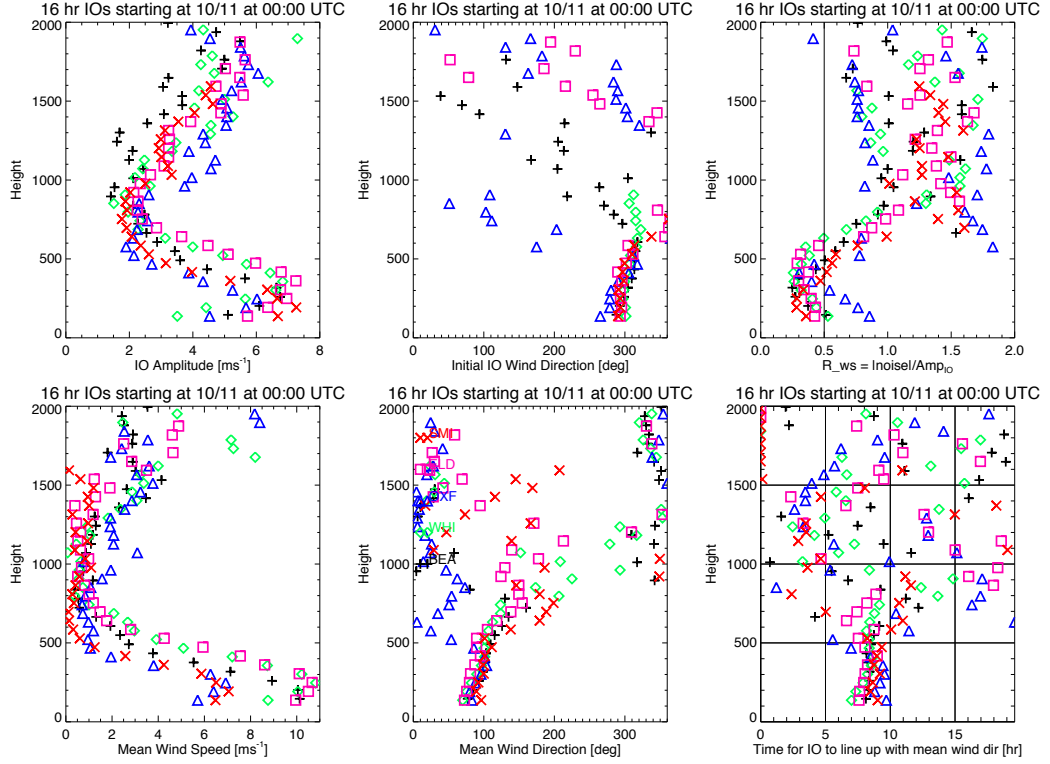


Figure 3.9: Characteristics of inertial oscillations starting at 0000 UT 11 Oct. Each symbol (+,diamond,triangle,square,x) corresponds to one profiler station (Beaumont, Whitewater, Oxford, El Dorado, Smileyburg). From left to right, the figures depict (a) the amplitude of the best-fit inertial oscillations, (b) the initial wind direction of those inertial oscillations, (c) the R_{ws} value, (d) the mean wind speed, (e) the mean wind direction, and (f) time in hours that it will take the inertial oscillations to align with the mean wind direction.

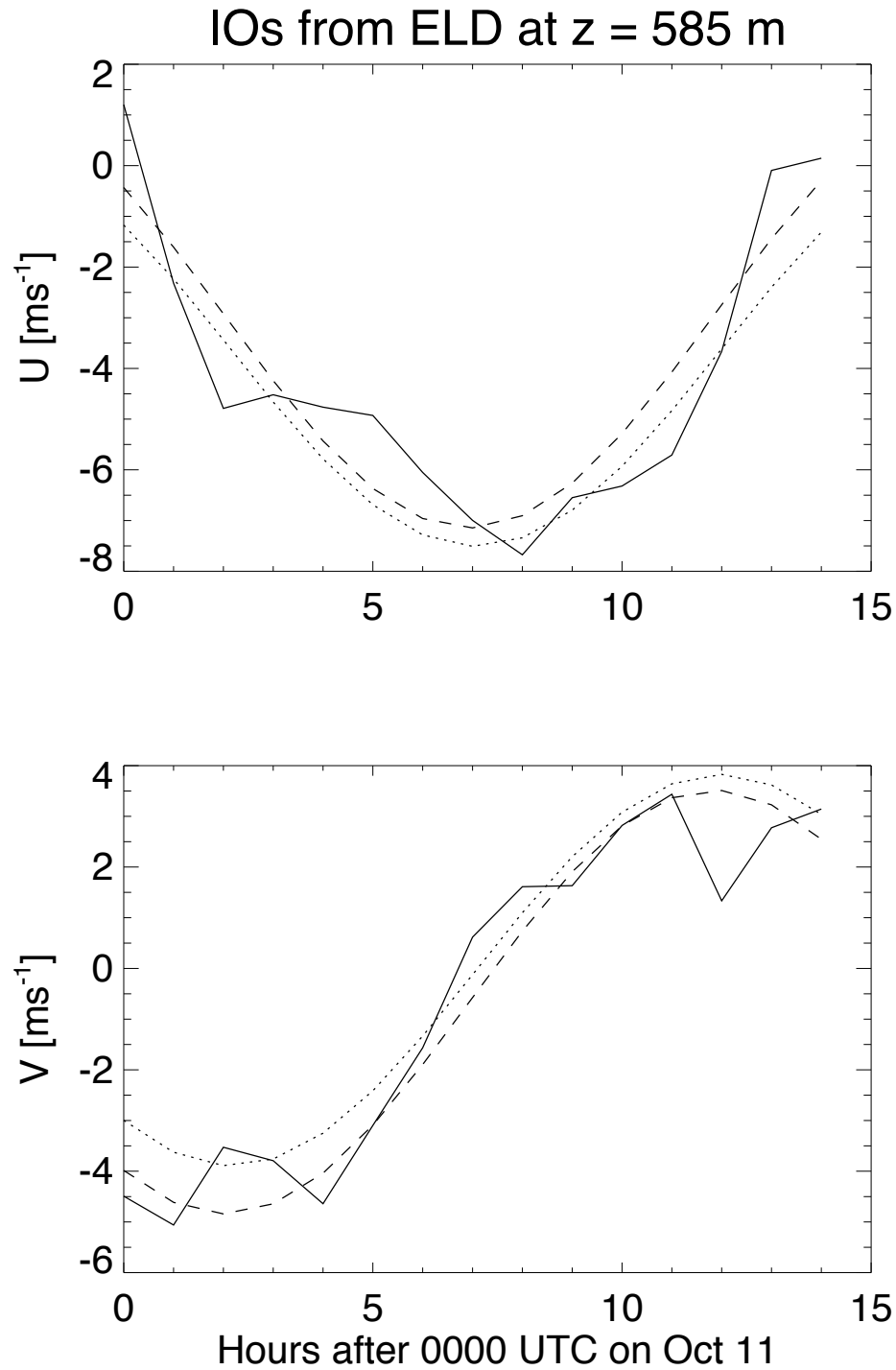


Figure 3.10: Data and fit inertial oscillations for 14 hours of wind data from the 585 m level of the El Dorado profiler, starting at 0000 UTC 11 October. The observations are represented by the solid line. The dotted line represents the old 4-parameter curve fit (4PCF) technique, which has an RMS error of 1.11 ms^{-1} . The dashed line shows the new technique, which has an RMS error of 0.97 ms^{-1} . For this case, $R_{ws} = 0.30$.

varying durations d_k ($k = 10, 11, 12, \dots, 23$ hours) and initial times s_l ($l = 0000, 0100, 0200, 0300, \dots, 2300$ UTC). I vary d_k to avoid presuming that an inertial oscillation lasts for the lifetime of the stable boundary layer. The initial times are varied sequentially throughout the month under consideration. If more than half of the data points in a particular time series are interpolated, then that time series is discarded from consideration.

According to Blackadar’s hypothesis, inertial oscillations are expected to commence at times close to when the surface flux changes sign, typically in the late afternoon. The inertial oscillations generated by the decrease in turbulent stresses are expected to last approximately 14 hours, the lifetime of the stable boundary layer. Figure 3.11, from Blackadar’s original work, shows the expected structure of inertial oscillations above the surface inversion. On the left, an “initial” wind profile is shown, typical of a classical convective boundary layer wind profile as it was envisioned in 1957. Winds gradually increase with height, so that aloft, winds approach the geostrophic limit. The large turbulent stresses near the surface, due both to convective turbulence and the drag of the surface, keep the winds significantly subgeostrophic. The wind profile on the right shows Blackadar’s proposal for a profile half an inertial period after the surface inversion has developed. Above the inversion, where turbulent stresses have essentially disappeared, inertially oscillating winds are supergeostrophic by the same amount they were subgeostrophic earlier: the amplitudes of the inertial oscillations are largest closest to the surface, where winds were most subgeostrophic during the day, creating a jet profile. Additionally, the amplitude of evening-decoupling-generated inertial oscillations is expected to be larger at levels closer to the surface inversion (or low-level wind speed maximum) and smaller at higher levels (Blackadar, 1957; Thorpe and Guymer, 1977; Singh et al., 1993), as illustrated in Figure 3.11. This structure in inertial oscillation amplitude applies only when the geostrophic wind shows no variation with height.

When the geostrophic wind varies with height, the ageostrophic component of the

wind, which provides the acceleration for the inertial oscillations, will also vary with height. The variation with the geostrophic wind with height can be caused by horizontal temperature variations, as seen in the thermal wind equations

$$\frac{\partial u_g}{\partial z} = -\frac{g}{fT} \frac{\partial T}{\partial y} \text{ and} \quad (3.15)$$

$$\frac{\partial v_g}{\partial z} = +\frac{g}{fT} \frac{\partial T}{\partial x}, \quad (3.16)$$

where u_g and v_g are the zonal and meridional components of the geostrophic wind, f is the Coriolis parameter, and T is temperature. If the geostrophic wind is southerly at the surface and warm air is to the west in the Northern Hemisphere, the geostrophic wind will decrease with height. If the geostrophic wind is northerly at the surface and warm air is again to the west in the Northern Hemisphere, the geostrophic wind will become more negative with height, or increase in magnitude. This variation of the geostrophic wind with height will affect the amplitudes of the generated inertial oscillations and the nocturnal wind profile. If the geostrophic wind increases with height, as shown in Figure 3.12, the wind profile will not necessarily show a jet structure atop the inversion, though the amplitudes of the inertial oscillations will increase with height. When the geostrophic wind decreases with height, a jet structure will form and the amplitudes of the inertial oscillations will decrease with height. Unfortunately, data are not available to define $\partial V_g / \partial z$, so one can only guess at the vertical variation of the geostrophic wind, estimating from the amplitudes of the inertial oscillations.

3.3.1 Inertial oscillation commencement times

Figures 3.13 and 3.14 show the initiation times for 14-hour time series which exhibit inertial oscillations at all heights observed during CASES-97 and CASES-99 for all stations summed and for the individual stations. (The duration of 14 hours was selected for closer inspection as that duration provided the most cases of inertial oscillations.) The onset of inertial oscillations has a strong diurnal variation, shown at

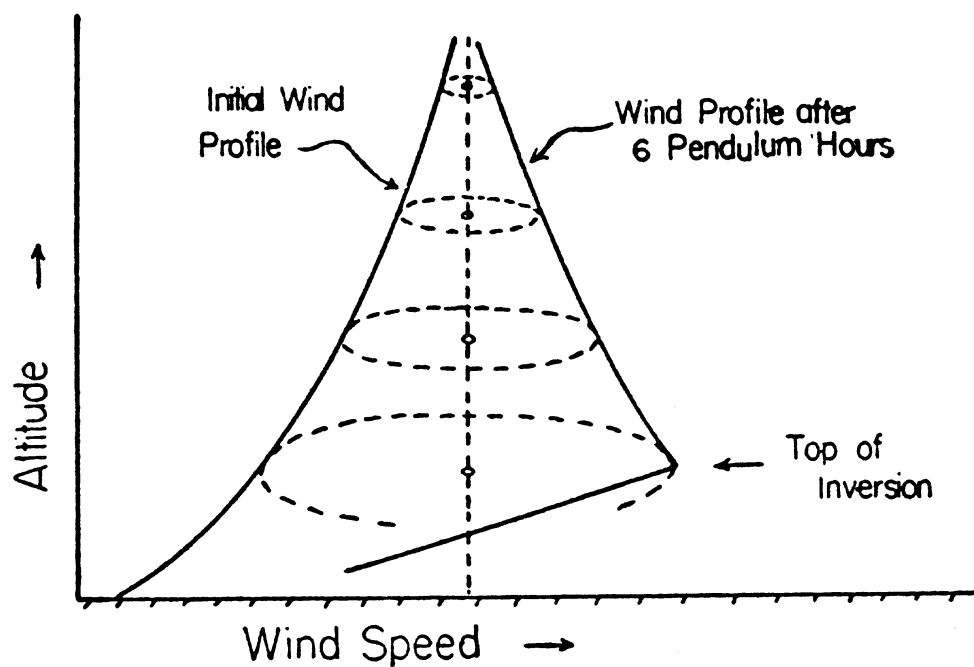


FIG. 10. Schematic illustration explaining the evolution of a boundary layer jet profile.

Figure 3.11: Blackadar's expected structure of inertial oscillations and the wind profile after half an inertial period when the geostrophic wind is constant with respect to height and time.

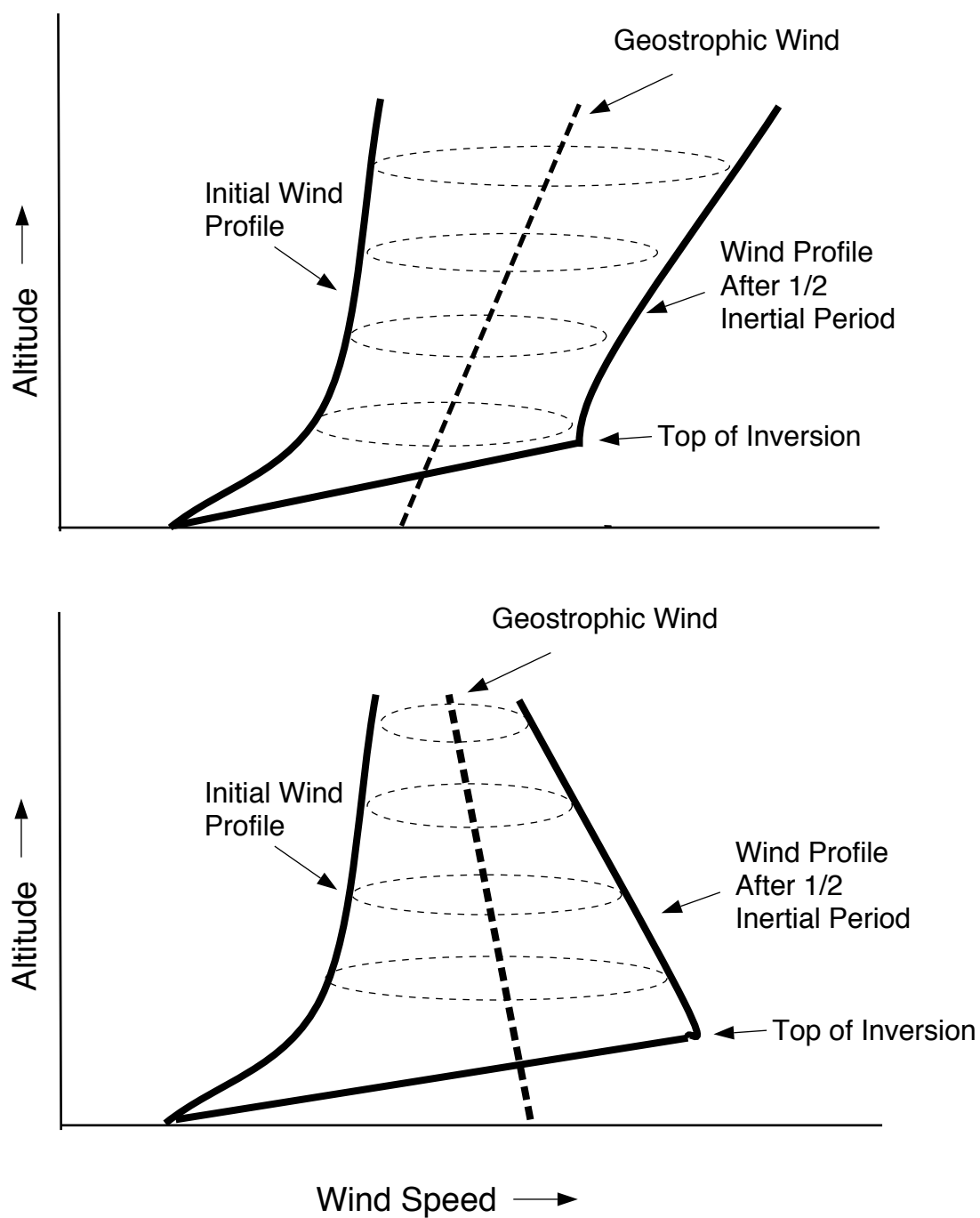


Figure 3.12: The structure of inertial oscillation amplitudes and the wind profiles after half an inertial period for cases when the geostrophic wind (dashed line) increases with height (top panel) and when the geostrophic wind (dashed line) decreases with height (bottom panel). After Blackadar's Figure 10 (Figure 3.11 in this work) from the 1957 paper.

all stations during both seasons. Inertial oscillations tend to begin between 2000 and 0300 UTC (1400 and 2100 LST); this peak is quite broad. Compared to this peak, half as many inertial oscillations tend to begin between 0700 UTC and 1600 UTC (0100 and 1000 LST), during the decay of the stable boundary layer. In the spring data, inertial oscillations appear to commence earlier than in the fall; the late afternoon peak is higher for the CASES-97 dataset. On 15 May 1997, surface heat flux changes sign at 0000 UTC, while on 15 October 1999, surface heat flux changes sign at 2300 UTC. More clouds were present during CASES-97 than in CASES-99, perhaps affecting the strength of convection in the late afternoon and allowing inertial oscillations to commence earlier.

To test Blackadar’s hypothesis more directly, I examine the characteristics of inertial oscillations on those nights with evidence of oscillations lasting from 0000 UTC (or 1800 LST, one hour after surface heat flux changes sign) to 1400 UTC (0800 LST). Of the 27 evenings examined from the CASES-97 dataset, eight nights show strong inertial oscillations commencing at 0000 UTC. These cases are summarized in Table 3.1. The CASES-99 data reveal similar behavior: eight of 27 nights show evidence of inertial oscillations commencing at 0000 UTC, one to two hours after the surface heat flux changes sign. These cases are summarized in Table 3.2. On many of these nights, detectable inertial oscillations do begin earlier than 0000 UTC and persist later than 1400 UTC.

3.3.2 Discerning the time of initiation of an inertial oscillation

As noted above, the CASES datasets show many nights with inertial oscillation activity spanning multiple 14-hour windows. For example, the night of 10-11 Oct shows inertial oscillations in the 14 hour windows spanning 2000-1000 UTC, 2100-1100 UTC, 2200-1200 UTC, 2300-1300 UTC, 0000-1400 UTC, 0100-1500 UTC, 0200-1600 UTC,

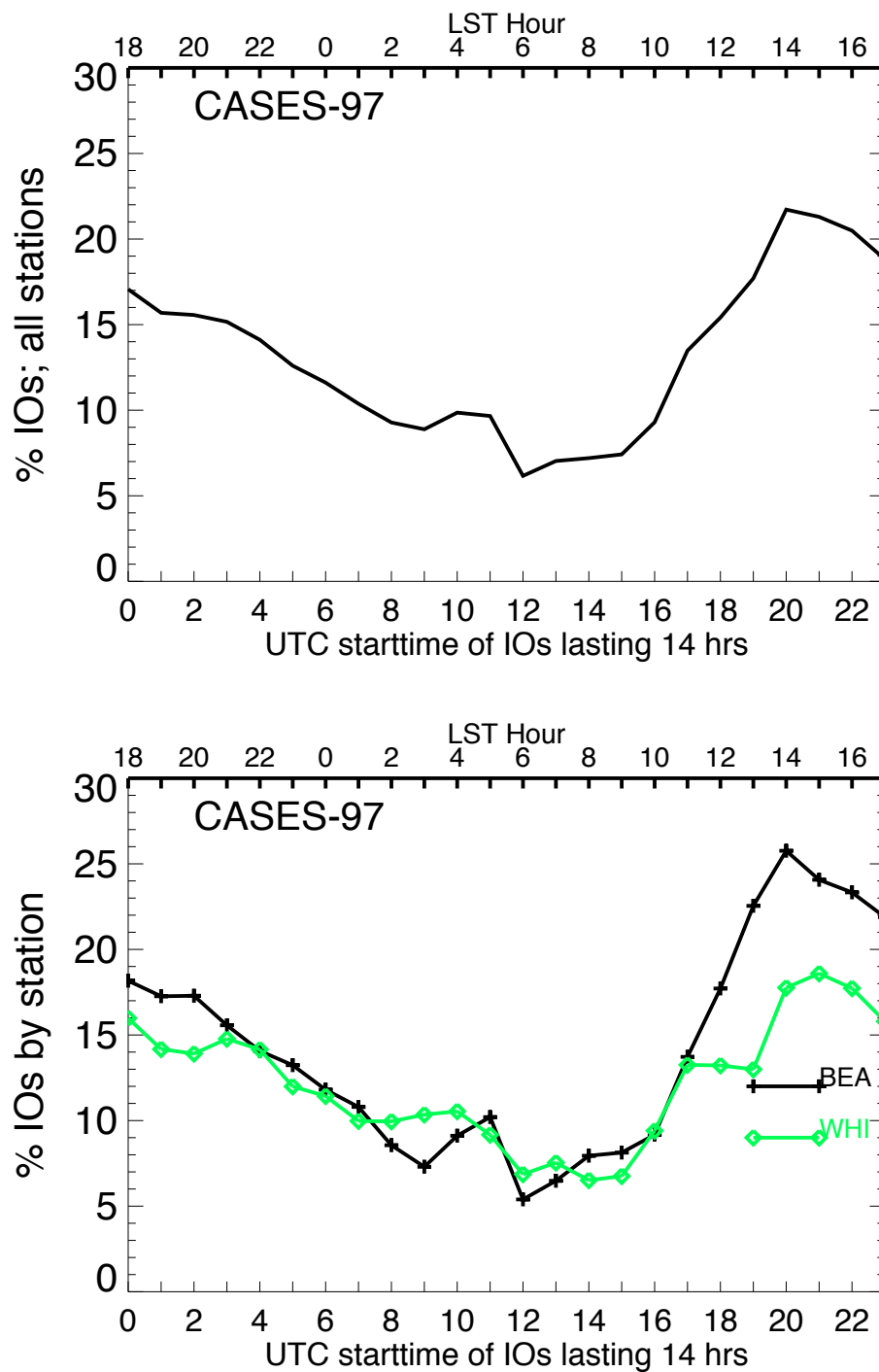


Figure 3.13: The commencement times of inertial oscillations at any height for CASES-97 wind profiler data. For each time of day, the percentage of “good” inertial oscillations out of all time series at all heights with adequate data are shown against the start time for the time series. The top panel shows all station; the bottom panel shows the variation between the stations.

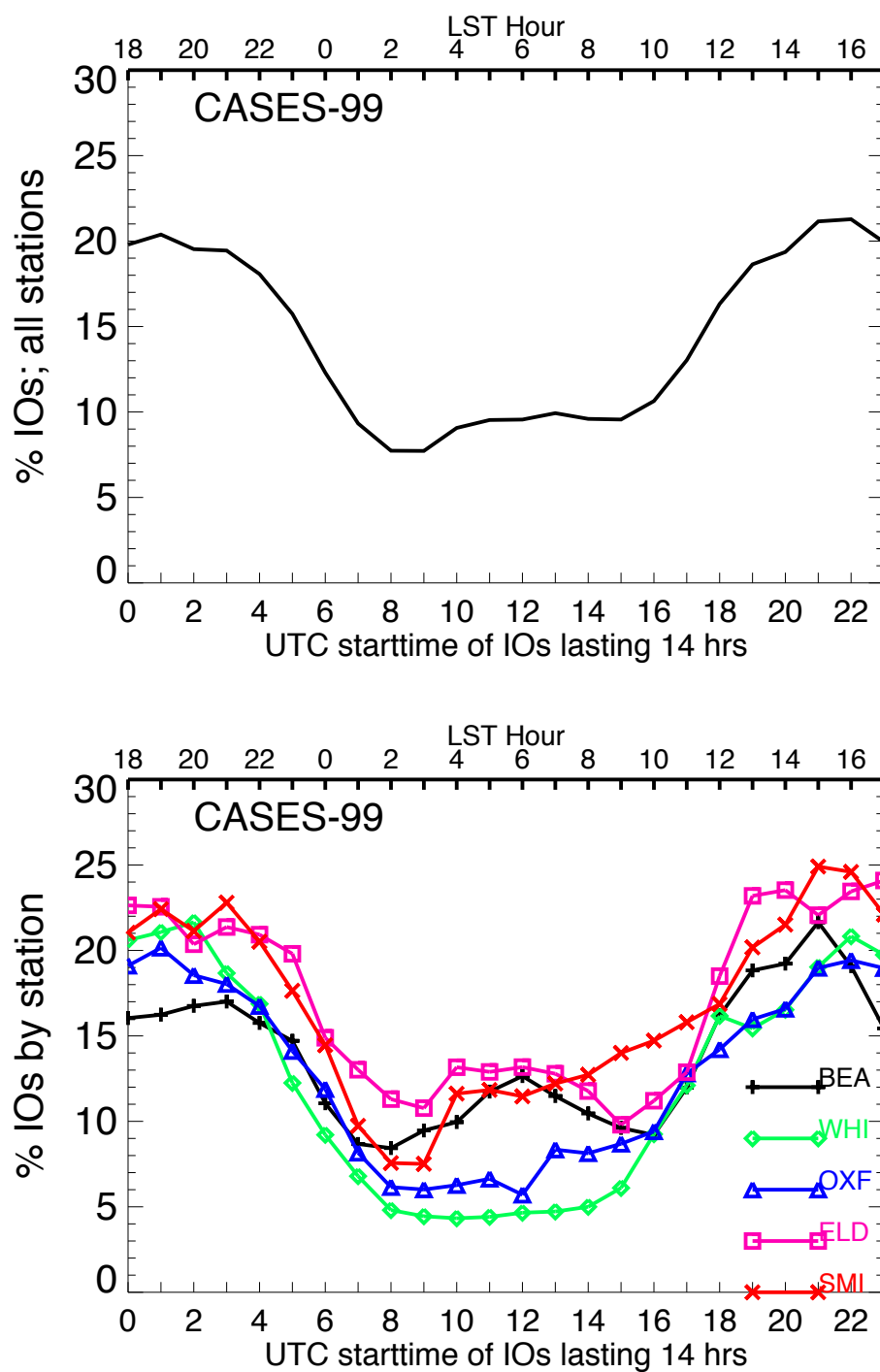


Figure 3.14: As in Figure 3.13 but for the CASES-99 dataset.

Table 3.1: CASES-97: nights with inertial oscillations beginning 0000 UTC and lasting for 12-16 hours; data are from the boundary-layer profilers at Beaumont (B), Whitewater (W), and Oxford (O).

Date	Sites	IOP	Soundings available?	Heights with good IO [m]
May 4-5	B,W,O	2	Yes	150-1500
May 6-7	B	-	No	700-1500
May 10-11	B,W	3	Yes	200-800
May 15-16	B,W,O	-	No	150-500
May 16-17	B,W,O	-	only after 1100 UT	150-700
May 21-22	B,W	-	only 0029, 0213 UT	200-800
May 22-23	B,W,O	-	No	150-1000
May 23-24	B	-	No	150-1300

Table 3.2: CASES-99: nights with inertial oscillations beginning 0000 UT and lasting for 12-16 hours at the boundary-layer profilers located at Beaumont (B), Whitewater (W), Oxford (O), El Dorado (E), and Smileyburg (S)

Date (Julian Day)	Sites	IOP	GLASS winds available	Heights with good IO [m]
Oct 4-5 (278)	B,W,O, E,S	1	no	500 - 1000
Oct 5-6 (279)	B,W,O, E,S	2	intermittently	1000 - 2000
Oct 7-8 (281)	E, S	-	yes	120-500
Oct 10-11 (284)	B,W, E, S	4	yes	100 - 600
Oct 11-12 (285)	W, E	5	yes	100 - 600
Oct 13-14 (287)	B,W, E, S	6	none but at 0300 UT ELD/SMI	100 - 500
Oct 22-23 (296)	B,O, E	10	only after 0700 UT	100 - 500
Oct 27-28 (301)	B,W,O, E, S	-	only at 1100 UT	200 - 800

0300-1700 UTC, and 0400-1800 UTC. This behavior could be explained in two ways. First, the atmosphere could be inertially oscillating for the entire period, from 2000 UTC 10 October - 1800 UTC 11 October. In this case, the fundamental characteristics of the inertial oscillations identified in each window would be similar: the mean wind speed, mean wind direction, the amplitude of the oscillation, and the time of alignment between the inertial oscillation and the mean wind would have a small variance between inertial oscillations starting at 2000 UTC and those commencing at 0400 UTC the next day. The alternative explanation would be that the assumption of (3.3) is incorrect. The winds are not composed of an inertially oscillating component and a mean component that evolves slowly compared to the inertial oscillation. The mean component could itself be changing in time at a rate similar to the rotation of the inertial oscillation, or there could be another influence on the winds during this window of time.

To determine which of these scenarios applies to the winds on a particular night, I compare the time of alignment between the inertial oscillation and the mean wind for all windows showing an inertial oscillation on a particular night. The alignment time can be calculated from the initial wind direction of the inertially oscillating component $\theta_{IO}(0)$, defined from the data using (3.10) and the direction of the steady-state component of the wind θ_{ss} defined from the data using (3.4):

$$t_{align} = \frac{2\pi}{f} (\theta_{ss} - \theta_{IO}(0)). \quad (3.17)$$

A range of alignment times over more than two or three hours indicates that some other forcing is causing conditions to change at a rate comparable to that of the inertial oscillation. Figures 3.15 and 3.16 show the times of alignment for inertial oscillations beginning between 2000 UTC and 0400 UTC on the nights identified as having inertial oscillation activity. Using the spread of the time of alignment variable as a guide, it is obvious from Figures 3.15 (CASES-97) and 3.16 (CASES-99) that the inertial oscillations on 15-16 May 1997, 4-5 Oct 1999, 5-6 Oct 1999, and 13-14 Oct 1999 are taking

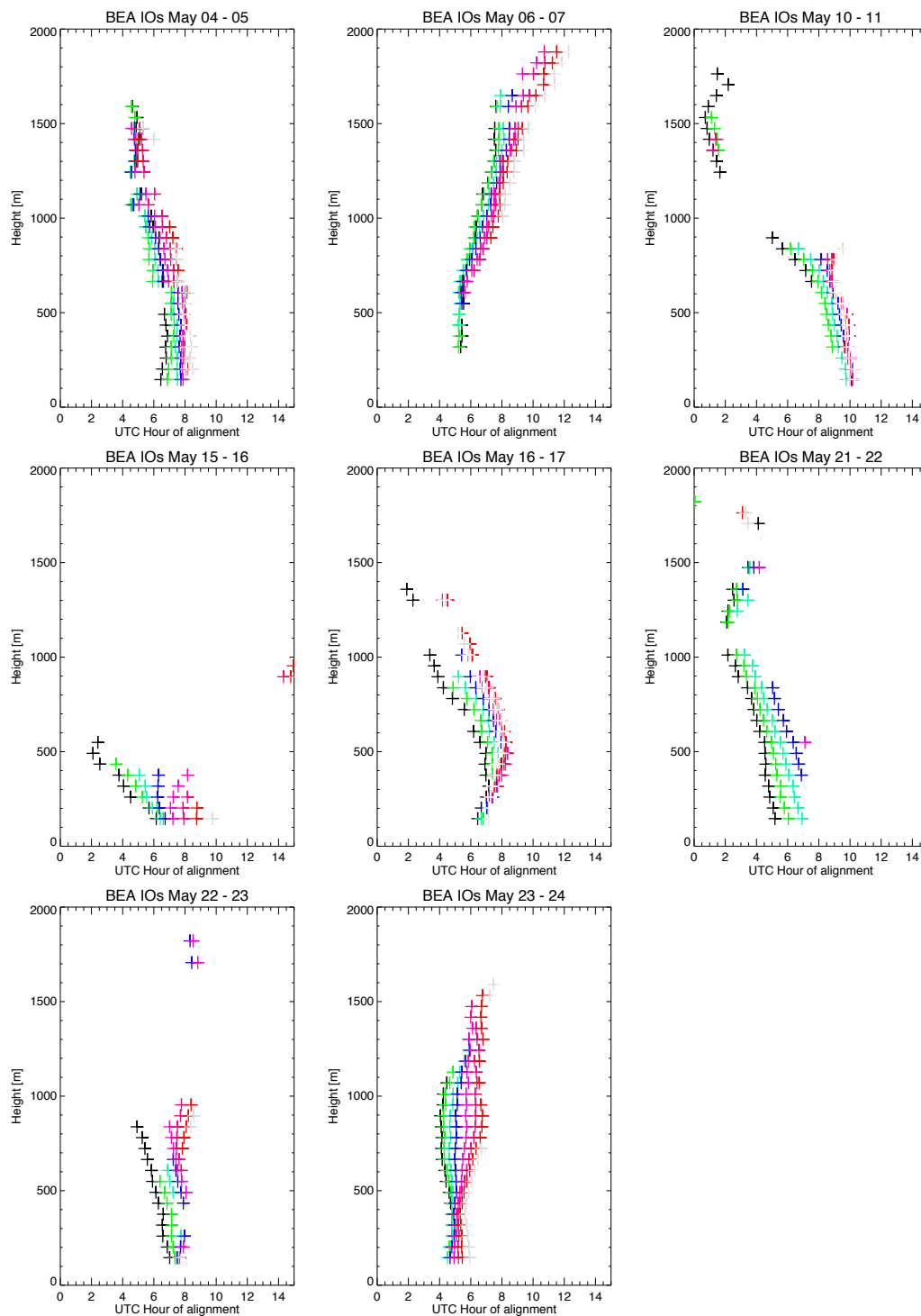


Figure 3.15: Alignment times for inertial oscillations and the mean wind from the Beaumont site during CASES-97. Each panel shows the inertial oscillations beginning between 2000 UTC and 0400 UTC on the night noted in the title. Times earlier in the night are black to green to blue, times later in the night are red. Inertial oscillations starting at 0000 UTC are in pink. A wide spread in these values indicates that the mean component of the winds on this night is changing in time or that the inertial oscillation is not the dominant influence on the winds on this night. A spread of 2 hours or less is expected due to averaging effects.

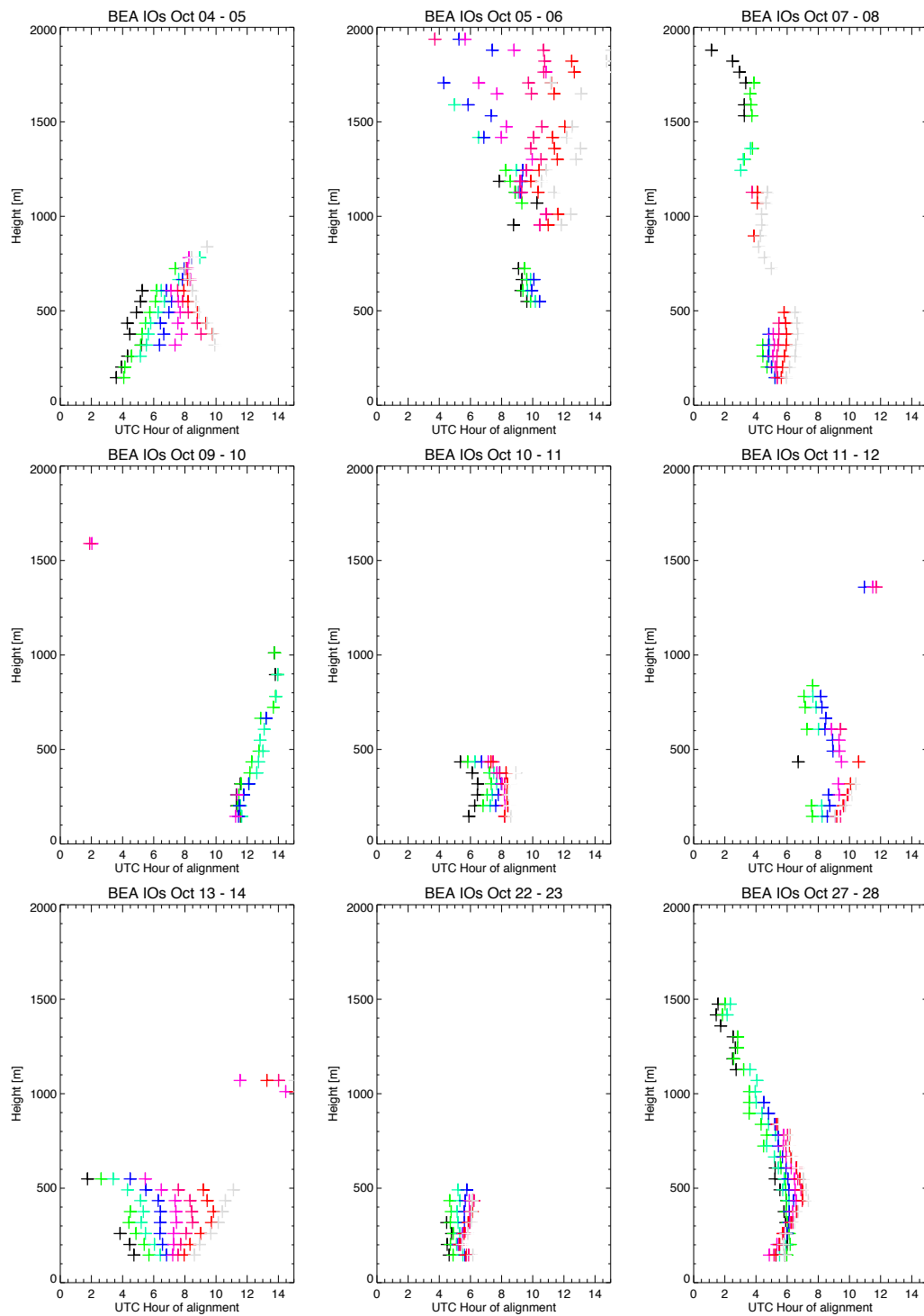


Figure 3.16: As in Figure 3.15 but for nights showing inertial oscillations from the Beaumont profiler during CASES-99.

place while the mean wind is also rotating. This implies that the evening transition is not the only, or even the primary, mechanism acting on the boundary layer winds on these nights. These nights are not suitable for direct comparison with Blackadar's model.

3.3.3 Comparison between observations and the Blackadar scenario

Of the sixteen nights in CASES-97 and CASES-99 that show inertial oscillations commencing around the evening transition, eight nights have adequate thermodynamic sounding data for a direct test of Blackadar's hypothesis as illustrated in Figure 3.11. Recall that Blackadar expected the winds above the surface inversion to oscillate inertially, and the largest amplitude oscillations would be closest to the surface. The nights of 4-5 October, 5-6 October, and 13-14 October are not considered here because the variation of t_{align} (discussed above) shows that inertial oscillations are accompanied by other forcings that modify the flow. For the remaining nights (4-5 May 1997, 10-11 May 1997, 10-11 October 1999, 11-12 October 1999, 22-23 October 1999), Figures 3.17- 3.21 compare the potential temperature profiles with the amplitudes of the observed inertial oscillations; these figures should be compared with Figure 3.11 for agreement with Blackadar's hypothesis. The amplitudes of the inertial oscillations should be largest at the inversion and decrease rapidly with height.

On the day of 4 May 1997, the convective boundary grows up to 1500 m; the nocturnal surface inversion reaches 200 m by 0347 UTC (2147 LST) on 5 May (4 May). Inertial oscillations are apparent from the lowest range gate through 1500 m at the Beaumont profiler, as seen in Figure 3.17. The amplitudes of these oscillations increase from the surface up to 350 m, the same height of the surface inversion layer at 0329 UTC. The amplitudes of the inertial oscillations are then constant with height throughout the residual layer instead of decreasing with height as Blackadar proposes for a geostrophic wind that is constant with height (Figure 3.11). For the Blackadar mechanism to apply

to this night, the geostrophic wind must vary with height so that the ageostrophic component of the winds is constant above 350 m and larger than that below 350 m (Figure 3.12).

Figure 3.18 depicts the data from the night of 10 - 11 May 1997. Inertial oscillations reach their largest amplitude at around 600 m, well above the surface-based inversion at 150 m. They are present throughout the lowest 1000 m, in the residual boundary layer. The structure of the amplitude of the inertial oscillations with height again indicates that the ageostrophic component of the wind varies with height.

On the night of 10-11 October 1999 (Figure 3.19, the surface inversion grows to a depth of 200 m by 0700 UTC. Though the profilers do not provide wind data much below this depth, they do show that the largest inertial oscillation amplitudes appear between 200 and 350 m above the surface. Additionally, the amplitudes of the inertial oscillations are only large enough to swamp the noise in the winds below 600 m. At 600 m, the 0700 UTC sounding shows a transition from a slightly stable layer to a residual mixed layer. For this case, enhanced stability allows the development of inertial oscillations in the stable boundary layer. These inertial oscillations follow the pattern expected if the Blackadar mechanism were the only influence on winds this night.

Again, on the night of 11 - 12 October 1999, (Figure 3.20) the boundary-layer wind profilers show inertial oscillations close to the surface. The amplitudes of the best-fit inertial oscillations are largest around 300 m, but noise in the signal interferes above 500 m. Radiosonde data (left panel of Figure 3.20) show that the surface inversion only grows to 150 m by 0700 UTC. The decay in the amplitude of the inertial oscillation shown in Blackadar's schematic is not apparent on this night, indicating a slight tilt in the geostrophic wind so that it slightly increases with height in accordance with Figure 3.12.

The night of 22 - 23 October (Figure 3.21) shows inertial oscillations between 150 and 500 m at three stations. The amplitudes of these oscillations actually increase

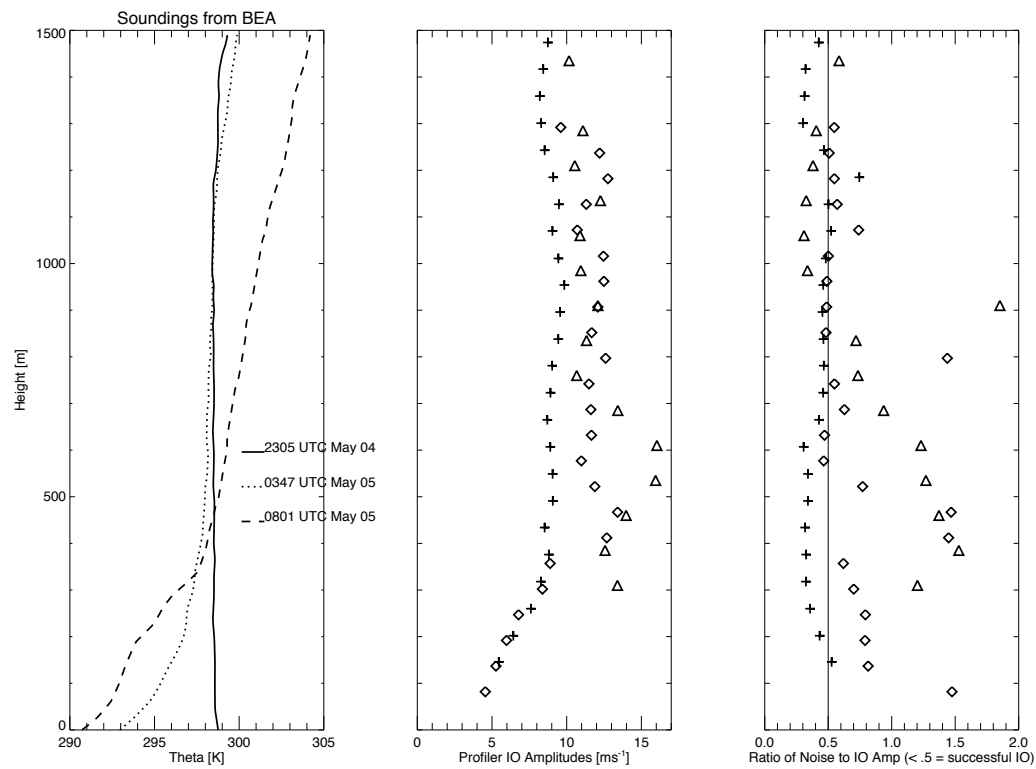


Figure 3.17: A test of Blackadar’s hypothesis for the night of 4-5 May 1997. The left panel shows profiles of potential temperature from the Beaumont site at nominal 2300, 0300, and 0700 UTC 4-5 October 1999. The middle panel shows the amplitudes of the inertial oscillations extracted from wind profiler data for the 3 profiler sites (Beaumont data points are represented with +, Whitewater with diamonds, and Oxford with triangles). The right panel shows the R_{ws} for the inertial oscillations extracted from the profiler data; values less than 0.5 indicate “good” inertial oscillations.

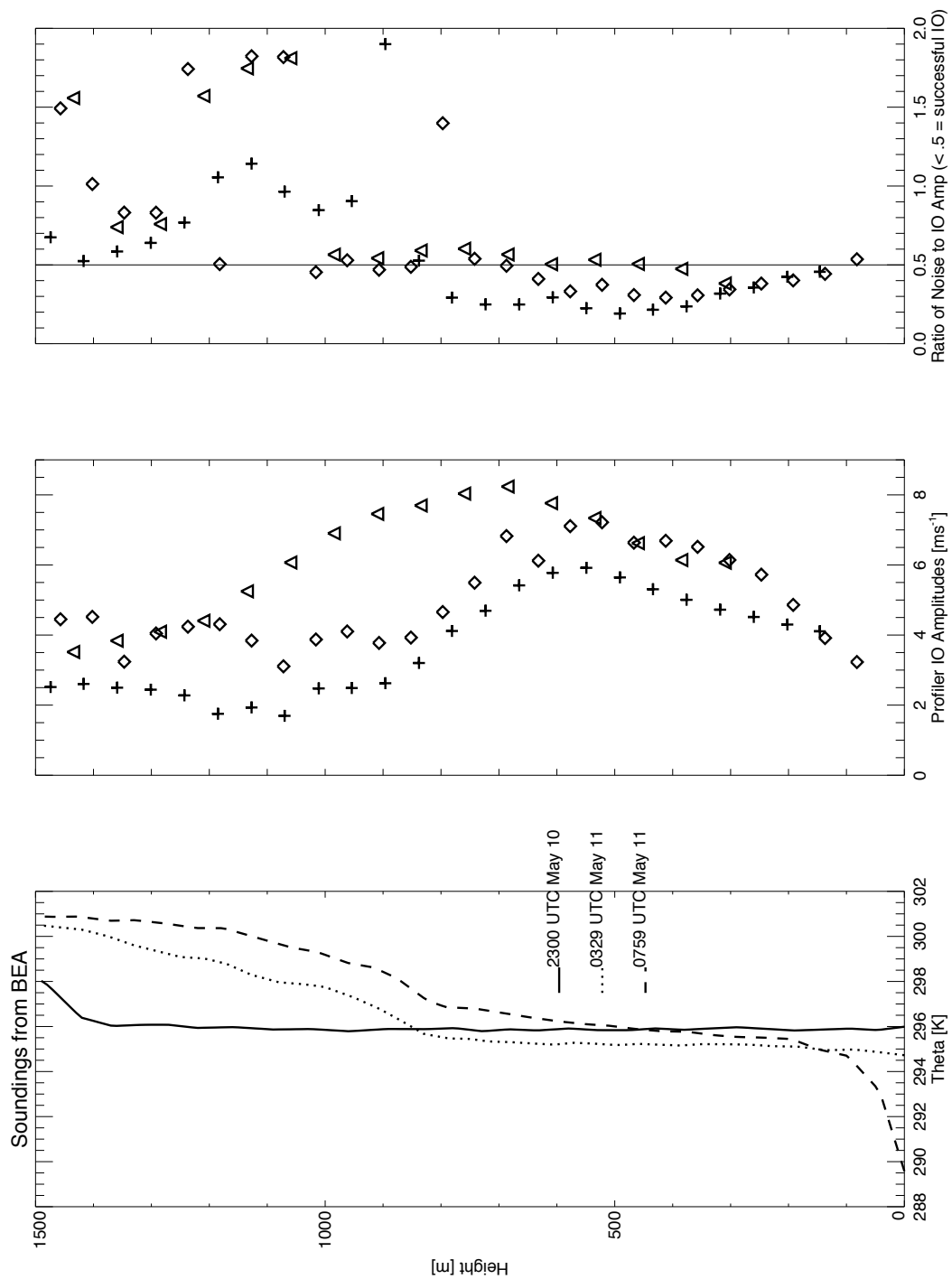


Figure 3.18: As in Figure 3.17 but with soundings from Beaumont on 10 - 11 May 1997.

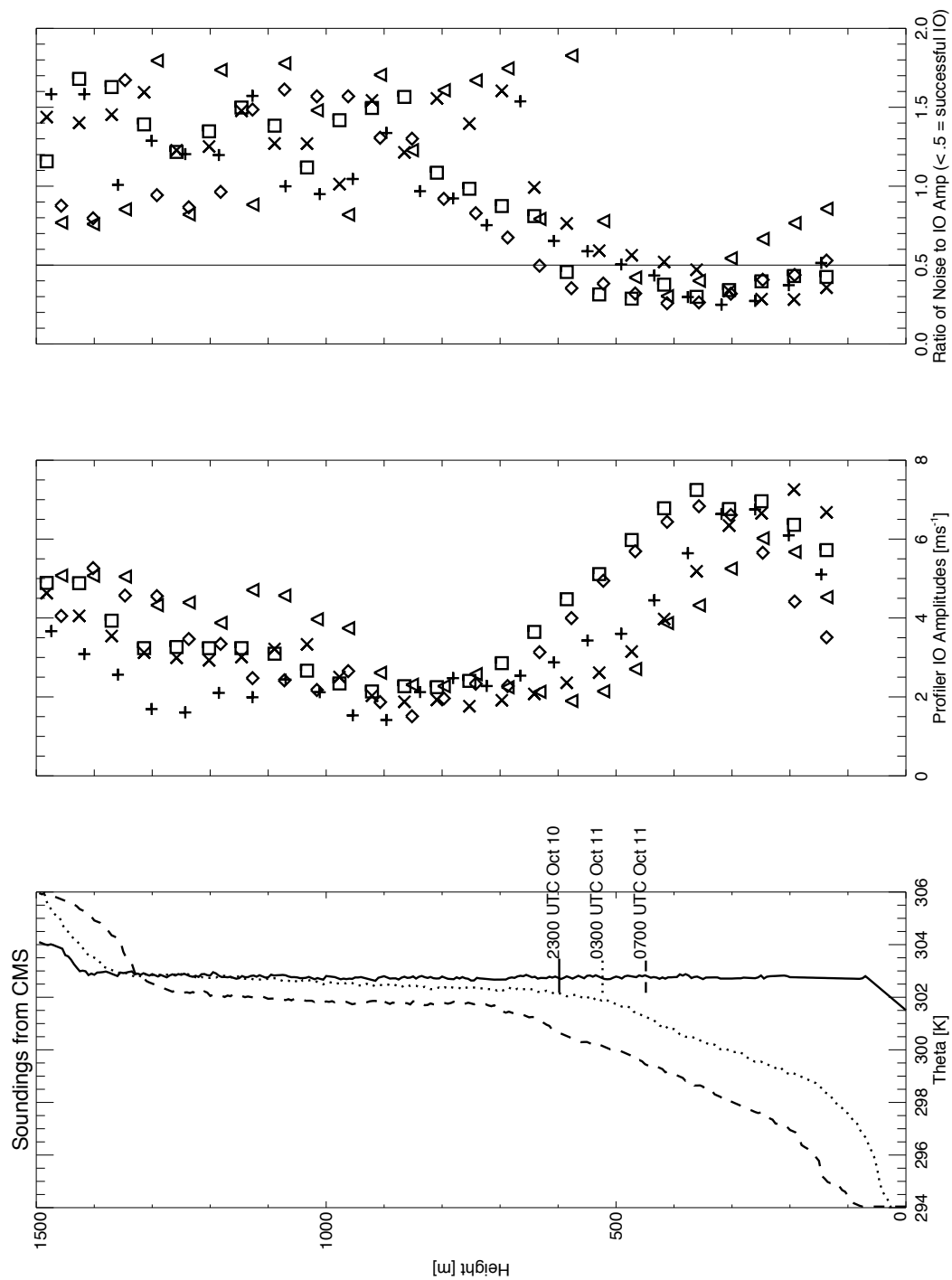


Figure 3.19: As in Figure 3.17 but with soundings from the Central Main Site on 10-11 October 1999. A data point from the El Dorado profiler is denoted with a square, and x represents data from Smileyburg.

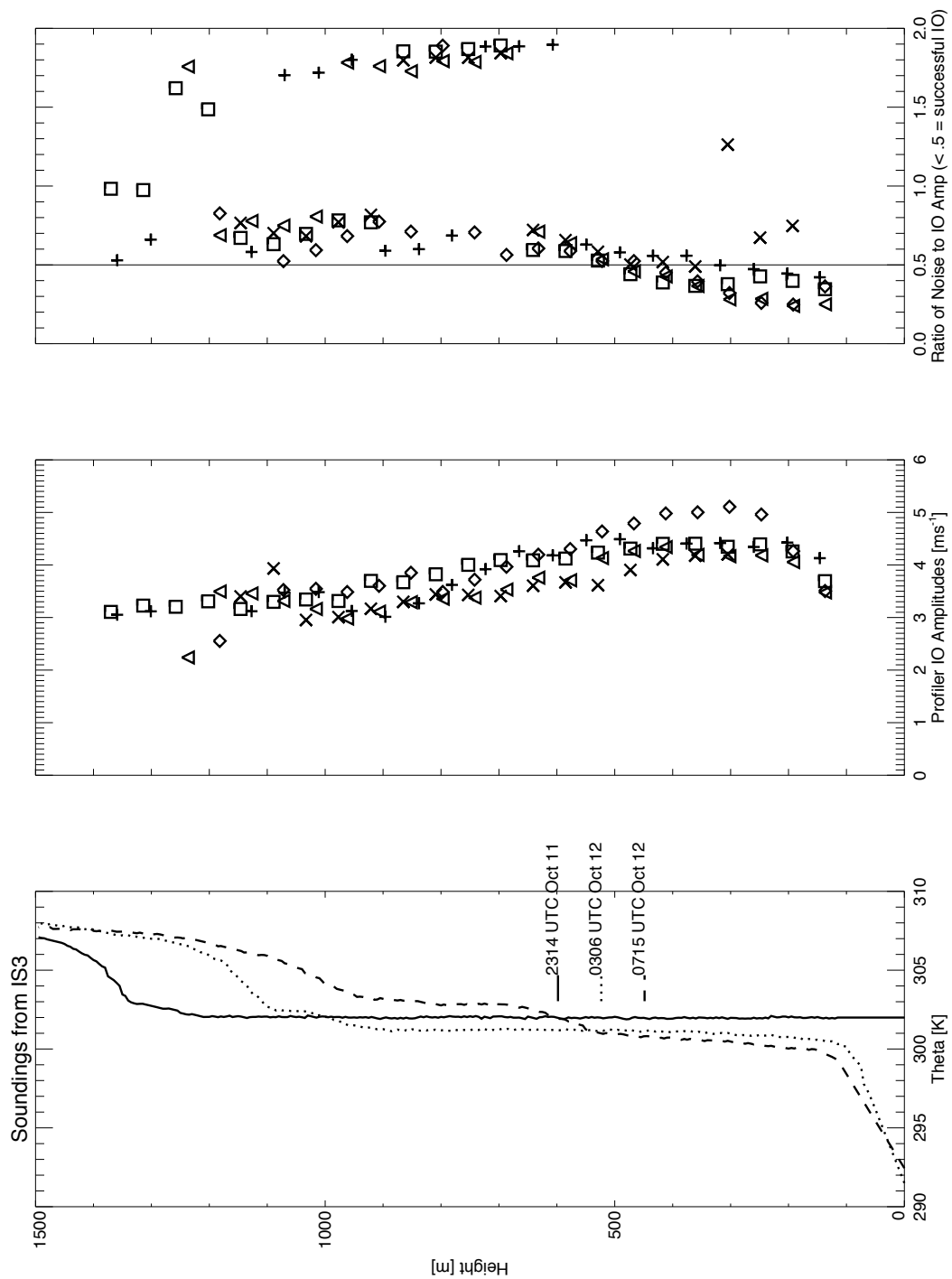


Figure 3.20: As in Figure 3.17 but with soundings from the Smileyburg site on 11-12 October.

with height, instead of decreasing with height as one would expect from a geostrophic wind with no vertical variation, again indicating that the geostrophic wind is strongly increasing with height.

The CASES-97 and CASES-99 datasets show few inertial oscillations that follow the proposed Blackadar mechanism. Instead, inertial oscillations are influenced strongly by other processes, including a changing geostrophic wind (both in height and in time). The fact that these inertial oscillations do begin at the evening transition of the boundary layer does suggest some mechanism similar to the one-dimensional low-level containment of frictional influence suggested by Blackadar. However, for complete understanding of the behavior of the observed inertial oscillations, other mechanisms must be considered on most nights. The one night exhibiting behavior similar to that expected by Blackadar will be explored in more depth, both in this chapter (with respect to whether or not inertial oscillations are the mechanism causing the low-level jet) and in Chapter 5, where this night will be simulated with a numerical model.

3.4 Interaction of inertial oscillations with the LLJ

The discussion in chapter 4 will argue that nocturnal low-level jets (LLJs) in both CASES-97 and CASES-99 are highly correlated with occurrences of inertial oscillations. In this section, I explore the relationship between inertial oscillations and the LLJ on the night 10-11 October 1999. I focus on this case study because it is simple enough for a direct comparison to the Blackadar model for LLJ initiation. The day of 10 October 1999 is marked by moderate winds, warm temperatures, few clouds, and the growth of a convective boundary layer. The mixed layer, sampled by a sounding at the main Leon site shown in Figure 3.22, grows to 1400 m by 2300 UTC on 10 October. The evening transition commences at 2230 UTC, when surface stations show that the surface heat flux changes sign. By 0100 UTC on 11 October, the wind profiles show an indication of LLJ development. Profiler data throughout the night are shown in Figure 3.23.

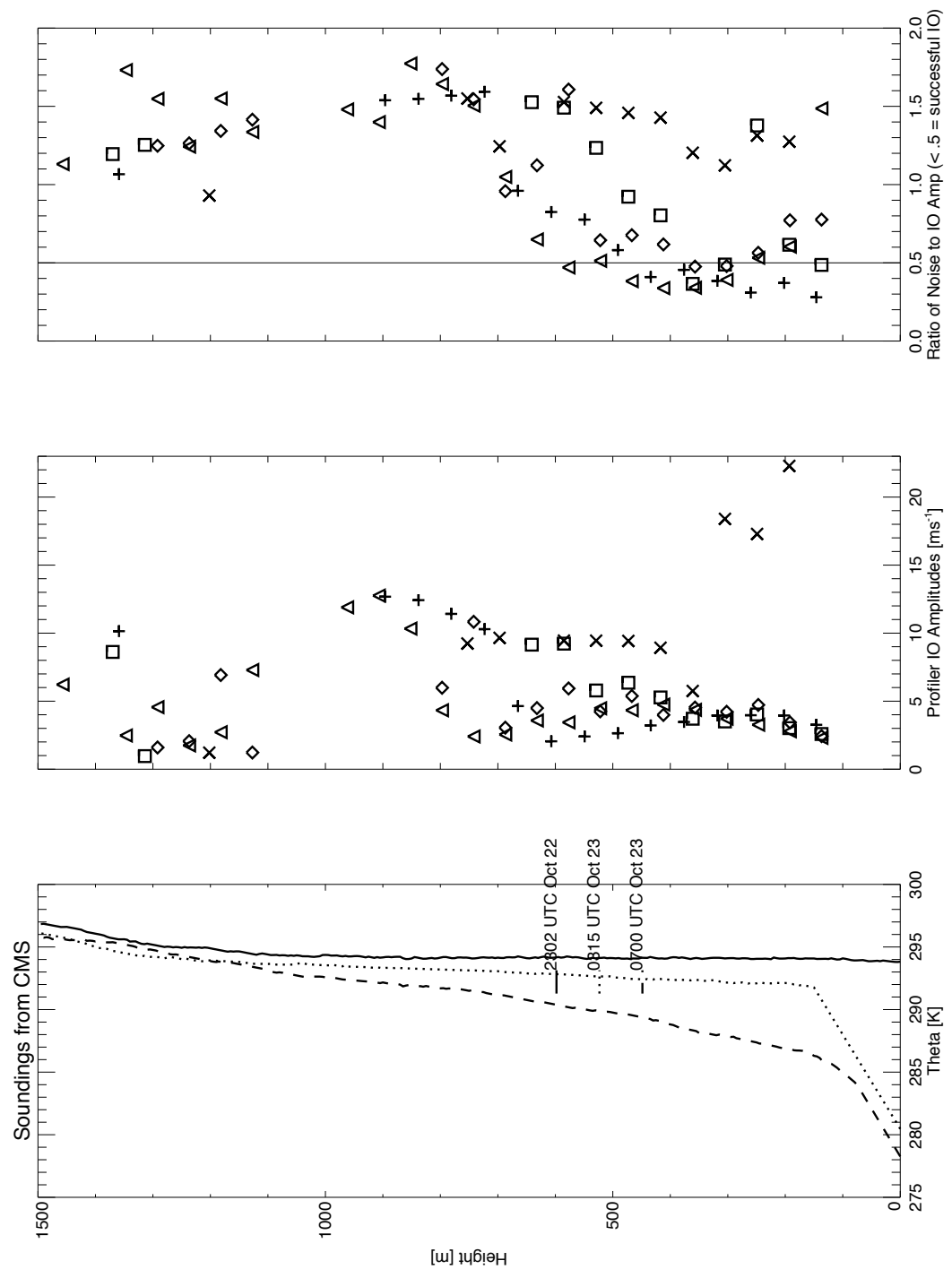


Figure 3.21: As in Figure 3.17 but with soundings from the central main site on 22-23 October.

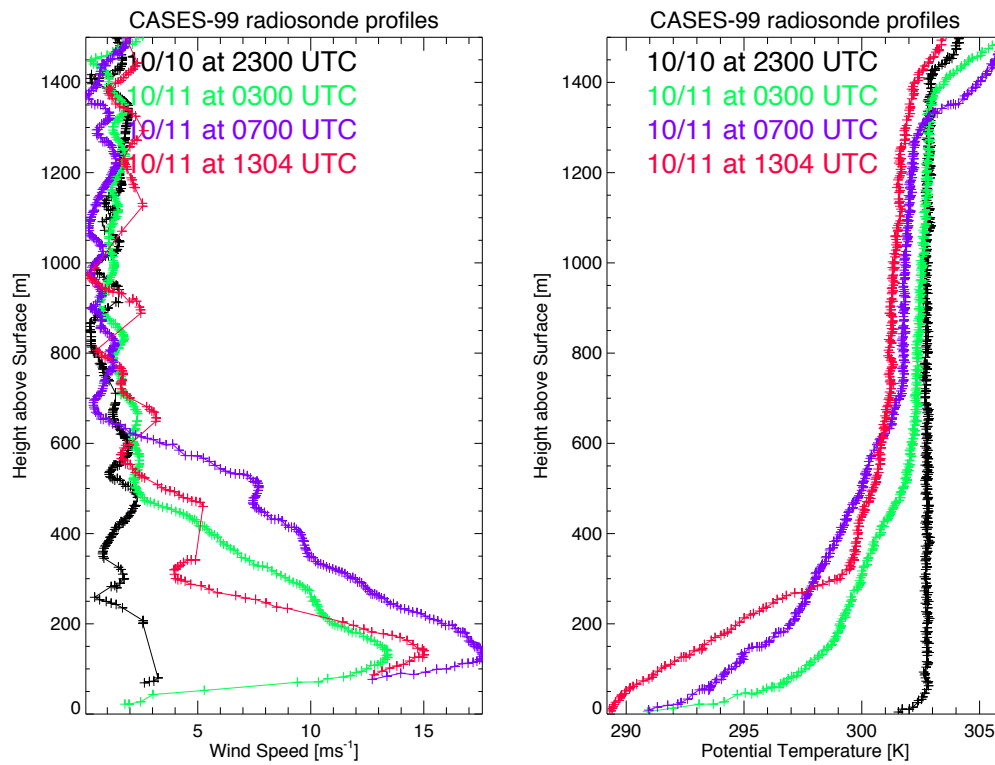


Figure 3.22: Profiles of wind speed and potential temperature from radiosonde data on the night of 10-11 October 1999. The + symbol indicates actual observations; some gaps exist in the winds portion of the dataset.

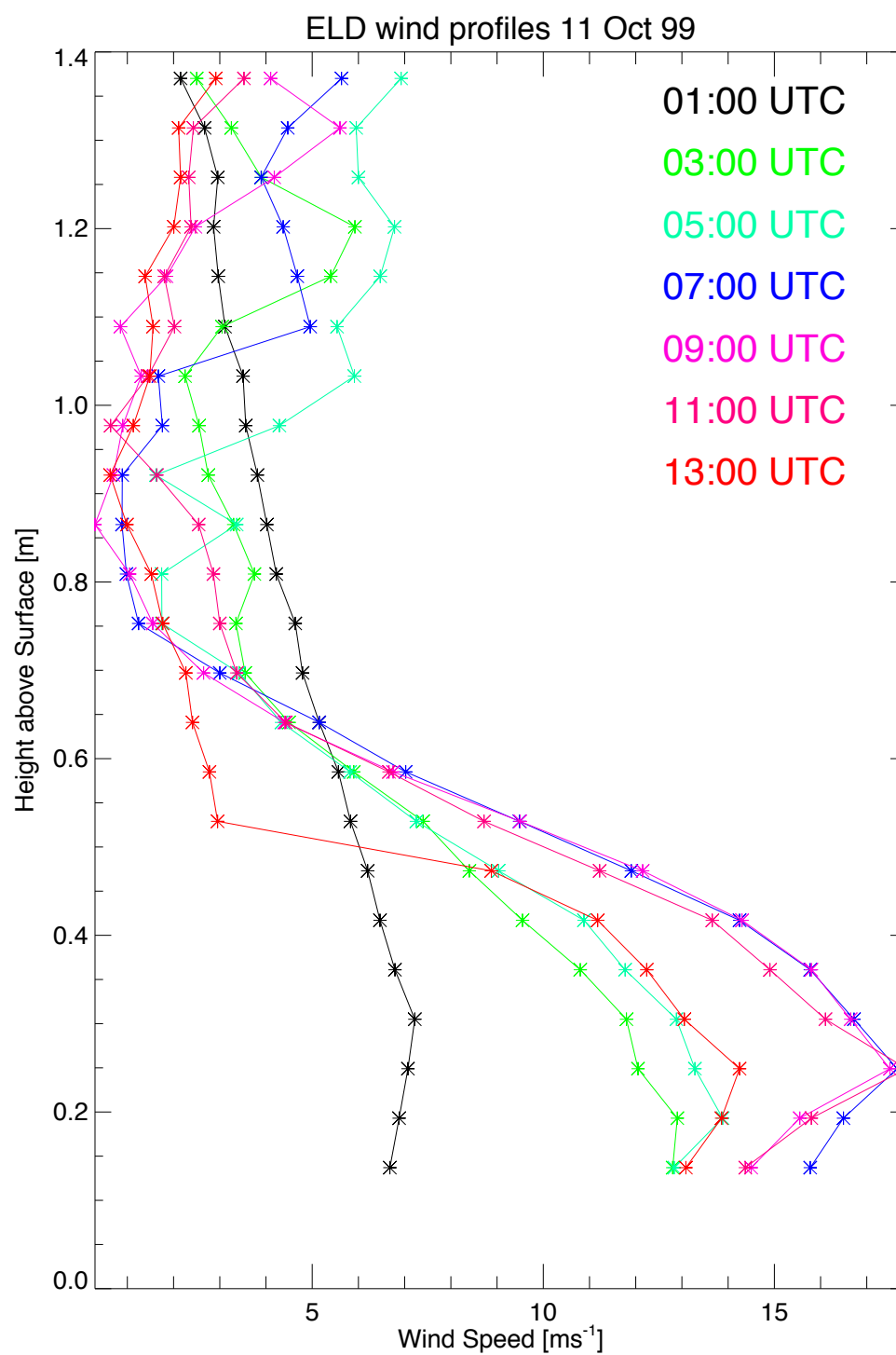


Figure 3.23: Wind profiler data from El Dorado at the indicated times of 10-11 October 1999

The 0300 UTC sounding and profiler data show a distinct jet structure in the winds of the lowest 600 m. This jet persists through the night, with the wind direction rotating through the east as the wind speeds in the jet increase to 18 ms^{-1} by 0700 UTC. The speed in the jet maximum remains constant through 1200 UTC, and then starts to diminish (as shown in Figures 3.22 and 3.23). Throughout the night, all jet development is confined to below 600m; profiler resolution does not reveal whether the jet maximum rises or descends over the course of the night. Thermodynamic data from the radiosonde does reveal that a strongly stable layer (7 K over 450 m) exists from 150 m to 600 m, atop a sharp surface inversion (4 K over 150 m). Above 600m, the residual mixed layer survives the entire night.

Analysis of the inertial oscillations on this night shows that they constitute a significant portion of the winds (see Figure 3.24). All five profilers show evidence of inertial oscillations beginning at 0000 UT; only below 600m do the inertial oscillations pass the $R_{ws} \leq 0.5$ test. Also, below 600m, we see that the amplitudes of inertial oscillations are large. Calculated amplitudes are largest between 200 and 400 m, tapering off above and below this range. Mean wind speeds at all stations have a maximum around 250 m. Inspection of the initial wind direction of the inertial oscillations shows that below 600m, there is coherence between all the stations and all levels; above, there is wide variance in the signal. This wide variance, coupled with small amplitudes and large R_{ws} values, indicates that inertial oscillations are not significant above 600m.

This analysis also shows that inertial oscillations in the highly stable layer below 600m constitute a large component of the LLJ. The inertially oscillating component of the wind, which begins its oscillation at 0000 UT, has an initial wind direction of approximately 290° , or from just north of west. As the bottom right panel of Figure 3.24 depicts, it will take seven hours for this component of the wind to align with the mean wind (which is easterly). These two components line up at 0700 UTC, the same time that the jet reaches its maximum amplitude. On this night, inertial oscillations in the

stable layer enhance the development of the maximum in wind speed in the lower levels of the boundary layer by 30 percent through the vertical extent of the jet (compare the amplitudes shown in Figure 3.24 with the profiles shown in Figure 3.23), although the jet profile would still be evident without the influence of the inertial oscillation. For this case, the thermal wind induced by the sloping Great Plains probably enhances the LLJ (Holton, 1967).

3.5 Summary

This chapter has discussed observations of inertial oscillations in radar boundary-layer wind profiler data. By applying a spectral analysis technique used by others (Gupta et al., 1997), I confirm that motions with periods near the diurnal-inertial peak influence boundary-layer winds in the CASES-97 and CASES-99 datasets. I then develop and verify a new technique for extracting the best-fit inertial oscillations from a time series of winds, and discuss the criteria for determining whether or not those inertial oscillations are significant or not.

This objective method of identifying inertial oscillations is an improvement over previous techniques. Many studies identify inertial oscillations only by looking for a qualitative veering in the wind, where this technique requires a veering rate determined by the local inertial period (Banta et al., 1998). By providing a measure of the magnitude of the inertial oscillation compared to the noise remaining after the inertial oscillation has been extracted, this technique will prove useful in future studies of undamped inertial oscillations.

I apply this technique to the CASES-97 and CASES-99 datasets and discuss the occurrences of inertial oscillations observed with this technique. Although several nights are identified with inertial oscillations commencing around the time the surface heat flux changes sign, only one night shows inertial oscillations consistent with the pattern predicted by Blackadar for a night with a constant geostrophic wind. On that

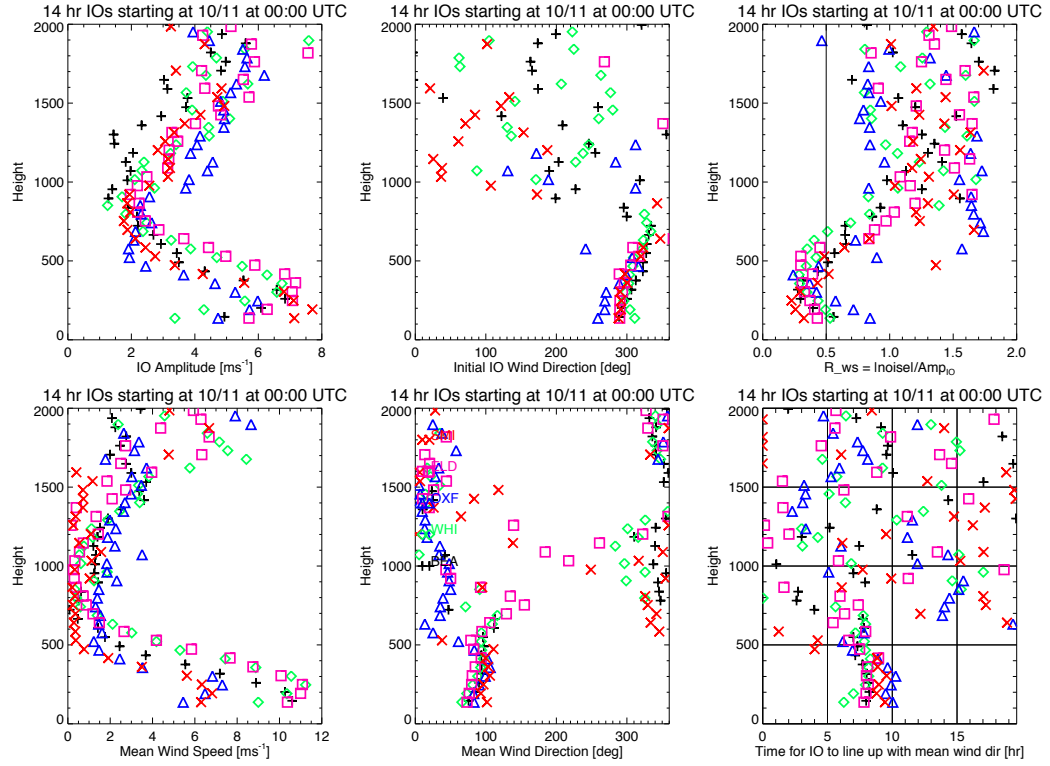


Figure 3.24: Characteristics of inertial oscillations starting at 0000 UTC 11 October 1999. Each symbol (+, diamond, triangle, square, x) corresponds to one profiler station (Beaumont, Whitewater, Oxford, El Dorado, Smileyburg). From left to right, the figures depict (a) the amplitude of the best-fit inertial oscillations, (b) the initial wind direction of those inertial oscillations, (c) the R_{ws} value, (d) the mean wind speed, (e) the mean wind direction, and (f) time in hour that it will take the inertial oscillations to align with the mean wind direction.

night, inertial oscillations are shown to be a significant component (30 percent) of the total winds, so that the low-level jet that forms would be greatly reduced without the presence of inertial oscillations. A broader discussion of the correlation of inertial oscillations with the LLJ follows in Chapter 4.

Chapter 4

Climatology of the Low-Level Jet

The duties of the Wind are few,
To cast the ships, at Sea,
Establish March, the Floods escort,
And usher Liberty.

The pleasures of the Wind are broad,
To dwell Extant among,
Remain, or Wander,
Speculate, or Forests entertain.

The kinsmen of the Wind are Peaks
Azof - the Equinox,
Also with Bird and Asteroid
A bowing intercourse.

The limitations of the Wind
Do he exist, or die,
Too wise he seems for Wakelessness,
However, know not I.

Emily Dickinson # 1137

The current focus on inertial oscillations has been justified on the grounds that inertial oscillations are proposed to be the mechanism for the generation of low-level jet structures in nocturnal winds. The exploration of inertial oscillations thus far has focused on the evolution of winds in time at one height. In this chapter, I examine boundary-layer winds from a different perspective, that of profiles of winds at all heights at one time, to develop a climatology of the low-level jet structures that inertial

oscillations are hypothesized to cause.

4.1 Background and Importance of the LLJ

Numerous studies indicate the importance of low-level jet (LLJ) structures in the wind profile to climate. These jets have been documented over all seven continents (Stensrud, 1996), and the exploration of the sources of LLJs has been an active area of research for at least the last 70 years.

Over the Great Plains of the United States, the LLJ dominates the summer moisture budget. The nocturnal LLJ is particularly important in transporting moisture from the Gulf of Mexico into the Great Plains and in generating mesoscale convective complexes that produce large amounts of precipitation (Rasmusson, 1968; Helfand and Schubert, 1995; Berbery et al., 1996; Arritt et al., 1997; Whiteman et al., 1997). During nights with LLJ activity, the low-level inflow of moisture from the Gulf of Mexico over the Great Plains increases by 45 percent, on average, over nocturnal mean values (Higgins et al., 1997). The LLJ has also been implicated in violent weather: 13-26 June 1993 were marked by “tornados, high wind, large hail, and heavy rain” that accompanied powerful thunderstorms (Wu and Raman, 1997). Coincident with these storms was an anomalously strong occurrence of the nocturnal LLJ (Mo et al., 1995).

Even during quiescent synoptic-scale flow regimes, LLJs can modify weather in dramatic ways, potentially threatening the safety of airport operations. Kaplan et al. (2000) argue that the diurnal LLJ represents the dominant low-level wind shear production mechanism during such conditions. The strong vertical wind shears and modified turbulence close to the earth’s surface generated by LLJs modify the drift and dissipation of wake vortices from aircraft. The change in atmospheric dissipation rates due to jet activity leads to a less predictable residence time for wake vortices. Determining the safe following range for aircraft becomes much more difficult during strong LLJ events.

The shear generated by this nocturnal LLJ, which often rests atop the growing

surface inversion (Kosović and Curry, 2000), also influences the growth of the surface stable layer (Mahrt, 1999). The shear at the top of the LLJ can produce a state conducive to the generation of Kelvin-Helmholtz billows, which can transfer heat and momentum throughout the boundary layer, the residual layer, and above (Stull, 1988). An improved understanding of the generation of the LLJ will enhance our understanding of the nocturnal boundary layer (Mahrt, 1999).

One of the barriers to understanding the generation of the LLJ is the plurality of possible sources of the LLJ leading to the plurality of types of LLJs. A wide variety of phenomena are included under the rubric “low level jet”. It is not clear that the nocturnal inversion wind speed maximum (Blackadar, 1957; Thorpe and Guymer, 1977; Kosović and Curry, 2000) is the same jet as the large scale Great Plains LLJ (Rasmusson, 1968; Helfand and Schubert, 1995; Berbery et al., 1996; Arritt et al., 1997; Whiteman et al., 1997). Even the phrase “low level” is ambiguous (Stensrud, 1996), as it has been applied to jets with wind speed maxima between 100 m (Källstrand, 1998) and 3000 m (Whiteman et al., 1997). Efforts have been made to clarify terminology, by distinguishing between 1) jets due to synoptic or quasi-geostrophic forcing and 2) jets generated to some extent by the Coriolis force (Stensrud, 1996) (these jets would include marine boundary layer jets as well as nocturnal inversion wind speed maxima). However, a review of the literature shows that the term “low-level jet” is usually applied based only on the presence of a jet structure in the vertical profile of the horizontal wind speed. Yet even that characteristic has often been ignored: Djurić and Damaiani (1980) for example, use “low-level jet” synonymously with “southerly wind in the mixed layer.”

Just as there is ambiguity in assigning the name “low-level jet”, there is debate over the sources of LLJs. Investigations into the development of the LLJ have found that inertial oscillations, clockwise rotation in the ageostrophic component of the wind profile in response to the Earth’s rotation, are present when the nocturnal LLJ forms and are presumed to constitute the primary mechanism for the generation of the LLJ

(Blackadar, 1957; Thorpe and Guymer, 1977; Zhong et al., 1996; Arritt et al., 1997; Whiteman et al., 1997). The Blackadar mechanism prescribes that a developing surface inversion will decouple winds aloft from the surface layer, generating inertial oscillations aloft. When these inertial oscillations align with the mean or geostrophic wind, a jet structure will form where the ageostrophic component of the wind was at its largest when the surface inversion developed. In a study that combines numerical simulation and coarse radar wind profiler observations, Zhong et al. (1996) conclude that over the Great Plains of the United States, the inertial oscillation is the most important mechanism for the formation of one observed nocturnal LLJ, while the strength of the LLJ is dependent on the geostrophic wind or synoptic-scale pressure gradient. Marine atmospheric boundary layer jets (Zemba and Friehe, 1987; Angevine et al., 1996; Källstrand, 1998) are theorized to share a common source with nocturnal inversion wind speed maximum jets. As warm continental air flows over cooler water, a decoupling occurs similar to the decoupling caused by the surface inversion generated at the evening transition. The warm continental air veers, as it would in an inertial oscillation, creating a jet profile some distance off the coast, where the distance is defined by the time it takes the inertial oscillation component to align with the mean wind. The exact time (or distance off-coast) for the marine boundary jet maximum depends on the angle between the geostrophic wind and the actual wind when the oscillation starts.

Although inertial oscillations that develop due to decoupling of the surface and winds aloft have been presumed to be the source for some jets, numerous other sources for jet development and enhancement have been proposed. Holton (1967) shows that even if eddy viscosity is constant and no decoupling occurs, the heating and cooling of the slopes of the steep Rocky Mountains and gently sloping Great Plains can produce diurnal oscillations that will create LLJs. Other sources include frontal motions (Ostdiek and Blumen, 1997) and synoptic scale pressure gradients (Uccellini, 1980; Djurić and Damaiani, 1980; Djurić, 1981; Mahrt, 1981; Arritt et al., 1997), land and sea breezes

(Källstrand, 1998), and differential heating from land surface uses (Wu and Raman, 1997).

Using data from the boundary-layer wind profilers in Kansas during spring (CASES-97) and fall (CASES-99), this portion of my study focuses on the connection between observed jet structures and occurrences of inertial oscillations. Although the established technique for a climatology of LLJs (Bonner, 1968; Mitchell et al., 1995; Whiteman et al., 1997) does not distinguish between jets due to inertial oscillations and those attributed to different sources, I can use the jets identified with that technique to document the connection between occurrences of some jet structures and inertial oscillations. In the first part of this chapter, I establish that the general characteristics of the jets observed with boundary-layer wind profilers in the CASES-97 and CASES-99 experiments are similar to those observed in previous climatologies of the LLJ which use data from different instrument platforms. I then combine the results from my inertial oscillation analysis of Chapter 3 and the LLJ observations using the established technique to document the correlation between inertial oscillations and LLJs.

4.2 LLJs observed in CASES-97 and CASES-99

Bonner (1968) establishes the first climatology of low-level jet structures in the continental United States. (Earlier, Farquharson (1939) summarizes observations of jet structures in the winds of tropical Africa without developing a categorization scheme for the jets he observes). Bonner does not seek to define the sources of the jets he observes, but rather to establish when and where they occur with what characteristics so that other studies on their sources can proceed. His primary questions are whether or not the jet is primarily a nocturnal event (which would support Blackadar's hypothesis) and whether it was constrained to the eastern slope of the Rocky Mountains (to support Holton (1967)). Bonner examines the twice-daily soundings (0000 and 1200 UTC) from 47 stations in the United States from January 1959 to December 1960. The low-level

resolution of his data is limited (he has access to data at the surface, 150, 300, 500, 1000, 1500, 2000, 2500, 3000, 4000 m, etc.), but he establishes the criteria by which jets in future studies (Mitchell et al. (1995), Whiteman et al. (1997), and the present work) are judged. Whether or not these criteria are the most suitable, given the present observational capacity, will be discussed below.

Bonner classifies each sounding into a LLJ category based on 1) the maximum wind speed between the surface and 1500 m, and 2) the decrease of winds between the maximum and the minimum wind speed between the maximum and 3000 m. He creates three categories, requiring for the weakest category (LLJ-1) that winds reach 12 ms^{-1} and drop off by 6 ms^{-1} aloft for a profile to be considered a LLJ. Later, Whiteman et al. (1997) supplements Bonner's categories with a weaker jet category (LLJ-0) that requires winds to reach 10 ms^{-1} (and drop off by 5 ms^{-1} aloft). Table 4.1 summarizes these criteria. Note that the categories overlap, meaning that a profile which satisfies the strict criteria for LLJ-3 would also satisfy the easier criteria for LLJ-0. Wind direction is irrelevant for identification purposes. Using these categories and the data described above, Bonner finds, regardless of the LLJ category considered, a pronounced maximum of jet frequency between 95° and 100° W, with a maximum within this zone located along the Oklahoma-Kansas border (the site of the present Great Plains ARM-CART site near Lamont, Oklahoma). The zone of maximum frequency is roughly parallel to the Rocky Mountains; LLJ frequencies drop off dramatically east of the Mississippi. Within this zone, 55-60 percent of all jet observations occur during summer months (April through September). The summer jets are overwhelmingly southerly.

Later, Whiteman et al. (1997) update Bonner's results with a study of two years of 4-times-daily research-quality soundings released from the Great Plains ARM-CART site. Table 4.2 shows the percentage of jets found by these two studies; Tables 4.3 and 4.4 show the same statistics for this study. The percentage of jet occurrences found here is similar to those found by Whiteman et al. and by Bonner for all jet categories

Table 4.1: Classification Categories for LLJ structures, as defined by Bonner (1968) and expanded by Whiteman et al. (1997)

Category	V_{max} ms ⁻¹	ΔV ms ⁻¹
LLJ-0	≥ 10	≥ 5
LLJ-1	≥ 12	≥ 6
LLJ-2	≥ 16	≥ 8
LLJ-3	≥ 20	≥ 10
no LLJ	-	-

in both the warm season (CASES-97) and the cold (CASES-99). The more detailed results found by Bonner (1968) and Whiteman et al. (1997) will be summarized along with my results for better comparison in the following sections.

I recognize that these categories are, to a certain extent, arbitrary: it would be more useful to categorize jets based on how supergeostrophic they are, for example. I am using these categories primarily as a way of showing that results with boundary-layer profilers are similar to those using data from radiosondes.

4.2.1 Diurnal variability of LLJs

In accordance with Blackadar's theory, previous studies (Bonner, 1968; Mitchell et al., 1995; Whiteman et al., 1997) find that LLJs are more frequent at night than during the day. Bonner finds that over most of the central (and eastern) United States,

Table 4.2: LLJ Frequencies (%) for the Bonner (1968) and Whiteman et al. (1997) studies

Category	Whiteman (1997) warm	Whiteman (1997) cold	Whiteman (1997) all	Bonner (1968)
LLJ-0 Jet	47.1	44.9	46.0	-
LLJ-1 Jet	34.4	34.6	34.5	28.0
LLJ-2 Jet	15.8	20.5	18.1	12.0
LLJ-3 Jet	6.7	10.7	8.7	4.0
No Jet	52.9	55.1	54.0	72.0

Table 4.3: % LLJ Classification of all good profiles

Category	CASES-97	BEA-97	WHI-97	OXF-97
LLJ-0 Jet	34.6	36.3	35.0	32.4
LLJ-1 Jet	28.4	31.2	28.2	25.8
LLJ-2 Jet	18.6	21.5	16.4	17.8
LLJ-3 Jet	10.5	14.8	6.95	9.93
No Jet	65.3	63.6	64.9	67.5

Table 4.4: % LLJ Classification of all good profiles

Category	CASES-99	BEA-99	WHI-99	OXF-99	ELD-99	SMI-99
LLJ-0 Jet	34.6	35.4	30.5	32.8	32.4	42.4
LLJ-1 Jet	29.3	30.8	26.1	27.5	25.9	36.7
LLJ-2 Jet	18.8	19.6	18.5	15.7	13.2	27.4
LLJ-3 Jet	9.24	7.82	8.56	6.76	5.85	17.7
No Jet	65.3	64.5	69.4	67.1	67.5	57.5

strong low level wind maxima are much more likely to occur at 1200 UTC (0600 CST) than at 0000 UTC (1800 CST), indicating a nocturnal mechanism for the jet. He finds the strength of the diurnal effect greater in summer than in winter, though that diurnal effect is obvious in both summer and winter jets. The Whiteman et al. (1997) study verifies that the nocturnal preference is enhanced during the warm season.

Both spring (CASES-97) and fall (CASES-99) boundary-layer profiler data confirm the diurnal variation of LLJ occurrence, as shown in Figures 4.1 and 4.2. All stations and all jet strengths show increased occurrences during the tenure of the stable boundary layer, from 0000 - 1500 UTC (1800 - 0900 LST), in excellent agreement with the previous studies. The present comparison between May and October data do not support the warm season nocturnal preference, even though both Bonner and Whiteman considered October to be part of the cold season. The CASES-97 (warm) and CASES-99 (cold) data are quite similar. As this present study is based only on two months of data in the shoulders of the warm and cold seasons, the slight disagreement with previous results should not be considered noteworthy.

The CASES tallies of LLJ frequency could be influenced by the common data outages during the night, which occur in both the CASES-97 and CASES-99 datasets (shown at the bottom of Figures 4.1 and 4.2). Many profiles at these times are rejected from LLJ consideration because the SAM algorithm (see Chapter 2) identified over 75 percent of data points within these profiles as bird-contaminated. To explore this possibility, we calculate correlations between the time series of data outages and the time series of LLJ frequency for each station. Tables 4.5 and 4.6 show how the correlation coefficients vary. We would expect large anticorrelation (values less than -0.5), indicating that when there are no data, there are also no jets. However, large positive values indicate that high LLJ frequencies occur during times that tend to have few data points: those jet occurrences are suspect because of the small number of profiles that are able to be considered. Only for the WHI-97 and SMI-99 datasets are the correlation values

Table 4.5: Correlation between data outages and llj frequency

Category	BEA-97	WHI-97	OXF-97
LLJ-0	-0.651551	0.716010	-0.628933
LLJ-1	-0.697706	0.755026	-0.533698
LLJ-2	-0.598036	0.792337	-0.504123
LLJ-3	-0.459270	0.692243	-0.344317
No Jet	0.651551	-0.716010	0.628933

large and positive. Although the results from these two data subsets could be considered suspicious due to this correlation, the other six datasets still reveal the nocturnal preference for jet activity indicated in the Bonner and Whiteman studies.

4.2.2 Variation of LLJ height with time of day

Blackadar’s theory of jetogenesis, supported by the analytical models of Thorpe and Guymer (1977) and Nieuwstadt (1984) and large-eddy simulations of Kosović and Curry (2000), would imply that as the height of the nocturnal surface inversion grows over the night before it reaches a quasi-steady state after an inertial period, the height of the LLJ (defined by the height of the wind speed maximum) should rise as well. Bonner’s 1968 study finds a correlation of 0.53 between the height of the jet at 1200 UTC (0600 LST) and the height of the thermal inversion. (Recall that his data are only available at the surface, 150, 300, 500, 1000, 1500, 2000, 2500, and 3000 m.) Whiteman et al. also explore the relationship between the jet speed maximum and the height of the temperature inversion for nocturnal jets. (They define the temperature inversion

Table 4.6: Correlation between data outages and llj frequency

Category	BEA-99	WHI-99	OXF-99	ELD-99	SMI-99
LLJ-0	-0.0549388	0.123821	-0.522328	0.227576	0.794472
LLJ-1	-0.136276	0.292829	-0.551113	0.222389	0.710084
LLJ-2	-0.235596	0.295055	-0.443562	0.367506	0.727051
LLJ-3	-0.233178	0.162916	-0.533757	0.326948	0.569029
No Jet	0.0549388	-0.123821	0.522328	-0.227576	-0.794472

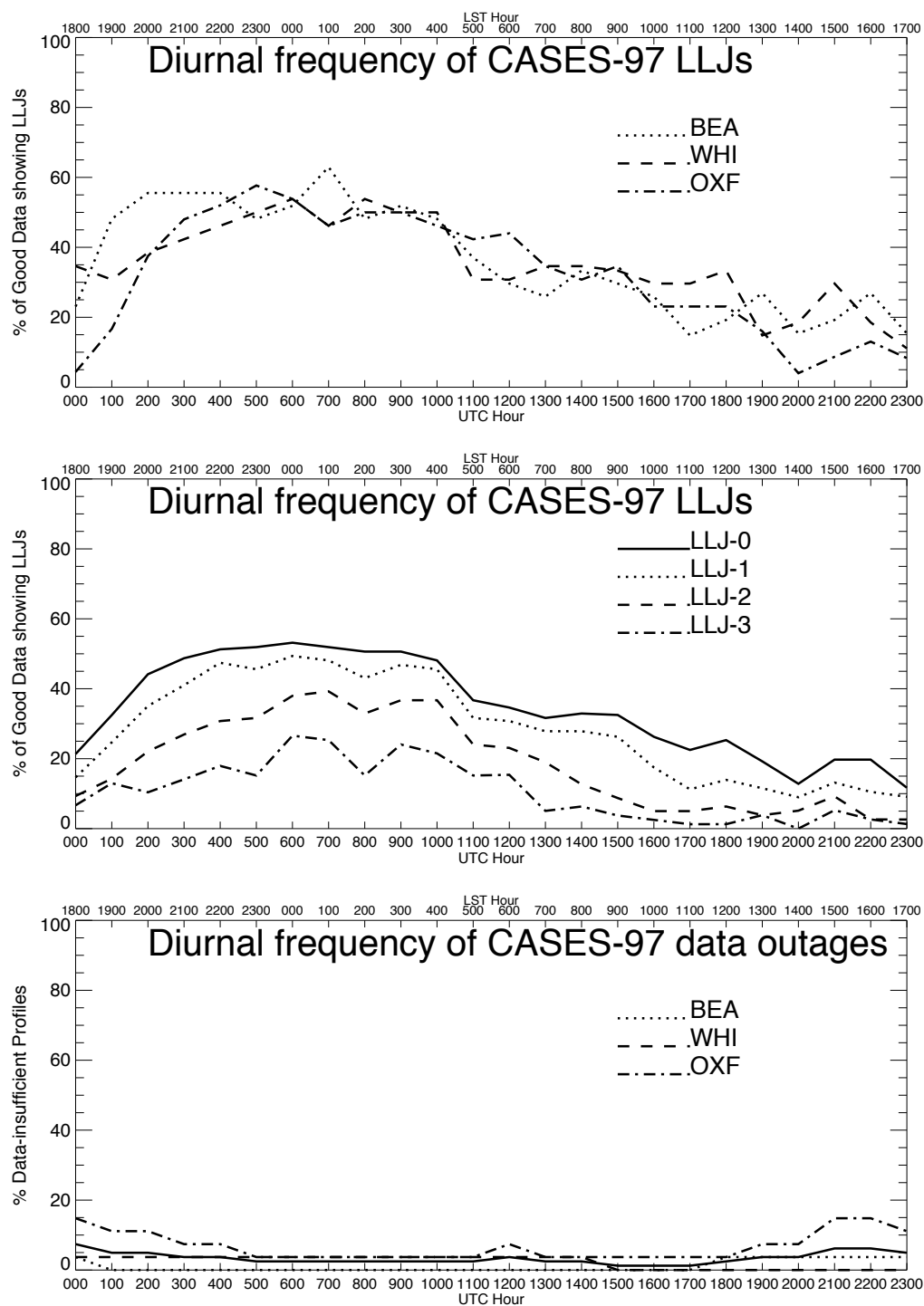


Figure 4.1: Diurnal frequency of CASES-97 LLJs. The top figure shows variation between the three boundary-layer profilers. The middle figure shows how the LLJ strengths vary with time of day, and the bottom figure shows data outages as a function of time of day.

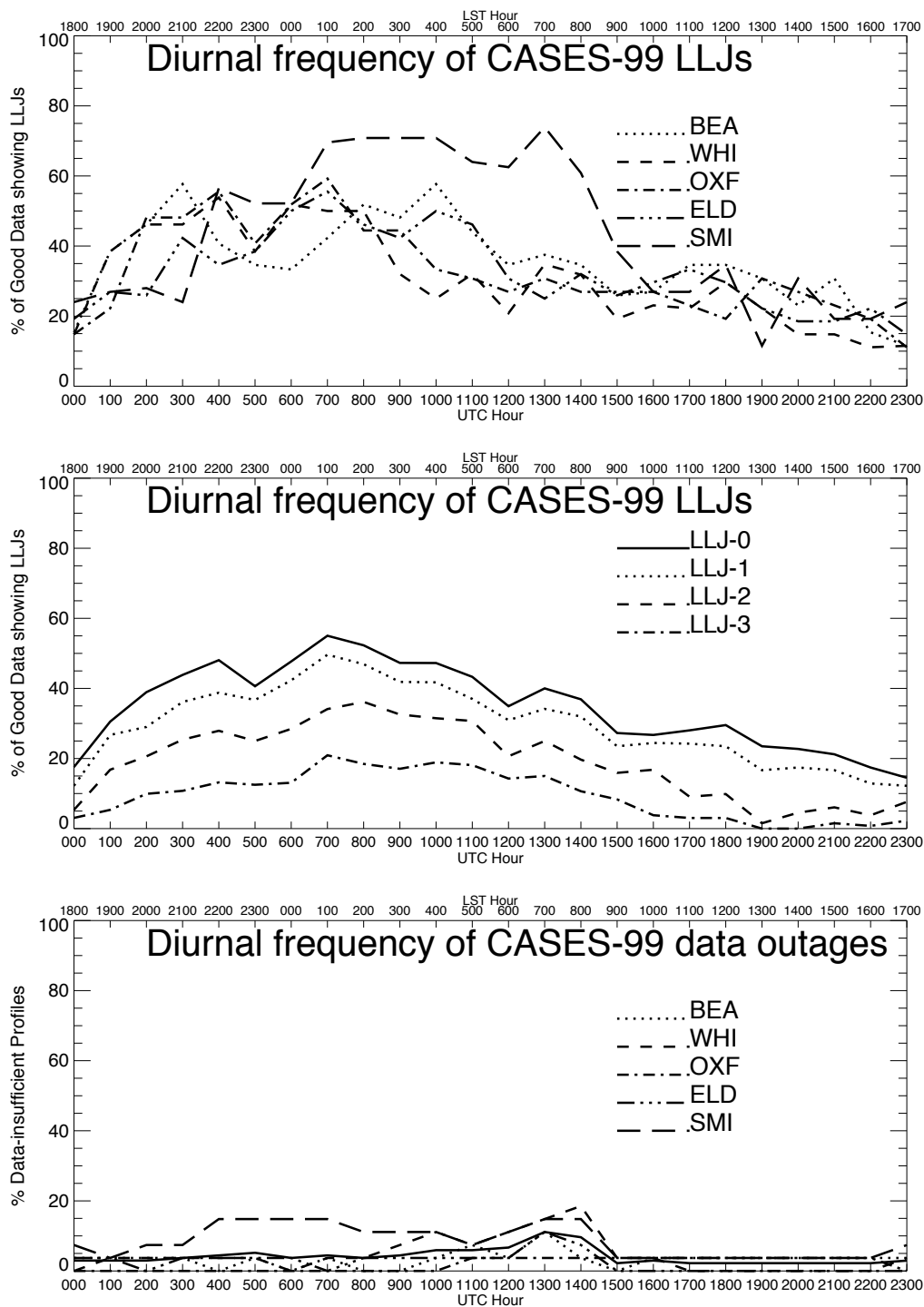


Figure 4.2: Diurnal frequency of CASES-99 LLJs. The top figure shows variation between the five boundary-layer profilers. The middle figure shows how the LLJ strengths vary with time of day, and the bottom figure shows data outages as a function of time of day.

to be the height of the temperature maximum between the surface and 3000 m AGL.) Though they do not distinguish between surface-based and elevated inversions, Whiteman et al. find that the jet heights tended to be above the height of the temperature inversions, with considerable variability. They also reveal that the median height of the jet maximum (452 - 456 m) varies little over the course of the night, while the median temperature inversion height grew from 137 to 373 m AGL.

As the present study relies on wind profiler data, we cannot directly test the relationship of inversion-top height and height of the wind speed maximum. However, we can inspect the diurnal variability of the height of the wind speed maximum knowing that the inversion is low at the beginning of the night and tends to grow through the night. Figures 4.3 - 4.4 show the mean and standard deviations for the height of the wind speed maximum at each time of day over the entire dataset for each station and each LLJ category. For both warm (CASES-97) and cold (CASES-99) conditions, there is a clear distinction between nocturnal and daytime conditions. During the night, the average jet height is lower, around 600 m. The advent of daytime brings a large increase in the height of a wind speed maximum, often up to 1200 m. However, there is no clear increase of height of the nocturnal jet as the night progresses (and in fact the mean level often decreases over the night instead of rising with the inversion as Blackadar's theory would dictate). The large values of the standard deviations of the height of the jet maximum indicate wide variation between nights. Additionally, the median jet heights tend to be lower than the means during the nocturnal hours, implying that a few outlying elevated jets influence the average.

4.2.3 Wind Speed and Wind Direction at jet max

Figures 4.5 and 4.6 show scatter polar plots of the wind speed and wind direction at the jet maximum for each inclusive jet category. Both warm and cold season show a strong preference for jets from the SW, in agreement with Whiteman's study, which

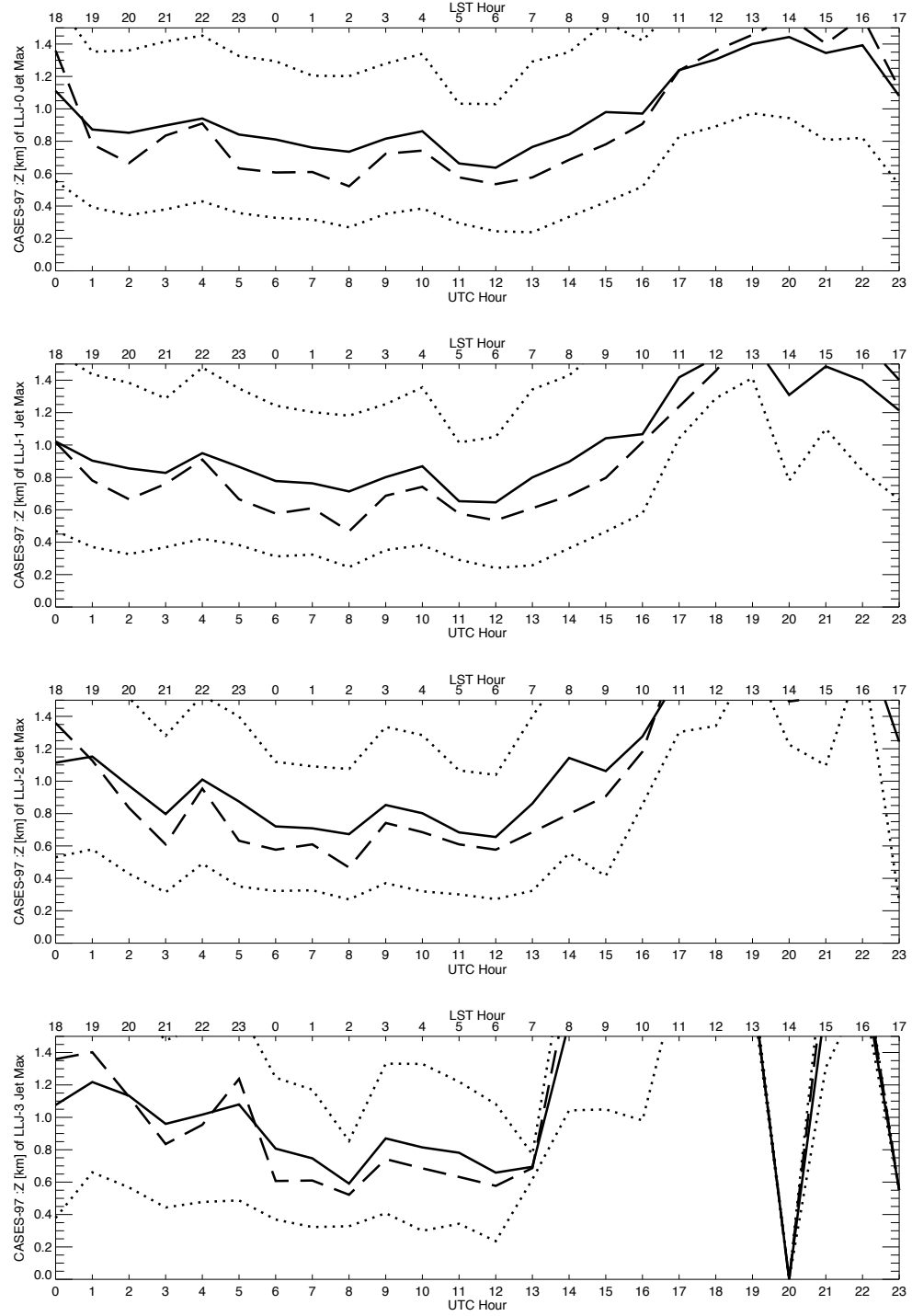


Figure 4.3: The mean (solid line), the median (dashed line) and the mean plus/minus standard deviation (dotted line) of the height of the wind speed maximum for the different LLJ categories for CASES-97. Top left, LLJ-0; top right, LLJ-1; bottom left, LLJ-2; bottom right, LLJ-3. The mean and median heights of the jet both show a decrease in jet height during the tenure of the nocturnal boundary layer. The standard deviations show that there is a wide range between each day in this dataset.

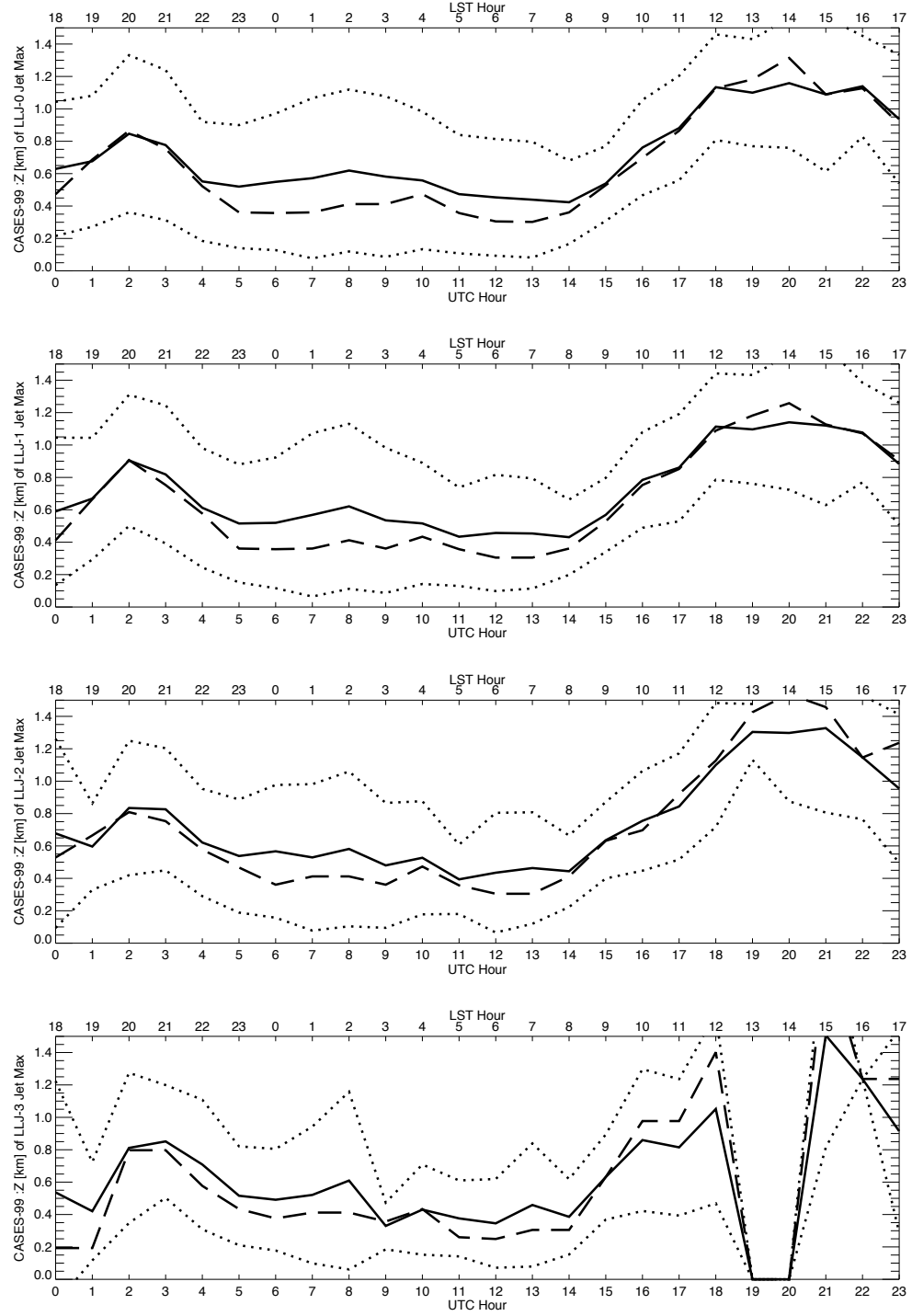


Figure 4.4: The mean (solid line), the median (dashed line) and the mean plus/minus standard deviation (dotted line) of the height of the wind speed maximum for the different LLJ categories for CASES-99. Top left, LLJ-0; top right, LLJ-1; bottom left, LLJ-2; bottom right, LLJ-3. The mean and median heights of the jet both show a decrease in jet height during the tenure of the nocturnal boundary layer. The standard deviations show that there is a wide range between each day in this dataset.

found that that warm season jets were confined to mostly southerly or southwesterly directions. Bonner's study also indicates a preponderance of southerly wind maxima. The CASES-99 dataset does show more northeasterly jets than revealed in either Bonner's or in Whiteman's study.

4.3 Correlation of LLJs and inertial oscillations

Constant amplitude inertial oscillations frequently appear at the level of the jet maximum (Sumner et al., 1983). Tables 4.7 and 4.8 show how often inertial oscillations, at any phase, are observable at the LLJ wind speed maximum. (The inertial oscillations are identified with the technique described in Chapter 3 and the LLJs are identified with Bonner's technique as described above.) Blackadar's theory indicates that inertial oscillations should be responsible for nocturnal LLJs and therefore observable at the height of the LLJ wind speed maximum during nocturnal LLJ cases but not at other times of day. Figures 4.7 and 4.8 show that at night, inertial-oscillation-influenced jets become much more numerous than during the day; a strong diurnal cycle is obvious. Figures 4.9 and 4.10 show that LLJs not associated with inertial oscillations have no such diurnal cycle. To determine the significance of this diurnal variation, a much larger dataset should be utilized.

A slight correlation is also apparent between occurrences of inertial oscillations at the jet maximum and the intensity of the jet. For most stations (WHI-97, BEA-99, WHI-99, OXF-99, ELD-99), inertial oscillations are less common as the jet intensity increases. This attrition is likely due to the fact that very strong jets are influenced by a number of mechanisms in addition to the inertial oscillation. These other influences include synoptic-scale motions (i.e. a jet leading a front) and the diurnally forced thermal wind due to the sloping Great Plains. Figures 4.9 and 4.10 show no diurnal variability in the number of LLJs that are not influenced by inertial oscillations. This behavior indicates that synoptic-scale motions or some other forcing mechanism independent of the diurnal

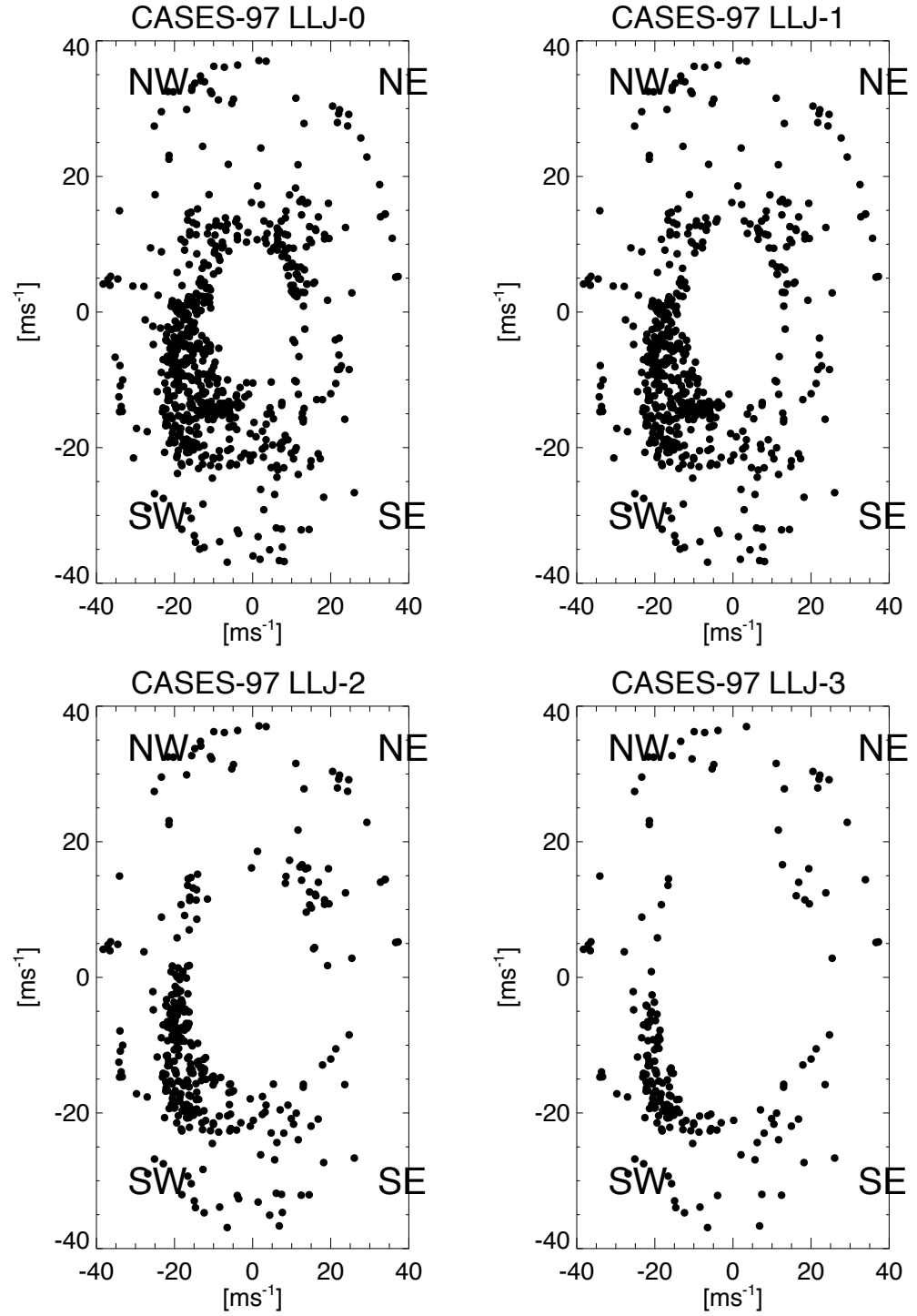


Figure 4.5: Polar plots of wind speeds and wind directions at the height of the wind speed maximum for the four LLJ categories for CASES-97.

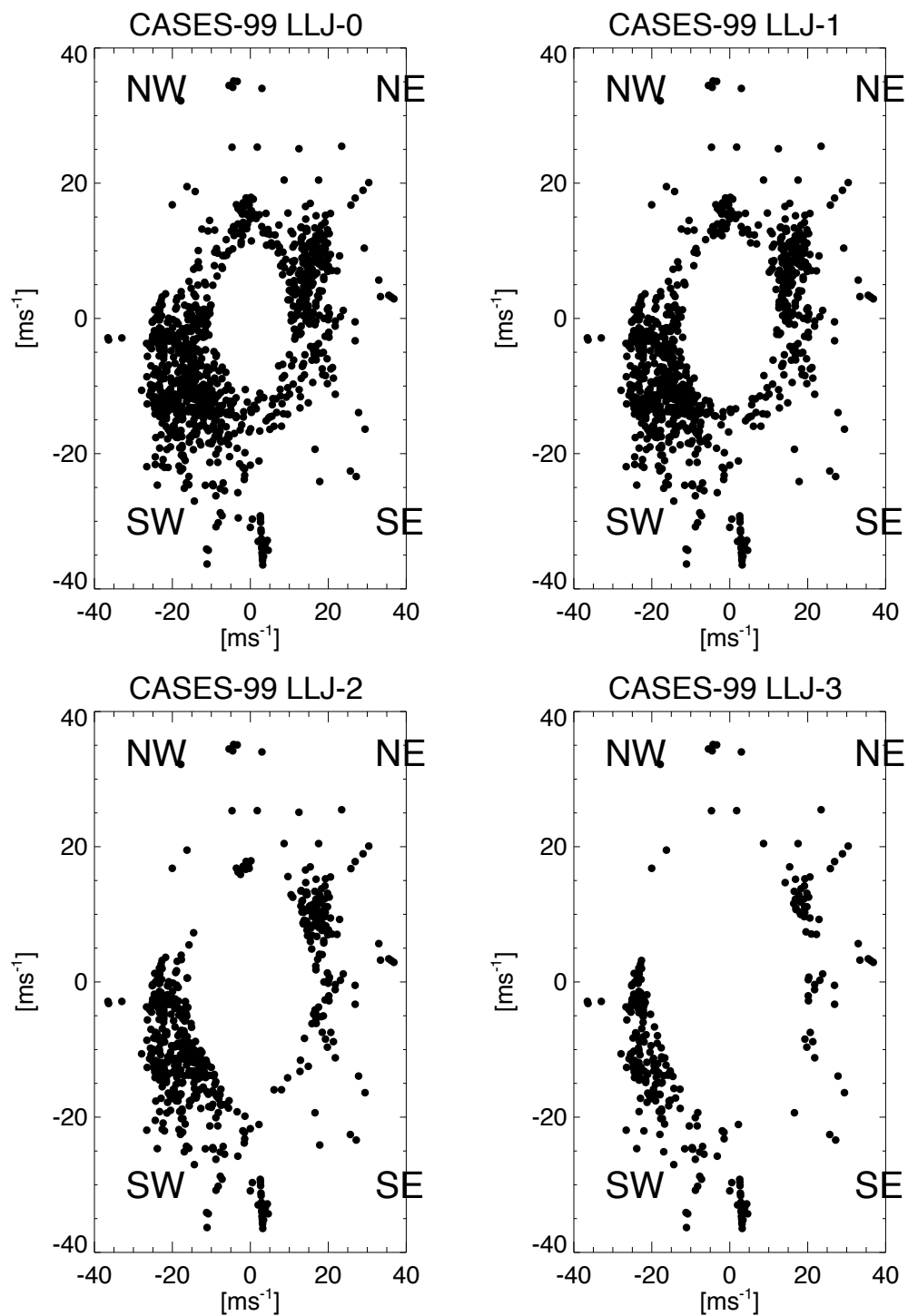


Figure 4.6: Polar plots of wind speeds and wind directions at the height of the wind speed maximum for the four LLJ categories for CASES-99.

Table 4.7: Frequency of IO occurrence during LLJ events at the height of the wind speed maximum, using CASES-97 (with respect to number of LLJ events)

Category	CASES-97	BEA-97	WHI-97	OXF-97
LLJ-0 Jet	82.7	81.9	80.6	86.2
LLJ-1 Jet	83.1	83.0	78.2	89.1
LLJ-2 Jet	83.7	84.7	75.9	89.8
LLJ-3 Jet	83.9	84.2	75.0	90.0

variation of the boundary layer modify these LLJs.

4.4 Summary of LLJ Climatology

This section discusses the results of a climatology of low-level jet structures observed in the CASES-97 and CASES-99 boundary-layer wind profiler data using the jet identification technique developed by Bonner (1968) and expanded by Whiteman et al. (1997). The general characteristics of this climatology are similar to those of previous climatologies based on data from different instruments. Low-level jet structures with maximum wind speeds of at least 10 ms^{-1} occur in approximately 35 percent of the hourly profiles with sufficient data to examine. LLJs identified with the Bonner (1968) method have a strong diurnal dependence in both spring and fall, with 50 percent of the nocturnal profiles exhibiting jet structures and 30 percent of the daytime profiles showing jets. The height of the wind speed maximum at night is lower than during the day, although that height does not rise over the night as one would expect if the

Table 4.8: Frequency of IO occurrence during LLJ events at the height of the wind speed maximum, using CASES-99 (with respect to number of LLJ events)

Category	CASES-99	BEA-99	WHI-99	OXF-99	ELD-99	SMI-99
LLJ-0 Jet	70.1	71.6	75.1	70.3	61.9	71.4
LLJ-1 Jet	69.7	72.5	74.0	69.1	60.9	71.1
LLJ-2 Jet	67.5	67.4	70.4	62.0	58.3	73.6
LLJ-3 Jet	65.5	65.3	62.2	55.8	43.2	79.0

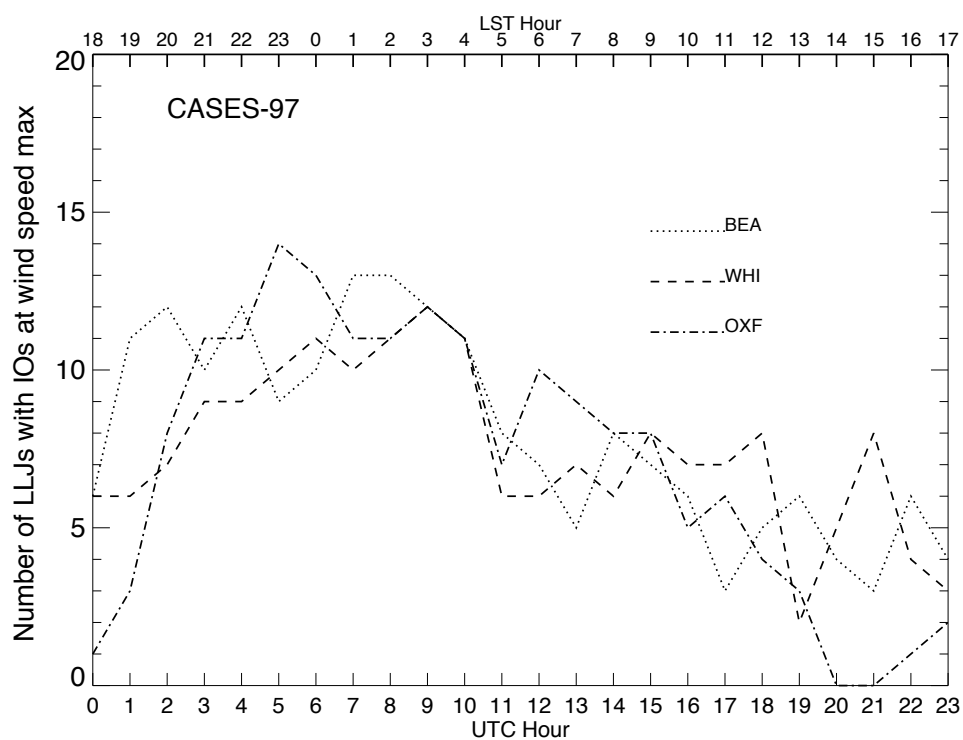


Figure 4.7: The diurnal variation of LLJs with inertial oscillations apparent within the wind speed maximum for CASES-97 wind profiler data.

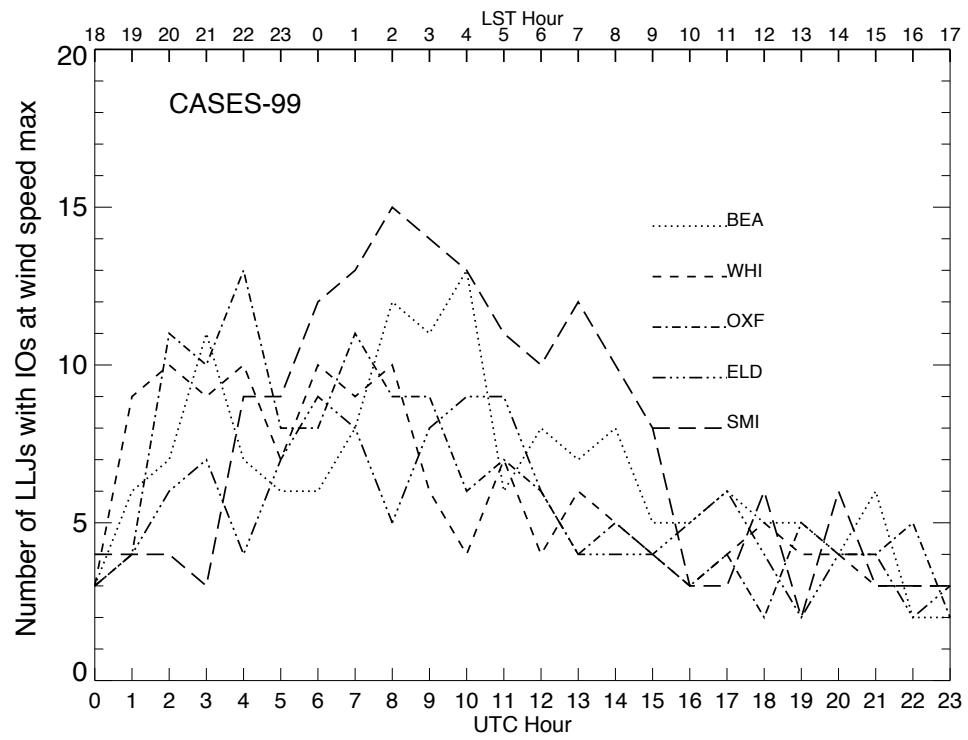


Figure 4.8: The diurnal variation of LLJs with inertial oscillations apparent within the wind speed maximum for CASES-99 wind profiler data.

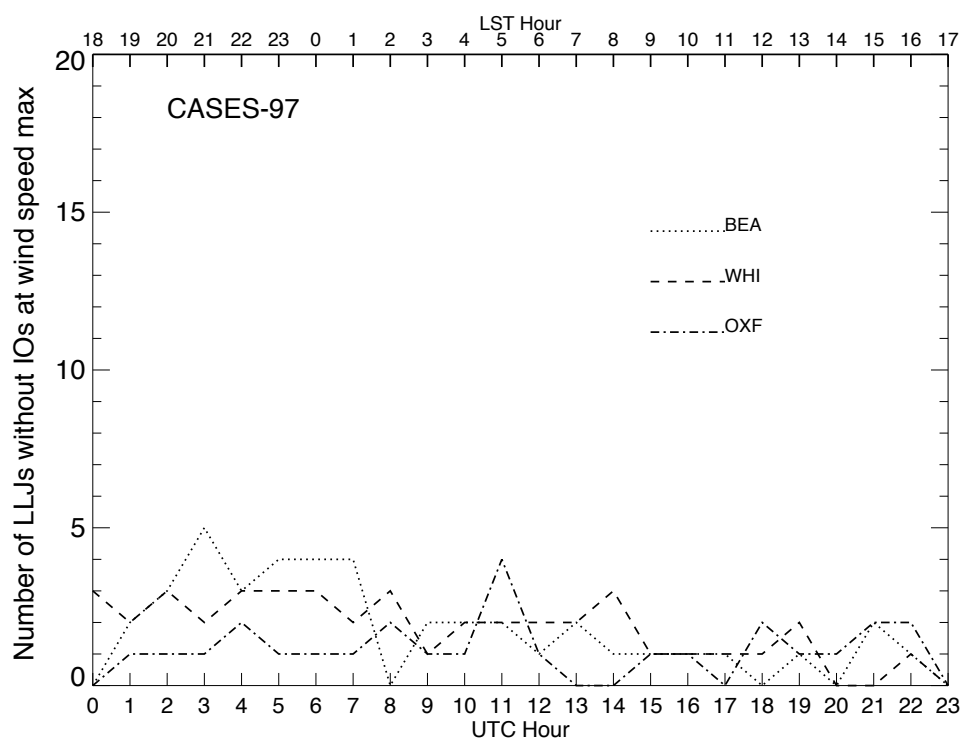


Figure 4.9: The diurnal variation of LLJs without inertial oscillations apparent within the wind speed maximum for CASES-97 wind profiler data.

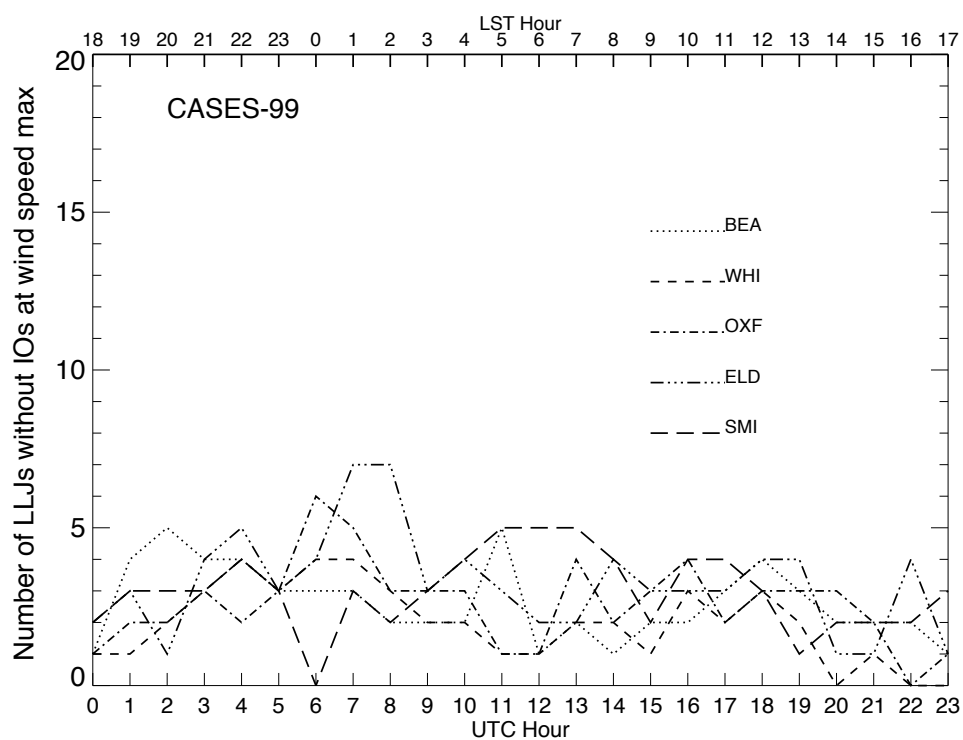


Figure 4.10: The diurnal variation of LLJs without inertial oscillations apparent within the wind speed maximum for CASES-99 wind profiler data.

jet existed atop the nocturnal surface, which rises over the night. These results are consistent with those found by Bonner (1968) and Whiteman et al. (1997) (though only Whiteman explored the correlation between the height of the jet maximum and the inversion as Bonner used only one sounding per night).

I expanded the traditional LLJ climatology by exploring how often inertial oscillations occur at the height of the jet wind speed maximum. Based on the inertial oscillation test developed in Chapter 3, nocturnal jets tend to have inertial oscillations at the height of the wind speed maximum, but jet structures during the daytime show inertial oscillations much less frequently. Correlation does not imply causation, of course, and more research is required to delineate which jets are caused by constant-amplitude inertial oscillations that develop due to evening decoupling above the nocturnal surface inversion.

Chapter 5

Inertial oscillations in a first-order 1D numerical model

Previous chapters have discussed observations of inertial oscillations that commence when boundary layer convection decays due to the reversal of the surface heat flux and the development of the surface inversion, as proposed by Blackadar (1957) and discussed earlier in this thesis. When convective turbulence decays due to the cooling surface, the vertical transfer of horizontal momentum starts to decrease. A temperature inversion, caused by the cooling surface, establishes itself. The ensuing decoupling or stratification of the boundary layer causes the acceleration of boundary layer winds into the gentle veering of the inertial oscillation (Blackadar, 1957; Thorpe and Guymer, 1977).

Despite the attractiveness of this theory (Stull, 1988; Garratt, 1992), the generation of inertial oscillations during the actual decay process has not been documented. Previous studies of inertial oscillations (Singh et al., 1993; Tan and Farahani, 1998) assume that inertial oscillations are generated at the evening transition without verifying that assumption either observationally or numerically. I seek to verify that the inertial oscillations I observe on 11-12 October 1999 (discussed in Chapter 3) could have been caused by the decoupling due to evening transition, and then to identify the controlling mechanism for that generation process. Towards this end, I develop a 1-D numerical model to simulate the evening decay process. The simplicity of the model allows only one parameter, the turbulent exchange coefficient k_m , to control the decay of turbulence.

5.1 Model Equations

To simulate the decay of convective turbulence and the generation of inertial oscillations, a simple first-order turbulence closure model was developed following Sorbjan (1989). The mean equations for the horizontally homogeneous atmospheric boundary layer are given as

$$\begin{aligned}\frac{\partial u}{\partial t} &= f(v - v_g) - \frac{\partial}{\partial z} (\overline{u'w'}) \\ \frac{\partial v}{\partial t} &= -f(u - u_g) - \frac{\partial}{\partial z} (\overline{v'w'}),\end{aligned}\tag{5.1}$$

where z is height, t is time, f is the Coriolis parameter, u and v are the zonal and meridional component of the wind field, u_g and v_g are the components of the geostrophic wind, and $(\overline{u'w'})$ and $(\overline{v'w'})$ are the turbulent momentum fluxes. The thermodynamic equation is given by

$$\frac{\partial \theta}{\partial t} = \frac{1}{\rho c_p} \frac{\partial R_n}{\partial z} - \frac{\partial}{\partial z} (\overline{w'\theta'}),\tag{5.2}$$

where θ is potential temperature, ρ is the density of air, c_p is the specific heat at constant pressure for air, R_n is net radiation, and $(\overline{w'\theta'})$ is the sensible turbulent heat flux.

To close these equations for modeling purposes, an expression for each of the turbulent fluxes is required. However, just as the mean quantities u , v , and θ depend on turbulent terms, each turbulent term is itself a function of the next higher moment term. The total statistical description of turbulence demands an infinite set of coupled equations that describe progressively smaller and smaller eddies. To avoid this infinite cascade, known as the closure problem (first identified by Keller and Friedmann in 1924 (Stull, 1988), numerous parameterizations have been proposed.

For so-called “first-order” closure or “K-closure”, as used in this model, equations are retained for the mean variables, and the turbulence fluxes are parameterized as functions of the mean variables:

$$(\overline{u'w'}) = -k_m \frac{\partial u}{\partial z}$$

$$\begin{aligned} \overline{(v'w')} &= -k_m \frac{\partial v}{\partial z} \\ \overline{(w'\theta')} &= -k_\theta \frac{\partial \theta}{\partial z}, \end{aligned} \quad (5.3)$$

where k_m and k_θ are respectively known as the turbulent exchange coefficients (or “eddy viscosity”) for momentum and heat. (This approach is based on Boussinesq’s 1877 hypothesis, that turbulent shearing stress can be related to the rate of mean strain (Sorbjan, 1989).) Some attention has been focused on whether or not to assume that

$$k_m = k_\theta, \quad (5.4)$$

but Estournel and Guedalia (1987) show that parameterizing them differently has little impact on first-order-closure model results for the stable boundary layer. Expressions for k_m are numerous and will be addressed below. I adopt first-order closure because I am most interested in global behavior, mean wind and mean temperature profiles.

Once first-order closure has been adopted, the momentum equations are rewritten in complex form

$$\frac{\partial W}{\partial t} = \frac{\partial}{\partial z} k_m \frac{\partial W}{\partial z} + S \quad (5.5)$$

where

$$\begin{aligned} W(z, t) &= [(u + iv) - (u_{g0} + iv_{g0})] e^{ift}, \\ S(z, t) &= if\Lambda z e^{ift}, \\ \Lambda(z, t) &= \frac{\partial u_g}{\partial z} + i \frac{\partial v_g}{\partial z}. \end{aligned} \quad (5.6)$$

Within these equations, u_{g0} and v_{g0} are the components of the geostrophic wind at the surface, assumed to be constant in time. Unless k_m couples this momentum equation, 5.5, and the thermodynamic equation (5.2), winds in this model are forced only by pressure gradients (through the geostrophic wind), the Coriolis force (represented through the Coriolis parameter f), and the variation of u_g and v_g with height through the thermal wind Λ .

To solve (5.5) numerically, all derivatives are approximated by finite differences:

$$\begin{aligned} \frac{W_j^{k+1} - W_j^k}{\Delta t} &= (k_m)_{j+1/2}^{k+1} \frac{W_{j+1}^{k+1} - W_j^{k+1}}{\Delta z^2} \\ &\quad - (k_m)_{j-1/2}^{k+1} \frac{W_j^{k+1} - W_{j-1}^{k+1}}{\Delta z^2} + S_j^{k+1}, \end{aligned} \quad (5.7)$$

where $z = z_j = j\Delta z$, $t = t^k = k\Delta t$, $j = 0, 1, \dots, n$, $k = 0, 1, \dots$, and $W(z, t) = W(z_j, t^k) = W_j^k$. This system of equations in time ($k = 0, 1, \dots, n$) can be solved by rewriting each time step as

$$\begin{aligned} A_0 X_1 + B_0 X_0 &= F_0 \quad \text{for } j = 0, \\ A_j X_{j+1} + B_j X_j + C_j X_{j-1} &= F_j \quad \text{for } j = 1, 2, \dots, n-1, \\ B_n X_n + C_n X_{n-1} &= F_n \quad \text{for } j = n, \end{aligned} \quad (5.8)$$

where $X_j = W_j^{k+1}$ and the coefficients are given by

$$A_j = -\frac{(k_m)_{j+1/2}^{k+1}}{\Delta z^2}, \quad (5.9)$$

$$B_j = \frac{1}{\Delta t} + \frac{(k_m)_{j+1/2}^{k+1}}{\Delta z^2} + \frac{(k_m)_{j-1/2}^{k+1}}{\Delta z^2}, \quad (5.10)$$

$$C_j = -\frac{(k_m)_{j-1/2}^{k+1}}{\Delta z^2}, \quad (5.11)$$

$$F_j = +\frac{1}{\Delta t} W_j^k + S_j^{k+1}. \quad (5.12)$$

We can assume that the solution at each timestep $k+1$ for the $n+1$ equations given in 5.8 takes the form

$$X_{j-1} = \alpha_j X_j + \beta_j \quad \text{for } j = n, n-1, \dots, 1 \quad (5.13)$$

Substitution of 5.13 into 5.8 gives

$$X_j = -\frac{A_j}{B_j + C_j \alpha_j} X_{j+1} + \frac{F_j - C_j \beta_j}{B_j + C_j \alpha_j} \quad \text{for } j = 1, 2, \dots, n-1 \quad (5.14)$$

which leads to

$$\begin{aligned} \alpha_{j+1} &= -\frac{A_j}{B_j + C_j \alpha_j} \\ \beta_{j+1} &= \frac{F_j - C_j \beta_j}{B_j + C_j \alpha_j}. \end{aligned} \quad (5.15)$$

The boundary conditions in 5.8 give

$$\alpha_1 = -\frac{A_0}{B_0} \quad \text{and} \quad \beta_1 = \frac{F_0}{B_0} \quad (5.16)$$

and an expression for X_n , the value at the top of the domain,

$$X_n = \frac{F_n - C_n \beta_n}{B_n + C_n \alpha_n}. \quad (5.17)$$

Equations 5.16 and 5.17 can be replaced with the known boundary conditions at the top and bottom of the domain. Winds at the top of the domain are set equal to the geostrophic wind; winds at the surface are zero. Potential temperature at the top is set equal to the initial value at that level, while potential temperature at the lowest level of the computational domain is taken from 5-minute averages of potential temperature observed at 2m.

By computing X_n from α and β , I am applying Gauss' elimination method, which is accurate and convergent for all positively definite arrays of moderate size, such as the arrays used here. (For all runs discussed here, $j = 10$ m and $k = 300$ seconds.) An iterative procedure is implemented, where, at each time step, t^{k+1} , the coefficients α_j and β_j are calculated from the A_j, B_j, C_j , and F_j values (functions of the wind speed at the previous time step k , the thermal wind forcing at $t = k + 1$, eddy viscosity at $t = k$, and space and time grid sizes) of the gridpoint just below. After α and β are calculated at all heights for time $k + 1$, X_n , the value at the top of the domain at time step t^{k+1} is calculated from 5.17. Moving down through the space domain, the remaining X_j values are calculated with 5.14.

The numerical scheme given above (5.7 through 5.17) also applies to a solution for 5.2, with the modifications that F_j and S_j would be replaced with

$$\begin{aligned} F_{\theta j} &= \frac{1}{\Delta t} \theta_j^k + S_j^{k+1} \\ S_{\theta j} &= -\frac{1}{\rho C_p} \frac{\partial R_n}{\partial z}, \end{aligned} \quad (5.18)$$

where R_n is the net longwave radiation, ρ is density, and c_p is the specific heat. The longwave cooling term is calculated using a narrow-band radiation model described in Section 5.2 below. Short wave radiation is not included as the model is initiated after solar radiation has ceased heating the surface.

5.2 Parameterization of longwave radiative cooling

Primary absorbers of both shortwave and longwave radiation in our atmosphere include water vapor, condensed water and ice particles in clouds, carbon dioxide, the water vapor dimer, ozone and solid particulates. Longwave radiative cooling becomes largest at or just beneath the top of a cloud (Garratt, 1992).

Although longwave radiative fluxes are swamped by turbulent heat fluxes in the convective mixed layer, the effect of longwave radiation on the evolution of the nocturnal boundary layer may not be neglected even when clouds are not present. When the inversion layer is considered as a whole, both observations and numeric models have shown (André and Mahrt, 1982) that turbulent heat flux divergence and clear-air radiative cooling make comparable contributions to the heat budget of the layer. Radiative cooling dominates the cooling of the layers closest to the surface and closest to the top of the nocturnal boundary layer, as shown by both Garratt and Brost (1981) and Tjemkes and Duynkerke (1989). Additionally, Gopalakrishnan et al. (1998) show that in quiescent conditions, radiative effects dominate the entire boundary layer, causing the inversion layer to grow in depth throughout the night instead of reaching a steady-state condition, as it does when turbulent effects are strong. Therefore, longwave radiative cooling must be parameterized as exactly as possible while minimizing computational expenses.

Not surprisingly, the number of models of longwave radiative cooling in the nocturnal boundary layer is equal to or greater than the number of researchers in the field. Two types of models dominate the discussion: the “line-by-line”, “narrow band”,

Table 5.1: Atmospheric mixing ratios [g/kg] of chemical species used in the MDTERP radiation model

Pressure [mb]	CO ₂ [$\times 10^{-1}$]	O ₃	CH ₄ [$\times 10^{-4}$]	N ₂ O [$\times 10^{-4}$]
1013	5.0099999	5.0336301×10^{-5}	9.6700003	4.2500001
902	5.0099999	5.5657965×10^{-5}	9.6700003	4.2500001
802	5.0099999	6.1421699×10^{-5}	9.6700003	4.2500001
710	5.0099999	7.0095361×10^{-5}	9.6700003	4.2500001
628	5.0099999	7.9972444×10^{-5}	9.6700003	4.2500001
554	5.0099999	9.1413684×10^{-5}	9.6700003	4.2500001
487	5.0099999	1.0620555×10^{-4}	9.6700003	4.2500001
426	5.0099999	1.2876237×10^{-4}	9.6700003	4.2500001
372	5.0099999	1.5133282×10^{-4}	9.6700003	4.2500001

and “k-distribution” models solve the radiative transfer equation exactly for each frequency in the longwave bands (Tjemkes and Duynkerke, 1989), while emissivity models (Sasamori (1972); Chen and Cotton (1983); Garratt et al. (1992); Delage (1988); Gopalakrishnan et al. (1998)) parameterize the variability of absorption coefficients in each band as a function of frequency. Emissivity-based models have been criticized for their inaccuracies of up to 20 W m² near the surface (Tjemkes and Duynkerke, 1989), and are therefore problematic for evening transition studies.

Because of the importance of longwave radiative cooling to the nocturnal boundary layer, a highly accurate narrow-band model (Ellingson and Gille, 1978) was utilized. This model assumes a plane-parallel atmosphere, local thermodynamic equilibrium, and a black surface. It considers the radiative effects of water vapor, carbon dioxide, ozone, methane, and nitrous oxide. The narrow bands are 305 spectral intervals from 0 cm⁻¹ to 3000⁻¹. For these calculations, the height levels were set at 10, 20, 30, 40, 50, 60, 70, 80, 90, 100, 110, 120, 130, 140, 150, 175, 200, 250, 300, 400, 500, 750, 1000, 1100, 1200, 1500, 1750, 2000, 2500, 3000, 3500, and 4000 m. Standard atmospheric values for carbon dioxide, ozone, methane, and nitrous oxide (Table 5.1) were used in these model runs, while water vapor concentrations were derived from soundings.

For verification, cooling rates were calculated for the Wangara Day 34, 0019 LST, a

case that has been simulated extensively in the literature (André and Mahrt, 1982), with excellent agreement. Figure 5.1 shows the water vapor mixing ratio and potential temperature profiles for a sample CASES-99 sounding, 5 October at 0310 UTC (2110 Local Standard Time).

5.3 Parameterizations of k_m , the turbulent exchange coefficient

This model relies on first-order closure, where turbulent fluxes are parameterized as a function of an eddy viscosity or turbulent exchange coefficient k_m and vertical gradients of the quantity under consideration:

$$\begin{aligned} \overline{u'w'} &= -k_m \frac{\partial u}{\partial z}, \\ \overline{v'w'} &= -k_m \frac{\partial v}{\partial z}, \\ \overline{w'\theta'} &= -k_\theta \frac{\partial \theta}{\partial z}, \end{aligned} \tag{5.19}$$

Estournel and Guedalia (1987) show that in first-order models of the stable boundary layer, $k_m = k_\theta$; this approximation is used in this model.

The simplest closure of (5.19) above allows k_m to be a constant value for all heights and times. However, as Derbyshire (1999) notes, it is generally acknowledged that, outside of a simple Ekman layer, any stable boundary layer modeling should allow turbulence to interact with local stability. Prescribing k_m as a constant forbids such interaction, and would in fact prevent any model from illustrating the evolution of the stable boundary layer when the turbulence term is significant compared to the other terms in (5.1) and (5.2). Observations show that appropriate values of the eddy diffusivity range from $100 \text{ m}^2\text{s}^{-1}$ during maximum convection in the convective boundary layer to less than $1 \text{ m}^2\text{s}^{-1}$ during the decreased turbulence of the stable boundary layer regime (Yasuda, 1988). In the stable boundary layer regime, Estournel and Guedalia (1987) find the eddy diffusivity coefficient to be of the order of $0.5 \text{ m}^2\text{s}^{-1}$.

With such wide variation of estimates of appropriate values for k_m in the stable

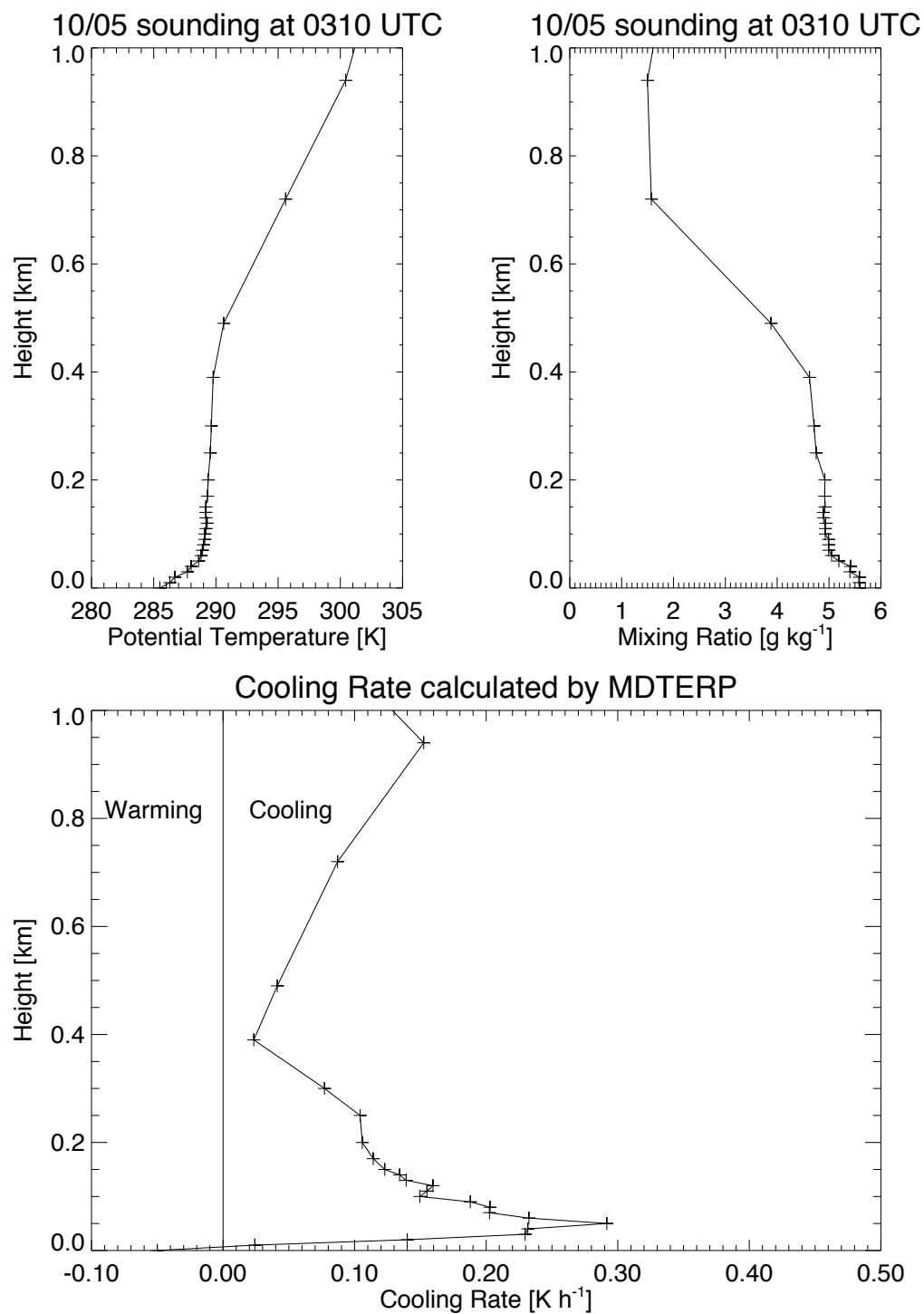


Figure 5.1: Cooling rate as calculated by the MDTERP narrow band model for a CASES-99 sounding, 5 Oct 1999, 0310 UTC

boundary layer and little information on how k_m would vary during the transition, it would be useful to know how k_m varies for this particular dataset.

The clear relationship between the eddy viscosity for potential temperature and the tendency equation for potential temperature, as expressed in equation 5.2 above, allows a determination what effective values of k_θ limit the behavior of potential temperature. Such an exercise in inverse modeling could provide an understanding of at least the range of expected values of k_θ . By referring to “effective” values of k_θ , we underscore the fact that advection, thermal wind, drainage currents, and all other process that influence the observations but are omitted from consideration in the model prevent a definitive calculation of k_θ .

We can rewrite the model equation 5.2 to express the eddy viscosity for potential temperature as a function of the observed potential temperatures θ and calculated longwave radiative cooling rate $(\frac{\partial\theta}{\partial t})_{LW}$:

$$k_{\theta j}^k = \frac{(\Delta z)^2}{\Delta t} \frac{(\theta_j^{k+1} - \theta_j^k - (\frac{\partial\theta}{\partial t})_{LW})}{(\theta_{j+1}^{k+1} - 2\theta_j^{k+1} + \theta_{j-1}^{k+1})} \quad (5.20)$$

where j is the height index and k is the time index. This expression assumes that turbulent heat exchange and longwave radiative cooling are the only processes relevant on this night.

We attempt this calculation using data from CASES-99 IOP 2 (5-6 October), a night with frequent (hourly - half hourly) and reliable soundings. The wind on this night is calm, dominated by a high pressure system. Skies are clear, so the radiative cooling should be adequately parameterized with our code.

From the 2200, 2330, 0000, 0030, 0100, 0200, and 0259 UTC soundings, potential temperature data are interpolated onto a regular grid ($\Delta z = 10$ m, $\Delta t = 300$ s) so equation 5.20 can be solved for k_θ . Contours of the gridded and interpolated potential temperature data are shown in Figure 5.2.

Figure 5.3 depicts the variation of diffusivity coefficients calculated in this man-

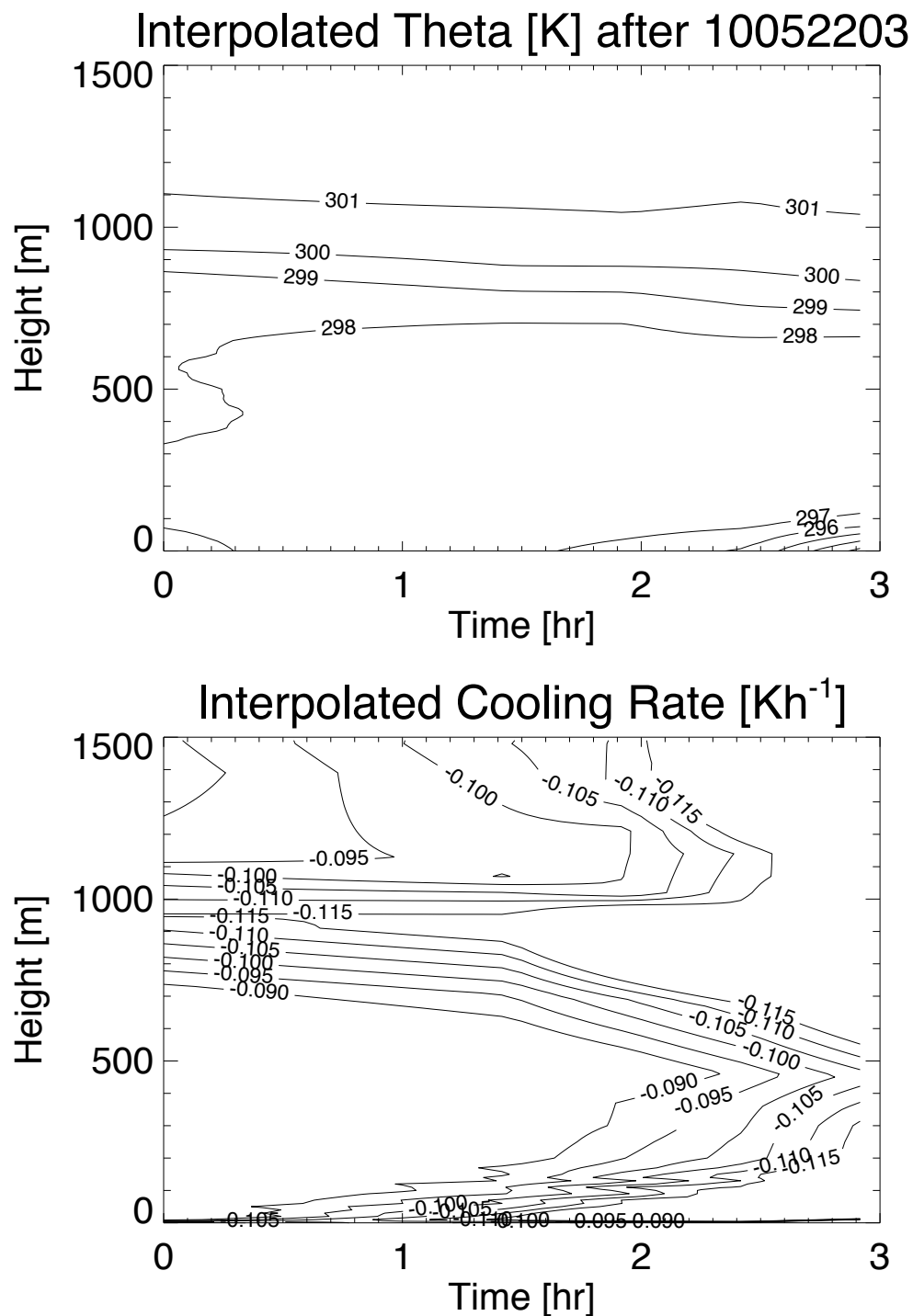


Figure 5.2: The top figure is a contour of values of potential temperature interpolated between soundings at the main site on the night of October 5-6. The bottom figure represents contours of calculated longwave radiative cooling using the MDTERP routine as described elsewhere in this volume.

ner. Because Equation 5.20 involves the very small differences of very large numbers, results are unsatisfactory even with varying grid sizes and smoothing of the input arrays. Although the information presented in Figure 5.3 does not support a constant k_m throughout the stable boundary layer for the night of Oct 5-6, it also indicates complete failure of this method of reverse-calculation of k_θ . This failure could be due to any or all of the following three influences. First, the precision of the radiosonde observations could be insufficient for this type of calculation, given that this calculation depends on the very small differences of very large numbers. Second, the vertical resolution of the data used for this calculation could be inadequate. Finally, other intermittent or vertically constrained phenomena could dominate the behavior of the atmosphere on this night.

We now return to a discussion of parameterizations of k_m . Early parameterizations assumed that turbulence within the stable boundary layer was similar to that in the convective boundary layer, so that fluxes at any level could be normalized with surface flux values and then expressed as a function of z/h , where h is the height of the entire nocturnal boundary layer (Monin and Obukhov, 1954). Numerous empirical functions of the normalized fluxes and z/h were proposed and tested. By normalizing fluxes aloft with surface values, one assumes an intimate connection between turbulence at the surface and turbulence aloft. However, the stability of the nocturnal boundary layer discourages eddies from penetrating the entire layer, so boundary-layer-wide scaling is inappropriate for the stable boundary layer (Sorbjan, 1997). In 1994, Derbyshire suggested that “perhaps the only prospect of avoiding closure issues in stable boundary layers is by exploiting Richardson number arguments.” Recent work has underscored the fact that turbulence within the stable boundary layer is produced locally (Wyngaard and Coté, 1972; Nieuwstadt, 1984; Estournel and Guedalia, 1987; Derbyshire, 1999) and should be parameterized as a function of the local flow.

Simple models using local-based parameterizations have been successful in rep-

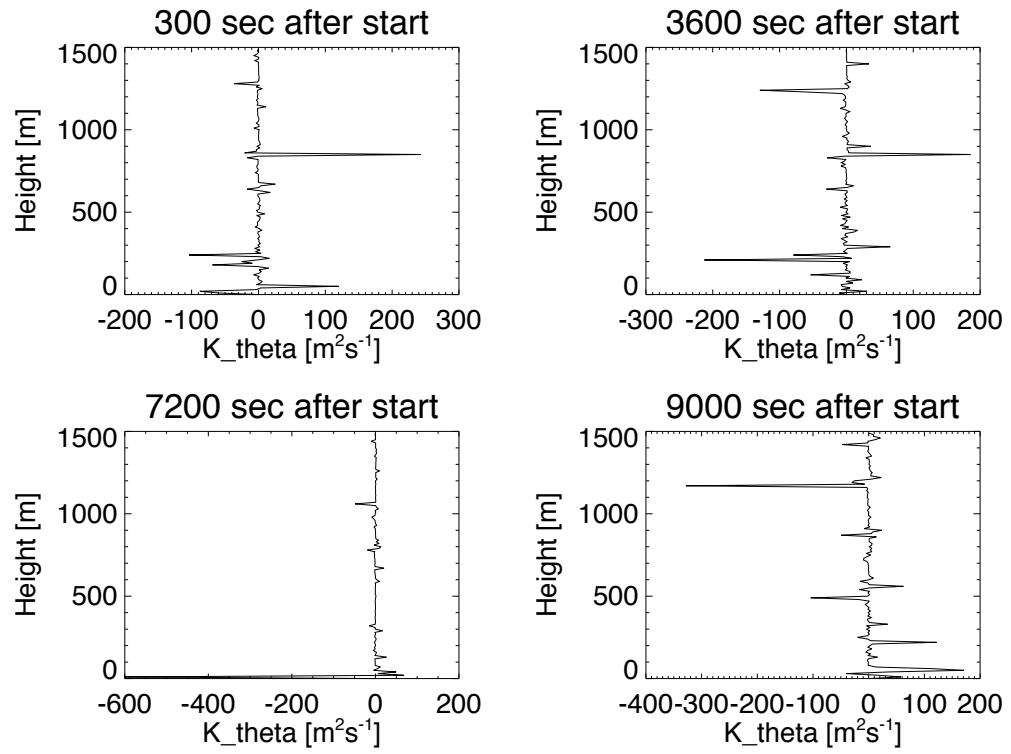


Figure 5.3: These profiles show the supposed time evolution of the coefficient of eddy viscosity for heat as calculated directly from gridded and interpolated sounding data.

representing moderately stable boundary layers. For instance, Tjemkes and Duynkerke (1989) use a 1D model which includes a narrow-band radiation model to find reasonable agreement with both mean and turbulent flux profiles from Cabauw data for two clear nights with moderate winds.

Almost all parameterizations of k_m as a function of the local Richardson number (see Mellor and Yamada (1982); Stull (1988); Sorbjan (1989); Garratt (1992) for summaries) take the form of

$$k_m = l^2 \left[\left(\frac{\partial u}{\partial z} \right)^2 + \left(\frac{\partial v}{\partial z} \right)^2 \right]^{1/2} f(\text{Ri}), \quad (5.21)$$

where l is a mixing-length, u and v are the components of the wind, and f is an empirical function of Ri, the local gradient Richardson number, given by

$$\text{Ri} = \frac{g}{\bar{\theta}} \frac{\partial \theta}{\partial z} \left(\left[\left(\frac{\partial u}{\partial z} \right)^2 + \left(\frac{\partial v}{\partial z} \right)^2 \right] \right)^{-1}. \quad (5.22)$$

In comparing nocturnal boundary layer data to numerical simulations utilizing different formulations of mixing-length l , Lacser and Arya (1986) find that the various formulae predict very similar mean profiles, though turbulent profiles are more sensitive to the l expression. For the purposes of this study, only mean profiles are important. We use Delage's (1974) parameterization, which Lacser and Arya found to give the best results for both mean and turbulent quantities as compared to Cabauw mast data. This parameterization was later modified by Estournel and Guedalia (1987) to allow for greater stability ranges; they find mixing length l well-represented as

$$l = \frac{1}{\frac{1}{kz} + \frac{1}{\lambda}}. \quad (5.23)$$

In this expression, k is the von Karman constant, z is height, and $\lambda = 0.0004Gf^{-1}$ (f : Coriolis parameter, G : the magnitude of the geostrophic wind).

To parameterize $f(\text{Ri})$, we also follow Estournel and Guedalia. Depending on the

value of the local Richardson number Ri ,

$$f(Ri) = \begin{cases} (1 - 5Ri)^2 & \text{for } 0 \leq Ri < 0.16 \\ \left(\frac{1}{1+41Ri}\right)^{1.68} & \text{for } Ri > 0.16 \\ 1 - Ri(1 + |Ri|^{1/2})^{-1} & \text{for } Ri < 0 \end{cases} \quad (5.24)$$

The function for $Ri < 0$ follows the suggestion of Garratt (1992) for the unstable parameterization of $f(Ri)$. For this model, I have assumed that $k_\theta = k_m$ as defined in (5.21) - (5.24). While Kondo et al (1978) have established that the ratio k_θ/k_m for strong stabilities decreases as the stability increases according to the following equation:

$$\frac{k_\theta}{k_m} = \frac{1}{6.873Ri + \frac{1}{6.873Ri+1}}, \quad (5.25)$$

Estournel and Guedalia concluded that the use of $k_\theta/k_m = 1$ or of (5.25) has only a weak influence on the results of simulations.

Instead of using the above formulations of k_m as a function of the Richardson number and of local shear, other researchers have represented eddy viscosity as a function of the turbulent kinetic energy (TKE) of the local flow. This one-and-a-half order closure (Garratt, 1992) allows the momentum and temperature variances to immediately affect the eddy viscosity. In this type of scheme,

$$k_m = \Lambda e^{-1/2} \quad (5.26)$$

where Λ is the local Monin-Obukhov length scale and e is the TKE, given by

$$\frac{\partial \bar{e}}{\partial t} = -\overline{u'w'} \frac{\partial \bar{u}}{\partial z} - \overline{v'w'} \frac{\partial \bar{v}}{\partial z} + \frac{g}{\theta} \overline{\theta'w'} - \frac{\partial \left[\overline{w'((p'/\rho) + e)} \right]}{\partial z} - \epsilon. \quad (5.27)$$

For the purposes of this study and for the dataset available, one-and-a-half closure is inappropriate. Although we have reliable initial conditions of the mean variables from a sounding, no profiles of fluxes above 55 m are currently available from either the CASES-97 or CASES-99 datasets.

Additionally, even if accurate initial values for fluxes could be defined, we would expect only slight improvements in model performance by parameterizing k_m based on

the TKE instead of the local Richardson number. From Nieuwstadt's (1984) quasi-steady theory, Derbyshire (1989) estimates the relative contribution of energy transport to the local TKE balance to be on the order of $0.1(\Lambda/h)^2$, where Λ is the local Monin-Obukhov length scale and h is the height of the boundary layer. The errors involved in guessing the initial values of TKE would be larger than this value.

5.4 Model Simulations

This model allows decaying turbulence and the developing stability commencing at the evening transition to influence the dynamics of the evolving stable layer. The evolving stability, reflected in the increasing Richardson number and the decreasing turbulent exchange coefficient k_m , is responsible for the observed structure of inertial oscillations.

As discussed in Chapter 3, data from the night of 11-12 October 1999 suggests the generation and persistence of inertial oscillations by the mechanism proposed by Blackadar. In order for observed inertial oscillations to have the structure described in Figure 3.11, the geostrophic wind must be constant with height, or even decreasing with height. The numerical model used here does not accommodate a nonzero $\partial V_g/\partial z$. Because the observational evidence indicates that on this night, $\partial V_g/\partial z = 0$ (the simple synoptic variation required for strict adherence to Blackadar's model), I choose this night to explore more closely.

On the night of 11-12 October, a high pressure system is located to the east and south of the CASES-99 array, while a weak front approaches from the west. The pressure gradients induce southerly-southwesterly geostrophic flow. Estimates of the geostrophic wind from radiosonde winds between 2000 and 5000 m, averaged between three sounding sites, show that the geostrophic wind increases slightly from 5.2 ms^{-1} (2300 UTC 11 October) to 6.7 ms^{-1} (0300 UTC 12 October) as the pressure gradients tighten over the night. Winds are from the southwest, and shift negligibly (8 degrees)

over the night.

Surface temperature (measured with a radiometer) decreases rapidly from 295 K to 284 K from 2300 UTC to 0200 UTC. The surface then warms slightly, remaining at 285 K until 1200 UTC 12 October. Figure 5.4 shows the time series of both surface temperature and potential temperature at 2 m for this time period. The potential temperature at 2 m is used as the bottom boundary condition for the model.

This rapid surface cooling leads to a strong surface-based inversion. Soundings from Beaumont, El Dorado, and Smileyburg disagree on the depth of the surface inversion. Figure 5.5 shows that by 0700 UTC, a surface inversion of 5 - 8 K extends over the lowest 50-150 m. (The variability between soundings can be partially explained by the variation in surface elevation. Beaumont, the highest of the three stations at 466 msl, shows the lowest surface inversion. El Dorado (420 msl) and Smileyburg (408 msl) show similar heights for the surface inversion.) At all stations, a slightly stable layer resides above the surface inversion, extending to 1000 m.

As discussed in Chapter 3, inertial oscillations on this night are present between the surface and 500 m. Above this level, the amplitude of the best-fit inertial oscillations gradually decreases, while the level of noise increases substantially, making the best-fit inertial oscillations fail the R_{ws} test. Inspection of the signal-to-noise ratios for this night (see <http://paradox.colorado.edu/~jkl/profilersnr.html>) does indicate bird contamination between 600 and 900 m on this night. Therefore, the lack of inertial oscillations aloft should not be taken as an absolute indicator that they are not present, but rather that the boundary-layer wind profilers are unable to detect them due to the contamination of the backscattered signal from migrating birds.

The model described above is run three times. Soundings taken at 2300 UTC (1700 LST) from Beaumont, El Dorado, and Smileyburg are used to initialize different runs. (The initial wind profiles, normalized with the geostrophic wind, are shown in Figures 5.6 - 5.8. Profiler winds are compared to the initialization soundings in Figures

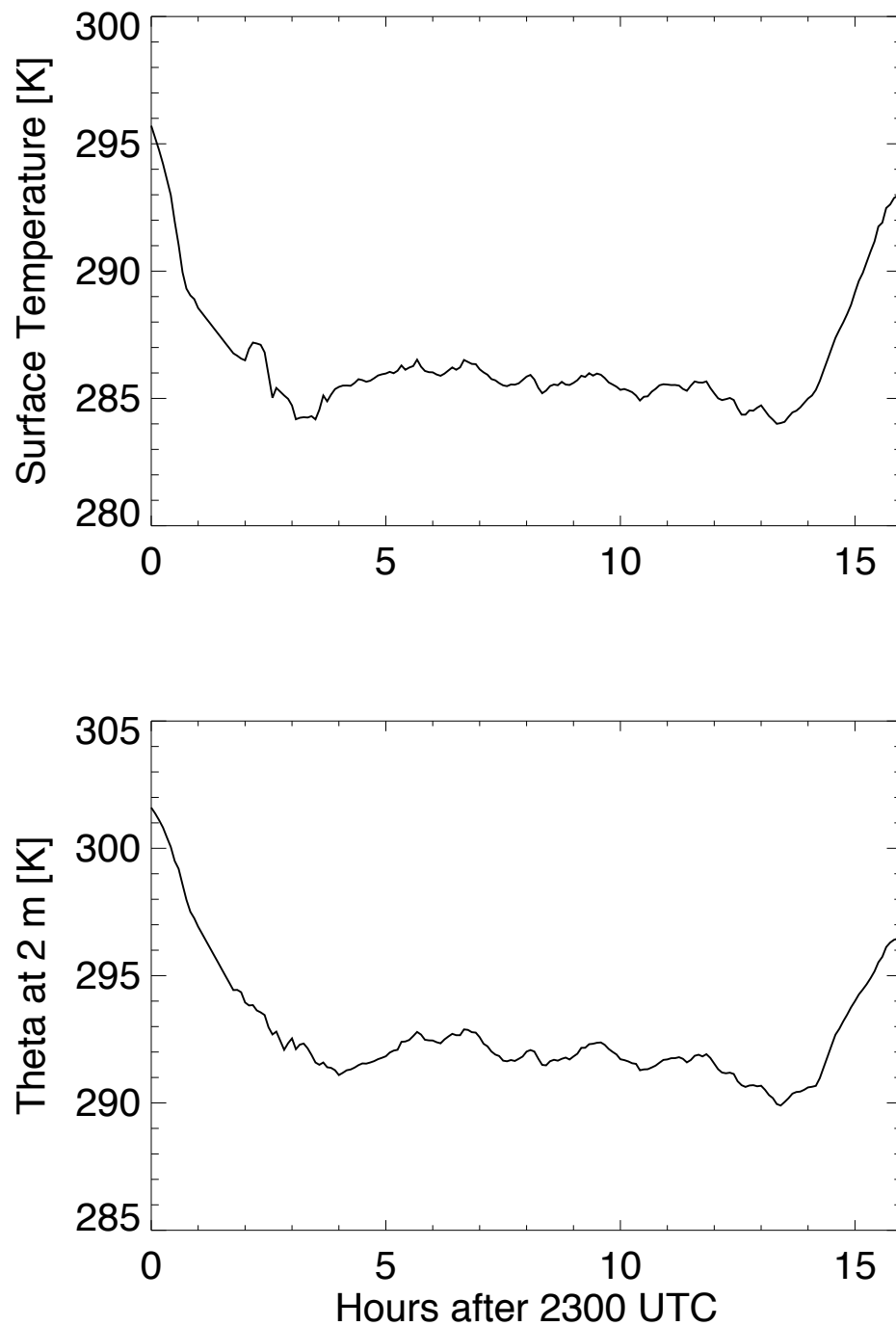


Figure 5.4: Time series of surface temperature and potential temperature at 2 m for the 16 hour period starting at 2300 UTC on 11 October 1999. Data were taken at the 55 m central tower of the CASES-99 array.

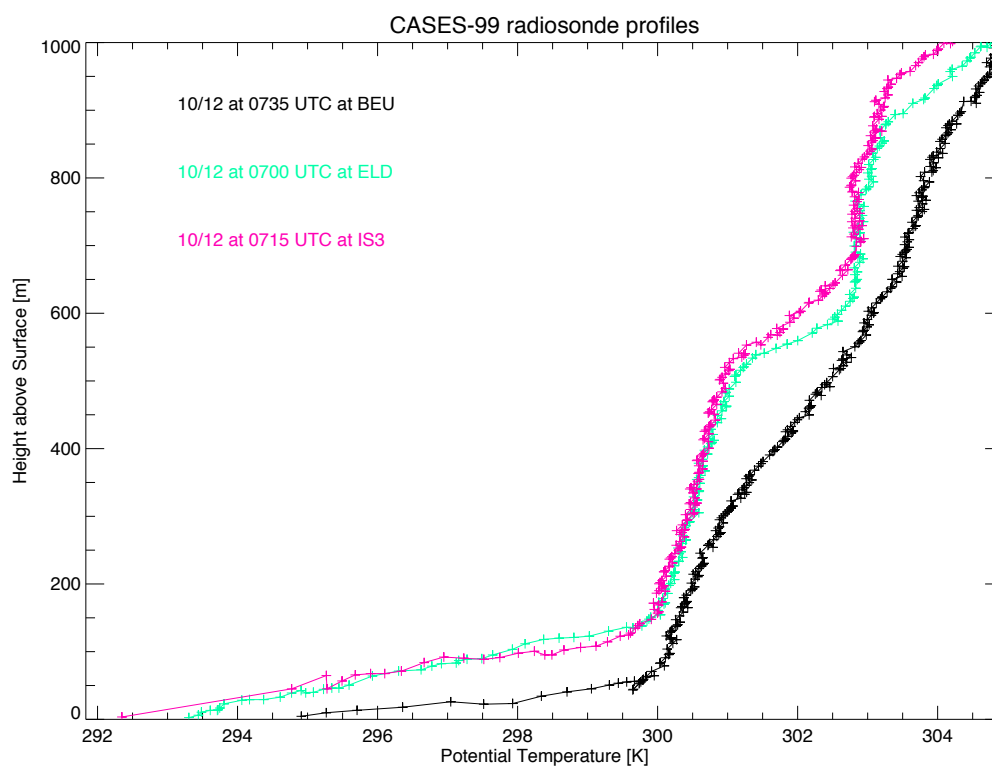


Figure 5.5: Potential temperature profiles at 0700 UTC 12 October 1999 for Beaumont (BEU), El Dorado (ELD), and Smileyburg (IS3).

5.9 - 5.11.) The runs are identical in terms of surface cooling (Figure 5.4) and the k_m parameterization. Because the geostrophic wind is calculated as the average of the winds between 2000 and 5000 m for each sounding, the geostrophic winds vary slightly between the three cases: The geostrophic wind for Beaumont is 4.6 ms^{-1} , El Dorado has 5.9 ms^{-1} , and Smileyburg has 5.2 ms^{-1} .

The mean profiles resulting from these model runs are compared to the 0300 and 0700 UTC soundings at the corresponding site. In all cases, the model is able to reproduce the shape of acceleration of the wind speed at and above the surface inversion (see Figures 5.6 - 5.8) although the wind speeds are smaller than observed. Though the model reproduces this acceleration of the entire residual layer, it does not produce the jet structure that is actually observed. Only for the Beaumont run (Figure 5.7) does the model produce a wind speed profile that tapers with height. Additionally, the model exhibits some rotation not apparent in the observations: wind directions above the inversion in the model for all three cases are approximately 30° larger (more westerly) than observed at all times.

The winds generated by each model run are examined for inertial oscillations using the technique described in Chapter 3; the amplitudes of the model's inertial oscillations are compared to the observations found from the boundary-layer profilers for this night. Figures 5.12 - 5.14 show that for the Beaumont and El Dorado runs, the model was able to reproduce the observed amplitudes of oscillations well within the anticipated error range of the observations. Even at those observation levels (600 - 900 m) where the best-fit inertial oscillations from observations fail the R_{ws} test because of large amounts of noise due to intermittent bird contamination, the model runs starting with Beaumont and El Dorado soundings replicate the observations quite well. The run initialized with a sounding from Smileyburg, however, predicts significantly weaker inertial oscillations than were observed.

Examining the difference between the Smileyburg (“unsuccessful”) and the El

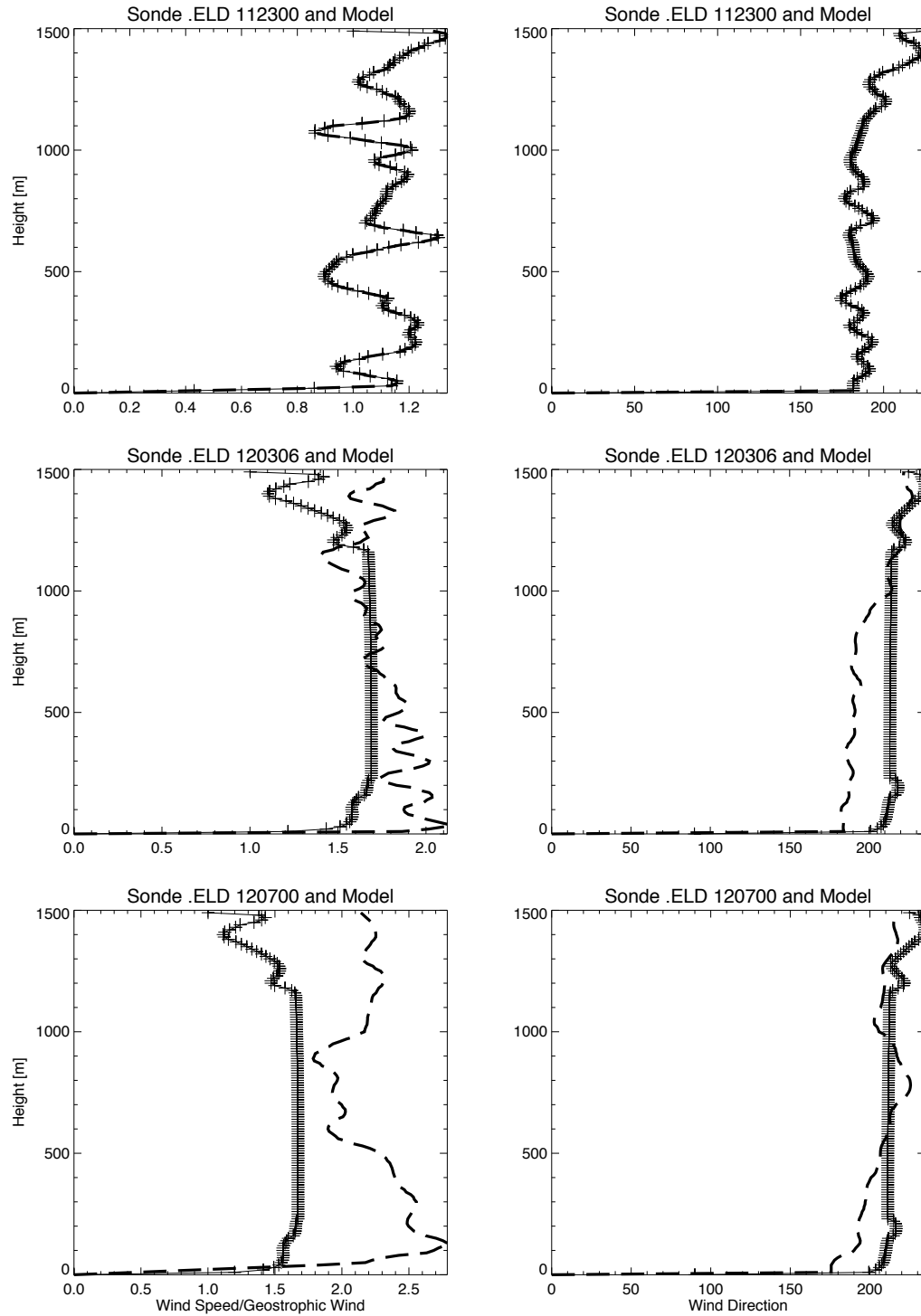


Figure 5.6: Comparison of model results (+) and radiosonde observations (dashed line) for nondimensional wind speed (normalized by the geostrophic wind), left panels, and wind direction, right panels. Three times are shown. The top two panels are for the initialization time of the model, 2300 UTC. The middle panels correspond to 0300 UTC; the bottom panels correspond to 0700 UTC. This model run was initialized with the 2300 El Dorado sounding of 11 October.

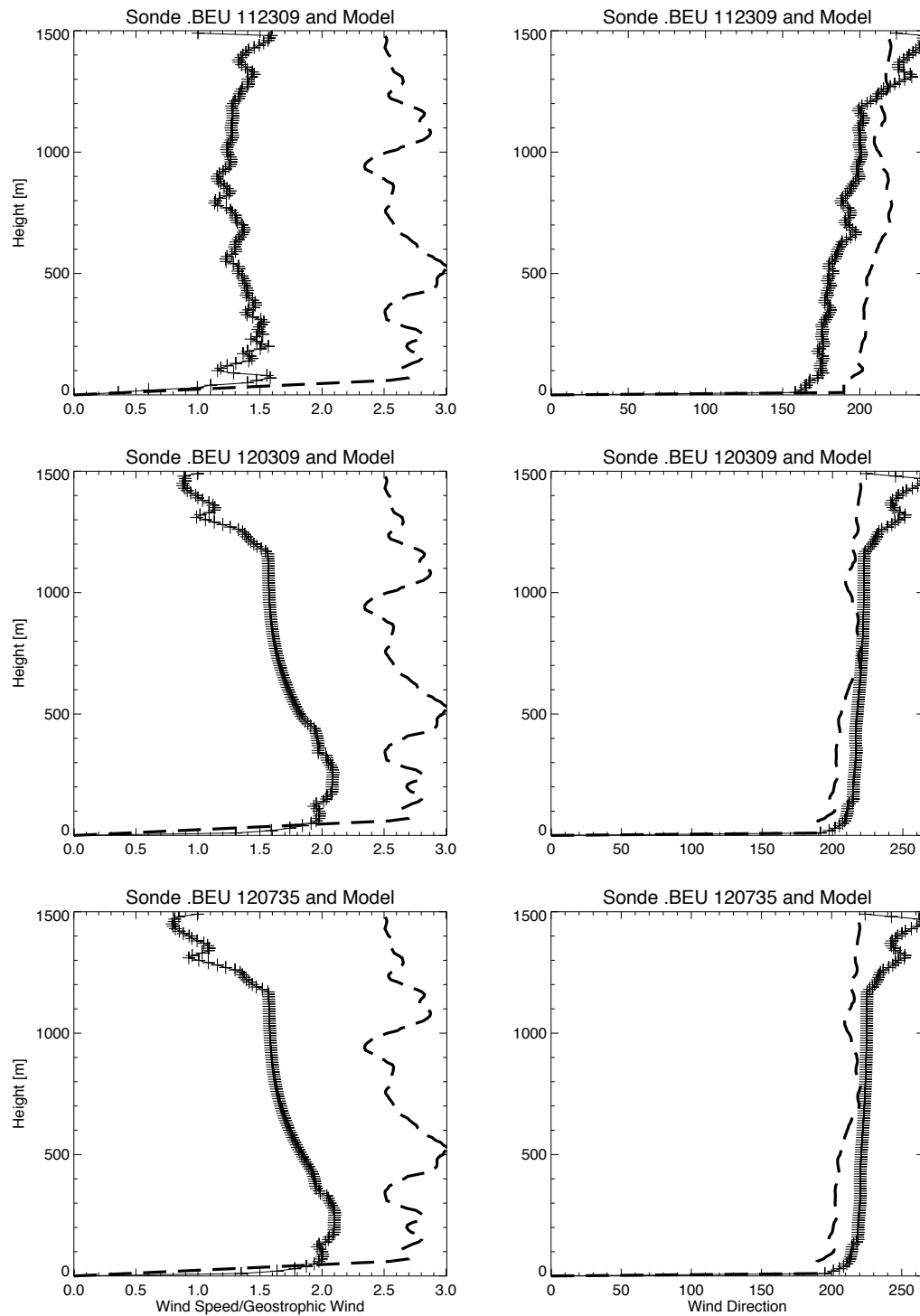


Figure 5.7: As in Figure 5.6 but for the model run initialized with the 2300 Beaumont sounding of 11 October.

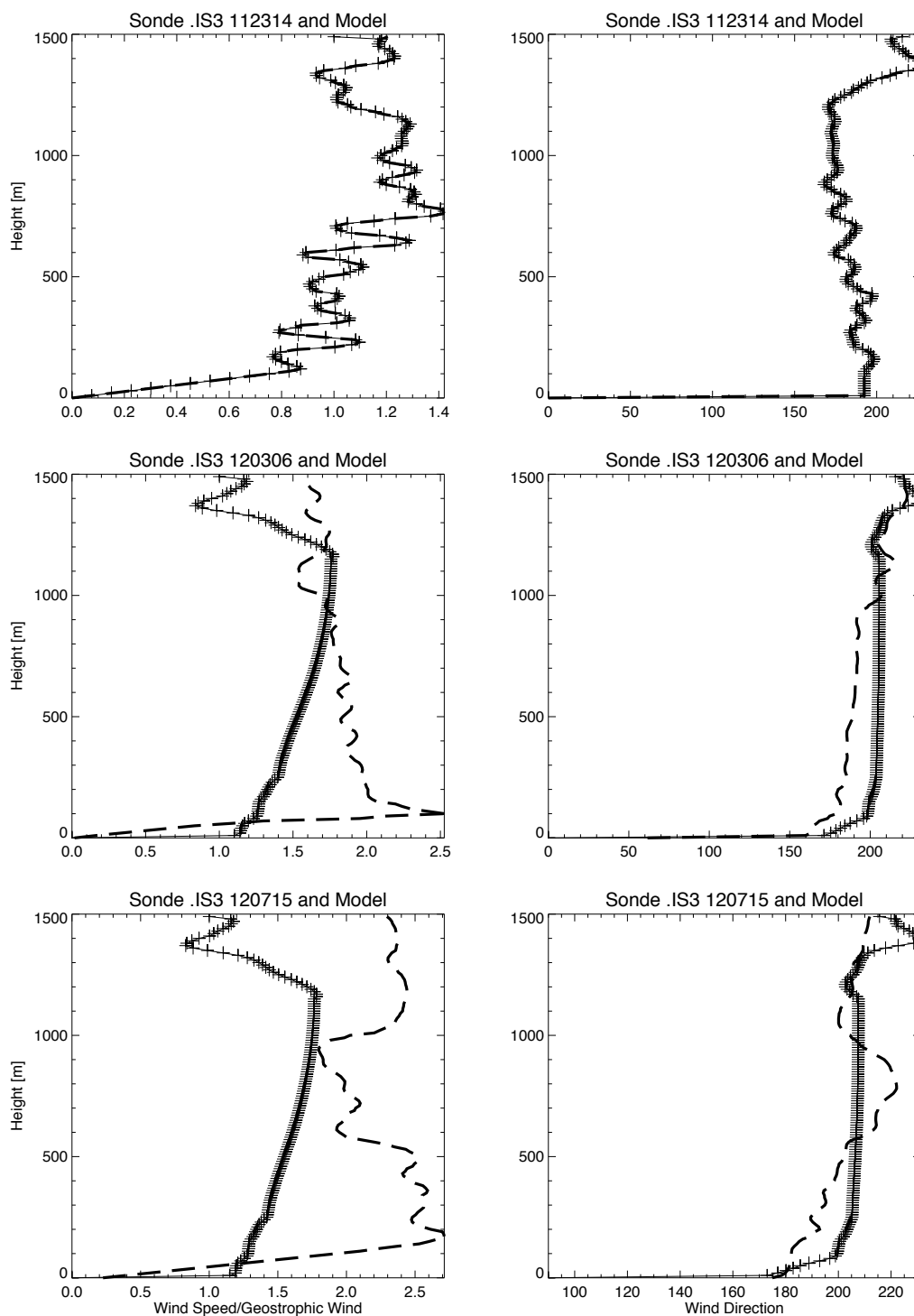


Figure 5.8: As in Figure 5.6 but for the model run initialized with the 2300 Smileyburg sounding of 11 October.

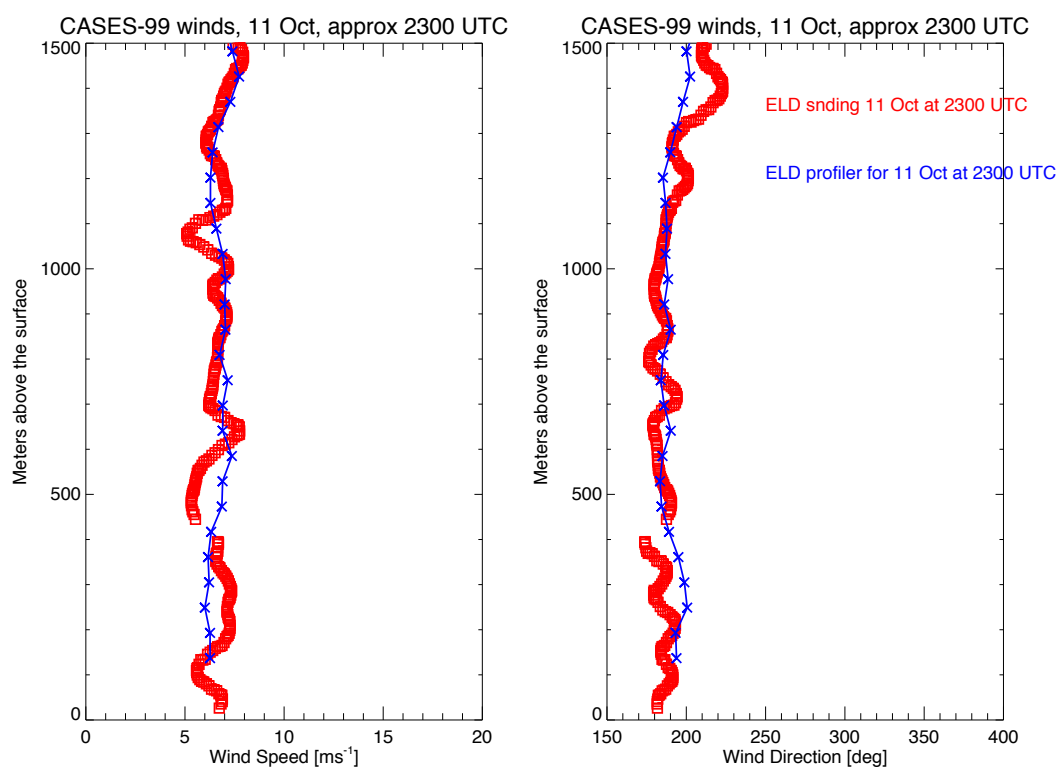


Figure 5.9: Profiler and sonde winds for 2300 UTC 11 October 1999 at El Dorado. The squares represent data from the sounding while the 'x' denote profiler data.

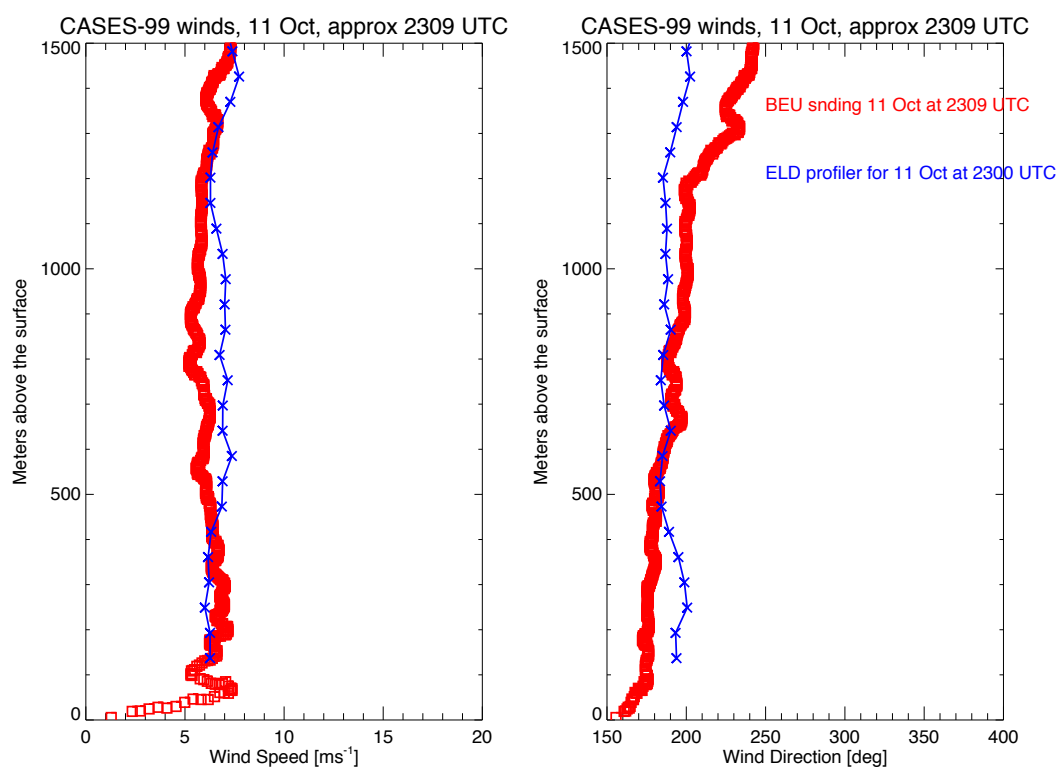


Figure 5.10: El Dorado profiler and Beaumont sonde winds for 2309 UTC 11 October 1999. The squares represent data from the sounding while the 'x' denote profiler data.

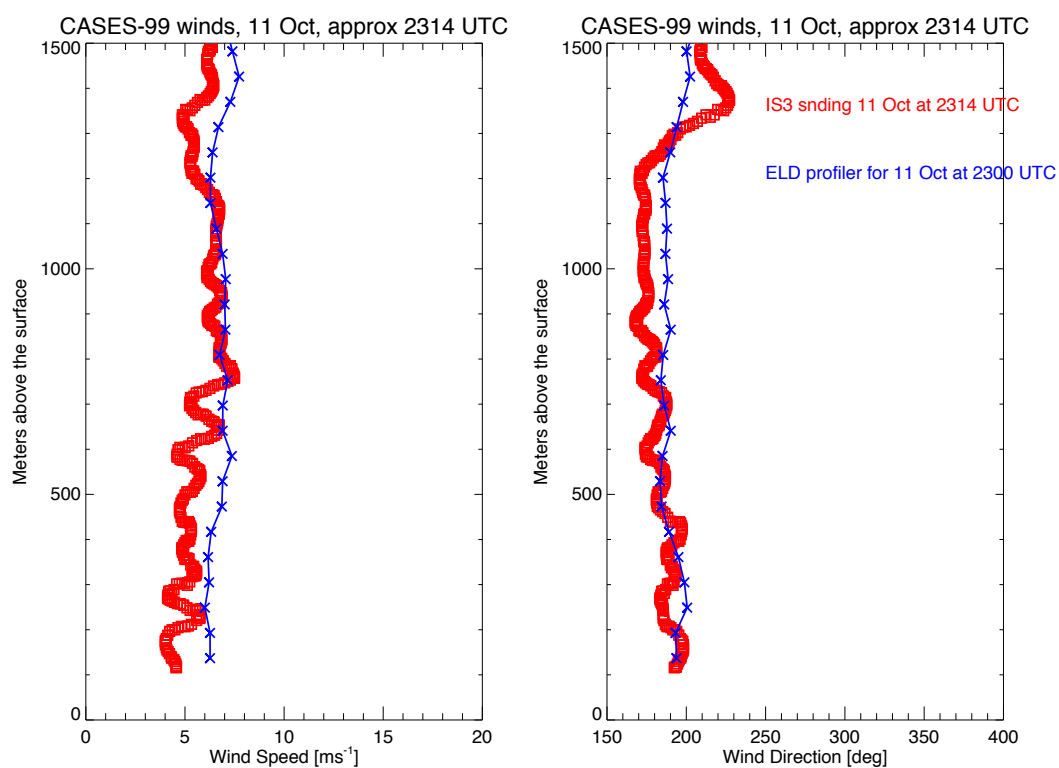


Figure 5.11: El Dorado profiler and Smileyburg sonde winds for 2314 UTC 11 October 1999. The squares represent data from the sounding while the 'x' denote profiler data.

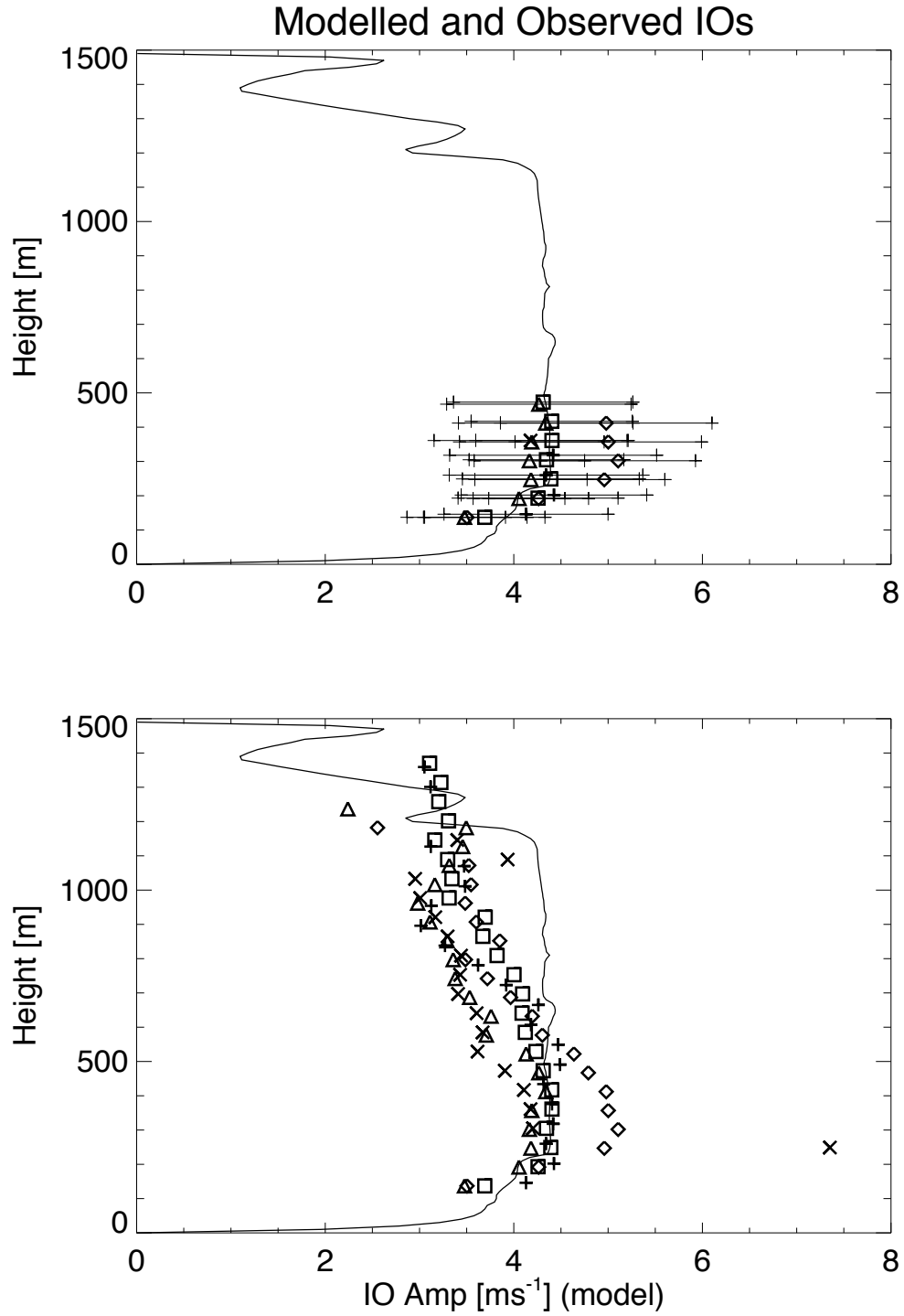


Figure 5.12: Comparison of the amplitudes of inertial oscillations generated by the model (solid line) initialized with the 11 Oct 2300 UTC El Dorado sounding and the amplitudes of the inertial oscillations observed with the 5 boundary layer profilers (symbols). The upper figure only includes those observed inertial oscillations that pass the R_{ws} test; error bars are included. The lower figure shows the amplitudes for all of the observed best-fit inertial oscillations, regardless of whether they pass the R_{ws} test.

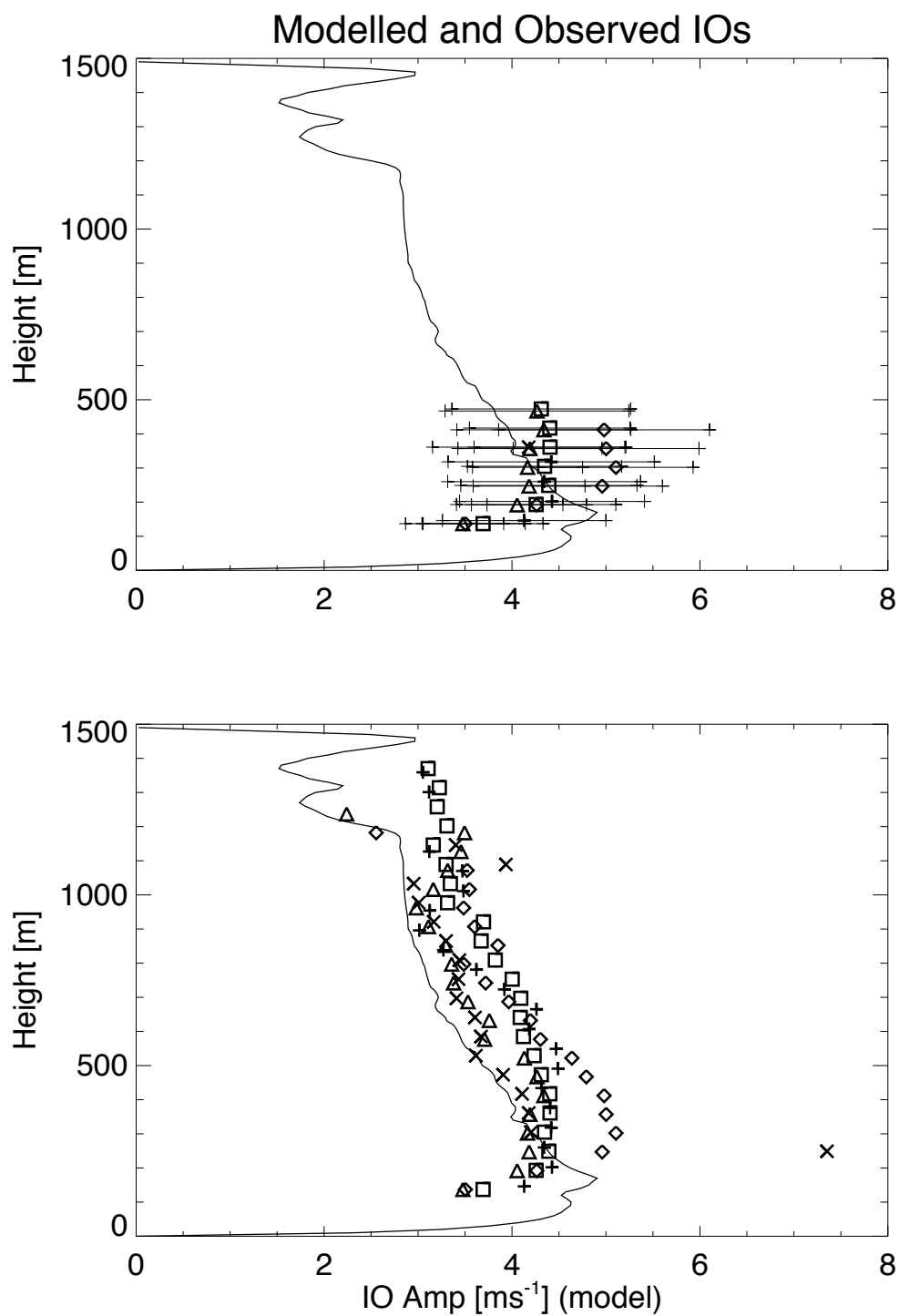


Figure 5.13: As in Figure 5.12 but for the model run initialized with the 11 Oct 2300 UTC Beaumont sounding.

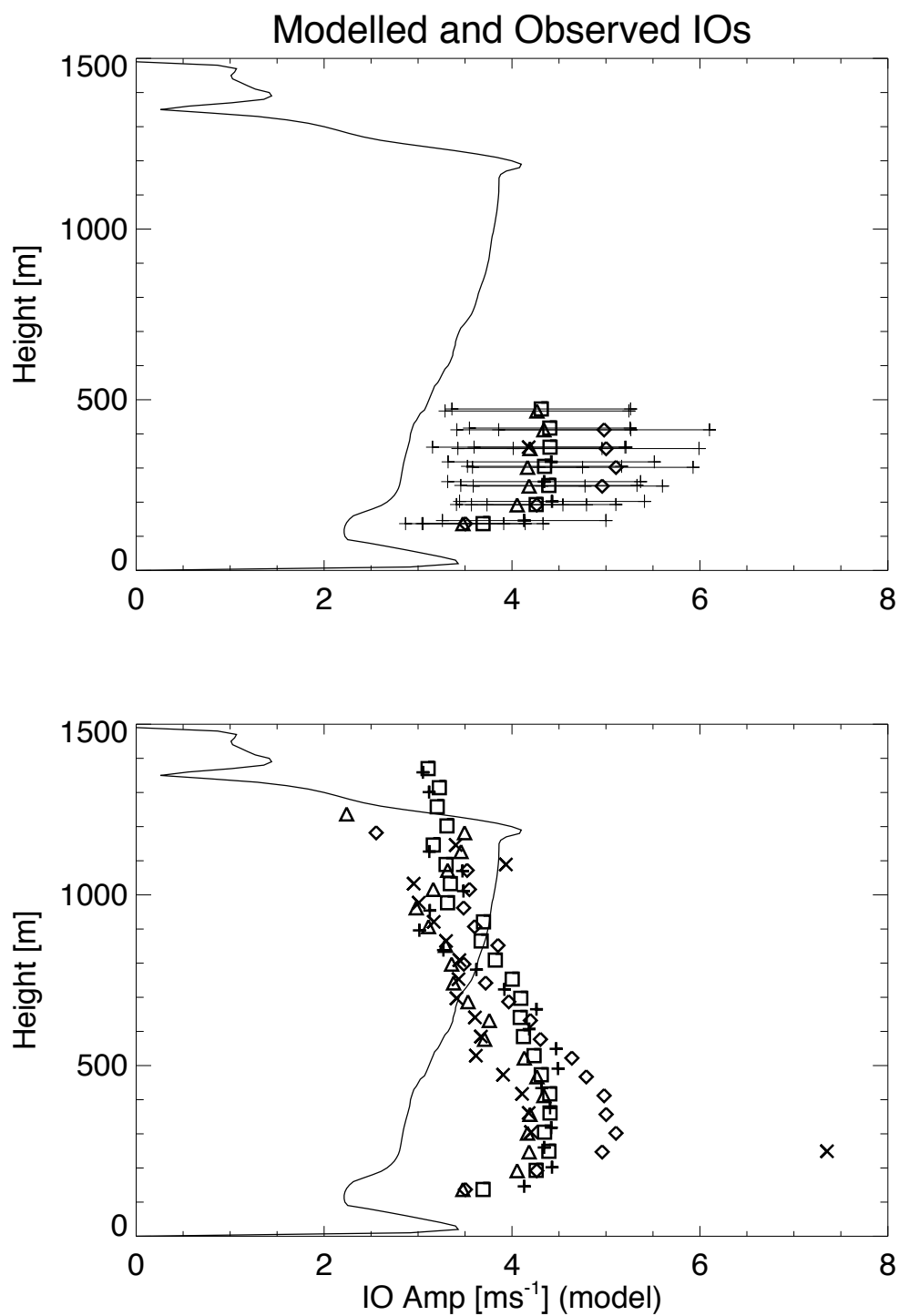


Figure 5.14: As in Figure 5.14 but for the model run initialized with the 11 Oct 2300 UTC Smileyburg sounding.

Dorado/Beaumont (“successful”) runs, we can isolate the processes with the most influence on the development of inertial oscillations. I consider the possibility that the variation between the geostrophic wind estimated by the initial soundings is responsible. However, Beaumont shows the weakest geostrophic wind (4.6 ms^{-1}), while the other successful run, El Dorado, uses a comparably large estimate of 5.9 ms^{-1} . These values bracket the Smileyburg estimate of 5.3 ms^{-1} . Therefore, the difference in the geostrophic wind (which is assumed to be constant in time and in height) between the three stations is not the primary factor determining the amplitude of the inertial oscillations generated in the model.

The actual values of k_m used in the model have little impact on the evolution of the wind and potential temperature fields in this model. As Figures 5.15 - 5.17 show, the ratio of the $\frac{\partial}{\partial z} k_m \frac{\partial u}{\partial z}$ and $\frac{\partial}{\partial z} k_m \frac{\partial v}{\partial z}$ terms to the ageostrophic terms fv_a and fu_a quickly becomes much less than one. To determine the values of these terms, I approximate the derivatives by finite differences (Roache, 1982), so that at a height j ,

$$\begin{aligned} \frac{\partial}{\partial z} k_m \frac{\partial u}{\partial z} &= \frac{(k_m)_{j+1/2}}{\Delta z^2} (u_{j+1} - u_j) - \frac{(k_m)_{j-1/2}}{\Delta z^2} (u_j - u_{j-1}), \\ \frac{\partial}{\partial z} k_m \frac{\partial v}{\partial z} &= \frac{(k_m)_{j+1/2}}{\Delta z^2} (v_{j+1} - v_j) - \frac{(k_m)_{j-1/2}}{\Delta z^2} (v_j - v_{j-1}) \end{aligned} \quad (5.28)$$

Initial values show that, at some levels, the turbulent terms (as calculated here) can momentarily be on the same order, or even larger than the ageostrophic terms (see the top left panels of Figures 5.15 - 5.17). These high values indicate strong residual mixing at those levels, in agreement with results from Large-Eddy Simulation as described by Sorbj an (1997). The turbulent terms quickly become smaller than the ageostrophic terms, though, so their influence on the flow is small and temporary. At other levels, particularly near the top of the model domain, the turbulent terms are initially at least an order of magnitude smaller than the ageostrophic terms; at these levels, the turbulent terms stay so small that they cannot influence the flow. These results are consistent between the three model runs, indicating that the details of the evolution of turbulence

are inconsequential to the development of inertial oscillations as long as the turbulent terms become small.

The data shown in Figures 5.18 - 5.20 confirm the hypothesis that the initial ageostrophic component of the winds is the most significant component determining the amplitudes of the inertial oscillations, even if the turbulent terms are temporarily large initially. Above the height of the surface inversion, the amplitude of the model inertial oscillations is almost identical to the ageostrophic component of the winds at the beginning of the model run. In fact, the correlation coefficients between these variables are 0.74 for El Dorado, 0.86 for Beaumont, and 0.82 for Smileyburg. These high values indicate that the initial ageostrophic profile has a strong influence on the eventual evolution of inertial oscillations. In contrast, correlation values between the initial k_m profile and the amplitudes of the inertial oscillations are .32 (El Dorado), 0.22 (Beaumont), and 0.21 (Smileyburg). The magnitude of the initial ageostrophic component of the wind is more significant to the evolution of inertial oscillations than the initial profile of eddy viscosity.

As Figures 5.18 - 5.20 show, the amplitude of the model inertial oscillations is not exactly equal to the initial value of the ageostrophic wind, although the amplitudes show similar variation with height. Although the turbulent terms, as calculated in this model, are small compared to the ageostrophic component, the turbulent terms are not zero. Therefore, the turbulent terms do modify the flow somewhat. In the regions where the turbulent terms remain large longer, (between 300 and 1000 m in Figure 5.15 and 5.18, for example), vertical transfer of momentum is encouraged by the relatively large k_m . Instead of the amplitude of the inertial oscillation at each level equaling the ageostrophic wind at that level, the amplitudes between 300 and 1000 m become an average of the ageostrophic wind within that region. Above this zone, k_m is very small and the inertial oscillation amplitudes follow the ageostrophic component almost exactly, as shown in Figure 5.18. The persistence of turbulence in the model smoothes

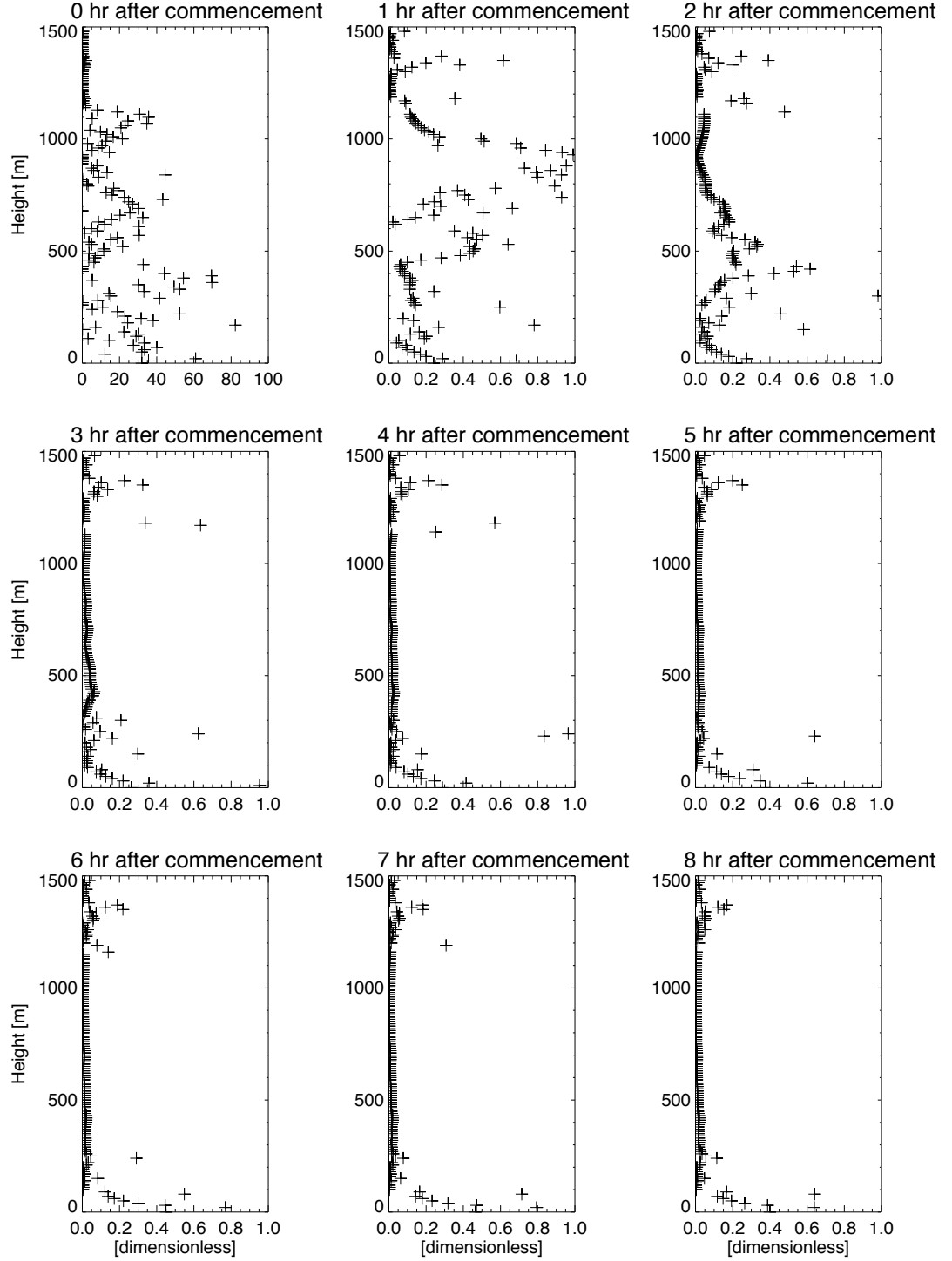


Figure 5.15: The ratio of the turbulent term $\left(\left(\frac{\partial}{\partial z}k_m\frac{\partial u}{\partial z}\right)^2 + \left(\frac{\partial}{\partial z}k_m\frac{\partial v}{\partial z}\right)^2\right)^{1/2}$ to the ageostrophic term $f\left((u-u_g)^2 + (v-v_g)^2\right)^{1/2}$ at various times in the model run initialized with the 11 Oct 2300 El Dorado sounding. Note that the abscissa for the initial time is two orders of magnitude larger than at later times.

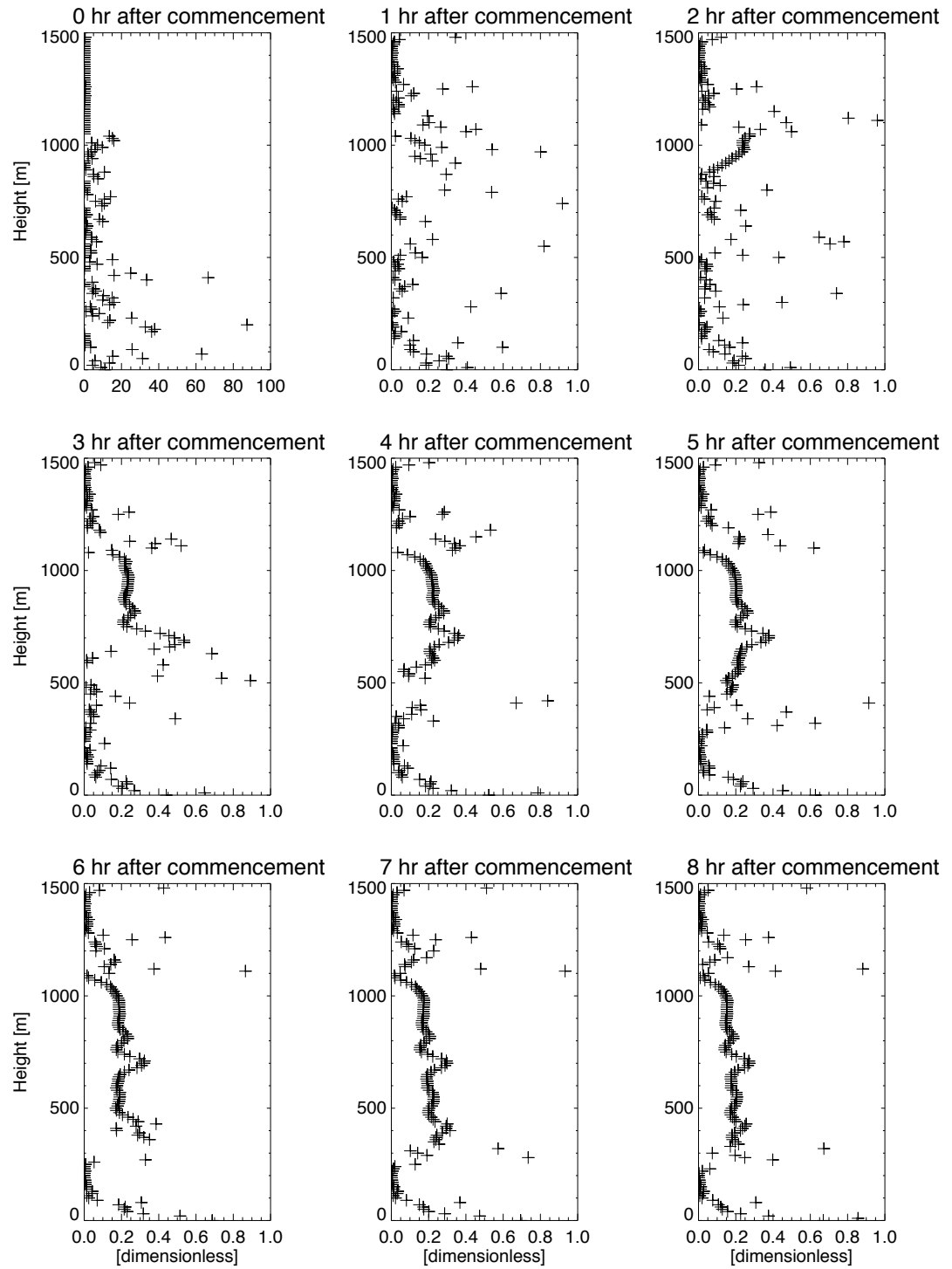


Figure 5.16: As in Figure 5.15 but for the model run initialized with the 11 October 2309 Beaumont sounding.

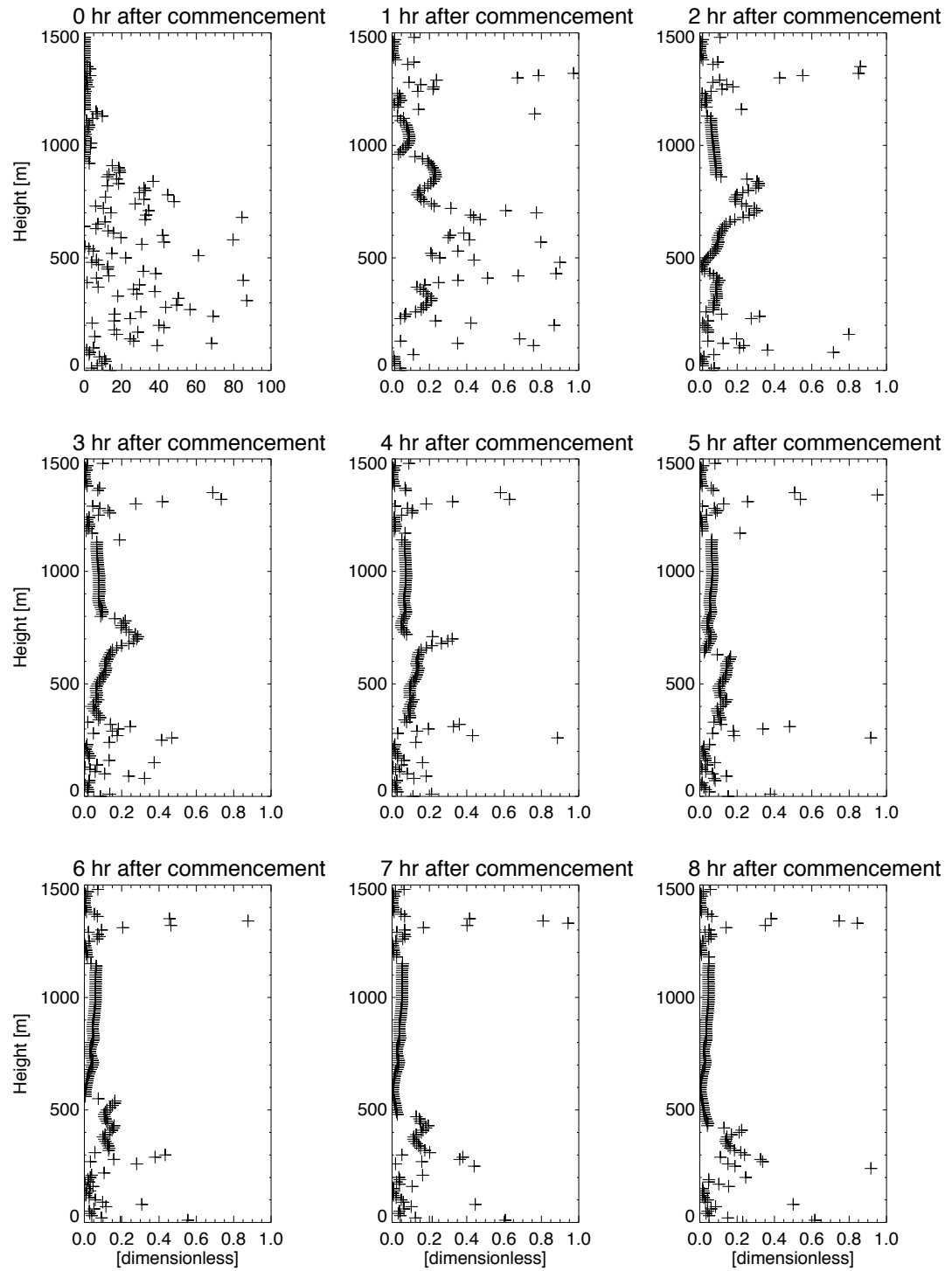


Figure 5.17: As in Figure 5.15 but for the model run initialized with the 11 October 2314 Smileyburg sounding.

the amplitudes of the observed inertial oscillations, compared to the initial ageostrophic profile, but the amplitudes are primarily determined by the magnitude of the initial ageostrophic component of the wind.

To further confirm that exact parameterization of k_m is inconsequential to the development of inertial oscillations, a model run with a different parameterization of k_m was run. This run follows the same form as (5.21) but defines the Richardson number dependence as $f(\text{Ri}) = (1. + 5\text{Ri})^{-2}$ instead of (5.24) above. The results of this simulation, initialized with the Beaumont 11 Oct 2309 UTC sounding are shown in Figure 5.21, which may be compared with Figure 5.7. Again, the model can reproduce the general pattern of acceleration atop the surface inversion, but cannot generate the strong jet atop the inversion. The jet was probably enhanced by a slope flow through enhancement of the thermal wind (Holton, 1967). Despite the different parameterization of k_m , this run shows the same dependence of the inertial oscillation amplitude on the initial ageostrophic component of the winds above the inversion seen with the previous parameterization (see Figure 5.22). The correlation coefficient between the initial ageostrophic component of the wind and the amplitudes of the inertial oscillations is 0.71 for this case.

5.5 Analytical verification of the importance of the initial ageostrophic wind

This dependence of the amplitude of the inertial oscillation on the initial value of the ageostrophic component of the wind is supported with results from Blackadar's (1957) simple theoretical model. (Data from the 2300 UTC 11 October sounding from El Dorado are used throughout this discussion.) As shown in Figure 5.15, the turbulent terms of (5.1) quickly become small compared to the ageostrophic terms. Because the turbulent terms are an order of magnitude smaller than the Coriolis terms, the turbulent

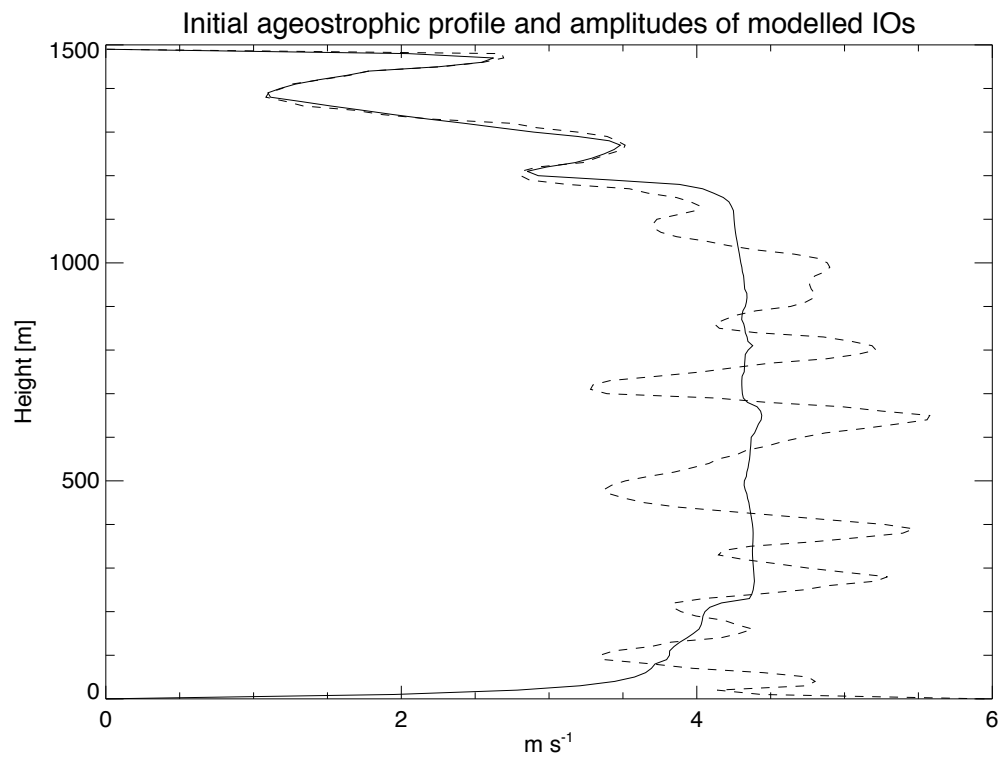


Figure 5.18: The amplitudes (ms^{-1}) of modeled inertial oscillations (solid line) and the initial ageostrophic components (ms^{-1}) of the wind (dashed line) for the model run initialized with the 11 October 2300 UTC sounding from El Dorado.

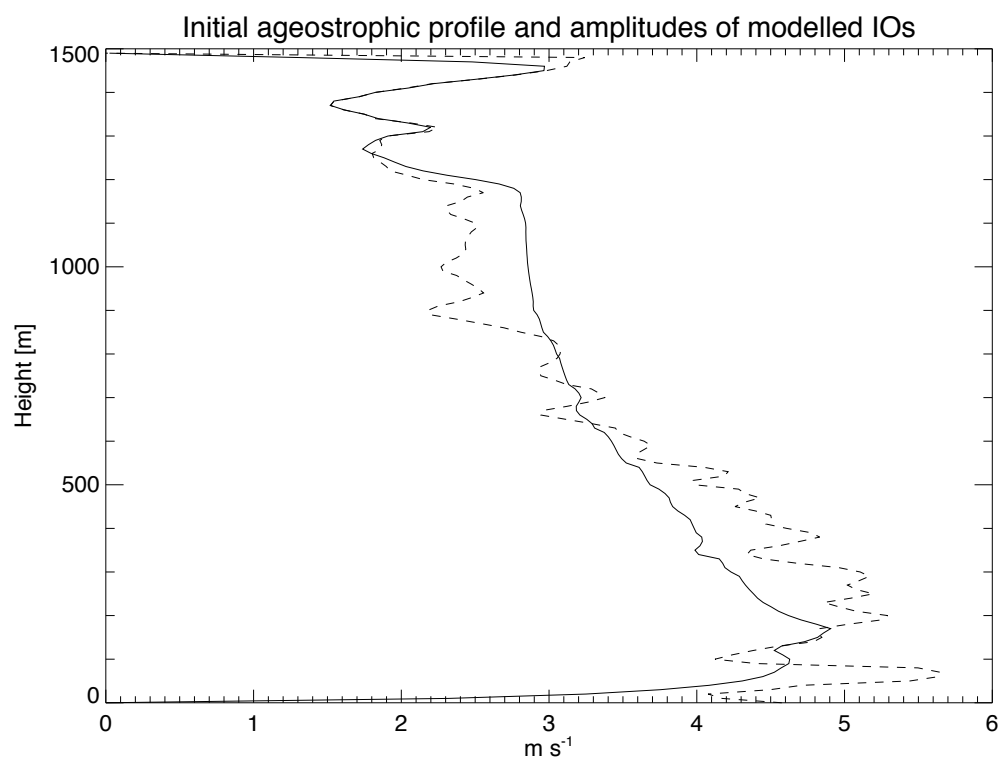


Figure 5.19: As in Figure 5.18 but for the model run initialized with the 11 October 2309 sounding from Beaumont.

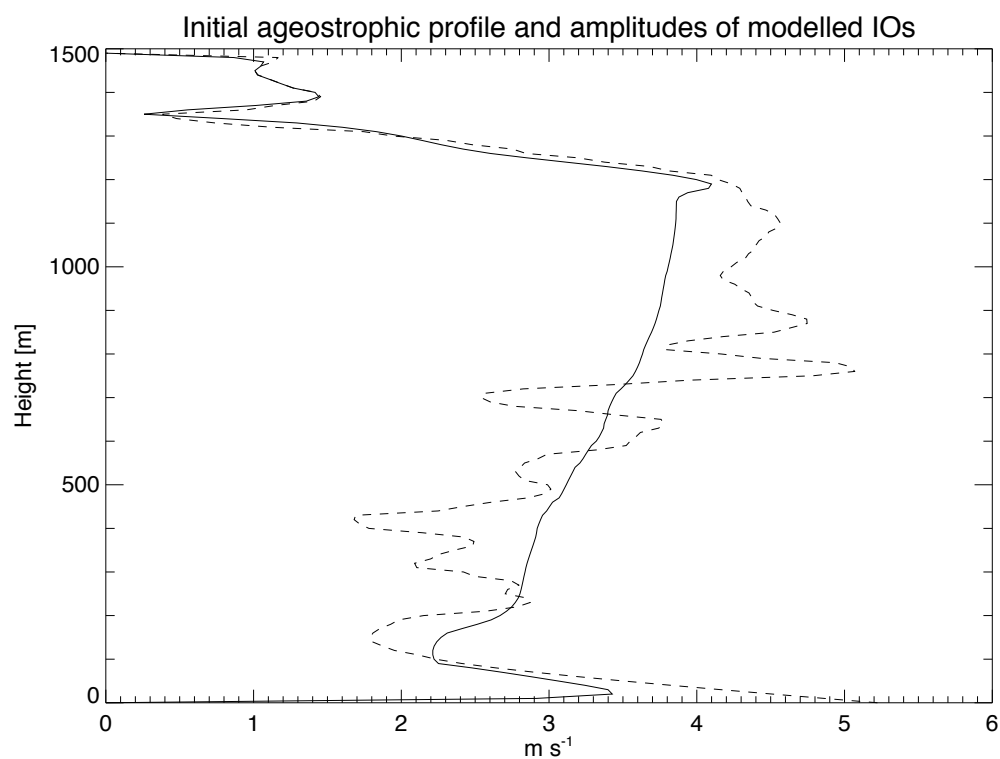


Figure 5.20: As in Figure 5.18 but for the model run initialized with the 11 October 2314 UTC sounding from Smileyburg.

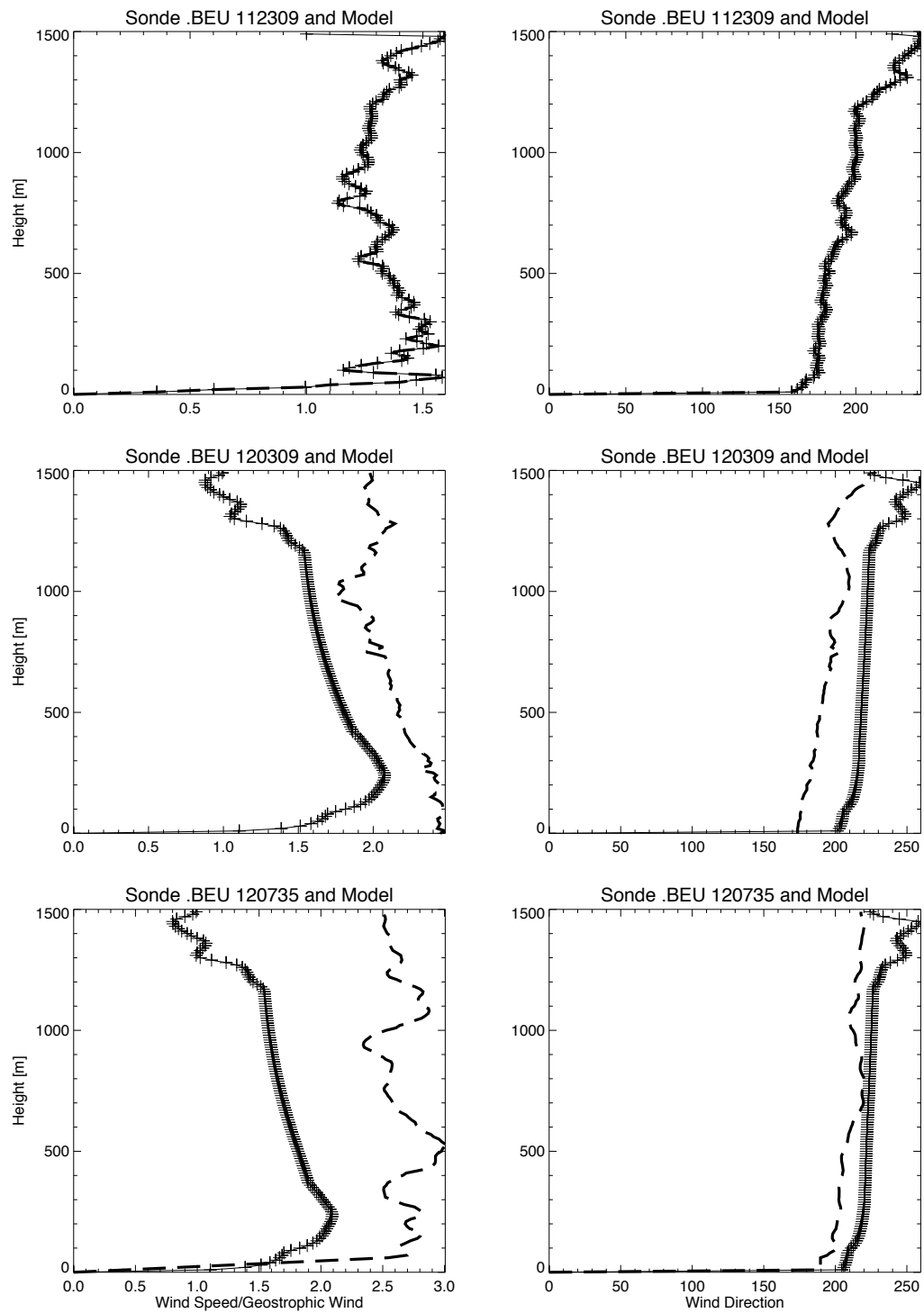


Figure 5.21: As in Figure 5.7, but eddy viscosity in the model run is parameterized with (5.4).

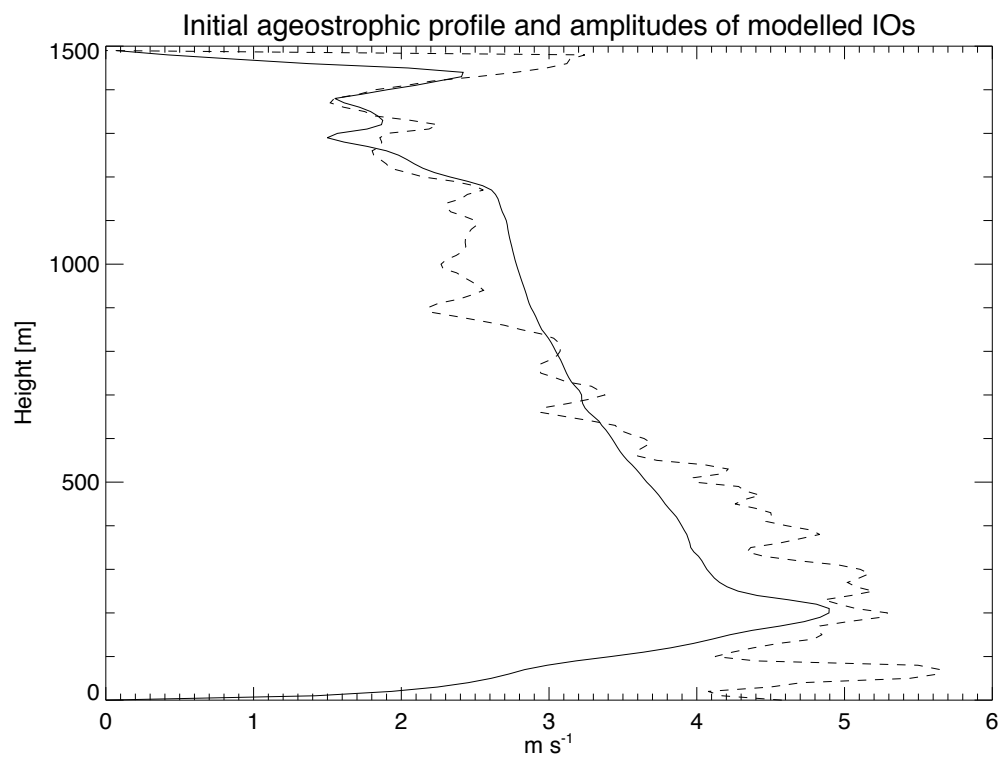


Figure 5.22: As in Figure 5.19 but for the model run initialized with the 11 October 2309 UTC sounding from Beaumont and parameterization of k_m according to (5.4).

terms are dropped, leaving the system of equations

$$\begin{aligned}\frac{\partial u}{\partial t} &= f(v - v_g), \\ \frac{\partial v}{\partial t} &= -f(u - u_g)\end{aligned}\tag{5.29}$$

for each height z . At each height z , winds are composed of a geostrophic and ageostrophic component:

$$\begin{aligned}u(z, t) &= u_g + u_a(z, t), \\ v(z, t) &= v_g + v_a(z, t).\end{aligned}\tag{5.30}$$

As with the numerical model discussed above, the geostrophic wind is set constant in height and in time so that at each height z , (5.29) simplifies to

$$\begin{aligned}\frac{\partial u_a(z, t)}{\partial t} &= f v_a(z, t) \\ \frac{\partial v_a(z, t)}{\partial t} &= -f u_a(z, t),\end{aligned}\tag{5.31}$$

which by substitution leads to

$$\begin{aligned}\frac{\partial^2 u_a}{\partial t^2} + f^2 u_a &= 0 \\ \frac{\partial^2 v_a}{\partial t^2} + f^2 v_a &= 0.\end{aligned}\tag{5.32}$$

The solutions take the form

$$\begin{aligned}u_a(z, t) &= A_u(z) \sin(ft) + B_u(z) \cos(ft) \\ v_a(z, t) &= A_v(z) \sin(ft) + B_v(z) \cos(ft)\end{aligned}\tag{5.33}$$

where $A_u(z), B_u(z), A_v(z), B_v(z)$ will be determined from the initial conditions as discussed below. It is useful to take the first derivative of (5.33),

$$\begin{aligned}\frac{\partial u_a(z, t)}{\partial t} &= f (A_u(z) \cos(ft) - B_u(z) \sin(ft)) \\ \frac{\partial v_a(z, t)}{\partial t} &= f (A_v(z) \cos(ft) - B_v(z) \sin(ft)).\end{aligned}\tag{5.34}$$

The solutions (5.33) may be rewritten in the form recognizable as that of the inertial oscillation

$$\begin{aligned} u_a(z, t) &= \sqrt{A_u^2 + B_u^2} \cos \left(ft - \cos^{-1} \frac{A_u}{\sqrt{A_u^2 + B_u^2}} \right), \\ v_a(z, t) &= \sqrt{A_v^2 + B_v^2} \cos \left(ft - \cos^{-1} \frac{A_v}{\sqrt{A_v^2 + B_v^2}} \right), \end{aligned} \quad (5.35)$$

where the amplitude of the inertial oscillation is given by $\sqrt{A_u^2 + B_u^2}$ or $\sqrt{A_v^2 + B_v^2}$; these terms will be shown to be identical below.

To determine the coefficients $A_u(z), B_u(z), A_v(z), B_v(z)$, the initial conditions are used to obtain

$$\begin{aligned} u_a(z, 0) &= B_u(z), \\ v_a(z, 0) &= B_v(z). \end{aligned} \quad (5.36)$$

By combining the initial condition for (5.31) and (5.34), we find that

$$\begin{aligned} \left(\frac{1}{f} \frac{\partial u_a}{\partial t} \right)_{t=0} &= A_u(z) = v_a(z, 0), \\ \left(\frac{1}{f} \frac{\partial v_a}{\partial t} \right)_{t=0} &= A_v(z) = -u_a(z, 0). \end{aligned} \quad (5.37)$$

The solutions (5.33) may now be expressed as

$$\begin{aligned} u_a(z, t) &= v_a(z, 0) \sin(ft) + u_a(z, 0) \cos(ft), \\ v_a(z, t) &= -u_a(z, 0) \sin(ft) + v_a(z, 0) \cos(ft) \end{aligned} \quad (5.38)$$

or

$$\begin{aligned} u_a(z, t) &= \sqrt{u_a^2(z, 0) + v_a^2(z, 0)} \cos \left(ft - \cos^{-1} \frac{v_a(z, 0)}{\sqrt{u_a^2(z, 0) + v_a^2(z, 0)}} \right), \\ v_a(z, t) &= \sqrt{u_a^2(z, 0) + v_a^2(z, 0)} \cos \left(ft - \cos^{-1} \frac{-u_a(z, 0)}{\sqrt{u_a^2(z, 0) + v_a^2(z, 0)}} \right). \end{aligned} \quad (5.39)$$

Note the one-quarter-period phase shift between $u_a(z, t)$ and $v_a(z, t)$.

According to this simple model, the amplitude and phase of inertial oscillations are strictly determined by the ageostrophic components of the wind at the time the inertial

oscillations commence. The inertial oscillations described in (5.39) have an amplitude equal to the magnitude of the initial ageostrophic wind at that level (Blackadar, 1957). This result is in agreement with the more complex numerical model results shown in Figures 5.18 - 5.20, where the amplitudes of the simulated inertial oscillations are compared with the initial magnitude of the ageostrophic wind. (In those figures, $u_a(z, 0)$ and $v_a(z, 0)$ at each z are determined from $u_a(z, 0) = u(z, 0) - u_g$ and $v_a(z, 0) = v(z, 0) - v_g$. Radiosonde observations of $u(z, 0)$ and $v(z, 0)$ are used in conjunction with the estimates of the geostrophic winds u_g and v_g from an average of winds between 2000 and 5000 m.) The influence of the initial ageostrophic component of the wind so dominates the turbulent term on the night simulated that actual evolution of k_m throughout the evening transition only slightly affects the generation of inertial oscillations above the stable surface layer. Within the stable surface layer, values of k_m , the gradient of k_m , and the gradients of the winds are expected to be larger, damping inertial oscillations within that surface layer. The strong dependence on the initial ageostrophic values only occurs when the turbulent term in the momentum equations is small compared to the $f u_a$ and $f v_a$ terms.

Because of this strong agreement, between the ageostrophic component of the winds at 2300 UTC, the amplitudes of the inertial oscillations generated in the numerical model, and the amplitudes of the inertial oscillations observed with the boundary-layer wind profilers, I conclude that on the night of 11-12 October 1999, the inertial oscillations observed do commence around 2300 UTC and are insensitive to the details of the evolution of k_m as a function of the stability. This case shows that when the geostrophic wind is constant in height and time, inertial oscillations can develop and persist whenever

$$\frac{\partial}{\partial z} k_m \frac{\partial u}{\partial z} \text{ and } \frac{\partial}{\partial z} k_m \frac{\partial v}{\partial z} \quad (5.40)$$

are small compared to the ageostrophic component of the wind. This development can

occur whether (5.40) becomes small during the evening transition, as observed here and proposed by Blackadar, or at another time. The terms in (5.40) can become small compared to the ageostrophic component of the wind when k_m decreases rapidly as convection dies out, or when the wind shear $\left(\left(\frac{\partial u}{\partial z}\right)^2 + \left(\frac{\partial v}{\partial z}\right)^2\right)^{1/2}$ becomes small. This explanation also accounts for observations of inertial oscillations that commence earlier in the afternoon in a layer with zero wind shear.

Mahrt (1981) remarks that “typical values of the daytime frictionally driven ageostrophic flow are far too small to account for the observed amplitudes of the nocturnal inertial oscillation.” In most of the inertial oscillations observed observed to commence near the evening transition during CASES-97 and CASES-99, the other mechanisms Mahrt discusses (baroclinicity due to sloping terrain, daytime baroclinicity, or temporal variations of the synoptic scale geostrophic wind) have been shown (Chapter 3) to interact with the inertial oscillation and modify the observed amplitudes. But on the night simulated here, 11-12 October 1999, the ageostrophic component of the wind at sunset does explain the amplitude of the observed inertial oscillations. On nights with more elaborate ageostrophic variation, only models that consider the variation of the geostrophic wind with height and time could be expected to encompass the observed variation of inertial oscillations and the influence of variations in eddy viscosity on those inertial oscillations.

Chapter 6

Conclusions and Suggestions for Future Research

“...perhaps the story you finish is never the one you begin.”

Salman Rushdie, “Midnight’s Children”, 1980

Inertial oscillations, circular patterns in the horizontal wind field, have a period equal to $2\pi/f$, where f is the Coriolis frequency. At a latitude of 45° , inertial oscillations have a period of 16.9 hours. These broad, gentle accelerations are generally understood as a common but unimportant consequence of the Earth’s rotation. However, nocturnal pollution dispersion cannot be understood or modeled accurately without consideration of inertial oscillations (McNider et al., 1988; Gupta et al., 1997; Banta et al., 1998). Additionally, inertial oscillations have been proposed as the cause of the formation of low-level jet (LLJ) structures in wind profiles (Blackadar, 1957; Buajitti and Blackadar, 1957; Thorpe and Gymer, 1977). These jets are important to climate (Stensrud, 1996) not only because of their prevalence over the Great Plains of the United States, but also because of their influence on large-scale convective storms (Rasmusson, 1968; Mo et al., 1995; Higgins et al., 1997; Wu and Raman, 1997). Through their shear-generated turbulence, LLJs can also disrupt airport operations (Kaplan et al., 2000).

The prevailing theory (referred to throughout this study as the “Blackadar mechanism”) explaining the presence of inertial oscillations in nocturnal winds identifies the decay of turbulence during the evening transition as the mechanism that allows the acceleration of winds into the inertial oscillation (Blackadar, 1957; Buajitti and Blackadar,

1957). Despite the relevance of inertial oscillations to the development of the LLJ, and though they are thought to be ubiquitous in the stable boundary layer, no quantitative climatology (until now) has documented the presence of inertial oscillations in the stable boundary layer. Additionally, until this study, no observation-based verification of their generation during the evening transition had been completed. Quantitative observations of inertial oscillations to date have focused on occurrences of inertial oscillations in the vicinity of fronts (Mori, 1990; Ostdiek and Blumen, 1997).

The dearth of observational studies of inertial oscillations is partially due to the limitations of observational technology: to resolve inertial oscillations in wind data, one must have a high-frequency (at least hourly) and long-duration (at least half a period, or eight hours in the mid-latitudes) dataset with several observation levels within the boundary layer. In this study, I use two months of data from boundary-layer wind profilers, which provide hourly observations of winds from approximately 150 m to 2000 m, to verify the presence of inertial oscillations and explore their generation. In Chapter 2, I discuss the limitations of the boundary-layer wind profilers. I also discuss the weaknesses of a low-level wind dataset from rawinsondes which derive wind speed and direction from Global Positioning System (GPS) satellites.

Chapter 3 offers observational evidence for the generation of inertial oscillations by the evening decoupling of the boundary layer. Using a new technique for identifying constant-amplitude inertial oscillations in time series of wind profiler data, I have documented the presence of inertial oscillations in boundary layer winds. This study shows that inertial oscillations are generated throughout the diurnal cycle, not solely in conjunction with the evening transition, as proposed by Blackadar (1957). They are also present during occurrences of low-level jet structures in the wind profile (Chapter 4) as proposed by Blackadar (1957) and Buajitti and Blackadar (1957).

In addition to establishing a climatology of inertial oscillations, I construct a climatology of the low-level jet using boundary-layer profiler data. The results of this

climatology are similar to those using rawinsondes, but I am able to confirm the theory that nocturnal low-level jets tend to have inertial oscillations at the height of their wind speed maximum.

On one night out of the two months of CASES data examined in Chapter 3, the vertical variation in amplitude of the observed inertial oscillations indicate that the one-dimensional Blackadar mechanism operates with little complication from other influences. (These other influences, which are shown to be active on other nights, include advection, subsidence, and a geostrophic wind changing in height or time.) Using a first-order closure numerical model, I simulate the observed development of inertial oscillations on that night, with good agreement between the model and the observations, as shown in Chapter 5 (Figures 5.12 - 5.14).

The amplitudes of inertial oscillations are found to be strongly correlated with the magnitude of the ageostrophic wind at model initiation, the time when surface heat flux changes sign, as shown in Figures 5.18 - 5.20. Though the correlations are high (greater than 0.75), the numerical model does not show exact correspondence between the initial ageostrophic wind and the amplitudes of the inertial oscillations, which are essentially averages over sixteen hours. This slight disagreement is shown to be a result of smoothing by persistent mixing at some levels: there are short periods of time where turbulent stresses are on the order of the ageostrophic component of the wind. Therefore, mixing periodically occurs, smoothing the vertical variation of the amplitudes. The amplitudes of the inertial oscillations are also found to be insensitive to the actual parameterization used for the turbulent stresses, which fluctuate on time scales much smaller than the period of the inertial oscillation (Figures 5.15 - 5.17). I return to Blackadar's simple analytical model to interpret the results of the numerical model. This model is based on the assumption that turbulent stresses are small compared to the magnitude of the ageostrophic wind. The results of this simple model confirm that the average acceleration in the winds is given by $f |\vec{u}_a|$, where $|\vec{u}_a|$ is the ageostrophic

magnitude of the initial wind profile.

Despite this agreement between the observations, the numerical model using first-order closure, and the simple analytical model on this night, the extension of these results to other nights and other situations is limited. First, the utility of the technique developed and used to identify inertial oscillations (discussed in Chapter 3) is restricted. It can only identify inertial oscillations that have a constant amplitude throughout the night. Garratt (1992) argues that the low-level jet atop the stable boundary layer arises due to a damped inertial oscillation because turbulent stresses never completely vanish. The increasing shear and turbulent bursts that have been documented within the stable boundary layer could further dampen the oscillations, perhaps to the point where this technique cannot identify them. Nights that do not pass the R_{ws} test, presented in (3.14) as a criterion for determining the presence of inertial oscillations, should not necessarily be considered “free” of inertial oscillations but rather free from constant-amplitude oscillations. Damped inertial oscillations are still expected to be able to modify plume dispersion (Banta et al., 1998) and cause a wind speed maximum at the top of the surface inversion (Thorpe and Guymer, 1977; Mahrt, 1981). Despite the limitations of this technique, applying it to more than two months of data could reveal seasonal variations in the occurrences of constant-amplitude inertial oscillations.

For future work on the influence of inertial oscillations on low-level jets, a more modern technique for jet classification and identification will be adopted. Although it is important to document agreement between state-of-the-art observations of LLJs and previous studies (as done here, in Chapter 4), current observational capacities far outstrip the data available to Bonner (1968), who established the current LLJ classification scheme used in this study and by Mitchell et al. (1995) and Whiteman et al. (1997). Bonner also noted that “it is not the intent of [his] study to examine the various kinds [of jets] in any detail - only to point out that they exist.” With better resolution both in height and in time, we should be to improve our exploration of the climatology

of low level jet structures. Because there is little variation in time of occurrence or direction between the jets in different Bonner categories, a classification scheme that could differentiate between sources of jets would be more useful. Rather than categorizing a jet based only on its maximum wind speed, we could quantify the level of supergeostrophicity of a jet structure. A criterion based on the height of the wind speed maximum relative to the surface inversion would also be useful to explore how often an inversion wind speed maximum occurs. Finally, the duration of the jet should enter into its classification. With such an improved classification scheme, the identification of the different flavors of low-level jet structures could promote a better understanding of their sources, especially their development from the inertial oscillation, which would lead to improved predictions.

This work has raised more questions than it has answered, especially with respect to the generation mechanisms of inertial oscillations. Although I have shown that the Blackadar mechanism, when applied to a first-order closure model, does describe the observed variations in inertial oscillations on one night, the interactions between inertial oscillations and other influences (as shown on the sixteen other nights discussed in Chapter 3) are still open for examination. Additionally, the numerous occurrences of inertial oscillations commencing at other times in the diurnal cycle should be explored. The model work has indicated that whenever the turbulent terms $\frac{\partial}{\partial z} \overline{u'w'}$ and $\frac{\partial}{\partial z} \overline{v'w'}$ terms become small compared to the ageostrophic term $f u_a$ and $f v_a$, inertial oscillations (with an amplitude equal to the magnitude of the ageostrophic component of the wind) can commence. Confirmation of this hypothesis through other modeling efforts that can capture more accurately the decay of turbulence, such as large-eddy simulation (Sorbján, 1997; Kosović and Curry, 2000), may prove helpful. Such simulations would require accurate profiles of winds, thermodynamic variables, and fluxes for initiation and comparison to observed inertial oscillations.

Further explorations of the generation of inertial oscillations should exploit the

numerous data sources from the CASES-97 and CASES-99 experiments. Although reliable profiles of winds between 55 m (the height of the CASES-99 tower) and 150 m (the lowest range gate of the boundary-layer wind profilers) are still difficult to obtain, other platforms (the Tethered Lifting System (Balsley et al., 1998), for example) can provide extensive datasets including high spatial and temporal resolution winds and thermodynamic variables. Future research programs, including the Vertical Transport and Mixing (VTMX) program (Doran, 2000), and platforms that are now in development for low-level winds, including lightweight remotely-controlled aircraft (Holland et al., 1992, 2001) will also offer opportunities to explore the generation and evolution of inertial oscillations and the decay of convective turbulence.

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Appendix A

Supplementary Spectral Analysis Figures

This study uses eight boundary-layer profiler datasets. In Section 3.1, power spectra of five of these datasets are shown. The power spectra reveal a diurnal-inertial peak in the wind data. The power spectra for the remaining three datasets are shown here for completeness.

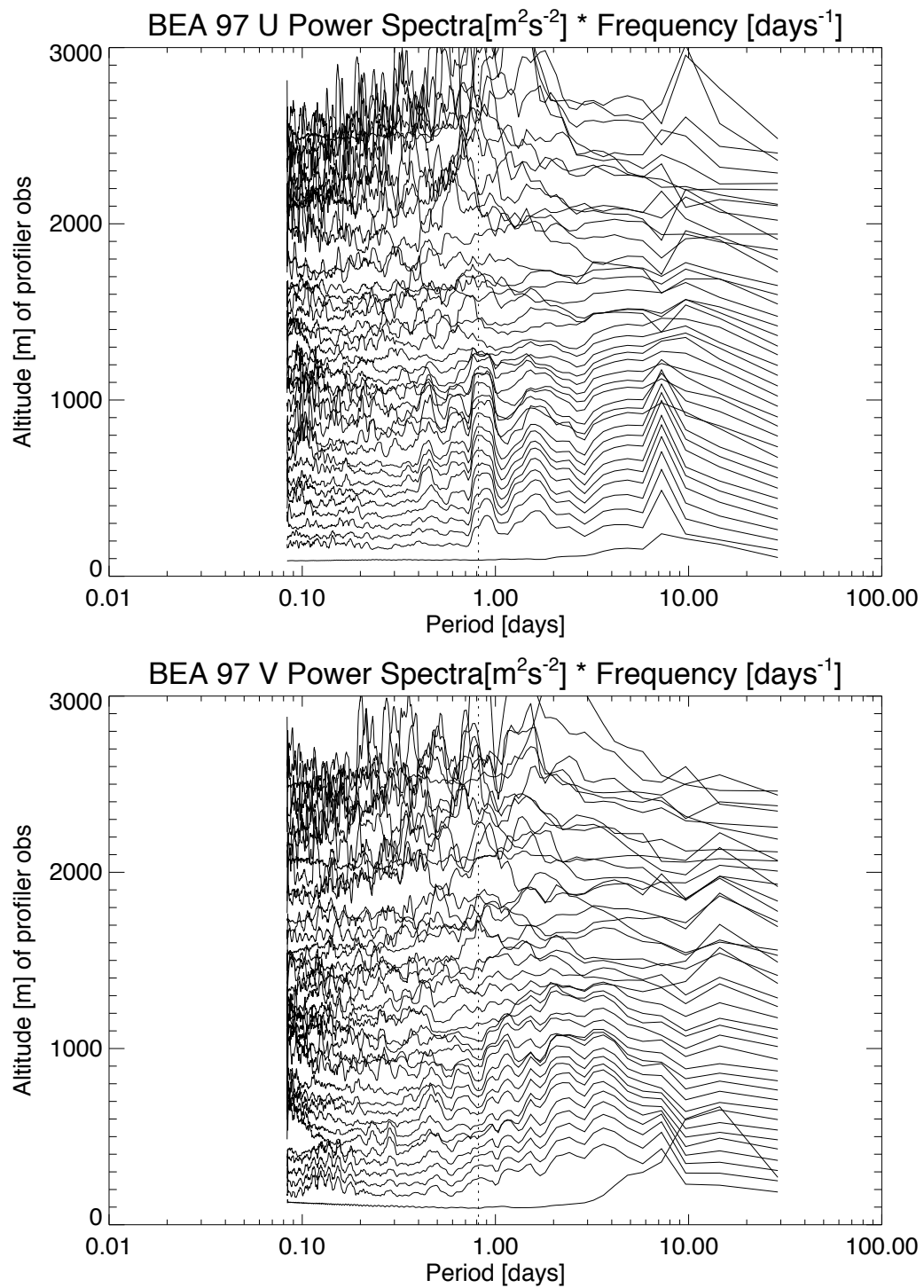


Figure A.1: Non-normalized power spectra for each level of data returned from the Beaumont wind profiler in May 1997. A constant has been added to each height to illustrate the vertical variation in the spectra.

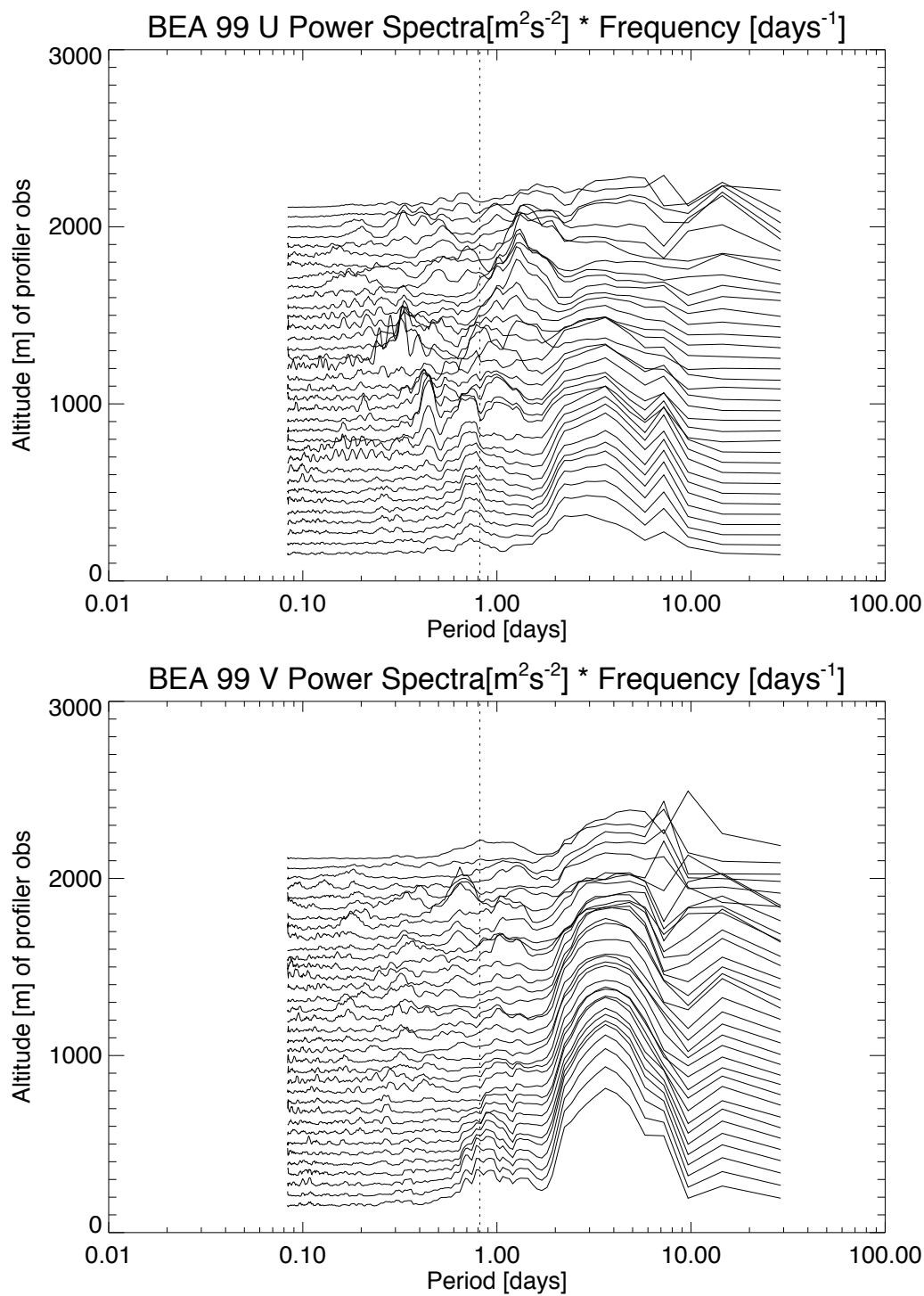


Figure A.2: As in Figure A.1 but for Beaumont, October 1999.

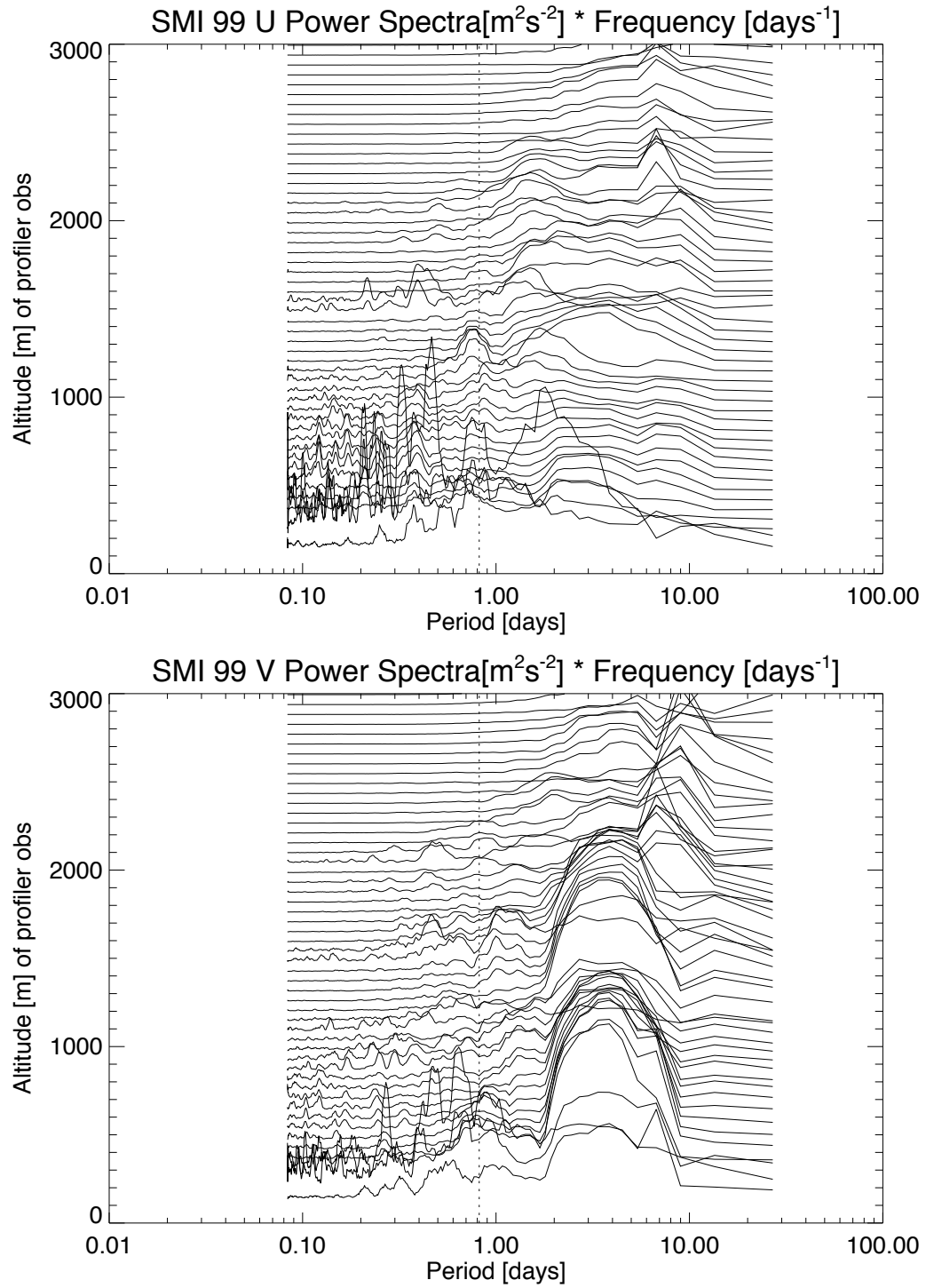


Figure A.3: As in Figure A.1 but for Smileyburg, October 1999.