Origin and maintenance of the stable boundary layer in a patchy landscape

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Origin and maintenance of the stable boundary layer in a patchy landscape

by

Luiz Eduardo Medeiros

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Abstract

Field observations made in Hudson Valley region, NY during the Hudson Valley Ambient Meteorology Study (HVAMS) are analyzed to examine how terrain and land cover influence nocturnal mixing in real-world landscapes. Important terrain features such as local topographic concavity and site sheltering are shown to exhibit systematic influence on turbulent intermittency and on the consequent nocturnal heat and momentum fluxes. Very local obstacles have their most important effects on mixing during strong winds (> 5m/s). Local terrain concavity was found to be the more important factor influencing surface fluxes than sheltering for all classes of winds.

Methodology is presented to define a regional bulk Richardson number $Ribr$ from operational and experimental soundings. Plots of $Ribr$ versus network averaged surface heat and momentum fluxes were obtained for the intensive operational and long-term operational phases of HVAMS. Results indicate that the need for extra mixing, above bulk Richardson number critical ($Ricr = \%$), in numerical weather prediction models (NWP), first invoked to avoid unrealistic cooling at surface and later hypothesized to be consequence of spatial averaging, is a consequence of spatial averaging of the local temperature and wind profiles in a heterogeneous landscape.

The area-site relative elevation ($mz - z$), a measure of terrain curvature, is a strong determinant of turbulent intermittency at the stations studied. For areas within one kilometer of a station this parameter is shown contain the same information as does the curvature of the least squares fitted local quadratic surface. The long-term data analysis
shows that the spatial variation of nocturnal surface temperatures depends on $mz-z$, a finding not apparent from data obtained during the six-week intensive operational period during 2003.

Elements of the regional stable boundary layer (SBL) heat budget were evaluated using measured network-averaged heat flux data, observed stable boundary layer (SBL) cooling rates, and modeled radiative flux divergence data. Results show that the radiative flux divergence plays the major role in SBL cooling (> 60%). The turbulent sensible heat flux divergence was found to play a smaller role (< 30% of the total cooling) but can be as high as 40% with increasing background wind speeds.
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1. Introduction

Motivation. Understanding the characteristics of mixing processes occurring in the stable boundary layer (SBL) has lagged behind that of the ones occurring in convective boundary layer (CBL) (Fernando and Weil, 2010), because the typical scale of the turbulence during stable conditions (tens of meters to a couple hundred meters) is smaller than that during convective conditions (on the order of a kilometer). The SBL usually occurs during the night overland, when the lower part of the CBL is cooled due to radiative flux divergence in the clear air and at the surface due to turbulent flux divergence, when turbulence stirs up the air near the surface causing it to loss heat the surface. Interpreting nocturnal turbulent fluxes is complicated when the mixing events are intermittent and spatially localized. One consequence of this ignorance is that predicting minimum temperatures and estimating the thickness of the SBL in realistic patchy landscapes, critical information for nocturnal air pollution prediction and for agriculture, remains a challenge.

In this thesis we examine turbulent intermittency both as a fundamental property of mixing in stable shear layers and as a consequence of sheared stable layers flowing over a real-world landscape made up of grassy clearings and forested patches. Modeling and laboratory studies are essential to study the fundamentals of idealized flows (e. g. Ohya et al., 2008), but field observations must be done to understand how relevant concepts based on the idealizations are to more realistic situations.

It is standard to use a dimensionless Richardson number to indicate the likelihood of turbulence and to determine whether or not a stably stratified shear layer is stable,
neutral, or convective. The gradient Richardson number $R_i_g$ is based on parameters external to the turbulence: $R_i_g = (g/\theta_{ref}) \partial \theta/\partial z/(\partial U/\partial z)^2$, where $\theta_{ref}$ is a reference temperature, $\partial \theta/\partial z$ the potential temperature gradient and $\partial U/\partial z$ the mean wind shear. When the derivatives are replaced by ratios of finite differences the ratio is known as the bulk Richardson number ($R_i_b$). Over the range $0 > R_i_b > -\infty$, turbulence is present, and it can be generated by convection and shear at same time. Strongly negative Richardson numbers indicate a situation for which turbulence is mainly generated by buoyant convection – the free convection regime. For $R_i_b \sim 0$ the layer is near neutral, and turbulence is mainly generated by shear and. For $0 < R_i_b < 1/4$, there is stable stratification and turbulence is generated by shear and destroyed by positive buoyancy. For $R_i_b \sim 1/4$ intermittent turbulence starts to occur; this is known as the critical Richardson number, $R_i_c$. Such intermittent turbulence can be generated as result of Kelvin-Helmholtz shear instabilities, by breaking gravity waves or by other instabilities forced by drainage flows. For $R_i_b >> 1/4$ the conventional notion is that turbulence is completely suppressed by the strong stability.

One practical issue is determining $R_i_c$. In theory and in wind tunnel studies, $R_i_c$ has been proven to vary around $1/4$ (Miles, 1961; Howard, 1961; Ohya et al., 2008). In the field, the requirement of making simultaneous local measurements of the gradients in time and space makes it difficult to determine the critical value. The $R_i_b$ criterion can fail when the turbulence is not generated locally, as in a situation with breaking gravity waves aloft or by shear instabilities forced by the low level flow aloft and is transported to where $R_i_b$ is evaluated. In this situation the local $R_i_b$ might not be coherent with local turbulence.
Problems with describing the SBL are reflected in persistent difficulties to parameterize the situation in meso- to large-scale models. Numerical weather prediction (NWP) models have difficulties on dealing with intermittent turbulence that occurs near the surface under very stable conditions (Fernando and Weil, 2010). In some cases, models produce too much cooling (Viterbo et al. 1999), while in others too thick a turbulent layer is predicted (McCabe and Brown 2007). This problem arises because NWPs adopt a classical concept of the SBL, in which there is continuous turbulence close to the ground. In many cases, a first order closure $K$-theory schemes are used to obtain the surface fluxes (McCabe and Brown, 2007). Turbulent diffusion coefficients required by such schemes are usually based on stability functions that depend on a bulk Richardson number.

Different kinds of stability functions are used to adjust the turbulent diffusion coefficients for momentum $K_m$ and heat $K_h$ in the sub-grid turbulence parameterizations in large-scale models. In modeling (Delage, 1997; Mahrt, 1987) it has been found necessary to allow mixing to occur even for grid-cell $Rib > Ricr$. In practice, this is done by using stability functions—relations that connect the Richardson number with local turbulent fluxes—with a “long-tailed” distribution, which allows infrequent extreme mixing events to occur at supercritical $Rib$. If the alternative is not to allow mixing under these conditions, then a function that decays fast with increasing $Rib$ (“short tail”) is preferred.

Absent these empirical corrections that have limited observational or theoretical support to date, models provide poor estimates of minimum temperatures and SBL thickness, which affects the predicted lifetime of low-pressure systems, and low-level jet height and strength can be misunderstood (Steeneveld et al. 2008). The short-tailed
distributions do not produce enough mixing, which leads to a weakly turbulent, shallow and cold SBL with insufficient drag. This leads to excessively deep and long-lived low-pressure systems. On the other hand, using the long-tailed distributions, by allowing more mixing, produces a warmer SBL with more drag, leading to better estimates of low temperatures and time durations of cyclonic systems. However, because of increased mixing, the simulated SBL is deeper and any low-level jet that is produced is weaker and at higher altitude than is observed. In aggregate, these problems limit the ability of such models to forecast dispersion of contaminants in the SBL.

Using a stability function that produces unphysical extra mixing for $Rib > Ricr$ has been justified by noting that unresolved subgrid surface heterogeneity causes turbulence to be localized spatially (McCabe and Brown, 2007). Mixing depends only on the local temperature and wind profiles. Landscape heterogeneities force spatial differences in the local profiles, and spatial averaging process attenuates such differences leading to a grid-cell $Rib > Ricr$. In this thesis, examining the degree to which spatial averaging accounts for the need of fluxes at ‘supercritical’ $Rib$ is one of our goals.

The observed SBL. In the conventional description of the stable atmospheric boundary layer under (usually nocturnal) in stable conditions with moderate to strong geostrophic wind, a turbulent layer extends from the surface up to tens of meters above the surface. Just above this turbulent layer, is where the vertical fluxes of scalars and momentum nearly vanish, and this height defines the thickness of the SBL (Brost and Wyngaard, 1978; Zeman, 1979). Other authors use the height of the potential temperature inversion $h_i$ to define the SBL thickness (Yu, 1978; Yamada, 1979; Nieuwstadt, 1980; Betts, 2006). Immediately above this layer, there is often a low level jet, from which the
turbulence extracts energy through shear production, while the negative buoyancy flux due to mixing in the presence of a stable density gradient is responsible for destruction of turbulent kinetic energy ($TKE$). Radiative flux divergence at the surface, throughout SBL, and in the layers above it, acts to increase static stability. Subsidence also strengthens the temperature inversion and therefore the stability. The mixing action of the turbulence counters these effects by redistributing the heat throughout the layer. The combined action of these agents defines the vertical profile of potential temperature ($\theta$), and therefore the stability. The wind profile is straightened by turbulent mixing but it can also be affected by baroclinicity, vertical changes in horizontal pressure gradient. The wind within the SBL is subgeostrophic due to friction but adjusts to near geostrophic at its top.

Such an idealized description of SBL can be approximately valid for flat homogeneous surfaces with small roughness subjected to continuous mixing. Such conditions are more likely to be found over oceans, flat deserts, and perhaps some ice caps with steady katabatic winds. Such special conditions are uncommon over land, especially on calm nights. The question that arises is: To what extent is the accepted model of the SBL relevant in complex heterogeneous terrain?

Over complex terrain, wind direction is not fully determined by the synoptic scale pressure gradient, the Coriolis force, and surface friction. Additional mesoscale pressure gradients (summing to the synoptic scale geostrophic wind) develop due to topographic and/or land cover contrasts. For example, it is well known that land-sea temperature contrasts can cause differences in the heating or cooling rate of the surface.
In valleys, wind can be forced by the topography to flow along the valley axis and as a result, the geostrophic wind may develop shear. Whiteman and Doran (1993) defined four processes that can explain the wind patterns within valleys. Considering only along-valley wind channeling, only two processes are most relevant to this study: *forced channeling, and pressure driven channeling*. In forced channeling, the geostrophic wind above the valley walls height is forced by the presence of valley-walls, to align and flow along the valley axis (Whiteman and Doran, 1993). In pressure driven channeling the pressure-gradient is aligned or has a component aligned to the valley axis that makes the wind to blow along valley axis (Whiteman and Doran, 1993; Fitzjarrald and Lala, 1989). Close to the valley surface, the wind field can be even more complicated due the presence of obstacles (trees, buildings, and etc.) that deflect the flow and can dominate very local mixing processes. Key to the following analysis is the *repeatability* of channeling in a valley. Research described below exploits the regularity of background winds on relatively calm nights to make observational ensembles of selected phenomena.

Under light winds and clear skies, many authors (e.g., Acevedo and Fitzjarrald 2003; Salmond and McKendry, 2002; Mahrt et al., 1998; Poulos et al., 2002, Nappo, 1991) have reported that there is a SBL with a strong temperature inversion very close to the ground, with intermittent or even vanishing turbulence. Under such conditions, continuous turbulence does not occur and another approach must be used to describe the SBL evolution and dynamics. Sometimes, when there is little turbulence, the near surface flow can remain disconnected from to the flow aloft (~ 100 m) (Banta et al., 2006; LeMone et al., 2003), and may respond to the effects of topography and land cover.
of the surrounding local terrain (Mahrt et al., 2001). Breaking internal waves (Finnigan et al., 1984; Einaudi and Finnigan, 1993; Sun et al., 2004), density currents associated with outflows from faraway storms or drainage flow (Sun et al., 2002), and Kelvin-Helmholtz (K-H) shear instabilities are all events usually associated to the generation of nocturnal turbulence. All these phenomena lead to flow instabilities that can grow with time. In some cases, the presence of a low level jet can be a trigger of shear instabilities (Salmond and McKendry, 2002). Nakamura and Mahrt (2005) used data from a network of six 5 m towers equipped with sonic anemometers, arranged in radii of 100m and 300 m around a 60-m main tower, and showed that intermittent turbulence can be localized spatially (e.g. vertical extension < 60 m and horizontal extension < 300 m).

The representativeness of wind, temperature, and other scalars observed at a given surface location is linked to topography and land cover/land use. These factors determine the buildup, placement, and exposure of stagnant air layers that have weak or no interaction at all with the background flow, and non-stagnant air layers that interact continuously with the background flow. Differences in the spatial scales of topography and land cover affect the SBL locally and non-locally. Local scales refer to features of the surface that have spatial extension smaller than the dominant feature of a region. In the case of a valley, for example, the dominant feature would be the valley’s average cross-sectional depression, whereas small bowls, hollows or hills within this are local features. Clearly there is some arbitrariness in the definition, since topographical patterns depend on how the scales are chosen, but these definitions can be made objective. For example, cold air pooling at the bottom of the valleys due to drainage of cold air from the valley sidewalls is a non-local consequence of the combined effects of the slope of the valley
sidewalls and radiative flux divergence close to surface. In this case, depending on the depth of the cold pool, the whole valley cross-section can be the most important feature of the topography regulating this phenomenon.

Acevedo and Fitzjarrald (2003, hereafter AF2003) addressed the problem of the spatial and temporal distribution of nocturnal turbulent fluxes for conditions of light winds and clear skies with patchy and intermittent turbulence. AF2003 related turbulent fluxes to station exposure measured by a transmission factor ($TF$; Fujita and Wakimoto, 1982). $TF$ is a measurable quantity of a network of observational sites that measures the degree to which a station is sheltered by surrounding obstacles. Input data are the wind direction and the wind speed in a network of stations exposed to a common wind field. In AF2003, $TF$ was augmented by considering topographic curvature near stations in a network of 26 automatic weather stations (PAM) stations, spread over an area 30 km x 30 km of complex terrain. They found that relatively open spots higher than surrounding hill tops in a radius of $\sim$ 3 km had more turbulent activity than ones relative sheltered and/or at places lower than its surroundings (bowls). In such conditions spatial heterogeneity of the surface leads to intermittent and localized mixing within a “real area” equivalent, to a grid cell domain, in which just a small percentage of the area has continuous or intermittent mixing while the remaining area has no mixing at all. An area average of the $Rib$ (which the models use) representative of a grid-cell domain is not appropriate to use in the stability functions because if it is above critical ($Ricr$), these functions will allow little or no mixing. AF 2003 compared fluxes obtained from area-average $Rib$ to direct area-average fluxes, and found that area-averaged $Rib$ approaches produce smaller fluxes representative of an area than would a direct area-averaged flux.
Another issue is the fact that much of the turbulence near the surface under these conditions could have been generated at higher levels as a consequence of a number of processes, such as breaking gravity waves, K-H shear instabilities, and density currents. If such is the case, a regular $K$-theory (Busch et al., 1976) cannot be the most appropriate model to describe turbulence near the ground (Sun et al., 2004).

Other studies have shown the complexity of the SBL and the phenomena that determine its structure. For example, Lemone et al. (2003) showed that when the Froude number $Fr \approx \left(\frac{1}{Ri_b}\right)^{0.5}$, $Fr << 3.3$ the flow close to the ground ($z \sim 2$ m) is decoupled from the one aloft. This can be a consequence of cooler air draining from higher surroundings and pooling in the low parts of a valley at night. This pooling determines the stability of the SBL from the surface up to some height above the valley floor where the air layer still decoupled. For $Fr \sim 3.3$, the turbulence close to the ground can be strong enough to keep the flow close to surface connected to the one aloft causing the air close to ground to flow dry adiabatically over the terrain (Lemone et al., 2003)) and therefore in this case turbulence activity is responsible for the stability of the layer. AF2003 anchored estimated fluxes at a site to its landscape features. However, because they lacked direct flux measurements they used empirical $Rib$ relations (Louis, 1979; McNider et al., 1995) to estimate the fluxes.

Even though these studies aimed to link observed regional and local flow regime to a background forcing (mesoscale pressure gradient), and to the landscape features, they were incomplete because they did not explicitly consider all the following aspects of landscape characteristics:
- **Topography affects flow patterns.** Examples include channeling in valleys and how very local curvature of the terrain – concave terrain (bowls) or convex terrain (hills) determines places where cold pooling may occur. Also, drainage flows along sloped terrain such as valley sidewalls are also likely to occur;

- **Land cover affects the radiative flux divergence.** Different types of surface have different absorptions/emissivity and reflectivity. Site exposure has a direct effect on mixing activity.

One aim of this thesis is to establish quantitative connections between the flow regime over an area with complex terrain and surface fluxes observed during stable conditions. We seek to understand better how the SBL develops in such realistic situations. To do this we develop a technique to estimate turbulent fluxes under stable conditions and indices to quantify turbulence intermittency, which in conjunction with topographic features of the local terrain, are used to analyze the distribution of turbulence over a network of observational sites under different stability conditions of a mesoscale forcing (e. g. mesoscale wind a cloud cover). The distribution of surface temperature with respect to the topographic features of network is also investigated to understand of the thermal structure of the SBL near the surface. In Chapter 2, we present details of the observations and modeled data used. In Chapter 3 our research plan to attack the problem is outlined, Chapter 4 examines the results of landscape influences on mixing and on temperature distributions. In Chapter 5, we evaluate mesoscale flow stability influences, through a developed regional Richardson number, on the network surface turbulence.
activity. Chapter 6 focuses on results of a bulk heat budget closure for the SBL. Chapter 7 treats the impacts of seasonal change in the state of the background flow on the turbulence behavior of a single site, and in Chapter 8 we discuss the potential overall results of this work. In the appendices are given details about instrumentation and locations of the experiments described in Section 2.1 of Chapter 2, results from a topographic assessment of all the sites, and other results not shown in the Chapters that were referred to in the thesis but not presented in the body of the text. References are given in Section 10.

1.2. Scientific Questions

*Main question* - How does the atmospheric stable boundary layer (SBL) form over realistic heterogeneous terrain?

*Follow-up questions*

- What is the role of topography and landscape structure in determining the turbulence activity and representativeness of local measurements?
- What does the background mesoscale conditions (wind and cloud cover) do to the regional turbulence activity?
- Which agent turbulent or radiative cooling is responsible for the bulk SBL cooling?

To address these questions we start from the premises that the overall dynamics of the SBL over the region under conditions of small major synoptic changes (no significant synoptic cold or warm advection or drastic synoptic pressure gradients changes) is fully controlled by the following external parameters (Van de Wiel et al., 2002):
• **Large-scale horizontal pressure gradient**;

• **Characteristics of the terrain surface** such as curvature, elevation, roughness, emissivity, surface and soil heat capacity, soil conductivity near the surface, each dependent on the land use/land cover, and topography;

• **Air emissivity**, which is a function of the temperature and moisture profiles.

• **Cloud cover fraction**; which alters surface upwelling longwave radiative fluxes.

Local physical processes (e.g., turbulence, gravity waves, K-H shear instabilities, drainage flows, low-level jets) that promote SBL mixing in the SBL are modulated by changes in these external parameters as they act together.

The approach is to identify regional and local (particular to surface station sites in the valley) modified bulk Richardson numbers ($Ri_b = \frac{g \Delta \theta \cdot h}{\theta_{ref} \Delta U^2}$), with height scale $h$, $\Delta \theta$ a characteristic temperature difference, and $\Delta U$ a scale wind speed difference between two levels. Other parameters are conventional. The hypothesis is that landscape heterogeneity causes sufficiently patchy mixing that area-averaged quantities differ substantially from in situ mixing indices. New regional indices are sought, ones based on properties of the background flow and along-valley pressure gradient (which enters in the $\Delta U$ scale), cloud cover (which enters in the $\Delta \theta$ scale), and features of the terrain surface (land use/land cover, and topography, which enter in the scaling of $h$, $\Delta U$ through the station exposure to the mean flow, and $\Delta \theta$ through the consequences of surface radiative cooling). Such non-local and local $Ri_b$ indicate the susceptibility of the whole valley to mixing and identify which regions are most likely to be mixed under given conditions of the background flow.
2. Data

2.1. *Hudson Valley Ambient Meteorology Study*

The Hudson Valley Ambient Meteorology Study (HVAMS) was an observational project whose intensive observational period (IOP) began in September 2003 with a broad network of observational sites and lasted until end of October 2003 with the complete network. A few surface sites remained in operation for several years following. The project was designed to understand the wind circulation and planetary boundary layer (PBL) in the Hudson valley region, NY. The study region is located between the cities of Poughkeepsie and Albany, NY. It corresponds to a north-south extension from approximately 42.8° N to 41.6° N, with an east-west extension from 73.5°W to 74.1°W (Figures 2.1, 2.2, and 2.5). This valley is about 300 km long from New York City up to Glens Falls NY. Just north from the city of Albany, the Mohawk River, flowing west to east, merges into the Hudson River. The Hudson Valley is typically 20 – 30 km wide with valley sidewalls varying from 200 m to 300 m high.
Figure 2.1: Perspective image of the Hudson Valley and environs. (NASA Shuttle radar Topography mission (13 February 2000) visible archive (http://visibleearth.nasa.gov/cgi-bin; Fitzjarrald and Freedman, HVAMS project description 12/13/02).
Figure 2.2 shows all the sites and instruments location during the HVAMS intense observational period (IOP). All numbers alone or with the “Pr” abbreviation, refer to Portable Automated Mesonet stations from NCAR (PAM, from now on referred by the word “station” and by its number (see appendix A.1 for details), each equipped with eddy-covariance flux system, that operated for ~1.5 months, during the IOP. “H” with a number at the right side refers to HOBO weather stations (hereafter HOBO and its number – see A.1) during the IOP, and after during the long-term observational period (LTOP). “G” refers to the Grieg Farm site, where a 20 m micrometeorological tower (station 10 hereafter) was installed during the IOP, and after during the LTOP. At this
site a tethered balloon (the NCAR Tethered Atmospheric Observation System - TAOS – see A.1) took several vertical profiles of wind speed, wind direction, temperature, relative humidity and pressure up to ~ 200m above ground during 11 nights of the IOP. These profiles usually were performed when there were conditions of clear skies with calm surface winds. When available, this balloon data together with the sodars, vertical wind profilers, and aircraft will be references for the wind aloft (background flow See Table 5.2 on Chapter 5) during the IOP.

During the IOP the King Air instrumented aircraft (University of Wyoming), capable of taking fast (25 Hz) temperature, wind, and pressure measurements, flew over the region and made special ‘close approach’ temperature and wind profiles with good vertical resolution at five small airfields between the cities of Poughkeepsie and Albany. Detailed profiles of the lower atmosphere were made at locations “1” (Alexander Farm), “8” (South Albany), “1B1” (Columbia county airport), “4” (Green Acres), and “M” (MIPS at Kingston Ulster). See Figure 2.2 for location on the map and Appendix A.1 for longitude, latitude and elevation. The aircraft measurements are used to complement TAOS data. The limitation of aircraft data is that the close approaches were done only at dusk and dawn transitions times, and therefore there is no information over the course of the night. For the LTOP, we must use sounding data during at 00UT and 12UT made at the Albany National Weather Service Forecast Office (ALB). A limitation of the standard soundings is that the vertical resolution of the atmosphere near the surface is poor and they are only available twice daily.

ALB refers to 00UT and 12UT soundings and hourly standard automated surface weather data (ASOS) at Albany airport, and POU to ASOS at Poughkeepsie airport. The
Albany ASOS also has ceilometer data, which are used to estimate the cloud cover fraction of the entire HVAMS region (See Table 5.2 on Chapter 5). SCH refers to the NOAA 915 MHz radar wind profiler at the city of Schenectady airport. “M” stands for the set of profiling instruments and one surface weather station from the University of Alabama-Huntsville that operated at the Kingston-Ulster airport during the IOP. This equipment included a 915 MHz radar wind profiler. A more detailed description including site location, elevation, instrumentation, periods of operation during the IOP, and as well after the IOP can be found in the appendix A.1.

2.1.1 Land use / Land cover and topography of the study region on the Hudson Valley

The 30 m x 30 m resolution land use/land cover and topography maps (Figure 2.2 and Figure 2.3) of the HVAMS domain region were obtained from USGS Global Land Information System. The land use / land cover maps were built for the entire region of the project, and as well as for 3 km x 3 km station study areas centered on each of the 10 flux stations. The surface types are classified in 9 main classes: Urban or Built-up Land, Agricultural Land, Rangeland, Forest Land, Water, Wetland, Barren Land, Tundra (outside HVAMS region), and Perennial Snow or Ice (also outside HVAMS) (Details of the subclasses derived from these can be found on Anderson et al. 1976). Table 2.1 contains the land use/ land cover partitioning amongst the classes for the whole HVAMS domain region and also for the ten station (all PAMs and station 10) study areas.
### HVAMS land cover / land use distribution

<table>
<thead>
<tr>
<th>Land cover / Land use types</th>
<th>HVAMS</th>
<th>St1</th>
<th>St2</th>
<th>St3</th>
<th>St4</th>
<th>St5</th>
<th>St6</th>
<th>St7</th>
<th>St8</th>
<th>St9</th>
<th>St10</th>
</tr>
</thead>
<tbody>
<tr>
<td>Water (%)</td>
<td>2.3</td>
<td>0.5</td>
<td>0</td>
<td>6.5</td>
<td>0</td>
<td>1</td>
<td>23.9</td>
<td>0.1</td>
<td>0</td>
<td>2.1</td>
<td>0.5</td>
</tr>
<tr>
<td>Urban (%)</td>
<td>11.5</td>
<td>6.3</td>
<td>11.2</td>
<td>7.4</td>
<td>8.4</td>
<td>7.3</td>
<td>11.2</td>
<td>10.6</td>
<td>24.5</td>
<td>5</td>
<td>14.1</td>
</tr>
<tr>
<td>Barren (%)</td>
<td>0.2</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>2.6</td>
<td>0</td>
<td>0.2</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>Forest (%)</td>
<td>59.6</td>
<td>47.4</td>
<td>36.6</td>
<td>67.4</td>
<td>36.5</td>
<td>51.3</td>
<td>22.5</td>
<td>50.9</td>
<td>22.7</td>
<td>37.9</td>
<td>27</td>
</tr>
<tr>
<td>Rangeland (%)</td>
<td>2.1</td>
<td>10.4</td>
<td>0.3</td>
<td>1.2</td>
<td>9</td>
<td>3.9</td>
<td>0.6</td>
<td>0.8</td>
<td>0.5</td>
<td>4.3</td>
<td>6</td>
</tr>
<tr>
<td>Agricultural (%)</td>
<td>16.5</td>
<td>31.1</td>
<td>20.8</td>
<td>13.8</td>
<td>45.1</td>
<td>35.6</td>
<td>15.9</td>
<td>3.2</td>
<td>31.1</td>
<td>49.9</td>
<td>52</td>
</tr>
<tr>
<td>Wetland (%)</td>
<td>7.8</td>
<td>4.2</td>
<td>31.2</td>
<td>3.8</td>
<td>1</td>
<td>0.8</td>
<td>23.3</td>
<td>34.5</td>
<td>21</td>
<td>0.8</td>
<td>0.4</td>
</tr>
</tbody>
</table>

Table 2.1: Percentage of land use / land cover of the whole HVAMS area domain (144.498 km x 55.213 km) and (3 km x 3 km) areas centered on each of the 10 sites of the HVAMS with fluxes measurements.
Topographic characteristics such as concavity, slope magnitude, and slope aspect of all the flux sites and for a central area (see Fig. 2.4 a) in the study region were obtained using USGS topographic map with a ~ 30m x 30m resolution (see Table 2.2.). The other topographic characteristic referred here as “mz - z” or relative elevation, which indicates whether or not a station’s elevation is higher or lower than the average elevation of the immediately surrounding area, was evaluated only for the station sites. In the limit of the four nearest points around the central point, mz - z can be interpreted as the discrete Laplacian of the topography:
\[
\frac{\Delta z(x,y)}{4} = \frac{z(x-h,y) + z(x+h,y) + z(x,y-h) + z(x,y+h) - 4z(x,y)}{h^2},
\]

where \( z \) is the elevation, \( x \) a distance along the west-east direction, \( y \) a distance along south-north, and \( h \) the lateral length of the area around the central point \((x,y)\) (Roache, 1976). In such a small limit, \( mz - z \) could be used to determine the site surface concavity because the Laplacian of a surface in 2-dimensions is the concavity. Here, instead of using the nearest four points around the station, we use the elevation average \((mz)\) of all the points within an area of \(\sim 2 \text{ km} \times 2 \text{ km} \) minus central point elevation (exactly the site elevation). Within this wider area range, the parameter only approximates the concavity.

The concavity or curvature was obtained by taking the Laplacian \((\nabla^2 z)\) of a quadratic surface \((z(x,y) = ax^2 + by^2 + cxy + dx + ey + f)\), which was fitted to an area of \(\sim 0.3 \text{ km} \times 0.3 \text{ km} \) around the stations. The fitting was done by the least-squares method through the R package routine function \(\text{surf.ls}\) implemented in the statistical software R-project (From:http://www.r-project.org) See Fig. 2.4 b and Appendix A.2 for fitted surface and fitting quality. The sign of the quantity \(4ab-c^2\), obtained using the fitted quadratic surface coefficients, gives an extra insight about local topographic. If \(4ab-c^2 > 0\) there is local minimum \((a \text{ and } b > 0)\) or maximum \((a \text{ and } b < 0)\), and otherwise neither minimum.

The local slopes were obtained by taking the magnitude of the gradient \((|\nabla z|)\) of the fitted surface to the local topography, and the aspect by the direction of this gradient \(\nabla z\) vector (using the meteorological direction convention). Concavity, slope, and aspect were evaluated at the center of the regions around the stations and are presented together
with $mz-z$, elevation, and $4ab-c^2$ on Tables 2.2 and 2.3. The central HVAMS study area, was also fitted a quadratic surface to obtain the main or large-scale valley concavity (Figure 2.4 (a) and (b)), but the west and east slopes were obtained in a different manner. To obtain these slopes the central HVAMS study area was divided exactly in middle, and the slope of both sides were obtained by a fitting a planar surface and taking $|\nabla z|$. 

We explore the influences of these site characteristics on the observed local flow in Chapter 4 by relating it to the turbulent fluxes, minimum and maximum temperatures.

![Figure 2.4: a) Topography of the central of the Hudson Valley study region. b) Fitted quadratic surface to the region in a). This approach was used as well for areas of ~ 0.3km x 0.3km around each station. The geometrical meaning of the concavity is as follows: convex (-), concave (+), and flat surface (plane). The magnitude of it indicates how concave or convex the surface is. The approximated surface (b) is concave (concavity $\approx 0.12$ m$^{-1}$), as expected.](image)
Table 2.2: Slope, Concavity, and elevation evaluated at the center of ~0.3 km x 0.3 km areas around the flux sites. Concavity, and \((4ab - c^2)\) parameter evaluated at the same area sizes, and \(mz - z\) parameter for area-sizes of ~2km x 2km, however representative for the entire area domain.

<table>
<thead>
<tr>
<th>St.</th>
<th>Slope (°)</th>
<th>Concavity ((x10^{-5} \text{ m}^{-1}))</th>
<th>Maximum/minimum ((4ab - c^2))</th>
<th>(mz - z) (m)</th>
<th>Elevation above sea level or Hudson River (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>2.30</td>
<td>49.66</td>
<td>No</td>
<td>20.67</td>
<td>156</td>
</tr>
<tr>
<td>2</td>
<td>2.71</td>
<td>-20.16</td>
<td>Maximum</td>
<td>1.39</td>
<td>47</td>
</tr>
<tr>
<td>3</td>
<td>1.22</td>
<td>-5.29</td>
<td>No</td>
<td>3.70</td>
<td>45</td>
</tr>
<tr>
<td>4</td>
<td>0.86</td>
<td>0.82</td>
<td>No</td>
<td>-2.86</td>
<td>94</td>
</tr>
<tr>
<td>5</td>
<td>3.99</td>
<td>-171.36</td>
<td>Maximum</td>
<td>-44.93</td>
<td>108</td>
</tr>
<tr>
<td>6</td>
<td>3.15</td>
<td>-52.81</td>
<td>No</td>
<td>-14.31</td>
<td>25</td>
</tr>
<tr>
<td>7</td>
<td>1.74</td>
<td>29.92</td>
<td>Minimum</td>
<td>13.74</td>
<td>133.2</td>
</tr>
<tr>
<td>8</td>
<td>0.43</td>
<td>-9.89</td>
<td>No</td>
<td>-7.56</td>
<td>53</td>
</tr>
<tr>
<td>9</td>
<td>1.07</td>
<td>-39.68</td>
<td>No</td>
<td>-14.67</td>
<td>76</td>
</tr>
<tr>
<td>10</td>
<td>0.44</td>
<td>0.88</td>
<td>No</td>
<td>-0.91</td>
<td>64.9</td>
</tr>
<tr>
<td>11</td>
<td>1.75</td>
<td>-20.81</td>
<td>No</td>
<td>3.85</td>
<td>202</td>
</tr>
<tr>
<td>12</td>
<td>1.58</td>
<td>-1.24</td>
<td>No</td>
<td>21.79</td>
<td>133</td>
</tr>
<tr>
<td>13</td>
<td>2.13</td>
<td>-9.02</td>
<td>No</td>
<td>47.16</td>
<td>559</td>
</tr>
<tr>
<td>14</td>
<td>2.02</td>
<td>9.32</td>
<td>Minimum</td>
<td>-4.33</td>
<td>414</td>
</tr>
<tr>
<td>15</td>
<td>2.78</td>
<td>-10.32</td>
<td>No</td>
<td>11.48</td>
<td>128</td>
</tr>
<tr>
<td>16</td>
<td>0.18</td>
<td>-17.55</td>
<td>Maximum</td>
<td>5.01</td>
<td>3</td>
</tr>
</tbody>
</table>

Table 2.3: Concavity, slope, and orientation for the central area (74.2°W – 73.6°W and 42.6°N – 41.8°N) of the study region.

- Slope (deg) | West side | East side | Central region |
- Slope aspect/orientation (deg) | 98.97 | 89.41 | - |
- Concavity \((x10^{-5} \text{ m}^{-1})\) | - | - | 0.12 |
2.1.2 Site Sheltering

Some authors Wieringa (1976; hereafter W1976) and Fujita and Wakimoto (1982; hereafter FW1982) have used measures of local site sheltering to correct surface measurements of wind from obstructed directions. In the W1976 approach, data from only one standard weather station is required. The correction is based on the hypothesis that wind gusts observed over certain period of time do not reflect local retardation effects due to isolated obstacles located upwind from the station, and therefore can be used in association to mean wind (averaged over the same time interval) to obtain the gust factor \( G(\theta) = \frac{\bar{U}}{U_{max}} \). The gust factor is empirically related to the variance of the horizontal wind speed \( \sigma_u \), and from such relationship is obtained an estimate for the upwind roughness length \( z_0 \) (based on gust factor). Wind speeds are corrected by hypothesizing a logarithmic wind profile and assuming that at a blending height \( H_b \) all wind profiles over the region converge to a unique wind speed. Such method is suitable for well-sited stations (e.g. typically airports with an upwind fetch of 150 m, at least a distance of 15 times larger than the highest obstruction in the upwind direction, for winds \( \geq 6 \text{m/s at 10 m above ground level (AGL)} \). By restricting the winds to a certain values, the condition of approximately neutral flow is achieved and the Monin-Obukhov similarity hypothesis wind profile relations perform well.

In FW1982 approach, the winds are corrected by using a transmission factor \( TF \), an empirical quantity that can be obtained when one has data from a network of weather stations. The \( TF \) is estimated for each station of a network by taking a station’s average wind speed from a given direction and dividing it by the maximum average wind speed over the network for that same direction. Thus, \( TF = 1 \) indicates an unobstructed site, and...
$TF = 0$ means that a station is completely blocked (e.g. FW1982). In both works cited above, the authors’ main focus was obtaining corrections for sheltered wind directions of a weather station that could lead to estimates of wind speeds without obstructions. In the present work we use their approach to determine the positioning of blockage nearby the stations (See Figure 2.5), and we investigate the influences of such blockage have on the spatial distribution of heat and momentum fluxes.

Figure 2.5: Topographic map of HVAMS region with the transmission factor ($TF$) for all flux and weather stations.

$TF$ was calculated for all the 10 flux sites using a wider network of stations than the flux stations. This wider network includes the ten flux stations plus eight conventional surface weather stations (Figure 2.5).
2.1.3 Experimental array during the LTOP

Data acquisition by PAMs, TAOS, sodar, MIPS, and the instrumented aircraft stopped following the IOP. Two HOBOs were moved from their original sites to replace two PAM sites, and a new HOBO was installed at Schodack Island. Table A.1.4 in the Appendix A.1 contains information about the dates of operation and locations of the weather stations that were relocated/installed. (See also appendix A.1 for instrument changes). During the LTOP only station 10 continued with direct measurements of turbulent fluxes and 4-levels profiles of wind, temperature, relative humidity (RH), and 5 HOBOs with basic 1-minute measurements of meteorological variables (Table A.1.2 lists the sensors). The rest of the sites SCH, ALB, and POU are permanent weather stations as listed in Table A 1.1.

2.1.4 Mesoscale model Meso-NH

During the HVAMS IOP model output data from simulations of the Meso-NH model are available for a sequence of four days and nights. Meso-NH is a non-hydrostatic mesoscale model developed by Laboratoire d’Aerologie (UMR 5560 UPS/CNRS) and CNRM-GAME (URA 1357 CNRS/Meteo-France) that can simulate the atmospheric dynamics from synoptic scale down to large-eddy scales (From: http://mesonh.aero.obs-mip.fr/mesonh/).

A general description of the model operation is given by Lafore et al. (1998)). We focus on the turbulence closure schemes, and radiation schemes of the model because we use modeled radiation and turbulent data. The radiation scheme takes into account absorption and emission for the longwave radiation, and reflection, scattering and
absorption for solar radiation in the atmosphere and surface. The shortwave radiation scheme uses a two-stream approach to solve the radiative transfer equation. The longwave radiation scheme assumes a broad band spectral emissivity method (Rodgers, 1967) to solve the radiative transfer equation (Morcrette, 1990) and it consider as absorbers H₂O, CO₂, O₃, CH₄, N₂O, Chlorofluorocarbons (CF C11 and CF C12), and aerosols. The absorption coefficients are pressure and temperature dependent. The effects of clouds are considered and the absorption/emission depends on the liquid water content of the clouds, however there is no scattering for long-wave. The radiation scheme is not called every time step because it demands great computational time, and is called at every 30 minutes, while the turbulence scheme is called every time step (2 seconds).

The turbulent closure is one-and-half order and the turbulent kinetic energy (TKE) is determined by a prognostic equation. So, there has to be conservation of TKE. The turbulent fluxes on the other hand are determined based on the TKE, and the gradients of the temperature and wind. The fluxes are diagnosed using a K-theory, where the eddy-diffusion coefficient is dependent on a mixing length, TKE, and stability function. The mixing length and stability functions are functions of TKE, and the potential temperature and wind gradients. A complete description of the model can be found on http://mesonh.aero.obs-mip.fr/mesonh/doc.html.

The series of simulations (See Table 2.4 and Figure 2.6) we use in this study was performed by the meteorology research center of the Universitat de les Illes Balears (UIB), in Mallorca, Spain by our colleagues Drs. Joan Cuxart and Maria Antonia Jiménez. Such model output is used to investigate the pooling of cold air at the bottom of the Hudson valley and to estimate in situ radiative flux divergence across the SBL. It is
useful to compare the total bulk rate of cooling of the SBL to the turbulent cooling and in situ radiative cooling, and estimate what percentage of the cooling of the layer is accomplished by mixing. The UIB website describing these simulations is available at http://turbulencia.uib.es/hudson_week/hudson_week_main.htm.

<table>
<thead>
<tr>
<th>Simulation period</th>
<th>Started at 1800UT on 10/05/2003 and ended at 1800UT on 10/09/2003. Four nights of simulation available</th>
</tr>
</thead>
<tbody>
<tr>
<td>Area domain size</td>
<td>400km (West-East) x 540 km (South–North)</td>
</tr>
<tr>
<td>Area centered at</td>
<td>74°W and 42.75°N</td>
</tr>
<tr>
<td>Horizontal grid cell size (horizontal resolution)</td>
<td>2km x 2km</td>
</tr>
<tr>
<td>Vertical grid cell size (vertical resolution)</td>
<td>3m in the first 104 m but stretches above (See Figure 2.6 for vertical resolution details). The lowest model level is at 1.5m (AGL).</td>
</tr>
<tr>
<td>Number of grid points</td>
<td>200 (West–East) x 270 (South–North) x 85 (Vertical)</td>
</tr>
<tr>
<td>Model time step</td>
<td>2s but the radiation schemes are updated only every 30 minutes.</td>
</tr>
<tr>
<td>Model schemes</td>
<td>Turbulent, radiative, surface and condensation</td>
</tr>
</tbody>
</table>

Table 2.4: Meso-NH model details (Adapted from J/ Cuxart and M. A. Jiménez, HVAMS simulations).
2.2. The Harvard Forest Environmental Measurement Site (HF - EMS)

The HVAMS region is a composite of urban clusters, open fields, and forested patches. During HVAMS, no observational site was located in deciduous forest. To correct the deficiency we include data from a well-documented long-running flux tower site at Harvard Forest, located in the type of forest common in the nearby Hudson Valley. The flux measurements at the Harvard Forest (“HF”) have been conducted by Harvard University, ASRC and other groups since 1989. HF sites are located in deciduous forest near Petersham, MA, USA. A more complete description of the site characteristics can be
found on Sakai et al. (1997) and at the website (http://www-as.harvard.edu/chemistry/hf/hfsite.html).

Though the HF data set is more than a decade long, we use only the data period that overlapped with the time that a NOAA 915 MHz wind and RASS temperature profiler was operating at the Orange MA airport, a few kilometers to the west of the flux tower site. The main observational site is the 30-m micrometeorological tower at the EMS site. Data available includes turbulent measurements (momentum, heat, water vapor, and CO₂ fluxes), meteorological and micrometeorological variables, however we use only momentum and sensible heat fluxes data. The 915 MHz data is used to obtain the regional Richardson number. In Chapter 5 is given a description how the regional Richardson number was calculated and its results.

Figure 2.7 shows the local topography around the EMS tower site. The local concavity was evaluated in a similar fashion that was done for the HVAMS stations and was found to be concave (0.27 m⁻¹) for an area-size similar to the ones for the HVAMS sites (~0.3 km x 0.3km).
Figure 2.7: Topography of a 2km x 2km area around the HF flux tower (EMS). The tower is located at (0,0). The vertical black bar shows the tower. Note that the vertical axis-z (elevation) is scaled for viewing. (Adapted from Staebler, 2003).
3. Turbulence breakdowns

3.1 Objective

The objective here is to illustrate the complexity of the turbulence under stable conditions, develop a technique to identify intermittent turbulent events and estimate fluxes during intermittent and calm periods. To do this, we use fast response data (10Hz) averaged down to 1Hz from 3D-sonic anemometers at the flux stations.

The behavior of the Nocturnal Boundary Layer (NBL) is still not well understood. One of the biggest challenges is to determine the turbulent fluxes of scalars (such as temperature, CO$_2$, H$_2$O) and momentum (Howell and Sun, 1999; Acevedo et al., 2006) in stable conditions. On nights with calm wind and clear skies, the eddy covariance technique might not work well because the turbulence is neither stationary (Baldocchi, 2003) nor continuous. In such cases, the turbulence might be intermittent, occurring in bursts. In these situations, distinct from daytime when the turbulence is continuous, the choice of the best time window to make Reynolds averages for the fluxes becomes extremely difficult. It can vary for every intermittent event depending on its time duration (Acevedo et al., 2006).

Here, however, we use the standard deviation of the turbulent vertical velocity to identify turbulent mixing periods that are responsible for an appreciable part of the nocturnal flux. In a two-step process we search first or periods when there are abrupt changes in this variable that signify the beginning of periods with turbulence. In the second step, a threshold is applied to the chosen variable to identify calm and turbulent periods, characterizing the latter as intermittent turbulent breakdown episode.
3.2 Observing elevated mixing events — TAOS

Using data from the NCAR tethered balloon measurement system TAOS during the IOP (Chapter 2), we can identify interesting characteristics of turbulence under stable conditions. Between ~1930H and ~2000H of 10/13/2003, the TAOS measurements at 50 m and 60 m show disconnection from the upper levels (>100m), where turbulence decreases. During this period the wind speed decreases and the wind direction is no longer close to the wind direction of the higher levels, which appears to be still connected to the mesoscale flow aloft. Time series of Rib for this night support this idea (Figure 3.3 (a)). During this period the Rib between the lowest and second lowest levels increase to values well above Ricr, while for the remaining levels Rib stays close to Ricr. Suddenly after this period between ~2000H and 2030H, a change in the wind direction of the lowest two levels increased the wind shear, provoking mixing that raised the potential temperature at the lowest two levels to a value close to that at the highest levels. During this and similar periods the Rib between the lowest two levels have instances for which conditions are subcritical (<Ricr), suggesting the presence of turbulent breakdowns. At the same time the Rib between the highest two levels also shows periods of intermittent turbulence, while the two middle levels stay supercritical (> Ricr). After this time, turbulence decreases at all levels, even though the wind speed increases except at the first level.

This example shows that the mechanism that generated turbulence above was not the same as that causing turbulence at the surface. Turbulence near the surface reflects interaction with the surface, while mixing above may have been due to another kind of shear instability, perhaps breaking gravity waves (Finnigan et al. 1984; Einaudi and
Regardless of the nature of the mechanism, our work is primarily interested in the mixing events that are connected with the surface, because only mixing close to surface can exchange momentum and scalars with the overlying atmosphere. Intermittent events above lead to internal stirring of the SBL and should affect neither bulk properties, (e.g. SBL bulk heat budget, Chapter 6) nor the surface temperatures.

Figure 3.2 illustrates a case exhibiting continuous turbulence, with higher winds. Under this circumstance the wind direction and potential temperature at all the TAOS levels were close. Comparing the temperature spread between the levels for this night 10/03/2003 with those for night of 10/13/2003 indicates that there was a well defined turbulent layer from the surface up to at least to the highest TAOS level (150m AGL). 

The identification of intermittent events, under these conditions, is of little use as turbulence at the surface was present at all times. It makes more sense to treat the whole period as one unique event. Even though the turbulence can be continuous, it can vary in intensity and the fluxes would have to be calculated during periods with stationary turbulence. In the following section, we develop a technique to identify intermittent and continuous turbulence.
Figure 3.1: Upper left panel – potential temperature time series, upper right panel – wind speed time series, lower left panel – wind direction time series, and lower right – sonde height (AGL) of the five TAOS levels. This flight occurred during the hours of 1735 and 2107 (LST) on 10/13/2003. For local hour is needed to subtract 4 hours from the universal time (UT). Note that a common color code refers to levels among the different panels. At each level a sonde measured wind, temperature and pressure (See Chapter 2). Data record frequency ~ 1Hz.
Figure 3.2: Upper left panel – potential temperature time series, upper right panel – wind speed time series, lower left panel – wind direction time series, and lower right – sonde height (AGL) of the five TAOS levels. This flight occurred 1735 – 2107 (LST) on 10/03/2003.
Figure 3.3: (a) – Four Rib time series obtained from the five TAOS soundings for night of 10/13/2003 between the hours of 1735 – 2107 LST. The 57m (blue line) represents the lowest two levels, the 84 m (green line) the second lowest and the middle levels, 125 m (red line) the second highest and middle levels, and 171 m (black line) the highest and second lowest levels. The 1Hz data was averaged down to 5 minutes to obtain the Rib’s.
Figure 3.3 (b): Four Rib time series obtained from the five TAOS soundings for night of 10/03/2003 between the hours of 1735 – 2107 LST. The 16 m (blue line) represents the lowest two levels, the 43 m (green line) the second lowest and the middle levels, 96 m (red line) the second highest and middle levels, and 137 m (black line) the highest and second lowest levels.
3.3 Technique to identify intermittent events

Periods of stationary turbulence were identified by selecting intervals in the perturbation vertical velocity ($w'$) time series for which its variance

$$\sigma_w^2 = \frac{1}{N} \sum_{i=1}^{N} (w_i - \bar{w})^2,$$

where $N$ is the total number of points in a chosen interval, $i$ is the $i$-th velocity element, $w'$ the velocity perturbation and $\bar{w}$ was approximately constant.

Howell and Mahrt (1995) developed this technique to identify these periods with constant $\sigma_w^2$. Originally this algorithm was designed to search for abrupt changes in a moving average of a time series of any variable using the variable blocking average (VBA) method. In our case, the method was used to search for times that mark abrupt changes in the $\sigma_w^2$ of $w'$ time series. The algorithm evaluates local maxima in the Haar wavelet transform of $w'^2$ for a fixed time interval scale for which $\sigma_w^2$ is calculated, identifying abrupt changes in $\sigma_w^2$. The Haar transform here is defined as:

$$Haar\text{\_\_\_transform} = \frac{1}{2m} \sum_{j=1}^{m} (w_{i+j}^2 - w_{i-m+j}^2),$$

where $m$ is equal to product of the chosen time window times sensor frequency (1Hz) (Adapted from Howell and Mahrt, 1995). By default, the beginning and end of the time series mark abrupt changes. A period with constant $\sigma_w^2$ in the time series, for example, corresponds to a time interval between two adjacent points. The period is usually larger than the chosen Howell’s time window (hereafter just time window) used to calculate the inner $\sigma_w^2$, used in the Haar transform, but if more than two maxima are indentified in a time interval shorter than this time.
window, only the largest maximum is kept (Howell and Mahrt 1995). The time window used to calculate $\sigma_w^2$ was chosen to be 20 minutes. It is likely that a shorter time window would identify more changes (vertical lines in Figure 3.4) in the time series than would a longer one. Consequently, there is a chance that longer periods will be split in sub-periods or that small periods will be swallowed into the bigger ones depending on the choice made.

Figure 3.4: In the upper and lower rows are, respectively, 1 Hz $w'$ time series for station 7 and for station 10, both during the night of October 24 -25, 2003. Every two consecutive vertical lines define a period. In a fixed row, from left to right are Howell’s time window parameters runs for 15, 20, 30, and 45 minutes. The periods marked with an “e” were classified as *periods of increased turbulence* because had their $\sigma_w \geq 0.1 \text{ m/s}$.

The shorter time windows identified more periods than did longer ones, due to two reasons: first just because in larger time windows the shorter distance allowed between two maxima are larger; and second because turbulence bursts with shorter duration than
the *time window*, which can produce temporary high fluctuations in $w'$ and therefore higher $\sigma^2_w$, are more dumped for longer *time windows* than for short ones.

The main objective here is not only to identify the periods with approximately constant $\sigma^2_w$ but rather to find periods that have sufficient mixing (turbulence intensity evaluated using $I = \sigma_u^2 + \sigma_v^2 + \sigma_w^2$), to affect the SBL development. In this study we believe that $\sigma_w$ alone is good enough to identify turbulent *periods* that contribute to the bulk momentum and sensible heat fluxes exchanges between the surface and the atmosphere. Taking into account this idea, we perform a second screening to find which periods had $\sigma_w \geq w'$ threshold ($\sigma_{th}$). In this sense, the words *periods of increased turbulence* will be reserved to describe those periods identified by Howell’s algorithm that satisfied the *threshold* condition (time intervals between two consecutive vertical lines with letters “e” in Figure 3.4), and the word *period* to the periods in general that were or were not classified as *periods of increased turbulence*. 
Figure 3.5: Total number of events (black line with dots with scale on the left y-axis), and periods of increased turbulence (red dotted line with pluses with scale on the right y-axis), at all stations detected during the whole HVAMS intensive observation campaign as a function of the time window parameter. The vertical line shows the chosen time window, used in the Haar transform, to assess the sudden changes in $\sigma_w^2$ that mark the boundaries of the turbulent period. The number of events was obtained by counting set of contiguous periods of increased turbulence ($\sigma_w \geq 0.1$ m/s).

Figure 3.5 presents results from the combined screening, for which Howell’s algorithm was applied first to select periods with stationary turbulence $\sigma_w^2$ and then a threshold of 0.1 m/s in $\sigma_w$ was applied to identify the ones that had a high turbulence level. The result of this figure includes all the stations during all nights of the IOP, and it shows that the number of periods, and therefore the number of periods of increased turbulence decreases with increasing time windows. In order to identify a turbulent event
or a turbulence breakdown, the \textit{periods with increased turbulence} (e.g. $\sigma_w \geq 0.1\, \text{m/s}$) that are contiguous are counted as single \textit{event}. In this way, if for example, during a certain night a station had all its periods with $\sigma_w \geq 0.1\, \text{m/s}$, than the entire night will be counted as a single event. On the other extreme, if all the periods had $\sigma_w < 0.1\, \text{m/s}$, then there will be zero \textit{events} for that night. For nights with more than one \textit{event/night}, it will be necessary at least to have more than one \textit{period of increased turbulence} followed by a \textit{period of weak turbulence}.

The chosen threshold, $\sigma_{th} = 0.1\, \text{m/s}$, was based on the finding that fluxes for \textit{periods} with $\sigma_w \leq \sigma_{th}$, did not contribute significantly to the overall momentum and heat flux average during the entire HVAMS campaign. In Figure 3.6, periods with $\sigma_w \leq \sigma_{th}$ had almost no contribution to the momentum and heat flux. The flux contribution starts to show only for periods with $\sigma_w > \sigma_{th}$. Using $\sigma_w = 1\, \text{m/s}$ means that almost all the \textit{important periods} contribution are considered in the average flux calculation (limit not shown in the Figure 3.6).
Figure 3.6: Cumulative average momentum and heat flux during the entire HVAMS campaign versus $\sigma_w$ threshold values using a 20-minute time window. The x-axis indicates an upper limit value for $\sigma_w$ of the periods. So, for example, the flux average value in y-axis corresponding to $\sigma_{th} = 0.1$ m/s, represents the contribution of periods for which $\sigma_w \leq \sigma_{th}$. The vertical dotted line shows the chosen threshold $\sigma_{th} = 0.1$ m/s, below this value no flux is observed.

Short time windows are more sensitive to abrupt changes in the $\sigma_w$, which makes better identification of the edges of events. The possible disadvantage is that single events can be identified as multiple events if the combined Howell and threshold screening fails to identify consecutive periods of increased turbulence as part of single events. Even if this occurs, the amount of flux added in the extra events selected using shorter time
windows does not significantly affect the nocturnal flux totals (See Figure 3.7). The choice of running mean window affects more the fluxes than does the time window. A 15-minute running mean window was chosen to remove the trend from the time series because the overall average fluxes had least variation for the different time windows tested (See Figure 3.7).

Figure 3.7: Nocturnal momentum (heat) flux versus time window for station 10 during November 2003. Thin solid line is using 15-minute running mean filter and the thick solid line results from using a 10-minute running mean filter.

To summarize, we chose a 20-minutes time window to indentify the periods. This choice is to some extent arbitrary, but does not affect the results of the study of the intermittency at the flux stations, since the same rules are used to identify the intermittent
events. The choice will only affect the counting of events somewhat. Because the rules are uniform, any shifting in the number of events per night should be approximately the same at all stations.

Doran (2003) applied a 0.015 K m/s threshold to kinematic heat flux calculated over 2 minutes to determine which periods within one 1-hour period were turbulent (had heat flux above the threshold). The weakness of applying a threshold to the heat flux is that, the choice of turbulent periods is partially dependent of the temperature fluctuations, which is not a measure of turbulence intensity. Despite the fact that turbulence intermittency is likely to happen under stable conditions, it is plausible to imagine a situation with a prolonged time of strong turbulence, where the initial temperature gradient and therefore the temperature fluctuations would be weakened overtime. In such situation, the Doran (2003) approach can miss those times with small heat flux due to weak temperature fluctuations but strong turbulence, while in the current approach this would not occur because the method does not rely on any temperature measurement. Another advantage of the current method is that the threshold was not arbitrary chosen, but was based on the finding that momentum and heat fluxes associated to periods with \( \sigma_w < \) threshold (0.1m/s) did not contribute to the overall nighttime fluxes. However, as was seen, the method is still carries some arbitrariness in the choice of the time window used to calculate \( \alpha_w^2 \).
4. Landscape, mixing and surface temperatures

4.1 Introduction

Importance of the landscape features on turbulence and surface temperatures - Sheltering, concavity, slope, elevation and surface cover.

The objective in this section is to identify which surface characteristics of the flux sites (e.g., sheltering, elevation, surface concavity and surface slope) are most associated with turbulent activity and temperature distribution across the flux sites network. For this we compare nocturnal turbulent fluxes, minimum and maximum temperatures to site transmission factor, surface concavity and slope. Another goal is to investigate the representativeness of wind, temperature, and other scalars observations made by simple weather station in a location characteristic of the sampled landscape. We do this by taking into considerations stations poorly sited by conventional standards, including situation with strong stability conditions, and relate the outcomes to the surface terrain features. Results from such an investigation would have an important value, for example, when linking regional minimum temperature to landscape.

Observed values of wind, temperature, and other scalars are representative of a certain air layer near the ground. Under stable conditions, turbulence can be patchy, sporadic (Stull 1988; Garratt 1992) and confined close to the ground. The typical turbulent length scale does not achieve values greater than a couple of tens of meters. The extent of this layer depends more on the influences of earth surface, such as topography, land cover/land use. These elements are also important during convective conditions, but the “weight” of the influences of the topography and land cover/land use is diminished
under such conditions. The typical length scale (~1 km) of eddies are larger during convective than they are under stable conditions and mixing can dilute more efficiently spatial contrasts in temperature, scalars, and wind.

Topographic elements such as hills, slopes, bowls, hollows and valleys, and surface cover elements, such as grass, water, buildings, trees determine the location for occurrence of drainage flows, accumulation of cold air (Clements et al., 2003), and at smaller scales the positioning of obstacles that block and/or deflect the flow. Accumulated cold air, drainage flows and sheltering (Gustavsson et al., 1998) can perhaps be ultimately tied to the mixing activity at the site. Drainage flows can generate mixing or accumulate cooler air at low places, increasing the local stability. Sheltering can slow down the wind reaching the station.

The true microclimate is determined when all properties of the site landscape (e.g., elevation, concavity, slope, sheltering, and surface cover type) are properly described. In more neutral conditions, which occur when the mean winds are stronger and/or when there is higher cloud cover, spatial contrasts between mixing and temperature are weakened.

4.2 Case studies

Landscape characteristics presented in Tables 2.1 and 2.3 and Figure 2.4 show the distinct elevation, concavity, slope, $TF$, and land cover type for each station. Figure 4.2 presents a case in which the turbulence ($w'$) is not equally distributed among the stations on a particular night. Stations 5, 4, and 10 have high continuous turbulence, stations 7, 6 and 3 have intermittent turbulent events followed by calm periods, and 1, 2, 8, 9 weak turbulence. The wind speed was higher at the more turbulent stations as well as during
periods of increased turbulence at the more intermittent ones, but the wind speed was lower (<1.5m/s) at the less turbulent stations. During this night, the mesoscale flow was along-valley southerly, increasing over the course of the night (Figure 4.1). This flow direction was felt at the more turbulent stations and during the periods of increased turbulence at stations with intermittent mixing. At the stations with less turbulence the wind direction was not persistent from any particular direction. This suggests that shear instabilities in the mesoscale flow might have been the principal trigger of turbulence at the turbulent and intermittent stations. However, at station 7, which shows intermittent behavior, it is not clear what type of forcing generated the turbulence during intermittent events, because the wind direction during (See Figure 4.4) such events was not persistent from south as it was at other stations. Perhaps the shear in the mesoscale wind generated turbulence, but because this station is the most sheltered in the network, the wind is always largely deflected and the observed wind direction is not close (mesoscale wind direction – 30° < station wind direction < mesoscale wind direction + 30°) to the mesoscale flow direction. Because it is located just east of the Catskill Mountains, the turbulence might also have been generated by shear in katabatic winds.

The momentum flux and heat flux were higher at the turbulent and intermittent stations, and at the less turbulent ones these fluxes were rather very weak or vanishing, as one would expect. The consequence of mixing followed by convergence of heat flux could be felt clearly during the intermittent events at the intermittent stations, when their potential temperature increased (See station 6, 7 and 3 in Figure 4.6). In Figure 4.6, it seems clear that the turbulent stations were the warmest stations in the network with close potential temperatures. At the intermittent stations, the turbulence breakdowns raised
their temperatures during the *events*. At the calm stations the potential temperature showed no appreciable jumps; these were the coldest stations.

All stations were exposed to approximately the same background flow and cloud cover fraction (see Table 5.2), but the surface result depends on the different local surface characteristics (see Tables 2.1 and 2.2). Therefore, it is reasonable to believe that inhomogeneities in the landscape could have forced such differences in the observed fluxes, wind and potential temperature.

October 30 – 31, 2003

![Time series of wind speed and wind direction](image)

Figure 4.1: Time series of wind speed and wind direction obtained from tethered balloon (TAOS) during the night of October 30 – 31, 2003 between the hours 1638 – 0300 LST. Blue 60m, green 100 m, red 120 m, light blue 140 m, and black 200 m AGL.
Figure 4.2: Time series at 1 Hz of turbulent vertical velocity ($w'$) of all flux stations during the night of October 30-31, 2003. Vertical red lines indicate periods of increased turbulence, referred to as intermittent turbulent events, referred to in the text as events.
Figure 4.3: Time series (1Hz) of wind speed at all flux stations during the night of October 30-31, 2003.
Figure 4.4: Time series (1Hz) of wind direction at all flux stations during the night of October 30-31, 2003.
Figure 4.5: Time series of momentum (lines with circles) and sensible heat fluxes (solid lines “+”) at all flux stations during the night of October 30-31, 2003. Right y-axis is the heat flux in $W m^{-2}$, left y-axis is the momentum flux in $kg m^{-1} s^{-2}$, and x-axis is the hour in UT. The trends in temperature and horizontal wind were removed by using running means of 15 minutes and fluxes of momentum and heat were calculated for periods of constant variance in the flux product $w'T'$ as discussed in Chapter 3.
Figure 4.6: Potential temperatures at station 1 (black with squares), 2 (red with circles), 3 (green with triangles), 4 (blue with “+”), and 5 (light blue “x”), 6 (pink with diamonds), 7 (yellow), 8 (gray with squares), 9 (black with “*”), and 10 (black with open squares) during October 30 – 31, of 2003. The sampling frequency represented here at all stations was 0.017 Hz.

The frequency distribution of minimum and maximum temperatures over the network during four IOP nights selected for strongly stable conditions (Table 4.1) is more evidence that landscape features influence the local temperatures. If the surface concavities of the sites are compared to Table 4.1, it is clear that stations located in a
convex surface (negative concavity), such as station 5, tend to be warmer than places with concave or flat surface (positive concavity), such as stations 1, 2, and 3. However, data from only four nights is insufficient to make firm conclusions. $TF$, elevation, slope and surface cover type are other parameters that should influence not only the local temperatures, as mentioned, but the wind vector and the turbulent fluxes as well.

<table>
<thead>
<tr>
<th>Station</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
<th>7</th>
<th>8</th>
<th>9</th>
<th>10</th>
</tr>
</thead>
<tbody>
<tr>
<td>Percentage (%) of time that a station had the maximum temp. over the network</td>
<td>0</td>
<td>0</td>
<td>1.25</td>
<td>9.4</td>
<td>38.4</td>
<td>15.0</td>
<td>3.6</td>
<td>1.7</td>
<td>0</td>
<td>30.6</td>
</tr>
<tr>
<td>Percentage (%) of time that a station had the minimum temp. over the network</td>
<td>19.7</td>
<td>36.7</td>
<td>3.2</td>
<td>4.5</td>
<td>0</td>
<td>0</td>
<td>1.7</td>
<td>13.0</td>
<td>20.6</td>
<td>0.2</td>
</tr>
</tbody>
</table>

Table 4.1: Occurrence frequencies of network maximum and minimum temperatures during four nights of the IOP when station 10 was already operating.

The nocturnal average fluxes during the entire IOP are stronger evidence that turbulence is spatially inhomogeneous across the landscape (see Figure 4.7). Under the same background conditions—e.g., mesoscale flow and cloud cover fraction—the flux value responds to the local surface characteristics. Station 5 fluxes dominate the turbulent transfer over the network, while places like station 2 contribute little. Therefore, one cannot assume that the occurrence of turbulence is spatially random in a heterogeneous landscape. We do not deny any possible importance of generation of shear instabilities, such as K-H shear instabilities or breaking gravity waves, (e.g. Finnigan et al. 1984; Einaudi and Finnigan, 1992, Sun et al., 2004), or density currents (e.g. Sun et al. 2002) to create turbulence. However, we believe that the landscape modulates the mechanisms that trigger turbulence as it is perceived at the surface. Recall that our objective is not to
find the mechanisms that triggered the generation of turbulence but rather to link its occurrence to the landscape features.

Figure 4.7: Half hour bin-averaged periods for momentum and heat flux during the IOP (50 nights). Black solid line is the momentum flux with units (kg m$^{-1}$ s$^{-2}$) on the left y-axis, and red dotted line heat flux with units (W m$^{-2}$) on the right y-axis. Horizontal axis indicates hours after sunset. Note that station 10 is not shown because of a smaller sample of nights for the IOP.

*Hypothesizing the influences of TF, concavity, slope and elevation*

From this point on, we assume that surface concavity, slope, elevation, sheltering, and land cover type fully describe the landscape characteristics. Their influences on the observed local flow are hypothesized:
• **TF should have a greater influence in momentum and heat flux exchanges during windy conditions and high values of TF should lead to high values of momentum and heat flux.**

Under windy conditions the local wind observed at the sites would be responding to a mesoscale pressure gradient with size comparable or larger than the network size. In this condition the different levels of sheltering from the directions around a station will cause large differences in the winds coming from different directions reaching the station, which ultimately through shear will provoke distinct levels of mixing for the different directions. So, if the wind has no preferential direction to occur, places with a large number of open directions (high TF for more directions) are expected to have larger fluxes on average than places with fewer open directions. For calm wind conditions TF would be less important, because winds reaching a station coming from any direction would be weaker with weak shear associated, and the resulting fluxes would be small with small differences.

• **Curvature, and slope effects should be more apparent during calm wind and clear skies. Stations in locations with positive terrain curvature, such as hollows or bowls, are expected to be cooler than the ones located over convex surface. Places located on sloped surface can have occurrence of drainage flows.**

Bowls and hollows need higher wind shear to break the high stability due to the accumulated cold air, which have resulted from drainage flows. In the other extreme, sites located on surfaces with negative curvature, such as hills, would need less wind
shear to mix because the stability would be weaker due to downslope loss of cold air (e.g. AF2003).

In overcast conditions the difference in the stability close the surface for concave, flat and convex terrain should be smaller. Radiative flux divergence near the ground and across the SBL is weakened and cold air sitting on slopes promoting drainage will not be so vigorous. For windy conditions the wind shear is expected to be stronger enough to break any stability across the landscape promoting more spatially homogeneous mixing across the landscape.

- **Elevation effects - places with elevations higher than 200 m do not experience effects of the cold pool.**

It is assumed that during nights with weak mesoscale flow and clear skies the bottom center of the valley floor fills up with cold air due to the drainage of cold air from higher places at the valley sidewalls. The main central depression (Figure 2.4b) is larger in horizontal scale than hills, bowls, and hollows embedded within it, and there is the possibility that the temperature, wind, and fluxes measured at different location in the central depression are not the same because of the influences of such smaller scale topographic features. Here it is being assumed that the different scales of the elements that compose the topography and surface cover can affect the temperature measured at a specific site.
4.3 Results - TF, concavity, slope and elevation

4.3.1 Intermittency and concavity

Turbulent behavior at a single station is not wholly random, but also depends on the site surface characteristics. Figure 4.8 shows that places with positive curvature or concavity (e.g., $\nabla^2 z(x,y) > 0$), which means a place with concave surface, have more intermittent turbulence because they have high number of events per night and the events are short-lived (high intermittency index). At places with strong negative concavity—(convex surface $\nabla^2 z(x,y) < 0$)—turbulence is more continuous or less intermittent with fewer long lasting events. It is important to note that higher is the intermittency index (left axis on Figure 4.8), less time of the nighttime period is filled with turbulence. This index was defined as being 1 minus the ratio of the sum of the duration of all intermittent events to the total nighttime duration. In this sense, an intermittency index = 0, means all the nighttime period is turbulent, intermittency index = 1 no turbulence during the entire nighttime, and any other value between 0 and 1 partially turbulent. For sub-ranges of concavity ($\nabla^2 z(x,y)$) within the extremes, concavity seems not be able to separate well the number of events per night, however the overall the number of intermittent events decreases with decreasing $\nabla^2 z(x,y)$.

The intermittency was measured here using the number of events per night. In contrast, Kondo et al. (1978), defined intermittency by the ratio of time that the flow was considered turbulent over 30 minutes. They used thresholds for the standard deviation of the temperature ($\sigma_T$), which was calculated for 30 s timescale. A period was considered
turbulent if $\sigma_T$ was larger than an established threshold or otherwise was consider quiet. Here, a threshold in the standard deviation of the vertical velocity ($\alpha_w$) is used instead (Chapter 3). Kondo et al. (1978) method might miss events when $\sigma_T$ is small due to the presence of weak vertical temperature gradients, while in the current approach this should not be the case because we use directly measure of the turbulence intensity ($\alpha_w$). In general, their event durations (< 30 minutes) were smaller than in the present work, because the smallest scale event has at least the size of the time window chosen (20 minutes) for Howell’s algorithm (Chapter 3). In the present study we not only have the time fraction the flow was turbulent (1-intermittency index), but there is also the number of events per night, which when combined gives extra insight in the sense we can know how “broken is the turbulence” (intermittent) and what is the typical length duration of the intermittent events (1-intermittency index)/number of events/night).

So using these indices we find that typically there is a low number of events per night over places with high negative $\nabla^2z(x,y)$, which last ~ $\frac{1}{2}$ of the nighttime, while over places with intermediate or high positive $\nabla^2z(x,y)$, there is a high number of events, which last ~ $\frac{1}{4}$ of the night period. Despite the reduced range of surface concavity found in the network, this result shows that turbulence behavior, e. g. event duration and number of events per night, depends of surface concavity.
Figure 4.8: Ensemble average of *intermittency index* as function of site concavity (line with squares - scale on the left y-axis) during the entire IOP (50 nights). Concavity was evaluated by adjusting a quadratic surface to an area of $\sim 0.3 \text{ km} \times 0.3 \text{ km}$ centered at the stations. Ensemble average of *number of intermittent events per night* as function *concavity* (thick line with pluses with scale on the right y-axis) during the same period. Numbers at top of the frame refer to the stations 1 - 9 (only PAM stations).
4.3.2 Concavity, TF, slope, and turbulent fluxes

Figure 4.9 (a): Nocturnal average momentum flux averaged for different mesoscale wind and cloud cover fraction conditions during the IOP as a function of site concavity evaluated by adjusting a quadratic surface to an area of \(~0.3 \text{ km} \times 0.3 \text{ km}\) centered at the stations. Green for \textit{windy and clear nights}, blue \textit{windy and overcast}, light blue \textit{calm and clear}, and pink \textit{calm and overcast}. Numbers refer to the stations (only flux stations). The colored lines represents a 3\textsuperscript{rd} order polynomial \((ax^3 + b)\) fitted by non-linear least square method to the four different wind and cloud cover conditions. Note that the coefficient of determination \((r^2)\) is given all four atmospheric conditions.
Figure 4.9 (b): Nocturnal average heat flux averaged for different mesoscale wind and cloud cover fraction conditions during the IOP as a function of site concavity. Green for windy and clear nights, blue windy and overcast, light blue calm and clear, and pink calm and overcast. Numbers refer to the stations (only flux stations). The colored lines represent a 3rd order polynomial ($ax^3 + b$) fitted by non-linear least square method to the four different wind and cloud cover conditions.

The momentum and sensible heat flux follows the local site surface curvature for an area of approximately 0.3 km x 0.3 km centered at the stations. In both Figures 4.9 (a) and 4.9 (b), the fluxes decreases (become more negative) with decreasing concavity. At the
extreme values of the concavity range this relation seems clear. However, within the sub-ranges of concavity values this behavior is less clear. This relation gets less clearly defined going from calm to windy conditions. During conditions of calm winds and clear skies, pooling of cold air may exacerbate the local stability of the atmosphere at places with positive concavity (concave). While during windy and/or overcast conditions the strong wind levels and reduced surface radiative flux divergence can, respectively, sustain strong shear to overcame the local stabilities or reduce it, and both can inhibit the drainage of cold air to low places. When comparing clear sky conditions with overcast conditions under relatively fixed wind conditions, it is not clear that the flux dependence with concavity is higher (higher coefficient of determination \( r^2 \) for the 3\(^{rd} \) order polynomial \( ax^3 + b \)) for clear sky than for overcast sky. Here, it has to be noted that the cloud cover fraction was calculated using cloud ceiling data collected at Albany airport, but the flux stations were located south of Albany between the cities of Albany and Poughkeepsie, NY. The cloud cover fraction was estimated by taking the ratio of number of times clouds are found within the first 3-ceilometer levels by the total number of samples over the night (Freedman et al. 2001). Hence, because the distance range is about 100km, it is possible that the cloud cover at Albany is not entirely representative of the cloud cover observed at the network.

To evaluate the dependence of momentum and heat fluxes with \( TF \), the wind direction is tracked to determine the \( TF \). To do such, the mean wind direction during periods of increased turbulence and as well as periods of weak turbulence are used to obtain the \( TF(wind \ direction) \) of the particular wind directions accessed during the entire IOP (50 nights). Once it is done, the nights are sorted out in terms of background wind
and cloud cover conditions, and $TF$ and the momentum and sensible heat fluxes are averaged for the different conditions. Note that $TF$ is not unique for a single station, because it is used only the $TF$ values corresponding to the wind directions that occurred within a specific background flow and cloud cover class.

![TF and momentum flux](image)

Figure 4.10: (a) Nocturnal average momentum flux averaged for different mesoscale wind and cloud cover fraction conditions during the IOP versus averaged $TF$. Green for windy and clear days, blue windy and overcast, light blue calm and clear, and pink calm and overcast. Numbers refer to the stations. The colored lines represents a 3rd order polynomial ($ax^3 + b$) fitted by non-linear least square method to the four different wind and cloud cover conditions.
Figure 4.10 (b): Nocturnal average sensible heat flux averaged for different mesoscale wind and cloud cover fraction conditions during the IOP versus averaged $TF$. Green for windy and clear nights, blue windy and overcast, light blue calm and clear, and pink calm and overcast. Numbers refer to the stations. The colored lines represents a 3rd order polynomial ($ax^3 + b$) fitted by non-linear least square method to the four different wind and cloud cover conditions. The color scheme for the lines is the same as the one used for the numbers.

Momentum and heat flux have higher dependence with $TF$ for windy than calm conditions (Figures 4.10 (a) and (b)). The fluxes increase with increasing $TF$—they tend to be higher for winds coming from more open directions. During calm nights this
dependence is not clear. This confirms earlier results from AF2003 that on calm nights the fluxes are overall smaller and fluxes differences amongst the directions are small. Less obstructed directions are associated to higher flux values than more obstructed ones. One possible weakness of using \( TF \) to evaluate sheltering around stations, is that the condition that all stations are subjected to a common mesoscale “unobstructed wind”, is not well satisfied for the HVAMS. The unobstructed wind reaching the different stations could be originated from different pressure gradients because the network array size is large (~100km), or even if the pressure gradient is the same for the entire region, obstacles larger than trees and buildings, like hills, could deflect the wind. Taking into account this last possibility, we used diurnal network wind to determine the station’s \( TF \), because the depth and vigor of the mixing during the day is stronger than during the night. The wind is better homogenized during the day than night. Another practical difficulty is that the range of stations \( TF \)’s for the bin average wind conditions does not span entirely the theoretical range (0, 1).

Figures 4.11 (a) and (b), suggest that there is an apparent dependence of momentum and sensible heat flux on the local site slope. Nevertheless, this dependence is rather a consequence of concavity than local slope. The highest slopes were associated to the stations with highest concavity, as shown in Figure 4.11 (c). The slope was evaluated exactly at the site (0, 0) and it may represent the very local slopes with horizontal scales smaller than a couple hundred meters. Because of its reduced scale such slopes might not force drainage flows strong enough to produce shear instabilities. Perhaps, the larger scale slopes do have an influence in the fluxes, but to investigate it, we would need wind measurements at several locations within a larger area combined with smoke release to
track the trajectory of the drainage flows (Mahrt et al., 2001), which were not present at
the time of the HVAMS experiment.

Figure 4.11 (a): Nocturnal average momentum flux averaged for different mesoscale
wind and cloud cover fraction conditions during the IOP versus local slope. Green for
windy and clear nights, blue windy and overcast, light blue calm and clear, and pink
calm and overcast. Numbers refer to the stations. The colored lines represents a 3rd order
polynomial \((ax^3 + b)\) fitted by non-linear least square method to the four different wind
and cloud cover conditions. The color is scheme for the lines is the same as the one used
for the numbers.
Figure 4.11 (b): Nocturnal average sensible heat flux averaged for different mesoscale wind and cloud cover fraction conditions during the IOP versus local slope. Green for \textit{windy and clear nights}, blue \textit{windy and overcast}, \textit{light blue calm and clear}, and pink \textit{calm and overcast}. Numbers refer to the stations. The colored lines represents a $3^{rd}$ order polynomial ($ax^3 + b$) fitted by non-linear least square method to the four different wind and cloud cover conditions. The color is scheme for the lines is the same as the one used for the numbers.
Figure 4.11 (c): Site slope versus site concavity. The concavity is evaluated by applying the Laplacian to a quadratic surface fitted by least square method to an area of 0.3 km x 0.3 km centered at the stations. The slope was obtained by taking the magnitude of gradient of the quadratic surface at the center of area (0, 0).

The spread in the fluxes with concavity, $TF$, and slope existed because influences of these local landscape elements could not be completely isolated. All of them influenced turbulence and therefore mixing together, in a way that only at the extremes of the concavity, and $TF$ range values for distinct values of background wind, their
influences emerged. For local slope, at the extreme values (high convex surface) the concavity overruled the slope. The influence of the local slope is severely complicated when the study area is not defined by an unique slope, which is the case in region of complex terrain like HVAMS. In the HVAMS case, it is likely that the horizontal scale (<< 0.3km) for which slopes can be approximately constant has smaller or comparable size scale to the typical turbulence scale found. Recalling the result from Nakamura and Mahrt (2005), which showed that intermittent turbulence near the surface can have horizontal scale on the order of few hundred meters or less, and bringing to attention that $TF$ is directly dependent on the area size of the clearings around the HVAMS flux stations, and that it is related to the turbulence (fluxes), we assume that the turbulence scale is on the order of the clearing sizes (∼ 100m wide).

4.3.3 Surface lapse rate and pooling of cold air

Experimental data and modeled data show that during nights with weak mesoscale flow (∼< 5m/s) and clear skies (cloud cover fraction ∼< 0.5) cold air accumulates in the main valley central depression (See Figures 4.12, 4.13, and 4.15). For example, during the early mornings of 10/06/2003 and 10/08/2003 (See Figures 4.12 and 4.13) the lower part of the vertical potential temperature profiles, obtained by the aircraft at the different airports during dawn, approaches to the surface potential temperature obtained from the surface stations at ~ 150m above sea level. Above this height the aircraft profiles and the surface potential temperatures diverge. (Note that the aircraft profiles have two legs, which correspond to the descending and ascending periods during the close approaches.) The aircraft profiles have a stronger gradient with height than the
surface potential temperature with elevation. Above the inversion top of the aircraft profile there is the lapse rate of residual layer (RL) remaining from the convection of the late afternoon of the day before. Vertical profiles of potential temperature obtained from the Albany-12UT soundings and from the microwave passive radiometer (MPR) at Kingston Ulster airport, are presented as well for comparison. As we can see, at some extension, they collapse within the aircraft profiles, and they also sow the presence of RL above the inversion. Below elevations of ~ 150m, the MPR profile approaches to the surface network potential temperatures. This pattern is clear for the 10/06/2003 than for 10/08/2003. On the early morning of 10/08/2003, when the profiles were taken at Kingston Ulster and Columbia County airports, part of the stability of the former stable layer had already been destroyed by the beginning convection and the layer was mixed rather than stable.

The range of elevations where the vertical profiles approaches to the surface potential temperature, is the layer where there is influence of cold air pooling at the bottom of the valley, and we define it as the “cold pool”. Meso-NH simulation for the night of 10/07 – 08/2003 indicates the presence of cold pool at the valley central depression from surface up to 300 m above sea level (See Figure 4.15). Four consecutive days, 10/05/2003 to 10/09/2003, were simulated (See Chapter 2), but only during the night 10/07-08/2003 a cold pool was clearly established.

For the remaining analysis, mostly to determine the height of the SBL in Chapters 5 and 7, we use as the top of the cold pool the inversion height obtained from the Albany sounding or aircraft soundings whenever available. Therefore, it accommodates the layer
in which the vertical potential temperature profile comes close to the surface station lapse rate (essentially the cold pool thickness), and the temperature inversion above it.

The formation of this cold pool affects the surface temperature lapse rate. In order to average multiple nights of surface potential temperature profiles and keep their structure, we use potential temperature difference normalized by the spatial network extremes, \((\theta_{st} - \theta_{min}) / (\theta_{min} - \theta_{max})\), which we define as being the ‘temperature spread factor’ (SF); where \(\theta_{st}\) is the station potential temperature recorded at minute intervals, and \(\theta_{min}\) and \(\theta_{max}\) are, respectively, the minimum and maximum potential temperature found in the network over the same minute. Figure 4.14 (a) shows, the obvious result, that stations within this cold pool tend to be colder than those ones above it, but also that cold air pooling is a common feature of the Hudson Valley during nights with calm winds and clear sky. The maximum and minimum temperature over the network shows a similar results (Figure 4.14 (b)).

The Hudson Valley SBL corresponds to the cold pool, and according to the night case studies, 10/05 -06/2003 and 10/07 - 08/2003, the height of the SBL is not the same above ground level (AGL) everywhere, places with higher elevation have shallower SBL than places with lower elevation. It is equivalent to say that the SBL height is fixed with elevation but not with height above ground. In Chapter 5, it is shown that temperature gradient across the cold pool (SBL) and the strength of the flow above it, regulates the dynamic stability of the regional bulk SBL, which ultimately affects the surface turbulence activity of the HVAMS region.
Figure 4.12: Solid black lines with “+” indicate the surface lapse rate found using the averaged potential temperature of the 2 hours just before sunrise at stations 1 – 9, and HOBOs 1 – 5 during the night of 10/05/2003 -10/6/2003. Solid lines with circles are the MPR average potential temperature at Kingston Ulster airport 2 hours before the sunrise during the same day. The thick light blue line, thick pink line, and thick green line are, respectively, the early morning potential temperature profile obtained during the aircraft sounding at the Kingston Ulster airport, Green Acres airport, and Columbia airport (station 4), during the same day. The dotted line is the potential temperature profile obtained from the 12-UT Albany sounding during the same morning. The arrow indicates the vertical extension of the cold air pool layer that forms at the valley floor.
Figure 4.13: Solid black lines with “+” is the surface lapse rate using the averaged potential temperature over the 2 hours just before sunrise at stations 1 – 9, and HOBO 1 – 5 during the night of 10/07/2003 - 10/08/2003. Solid lines with circles are the MPR average potential temperature at Kingston Ulster airport 2 hours before the sunrise during the same day. The thick light blue line, thick pink line, and thick green line are, respectively, the early morning potential temperature profile obtained during the aircraft sounding at the Kingston Ulster airport, Green Acres airport, and Columbia airport (station 4), during the same day. The dotted line is the potential temperature profile obtained from the 12-UT Albany sounding the same morning.
Figure 4.14: a) Average temperature spread factor as a function of elevation. b) Stations’s percentage of time as maximum and minimum temperature over the network as a function of elevation. Numbers indicate station numbers: 1 – 9 (PAMs), 11 - 15 (HOBOs). The percentages of maximum and minimum temperatures in the network were obtained using the 1-minute potential temperature data. Only nights with calm and clear sky condition during the IOP were used (17 nights).
4.3.4 Surface temperature distribution related to concavity, $TF$, and slope

The occurrence of maximum and minimum temperature considering only the stations within the cold pool or stations with elevation $< 200$m, does not seem to follow the terrain curvature (Figure 4.16 (a)) neither the relative elevation $mz - z$ (result not
shown), nor the local slope (Figure 4.18) for calm nights with clear skies. The same is true for temperature spread factor (Figure 4.16 (a)). However, for $SF$ we can see that for the most convex, site station 5, surface concavity is dominant topographic feature affecting its temperature. Perhaps the amount of nights available (17) was insufficient to obtain less noisy curves for $SF$ and the percentage of maximum/minimum temperatures.

Figure 4.16: a) Average minute temperature spread factor ($SF$) as function of concavity. b) Stations’ percentage of nocturnal period as network maximum temperature (red line) and network minimum temperature (blue line) as a function of site concavity. Numbers indicate station numbers: 1 – 9 (PAMs), 11, 12 and 15 (HOBOs). Percentages of maximum and minimum temperatures were obtained from the one-minute potential temperature data. Only nights with weak mesoscale flow ($\sim 5$ m/s) and clear skies (cloud cover fraction $\sim 0.5$) and stations with elevation $< 200$ m were included in (a) and (b).
Figure 4.17: Temperature spread factor versus $TF$ (green numbers = average temperature spread factor and $TF$ for windy and clear skies, blue numbers = average for windy and overcast, light blue numbers = average for calm and clear skies, and pink numbers = average for calm and overcast). The temperature spread factors and $TF$ shown were obtained from averaging all these quantities during periods/events. Note that $TF$ is not unique for a single station for the different classes of background flow. It was used only those $TF$ values corresponding to the wind direction that occurred at the station within a specific background flow class. Thin line represents the smoothed points for both calm conditions, *clear* and *overcast*, and thick line the same but for windy conditions.
The temperature spread factor does not show simple dependence on the $TF$ for all classes of mesoscale flow and cloud cover (Figure 4.17). It would be expected that higher $TF$ would be associated with warmer temperatures, because high $TF$ values tend to have high turbulence levels (heat flux) (See Figure 4.10 (b)) and warm air from aloft can mix down keeping the station warmer. This assertion is likely to be valid during the occurrence of an intermittent event breakdown, but there is no guarantee that the number of turbulence breakdowns per night will be high enough to affect the nocturnal temperature average (See specifically station 6 on Figures 4.1 and 4.5). If this the case, cold air coming from higher places than the surroundings and local radiative cooling might be affecting the temperature more than intermittent events. This raises an interesting point that the scale sizes of the turbulence and all the other processes (drainage and radiative cooling) controlling local temperature might not be the comparable.

Gustavsson et al. (1998) showed that small valleys with forest are colder than open larger valleys, and argued that pooling of cold air can be a consequence of topographic sheltering and/or sheltering caused by trees/buildings, which would inhibit mixing allowing strong in situ cooling of the air near the surface. However, the fact that station 7 has larger a larger $SF$ than station 2 (Figure 4.17), even being the most sheltered station in the network, does not support this idea. One of the reasons for the conflict between idea and result, comes from the definition of sheltering. Here, we assume sheltering as being caused by obstacles like trees and buildings, and not to local topography. Local topography is related to surface to concavity, instead. In Gustavsson et al. (1998) there seems to be no distinction between sheltering caused by local topography
and sheltering caused by trees/buildings. The difficult then arises, because topography not only cause flow blockage but also drainage, and when relating it to sheltering, both effects are superposed. Here, there is no such a problem because we consider sheltering as being caused only by trees/buildings (e. g. AF2003), and topography is related to concavity and slope. In this sense the concept of sheltering in the present work, is unique tied to flow blockage/deflection, and relating $SF$ with $TF$, is a cleaner way to infer that sheltering can or cannot cause cold air pooling, however, because other local surface features (e. g. slope, concavity, and elevation) of the areas around the stations were variable, the fact that $SF$ is not dependent on $TF$, is not a strong result.

No dependence of the $SF$ on the local slope was found (Figure 4.18). Despite, it is physically plausible that drainage flows can promote turbulence and temperature advection at any measurement site they reach or pass by, they might not be responding to the local site slopes but rather to larger scale slope, as pointed out before when relating turbulent fluxes to local slope.
Figure 4.18: Average minute temperature spread factor as function of local slope. SF was obtained from the averaging the minute-by-minute spread factor. As for concavity, note that only nights with weak mesoscale flow (~< 5m/s) and clear skies (cloud cover fraction ~< 0.5) were included and stations that have elevation < 200m. Numbers indicate stations numbers: 1-10 flux stations, and 11, 12, and 15 weather stations.

4.3.5 Influences of land cover/land use on local turbulence and surface temperature

The land cover / land use type affects the surface, in the sense that it determines a typical roughness length (usually associated with wind and temperature logarithmic
profiles from M-O similarity hypothesis under neutral conditions), which will affect the momentum and heat exchange between the surface and atmosphere. It also is important for surface radiation absorption and emission, which should affect the surface net radiation, and soil properties (e.g. heat capacity and conductivity), which will affect the soil heat flux and therefore the surface energy balance. This idea is valid under homogeneous surface conditions, which are found when there is flat surface and large enough fetch around the station (e.g. the distance from the nearest obstacles should be at least 10 – 15 times larger than obstacle size (Wieringa, 1976)). Such conditions were obviously not found for the HVAMS sites, and the idea of typical roughness lengths for heat and momentum, unique surface radiation emission and absorption, and unique soil heat flux are not met for this environment.

The analysis of the land cover /land use data done in Chapter 2 (See Table 2.1), revealed that for areas of ~ 3km x 3km around the stations, the three dominant land cover types found were forest, agricultural and wetland, respectively. The rest of the land cover types (e.g. water, urban, barren and rangeland) played a minor role. Station 6, which was the closest to the river, even had most of its surrounding area covered by water. Even though there were dominant land cover classes, no one site could be uniquely describe by only one land cover type in a sense that for almost all stations these classes did not represent more than ½ of the total cover. Only station 3 had ~ 70% of its area covered by one class (forest). In addition, all the stations were located on clearings/openings composed of grassland with very sparse trees and/or buildings in some cases, or a mix of grassland and orchard. The gap sizes were variable in size but had a typical area size of
0.1 km x 0.1 km for most of the sites. The distance of nearest obstacle varied between ~10m to more than 100m (These distances were determined based on aerial photos.).

We conclude that the land cover / land use at the HVAMS sites were heterogeneous but similar amongst the network station sites, we could not associate unique land cover types with characteristic roughness lengths, surface emission and absorption, and soil heat properties to the sites, and it was not possible to isolate and investigate land cover / land use influences in the local turbulence and surface temperatures.

4.4 Summary

Local surface concavity (or curvature) is the most important factor in determining heat and momentum fluxes under conditions of calm winds and clear skies ($r^2 \sim 0.8$ for momentum and $r^2 \sim 0.9$ for heat) and less important for windy and overcast conditions ($r^2 \sim 0.5$ momentum and $r^2 \sim 0.3$ for heat). The local stability is more heterogeneous but generally stronger during calm winds with clear skies; mixing activity can be shut down at places where there is strong stability.

In contrast to concavity, $TF$ is more influential in windy conditions ($r^2 \sim 0.3$ for heat and momentum). During windy conditions the overall fluxes are higher and differences in the fluxes due to obstructed and open direction are therefore higher too. These results perhaps would be clearer if there were a wider range of $TF$ and all the sites had similar local surface concavity and slope.

The larger scale concavity of the valley central depression (positive concavity) exerts a bigger role compared to the small-scale concavity (~0.3 km x 0.3km) or $mz-z$ (~2km x 2km) on the maximum/minimum temperatures in the network over the night.
Despite the large variability due to various local aspects acting together (slope, local curvature, and $TF$), the spread factor ($SF$) tended to increase with increasing elevation ($z$). The same was true for the percentage of network maximum potential temperatures, which showed an increase with elevation, and for minimum potential temperatures, which showed a decrease with elevation. However, when considering stations within the cold pool (elevation below ~200m), the larger scale concavity of the valley is unimportant. Within this reduced elevation range, the small-scale concavity, slope, and $TF$ are the dominant landscape features affecting the temperature spread factor ($SF$). The land cover/land use affected lesser $SF$ because they were similar for the majority of the sites. Because the network of weather stations was restricted to a few stations (12 stations in the IOP and 6 in the LTOP), and there was no selection in terms of specific site surface characteristics prior to the installation of such stations, the majority of them did not have a true local minimum for concave surfaces or maximum for convex surface (Table 2.2 in Chapter 2). In these situations the surface resembled more a saddle point. Thus, local pooling of cold air or “run way” of cold air, which should increase or decrease, respectively, the local stability is not possible. Under this circumstance surrounding slopes might affect turbulence and local temperature more than concavity.

Ideally, to isolate the effects of local surface curvature it would be necessary to have sites located at hills and bowls with different curvatures but with similar sheltering and land cover/land use. On the other hand, for isolation of sheltering effects, it would be necessary to have a collection of sites with similar curvatures, slopes, and land cover. In summary, to evaluate the influences of any landscape feature, all the other landscape
parameters would have to be kept constant. At the time of HVAMS having enough stations properly to cover all these terrain conditions was not technically feasible.

The surface temperature spread factor \( (SF) \) did not show simple dependence on local landscape characteristics, e.g. curvature, \( mz-z \), \( TF \), and local slope during the IOP. The possible reasons are: the data record was too short, and the number of stations was not very large (12 for curvature and \( mz-z \), and 9 for \( TF \)). In Chapter 7 using a longer data record, we are able to show that \( SF \) exhibits a clear relationship to \( mz - z \).

The scales of the local curvature and \( TF \) are comparable with the typical turbulence scales found at the sites, but not to all other processes controlling the local temperature, such as drainage flows and in situ radiative cooling, which for certain instances might have been dominant. If this were the case, curvature and \( TF \) would have minor effects on \( SF \) and maximum/minimum temperatures. Nevertheless, the larger scale \( mz - z \), which might have better accounted for less localized effects, also did not show clear relationship with \( SF \) during the IOP.

AF2003 showed that the surface temperature distribution and turbulence (sensible heat and latent heat) have a relationship with \( mz - z \) parameter (in their work presented as \( z - mz \)) in area sizes of \( \sim 3\text{km} \times 3\text{km} \) and with \( TF \). They found that at some extent turbulence (heat flux only) and surface temperature increase with decreasing \( mz - z \) or increasing \( TF \). They considered that \( TF \) influences represent effects of the fine texture of the landscape, and therefore its scale of influences are more localized than \( mz - z \), which we also believe to be the case here. In the present work, however, we believe that the effects of real local curvature on turbulence and surface temperatures are clearly investigated because of two reasons: First- in section 2.1.1 of Chapter 2, we showed that
\( mz - z \) approaches to local surface curvature (\( V^2z(x,y) \)), only when \( \lim_{\text{area} \to 0} mz - z \). In the current work, however, curvature was also determined by taking the Laplacian of a quadratic surface adjusted to the area. Some uncertainty will come from the quality of the fitting, which showed to be the best for areas of \(~0.3 \text{ km x 0.3 km}\) (See Figures A.2.1 and A.2.2 on Appendix A.2), and are smaller than the area size chosen in AF2003. Secondly, because in the current work there was direct turbulence and flux measurements, intermittent events and fluxes were not determined indirectly, based on a connection height (\( H_c \)) as in AF2003, and local temperatures and winds. In addition, by having direct turbulence measurements we not only counted turbulent breakdown events but also quantified the fraction of a nighttime period that a station had turbulence, which combined with the former, can give the average duration of each event.

In this chapter, it was shown that intermittency is tied to local site curvature. Besides some scatter due to reasons discussed earlier, the number of intermittent events and intermittency index increases when going from convex to concave surfaces. At the most convex sites the number of events per night and intermittency index are minimum while at the most concave they are maximum. At these two extremes, respectively, there is a regime of a few long-lasting events and a regime of several short-lived events. At the more flat surfaces (low curvature) these two quantities did not show a clear pattern, perhaps because curvature differences were small and changes in slope and \( TF \) affected turbulence the most.

The influences of land cover / land use in the site turbulence and surface temperatures could not be investigated because the area around the sites were
heterogeneous but at the same time they presented similar structures amongst the sites, which did not permit distinguishing the sites by unique land cover / land use types.

5. Mesoscale influences

5.1. Introduction

Many studies have shown that turbulence in the stable boundary layer (SBL) can be sporadic in time and spatially discontinuous. Spatial turbulent variability occurs in the horizontal (Salmond and McKendry 2002; Acevedo and Fitzjarrald 2003) (Figures 4.2 and 4.7) or in the vertical (Mahrt and Vickers 2002) (See section 3.3 of Chapter 3.2).

The methods used to estimate fluxes under conditions of intermittent turbulence are better suited to studies of data from single stations (Acevedo et al. 2006) rather than those aiming to apply to broader areas. On the other hand, NWP models deal with grid cells much larger than the representative area of a single station. Effectively modelers must assume that there is an average amount of turbulence, which is continuous in time and space over the grid cell. Further, the amount of turbulent mixing varies with the stability of the layer (McCabe and Brown 2007).

A common difficulty is that when the stability criteria Richardson number,

\[ Rib = \left( \frac{g}{\theta_{nf}} \right) \cdot \left( \frac{\Delta \theta \cdot \Delta z}{\Delta U^2 + \Delta V^2} \right) \]

used in the models is stable, the classical critical value \( R_{icr} \approx \frac{1}{4} \) is too restrictive, allowing no turbulent mixing above \( \frac{1}{4} \). Perhaps this classical critical value makes more sense when analyzing the turbulence locally such at that seen in wind tunnels and tower layers, essentially single site observations, a situation for which the requirement of locality of processes that determine wind and temperature profiles is more likely. However, when analyzing a broader area, such requirements of
locality for temperature and wind profiles might not be feasible. Representing spatial averages sometimes leads to the area-average layer to have \( Rib > Ricr \) (Mahrt, 1987). So, to overcome this difficulty, NWP models are forced to keep some level of mixing even when \( Rib > Ricr \) (Viterbo et al. 1999).

Our main goal here is to address the following question: Can the apparently unphysical extra mixing in models under very stable conditions (\( Rib > Ricr \)) be attributed to real extra mixing observed under the same circumstances of stability at certain parts of the domain? We seek observational support for the modelers’ empirical practice of allowing mixing above \( Ricr \).

The main objective of developing a regional \( Rib \) is not to address the issue of what causes intermittent and localized turbulence, but to allow field data, collected over a region with heterogeneous terrain, to define the basic mesoscale state in which the intermittent mixing occurs. Then we look to understand how intermittent and localized turbulence may cause the spatially averaged flux to depart from that seen at single stations. To do this, we will first combine features of the background state of the flow such as wind speed, temperature jump across the SBL top and the SBL height, which we express as a ‘regional bulk Richardson number (\( Ribr \))’, that can be related to area-averaged (or more precisely to station network average) turbulence activity over a region. How are the mesoscale background conditions related to the network-averaged fluxes or regional fluxes? Below we outline the steps that were used to define \( Ribr \) and to identify observed mixing, particularly when \( Ribr > Ricr \).
5.1.1. Regional bulk Richardson number obtained from the HVAMS data

Eleven different configurations were tested to define a regional bulk Richardson number ($R_{ibr}$) using the data available for the intensive observational period (IOP) of the HVAMS project (see Appendix A.1). Required Richardson number parameters were determined alternately using the upper air measurements from microwave passive radiometer, wind profiler, tethered balloon (TAOS), weather soundings data, and lower atmosphere measurements from the network of flux stations. The key difference among the different configurations is how $\Delta U$, $\Delta \Theta$, $\Delta z$, were obtained from the data. Table 5.1 below summarizes the format of all the configurations and the data used for each. The rest of this section discusses in detail how $R_{ibr}$ was calculated for each configuration.
\[ \text{Rib} = \left( \frac{g}{\theta_{\text{ref}}} \right) \cdot \left( \frac{\Delta \theta \cdot \Delta z}{U^2} \right) \]

<table>
<thead>
<tr>
<th>Config.</th>
<th>Data used</th>
<th>Delta-(\theta)</th>
<th>Wind speed</th>
<th>Delta-z</th>
</tr>
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<tr>
<td>1</td>
<td>MPR, 915 wind profiler and network</td>
<td>(\theta_{\text{mpr100}} - \theta_{\text{network min}})</td>
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<tr>
<td>4</td>
<td>&quot;</td>
<td>(\theta_{\text{mpr100}} - \theta_{\text{surface}})</td>
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<tr>
<td>5</td>
<td>&quot;</td>
<td>(\theta_{\text{mpr200}} - \theta_{\text{network min}})</td>
<td>(U_{915-200m})</td>
<td>(z_{\text{mpr200}} - \bar{z}_{\text{network}} = 155.4)</td>
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<tr>
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<td>&quot;</td>
</tr>
<tr>
<td>9</td>
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<td>(U_{\text{taos highest}})</td>
<td>(z_{\text{taos highest}} - \bar{z}_{\text{network}} = ) variable (110m to 170m)</td>
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<td>(U_{\text{sounding}})</td>
<td>(z_{\text{sounding}} - \bar{z}_{\text{network}} = ) variable</td>
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<tr>
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<td>Albany soundings, 915 wind profiler, and network</td>
<td>&quot;</td>
<td>(U_{915-\text{fit}})</td>
<td>&quot;</td>
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</table>

Table 5.1: Eleven alternates to obtain \(\text{Ribr}\) using data available during the HVAMS IOP.  
\(\theta_{\text{ref}}\) = average potential temperature between the top level and bottom level.  \(g\) = gravitational acceleration (9.8 m/s\(^2\)).  \(\Delta z\) = Layer thickness measured as elevation difference between two levels.  \(z_{\text{mpr100}}, z_{\text{mpr200}}, z_{\text{taos highest}}, z_{\text{sbl top}}\) = local elevation plus height above ground, respectively, at microwave passive radiometer site for level 100m, 200m, at TAOS site for highest level, and for the top of SBL at Albany sounding site.  
\(\bar{z}_{\text{network}}\) = average elevation of the network (90m).  \(z_{\text{sbl top}}\) = height above ground of the top of the SBL plus local elevation of Albany sounding site (\(z_{\text{alb}}\)).  \(U_{915-100m}, U_{915-200m}\), \(U_{\text{taos highest}}, U_{\text{taos lowest}}\) = Wind speed at 915 wind profiler for 100m and 200m, and wind speed at TAOS for highest and lowest level. \(U_{\text{sounding}}\) = wind speed measured by the Albany 12 UT at the top of the SBL. \(U_{915-\text{fit}}\) = wind speed measured by the 915 MHz wind profiler located at Kingston Ulster airport. The wind speed used corresponds to the level (elevation) that better matched the level of the SBL top (\(z_{\text{sbl top}}\)).  \(\theta_{\text{mpr100}}, \theta_{\text{mpr200}}\) = potential temperature at the MIPS microwave wind profiler for 100m and 200m levels.  
\(\theta_{\text{surf acc}}, \theta_{\text{network min}}, \theta_{\text{network mean}}, \theta_{\text{network max}}\) = black body surface potential temperature minimum, mean and maximum potential temperature over the network of flux stations.
For configurations 1 to 8, $Ribr$ was calculated using the 6-hourly average of upper and lower air potential temperature and wind speed, available from microwave passive radiometer (MPR), 915 MHz wind profiler, and the network of flux stations. The potential temperature from the network was used as lower boundary condition and potential temperature from MPR at 100m and 200m above ground level (AGL) as upper boundary condition to obtain delta-theta ($\Delta \theta$). For delta-wind ($\Delta U$) the wind speed from 915 MHz wind profiler at 96 and 206m AGL were used as upper boundary condition and the wind speed at the ground, set as zero, was used as lower boundary condition. $\Delta z$ was defined as (local elevation of MPR and 915 MHz wind profiler site plus level of measurement) minus (the network average elevation plus measurement level).

For configuration 9, hourly averages were used because the tethered balloon flights were not available for all nights of the IOP and their durations were variable, usually with time duration shorter than the length of a nighttime period (~12h). In this case, the highest level (~150 – 200m AGL) of TAOS temperature and wind speed measurements was used as an upper boundary condition for potential temperature and wind speed. The network mean potential temperature served as lower boundary condition. $\Delta z$ for this configuration was defined as (the height AGL of the highest measurement level plus local site elevation) minus (network average elevation plus measurement height). In this case $\Delta z$ was variable because the flight height varied from night to night.

For the last two configurations 10 and hybrid, only one $Ribr$ value is available each night. In both $\Delta \theta$ is calculated using the 1200 UT sounding data from early morning, and the minimum network potential temperature averaged for the final last 2 hours before
sunrise. For configuration 10, \( \Delta U \) is obtained from the 12 UT Albany sounding, which measured the wind near the top of the of SBL, and for the hybrid from the 915 MHz wind speed level that most matched the SBL top (See Figure 5.1 for SBL top definition). In case of the hybrid configuration, the wind speed was also obtained from an average of the final 2 hours before sunrise. The upper potential temperature was chosen by visual inspection of the vertical profile of potential temperature. It was picked to be the height above the first significant potential temperature jump, and the height AGL used to calculate \( \Delta z \) was chosen right below the temperature jump (see Figure 5.1). \( \Delta z \) was defined as (the chosen height AGL plus local site elevation) minus the (network elevation plus measurement height). \( \Delta z \) also varied here, however in a broader range of values than for configuration 9 because the elevation of the first temperature jump widely varied (152m-613m AGL against ~150m -- 200m AGL of configuration 9) from night to night (See Table 5.2).
Figure 5.1: Early morning Albany sounding (12 UT) on 10/02/2003. Lower arrow shows the height chosen to be the SBL height. Second higher arrow shows the level corresponding to the residual layer potential temperature chosen to be the upper boundary potential temperature for Ribr configuration 10. Note that the height above ground is used instead of elevation (Albany = 89m above sea level).

In this case the upper level used to obtain the $\Delta z$, corresponds to the height above ground where there is still mixing (Figure 5.1), when mixing is present, or the elevation just above the temperature inversion. When the mixing is weak or nonexistent, radiative flux divergence dominates the cooling of the SBL. The approach of using a temperature jump will not work well when there is warm or cold advection at different levels of the SBL and the residual layer (RL), or when the cloud cover fraction is high. In the latter case, the layer would be close to isothermal (the vertical net radiation ($R_n$) is reduced), and the potential temperature will have a constant vertical gradient. For example, the
night of the “day of the year” (DOY) 270 (See Table 5.2) with $\Delta z = 702m$ (SBL elevation) – 90m (network average elevation) = 612 m, average wind speed (5.8m/s), and if we assume that the layer is approximately isothermal $T = 10^\circ$ C, we would have a $\Delta \theta \approx 6K$, and $Ribr \approx 3.7$. In such a case, there is a possibility of non-vanishing heat fluxes while $Ribr$ is well above critical ($\frac{1}{4}$). To conclude, we can say with conditions close to isothermal, potential temperature jumps are not clear and any chosen temperature jump may not represent the height above the turbulent layer.

From the results presented in Table 5.2, the largest estimates of SBL occurred during overcast nights, which may be well above the turbulent layer. There are also cases with high cloud cover fractions and low heights. In these cases, warm or cold advection at different levels, associated with wind channeling in the valley or to local baroclinicity, may have produce a temperature jump that is not associated with the action of regional heat flux divergences or regional radiative flux divergences or both combined.
<table>
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<th>SBL Height above sea level</th>
<th>Upper potential temperature (K)</th>
<th>Wind speed</th>
<th>Cloud cover fraction</th>
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Table 5.2: First: DOY; second SBL height above sea level at Albany; third: \(\theta\) just above the SBL (12UT sounding); fourth: nocturnal average wind speed at Kingston Ulster Airport at 206m above ground; and fifth: nocturnal average cloud cover fraction at Albany Airport. The SBL height standard deviation was 125.25m. Cloud cover fraction was calculated using the Albany Airport ceilometer. It was defined as the ratio of number of times clouds are hit within the first 3-ceilometer levels by the total number of shots over the night (Chapter 4).
5.2. Results

Configurations 9, 10 and hybrid have higher positive values than configurations 1-8 (See Figure 5.2 and 5.3). There are some negative values of Ribr for configuration 3, because they use the maximum temperature of the network for the $\Delta \Theta$, which sometimes was higher than the upper layer temperature. Note that $\Delta z$ used in all configurations, was taken as being the elevation difference between the elevation of the measured top $\Theta$ and network average elevation (90m). Thus, there is a chance that the configurations that used the maximum network $\Theta$, refer most to the time to stations 5 temperature (See Tables 4.1, and Figures 4.16 (a) and (b) on Chapter 4), which tended to be the warmest station. If this is the case, the $\Delta z$ would be $< 55.4$ m (See Table 5.1 for used $\Delta z$), and it would not be hard to imagine that temperatures with such close elevation could have close values producing a negative $\Delta \Theta$. In contrast, configurations that used minimum network $\Theta$, will most likely to have a larger $\Delta z$ than the ones that used maximum or mean network temperature because station 2 tended to be the coldest, and it has lower elevation than the network average elevation.

The nocturnal average of the cluster of points of all Ribr configurations shows dependence on the external parameters, upper wind and cloud cover fraction (See Figure 5.2). For nights with stronger winds, or with more overcast conditions, or with both conditions present at the same time, the average tends (See dashed line on Figure 5.2) to stay at the smaller values of Ribr. When the wind is weaker and the sky has low cloudiness, the average tends to stay at higher values. This behavior illustrates the fact that upper wind and cloud cover partially control the temperature and wind speed gradients within the SBL. The rest of the control of these two quantities must come from
the landscape characteristics such as surface cover and topography. It was sown in
Chapter 4 that under fixed conditions of background wind and cloud cover any difference
in surface temperature is caused by contrasts in the topography.
Figure 5.2: Upper panel: time series of $R_{ibr}$ for all 10 configurations (symbols) and the hybrid configuration (thick solid red line) for the HVAMS IOP (09/26/03 – 10/31/03) for 6pm (2200 GMT) to 6 am (1000 GMT). For clarity, outliers are not shown. Note that there are fewer nights for configuration 9 because the TAOS flights were performed only during light wind, rain-free conditions. For configurations 1-8 there are two points nightly because half night (~6 hours) averages are used; for configuration 9 there are hourly points because 1-hour averages were used; for configurations 10 and hybrid there is only 1 point nightly because whole-night averages were used. Blue dashed line is the nocturnal average $R_{ibr}$ of the first 10 configurations (average of all symbols). Lower panel: time series of nocturnal average wind speed (lines with squares) using 206m level of 915 wind profiler data and cloud cover fraction (lines with pluses) using Albany ASOS ceilometer data. Note: There is only one wind speed and cloud cover fraction value because they are averages over the night.
Figure 5.3: Rib histograms for all configurations. Note the differences in the number of data points for the sets of configurations 1-8, 9, 10, and hybrid. These differences were caused because of sensor failures or missing data. The points of configuration 1-8 and 9 were averaged for comparison with configurations 10 and hybrid. For configurations 1 - 8, the two points were averaged to obtain the night Ribr, for 9 the 1-hour points were averaged, and for 10 and hybrid there was only one per night. There are fewer nights for configuration 9 because TAOS flights were done only for the fair weather nights, while for other configurations all nights are available.

Ribr tends to less stable values (decreasing positive values to negative values), as the bottom potential temperature is changed from minimum to mean, and then mean to maximum over the network (changing from configuration 1 to 3). The same idea is valid when we change from configuration 5 to 6, and then 7. One possible explanation for this may be that the minimum temperatures represent the temperatures of “local layers”
located at places disconnected from mean regional flow aloft (~150m of elevation around the center of the valley) with intermittent or nonexistent turbulence, while maximum temperatures represent places connected to the flow aloft that have more continuous turbulence. According to AF2003 study, places with concave surfaces and/or sheltered from the mean flow, can potentially have less turbulence than relatively exposed ones with convex surface. In fact, we observed this behavior for HVAMS (See Figures 4.8, 4.9 (a), and 4.9 (b) on Chapter 4). For configurations that use mean temperature we have an intermediate situation, with an air layer not constantly connected nor constantly disconnected from the background flow. These are places where the turbulence is more intermittent.

The amount of kinetic energy that must be extracted from the kinetic energy of mean flow to overcome the potential energy barrier and mix the air layers at disconnected places is higher than in connected ones. Therefore, if a regional SBL is composed of a range of air layers varying from very stable stagnant to weakly stable layers with fast moving air, Ribr’s obtained from configurations that use minimum potential temperature will tend to be more representative of an entire region than the ones that use mean or maximum temperature.

A way to interpret the Richardson number physically is that at the critical Richardson number value (Ricr), the amount of kinetic energy that must be removed from the mean flow should at least balance the potential energy barrier created by the density profile which is directly related to the potential temperature profile. The potential energy (V) is proportional to g ΔΘ and kinetic energy (K) is to the square of the mean wind speed difference across the layer. Using an example of two air layers, initially with different
wind speeds and temperatures which mix producing a final state of one combined air layer well mixed, it is possible to show that in this case we have to have $K_{\text{initial}} - K_{\text{final}} \geq V_{\text{final}} - V_{\text{initial}}$, which leads to $Rib \geq \frac{1}{4}$ ($Ricr=\frac{1}{4}$). Mixing rearranges the potential temperature distribution across the layer (stirs up the layer) and places initially with stagnant air and at lower temperatures should be warmed more than anywhere else because $\Theta_{\text{Top}} < \Theta_{\text{max}} > \Theta_{\text{mean}} > \Theta_{\text{min}}$. The thickness ($\Delta z$) of the air layers is also relevant. For example, if we have one layer bulk model composed by two internal layers with fixed $\Delta \Theta$ across the entire bulk layer (see Figure 5.4 below), we can see that at places with larger $\Delta z$ more kinetic energy must be drawn from the mean flow to stir up the air than at places with smaller ones. Our observational problem is that we do not have vertical density profiles that are continuous in time, even at the most instrumented sites.

![Figure 5.4: Schematic of an idealized one-layer bulk model with two internal layers. The energy needed to mix the entire bulk layer is drawn from the kinetic energy of the mean flow aloft.](image)

Ideally, to evaluate the potential energy barrier across the bulk layer at any place, we would need to know the potential temperature profile of the layer.

For configurations 4 and 8, which use as the bottom condition for potential temperature the mean radiative surface temperature, the situation is somewhat different
from the ones that use minimum temperature. This is because we are not using the air temperature at 10 m above ground level (AGL) but the earth’s surface emission temperature, and it is restricted to data from stations, 1, 2, and 3. The second reason is that the surface emission temperature depends on the choice of surface emissivity ($\varepsilon = 1$). Despite these differences, configurations 4 and 8 still have high stability values like 1 and 7, probably because the ground surface always tends to be colder than the air at 10m in any place.

The differences in the results among the set of configurations 1 and 5, 2 and 6, 3 and 7, and 4 and 8, is that configurations (5, 6, 7 and 8) which use $\theta$ at 200 m and $U$ at 206 m measurements have a wider $Ribr$ range than those configurations (1, 2, 3, and 4) that use $\theta$ at 100 m and $U$ at 96m (See Figure 5.3). This is because in the deeper layer configurations, $\Delta z$ was sometimes thick enough to include the SBL top values (See Table 5.2 for SBL top elevations).

The configuration 9 was the most skewed towards high positive values of $Ribr$, even though the top level measurement of the potential temperature and wind were between the higher and lower levels of MPR and 915 wind profilers, which would infer smaller $\Delta \theta$ and $\Delta z$. Three conditions might have caused such stable $Ribr$ values: First the potential temperatures from the MPR were usually lower than TAOS (see Figure 5.5), second the wind speed from 915 wind profiler were usually higher than TAOS (see Figure 5.5), and third the nights when the tethered balloon flew were the most stable because the winds were calm and there was no rain.
Figure 5.5: Left panel is 3 hours and 50min. long time series of potential temperature for the evening of 10/13/2003 for MPR (squares = 145.4 m, circles = 245.4 m, and triangles = 345.4m above sea level), for TAOS (black line =240m, light blue = 200m, pink = 150m, green = 115m, and blue = 90m above sea level), and for the network (red line = maximum temperature over the network with variable elevation according to elevation of the station that had the maximum temperature and blue line minimum = temperature over the network). Right panel: wind speed time series for all five TAOS levels in the same color scheme, the 915 MHz wind profiler (square=140.4 m and circles=250.4m) and the maximum network station temperature (red line). Note there are fewer points for the 915 wind profiler time series because this instrument outputs half hour averages.

Configurations 10 and hybrid had the widest Ribr range, probably because of a variable and thick $\Delta z$. In some extreme cases, $\Delta z$ was almost 6-times higher than the ones from the other configurations (Compare $\Delta z$ for nights 270 and 299 in Table 5.2 with Table 5.1). For these configurations $\Delta z$ was chosen by visual inspection using the early morning potential temperature to infer the top of inversion. So, in these cases most of the discrepancies between these two configurations and the others can be attributed to a thick variable $\Delta z$ and data used from three different sites.
5.2.1. Fluxes versus $Ribr$ using configurations 1 - 10

Figure 5.6: Plots of half–night average network momentum flux vs. $Ribr$ for configurations 1 to 8. The squares represent the windy ($U > 5\text{m/s}$) and clear skies (cloud fraction $< 0.5$), circles windy and overcast (cloud fraction $> 0.5$), triangles calm ($U < 5\text{m/s}$) and clear skies and crosses calm and overcast. There are two $Ribr$ values per night. All IOP nights are included. Recall that $Ribr = (9.8/\theta_{rg}) \times 55.4 \times \frac{(\theta_{\text{MPR}(100m)} - \theta_{i})/U_{915}(96m)^2}{915}$, where $i =$ network maximum, network mean, network minimum, and surface radiation temperature, respectively for configurations 1-4, and $Ribr = (9.8/\theta_{rg}) \times 155.4 \times \frac{(\theta_{\text{MPR}(200m)} - \theta_{i})/U_{915}(206m)^2}{915}$, for configurations 5 - 8.
Sensible heat flux vs. $Ribr$ (config. 1 – 8)

Figures 5.6 and 5.7 above illustrate that, for most of the configurations (except for 3 and 7), the shape of the $Ribr$ vs. network average momentum fluxes curves are very similar, as are the respective $Ribr$ vs. network average heat flux plots. Network-averaged turbulence activity over the study region has a dependence on the potential temperature and wind speed of the flow aloft (>150m – 300m of elevation above sea), which is the background ambient condition. For configurations 3 and 7, which use the network maximum temperature as lower condition, the situation is similar. Their shapes seem to
agree with each other. However, for these two configurations we are relating the amount of energy to mix only a few stations, which are more connected and less stable, to the overall flux average of the network.

When the background conditions are divided into classes as *windy and clear skies*, *windy and overcast*, *calm and clear skies*, and *calm and overcast*, 4 regimes or clouds of points are observed (see Figure 5.8).

![Figure 5.8: Ranges of $\Delta \theta$ versus ranges of $\Delta U$ for configurations 5-8. Squares = *windy and clear skies*, circles = *windy and overcast*, triangles = *calm and clear skies*, and crosses = *calm and overcast*. Y-axis represents the 6 hour averaged potential temperature difference between 200m level of microwave passive radiometer and network (maximum, mean, minimum, and surface radiation) potential temperature, and x-axis is the 6 hours averages upper wind (206m) from the 915 MHz wind profiler. The lines show the constant values of $\text{Ribr} = (9.8/\theta_{\text{ref}}) 155.4 (\theta_{\text{MPR}}(200m) - \theta)/U_{915}^2(206m)$, with $i = \text{network maximum, network mean, network minimum, and surface radiation temperature}$, respectively, for configurations 5-8.](image)
For windy and clear sky conditions the cloud of points stays at small positive \( Ribr \), corresponding to high momentum and heat fluxes. For windy and overcast conditions, the \( Ribr \) values are similar, but momentum values are somewhat higher, and heat fluxes are smaller. For calm and clear skies and calm and overcast conditions, the \( Ribr \) values are the most stable ones, with smaller associated fluxes. For momentum flux the clouds of points for both conditions overlap; for heat flux, overcast conditions tend to have the smaller values than do clear skies. The question that arises from these results is: Why is the separation of points between clear skies and overcast conditions better defined for heat flux than for momentum flux? If the flux-gradient relation
\[
( w'x' = -K_s \frac{\partial \bar{x}}{\partial z} )
\]
is a good approximation for both fluxes, under overcast conditions the vertical temperature gradient is weaker in the absence of any cold or warm advection because the radiative flux divergence across the layer is reduced. As consequence, because the heat flux is proportional to the vertical temperature gradient, it is smaller in overcast than in clear skies conditions. For momentum flux, the situation is not as well defined. The turbulent diffusion coefficient \( (K_x) \) can be higher in overcast conditions than in clear sky conditions because the stability is reduced. However, higher diffusion of momentum will make the vertical gradient of the wind flatter. The result of both effects—increased \( K_x \) and reduced \( \frac{\partial \bar{x}}{\partial z} \)—can compensate, leaving momentum flux nearly constant.

Plots of momentum and heat fluxes (not shown) using net surface radiation to distinguish between clear skier and overcast conditions presented very similar results to
those in Figures 5.6 and 5.7. Thus either parameter—net surface radiation or cloud cover—separates the background conditions into *clear skies or overcast*.

![Figure 5.9: Upper panel: momentum flux vs. Ribr; and bottom panel: sensible heat flux vs. Rib using configuration 9 (TAOS). Recall that Ribr = (9.8/θ_{ref}) Δz (θ_{TAOS(highest level)}-θ_{network(minimum)}) / U_{TAOS(highest level)}^2, and that Δz, was variable but with a typical range between 110m to 170m.](image)

We see from Figure 5.9 that configuration 9 tends to yield higher stability estimates than does the 1-8 set due to reasons discussed previously. This relation was plotted to investigate the background influences on the surface fluxes over periods shorter than half a night (6 hours) from configurations 1-8, using higher resolution measurements of wind and temperature. Momentum and heat fluxes tend to increase with decreasing stability (See Figure 5.9), which agrees with results based on half night basis. The problem with this configuration is that it is only available for light wind conditions,
narrowing therefore the range of background conditions \((Ribr)\), making the fluxes to be concentrated at small values. It is also more likely to not catch the SBL top.

![Graph showing network average momentum flux versus Ribr](image1.png)

**Figure 5.10:** Left panel plot of network average momentum flux versus \(Ribr\) for \(Ribr\) obtained configuration 10. The right panel is the same idea however for sensible heat flux versus \(Ribr\). The squares represent windy and clear sky nights, circles windy and overcast nights, triangles calm and clear sky nights, and pluses calm and overcast. Note that configuration 10 uses the 12 UT Albany sounding and network minimum temperature. All nights of IOP were included, when an upper wind aloft was available from the soundings. Recall that \(Ribr = (9.8/\theta_{ref}) \Delta z (\theta_{sounding}(SBL\ top) - \theta_{network}(minimum)) / U_{sounding}(SBL\ top)\), and that \(\Delta z\) was variable but with a wider range than configuration 9.

In Figure 5.10, which uses \(Ribr\) from configuration 10, the dependence of the network average momentum and heat fluxes in terms the \(Ribr\) is not as strong as for configurations 1, 2, 4, 5, 6, 8. This weak dependence is a consequence of poor estimates of the upper wind speed used for \(\Delta U\). In the majority of the soundings, there is no wind data available close to the top of the layer, and the wind speed at this level was obtained by linear interpolation of the two nearest points around the top of the layer. Such interpolations of the wind speed may have been compromised because for the majority of the nights these points were not evenly distributed around the SBL. This problem perhaps could also have affected the separation between windy and calm, because such distinction among these two wind categories were made using wind profiler data and windy/calm conditions from the profiler may have not matched windy/calm conditions.
from the soundings. In all others configurations this distinction is not compromised because wind speed data from the wind profiler is used to obtain $\Delta U$ and to distinguish windy and calm conditions. Despite the problems to obtain the upper wind speed in configuration 10, this configuration is used to evaluate the SBL flow stability during the LTOP (Chapter 7). It is implemented in the reduced network of stations -1 flux station and 5 weather stations (See Chapter 2 for details of network during the LTOP) at that period available during the LTOP. There is 2 years worth of sounding, flux and weather station data during LTOP, and it was possible to find enough nights with good wind data to draw a clear and stronger relationship between fluxes and Ribr (See Figure 7.1 on Chapter 7).

### 5.3 Reasons to use a hybrid Ribr

In configurations 1 – 8, which had the most continuous Ribr time series, $\Delta z$ was fixed and the evaluation of Ribr was restricted to a fixed air layer thickness, which is not representative of the flow stability of a real SBL with variable height. Configuration 9, which uses a variable $\Delta z$, is not the most appropriate either, because $\Delta z$ was chosen based on the highest level of the tethered balloon sonde, which does not necessarily agree with the SBL height. Configuration 10, which ideally should provide the best estimate of Ribr because it used a $\Delta z$ determined in terms of the SBL height, suffered of the lack of continuous upper wind speed data. Due to these technical and conceptual difficulties present in the previous Ribr configurations, we believe that the hybrid configuration of Ribr is the most appropriate to assess the regional flow stability during the IOP. As was shown, it has in its formulation a variable $\Delta z$ based on the SBL height obtained from
12UT Albany soundings, a $\Delta \theta$ obtained from the difference of potential temperature right above the SBL top and network minimum potential temperature, and continuous $\Delta U$ obtained from the 915 wind profiler. At the end of the night, the height or depth of the SBL will carry cumulative influences of nocturnal fluxes, and a night average of the fluxes should be the most suitable to be used.

5.3.1 How turbulent fluxes are related to the hybrid $Ribr$ configuration

Even though there were difficulties to obtain $Ribr$ in Figures 5.11 and 5.12 because the data came from three different sources (see previous section for details), there were data for almost all nights (35 nights out of total of 46 nights during the IOP) and it is possible to see that the network average momentum and heat fluxes have a clear dependence in terms of the hybrid version of $Ribr$ for the overall range $0 < Ribr < 45$.

For subrange $0 < Ribr < 1$, surprisingly the dependence is not clear. Both momentum and heat fluxes do not become more negative with decreasing $Ribr$. They rather show points that span for part of the range of momentum and heat fluxes. The overall results using hybrid $Ribr$ (“squares” on Figures 5.11 and 5.12) or the majority of others $Ribr$ configurations (Figures 5.6, 5.7, and 5.10), indicate that all of these stability flow indicators are capable of dividing the flow into basically two categories: one with almost zero fluxes at high $Ribr$ values and others with non-zero fluxes at a low $Ribr$ values. The differences among the configurations, is that the dividing point between high $Ribr$ values and low $Ribr$ values varies.

Aircraft data (“x” in Figure 5.11) is used as a final alternative to relate the network fluxes to regional $Ribr$. These data is available for only 7 nights (“x” on Figures 5.11 and 5.12), but it should be the most accurate one, and would be the closest to a
modeled grid-cell $Rib$. It is obtained from spatial averages of direct wind, and upper SBL potential temperature measured in the early morning during close approaches at the five local airports (Free Hold, Columbia County, Kingston Ulster, Green Acres, and South Albany – Table A.1.1 on Appendix A.1) in the HVAMS region. Note that the bottom potential temperature still comes from the minimum network potential temperature. The majority of the aircraft data points lay within the set of points of the hybrid configuration, and it therefore reinforces that the hybrid $Ribr$ is a representative estimate of regional flow stability.

Following Poulos and Burns (2003) (hereafter PB2003), theoretical curves for momentum and sensible heat flux are compared with the network average fluxes. The fluxes were estimated using $\left(\frac{w}{\chi}\right) = -l^2 \left(\frac{\Delta U}{\Delta \chi} \left(\frac{\Delta \chi}{\Delta \zeta}\right) f_a(Ribr)\right)$, where $l =$ mixing length, $\chi = U$ or $\Theta$ and $a =$ short-tail, long-tail, Louis79, and Delage97). From MacCabe and Brown (2007) $f_{\text{short-tail}} = (1 - 5Ribr)^2$ for $0 < Ribr < 0.2$, $f_{\text{short-tail}} = \left(\frac{1}{20Ribr}\right)^2$ for $0.2 < Ribr < 45$, for momentum and heat, and $f_{\text{long-tail}} = \frac{1}{1 + 10Ribr}$ for momentum and heat. From PB2003 $f_{\text{Delage97}} = \frac{1}{(1 + 12Ribr)^2}$ for momentum and heat, and $f_{\text{Louis81}} = \left(1 + \frac{10Ribr}{\sqrt{1 + 5Ribr}}\right)^{-1}$ for momentum, and $f_{\text{Louis81}} = \left(1 + 15Ribr\sqrt{1 + 5Ribr}\right)^{-1}$ for heat. We use a neutral mixing length $l = \frac{\kappa \zeta}{1 + \frac{\kappa \zeta}{\lambda}}$ (Blackadar, 1962; Mahrt and Vickers, 2003), averaged for the SBL depth which leads to $l = 0.114h$. Where in $l$, $\kappa =$ von Karman constant (0.4) and $\lambda$ is an adjustable parameter which was chosen as being one third of
the SBL height (Ballard et al. 1991). $l$ has asymptotic behavior assuming a constant value in the free atmosphere (e.g. Blackadar, 1962). To smooth the theoretical-flux curves, which were dependent on the night $Ribr$, SBL potential temperature gradient ($\Delta \Theta/h$) and wind shear ($\Delta U/h$), a non-linear least-square fitting was used to obtain empirical relations ($\Delta \Theta = 5.96Ribr^{0.13}; \Delta U = 5.97Ribr^{0.34}$) of these two SBL gradients in terms of $Ribr$. Note that the SBL height ($h$) cancels when multiplying by the square of the mixing length.
Figure 5.11. Momentum flux vs. $Ribr$ using the hybrid $Ribr$ (squares) that depends on sounding, network, and 915 MHz wind profiler data, as well as early morning aircraft soundings (“x”). Predicted theoretical momentum using sharp-tail, long-tail (MacCabe and Brown 2007), Louis81 (Louis et al., 1981), and Delage97 (Delage, 1997) (PB2003) stability functions (solid line, dashed line, dotted line, and dashed and dotted line). In the case of hybrid $Ribr$ each point represents five nights bin-average of nighttime network flux average.
Figure 5.12: Sensible heat flux vs. $Ribr$ using the hybrid $Ribr$ (squares) and $Ribr$ obtained from early morning aircraft soundings (“x”) at local airports. Predicted theoretical momentum using sharp-tail, long-tail (MacCabe and Brown 2007), Louis81 (Louis et al., 1981), and Delage97 (Delage, 1997) (PB2003) stability functions (solid line, dashed line, dotted line, and dashed and dotted line). In the case of hybrid $Ribr$ each point represents the five nights bin-average of nighttime network flux average.

The main differences between the current work and PB2003 are: $Rib$ is determined regionally in the current work and locally in PB2003, fluxes correspond to network averages in the current study and to a single location in PB2003, and
temperature and wind gradients refer to the bulk SBL in the current study, and to tower levels in PB2003. A minor difference is that our constant in the flux relation is 0.114, while in PB2003 their drag coefficient is ~ 0.002 – 0.003 (assuming a roughness length of 0.01m). This difference caused the theoretical curves in the current work to be displaced to higher values of Rib.

The main difference among the different stability functions is that the most restrictive ones, “sharp-tail” (McCabe and Brown, 2007) and “Delage97” (Delage, 1997) start to underestimate the fluxes for lower Ribr values than do the “long-tail” (MacCabe and Brown, 2007) and “Louis81” (Louis et al., 1981). In addition, for weak stable conditions, the usage of the former two, produce flux values closer to real ones while the latter two result in overestimated fluxes. More specifically, the predicted momentum flux using sharp-tail and Delage97 stability functions, is closer to real flux in the range Ribr ~< 2, while for sensible heat flux, the predicted values are partially overestimated in the same range. For more stable conditions (Ribr >~ 2), the usage of both functions, underestimate momentum and sensible heat fluxes. The long-tail predicts momentum fluxes closer to the real ones for Ribr > ~ 2 and extremely overestimates for Ribr ~<2, while Louis81 overestimates for the entire Ribr range. In the case of heat flux the situation is opposite. The predicted sensible heat flux using the long-tail is overestimated for the entire range, while for Louis81 the flux are close to real ones for Ribr >~ 2 and overestimated for Ribr ~<2.

PB2003 using flux and Rib data from 65m-tower in CASES-99, has shown that both Louis81 and Delage97 stability functions predict vanishing heat fluxes for Rib > 2 and Rib >~ 1, respectively. For momentum flux they found that Louis81 closely
reproduced the observed values and the Delage97 underestimates them. For the range $0.1 < \text{Rib} \leq 1$ both overestimate the sensible heat flux, and for the range $\text{Rib} < 0.5$ they overestimate momentum flux. Because our $\text{Ribr}$ is not locally defined it spans a wide range of $\text{Rib}$ values, but tends to be shifted to high values of $\text{Rib}$. Besides this difference the results of both studies agree, that overall, the stability functions have a region where they overestimate the fluxes (weak stable conditions) and another (strong stable condition) where they underestimate the fluxes.

To conclude, we see that there is a dependence of network average fluxes on the $\text{Ribr}$ for wide ranges ($0 < \text{Ribr} < 45$) for all 11 different configurations. The separation of background conditions into windy and calm allows us to separate the data into two regimes. For windy nights (night average of wind at 206m of 915Mhz wind profiler $> 5$ m/s) the $\text{Ribr}$ values tend to be small and the fluxes large (high negative values), while for calm nights ($U < 5$ m/s) $\text{Ribr}$ values are large and the fluxes values small (Figures 5.2 and 5.8). The separation into clear skies and overcast seems to work as well. Clear sky nights have higher positive values of $\text{Ribr}$, smaller values of momentum flux and higher values of heat flux compared to overcast nights.

A second result evident from all the $\text{Ribr}$ vs. flux plots is that there is appreciable network average flux when $\text{Ribr} > \text{Ricr}$. This may justify the usage of stability functions that allow mixing for $\text{Ribr} > \frac{1}{4}$ and also may illustrate that the $\text{Ricr} = \frac{1}{4}$ cutoff for turbulence is not suitable for flow over heterogeneous terrain.
Figure 5.13: Momentum flux versus $Ribr$ (left panel) and sensible heat flux versus $Ribr$ (right panel). “M” represents bin-average of nighttime average of the maximum flux anywhere in the HVAMS network and “F” is the nighttime flux average for Harvard Forest site (HF). In the case of HVAMS each bin corresponds to five nights bin-average of nighttime maximum network flux.

Because there was no flux site located at the forest in the HVAMS project, the predominant land cover of the region (See Table 2.1 on chapter 2), we use flux data from Harvard Forest (HF) site between July and September of 2002. The forest type of the HF region is similar to the one found in the HVAMS region. This period was selected because there was a NOAA 915 MHz wind and RASS temperature profiler running a few kilometers from HF flux site at the local airport in Orange, MA. The 10 nights from HF are used for comparison because they reflect the number of time we could identify a well defined stable boundary layer (SBL) top. The regional Richardson number ($Ribr$) for HF was obtained in a similar fashion as for HVAMS. The upper wind and potential temperature, and SBL height ($h$) were obtained from the NOAA 915 MHz profiler, while the bottom potential temperature was obtained from a single ASOS weather station.
located at the same airport. The main difference between the HVAMS and HF *Ribr*, is that the bottom potential temperature in the HVAMS case was taken as the minimum potential temperature found in the network, while in the HF we had to use potential temperature from a single weather station at the same site where the upper potential temperature and wind were taken.

The obvious result from such comparison (Figure 5.13) is that HF can sustain higher momentum and sensible heat fluxes under more stable conditions of the mesoscale flow. But why there is such difference? The difference in the land cover /land use of HVAMS sites and HF site is one of the causes of the different behaviors. Perhaps, the HF surface, which is forest, is rougher than that ones found in the HVAMS sites, which were clear gaps in the forest with at least 100m wide covered with grass. In a scale of ~ 0.3 km x 0.3 km, the HF site is concave (0.27 m\(^{-1}\)), and in larger area of ~ 2 km x 2 km, the relative elevation \(mz - z\) (8.42 m) indicates that the site is lower than the surrounding. According to the results of Chapter 4, these two topographic parameters indicate that the local topography does not favor turbulence. In a larger scale, the HF and HVAMS topography are different. For the HVAMS there is the effect of the valley, which forces the establishment of a “cold pool” (Chapter 4) and sustain the occurrence of wind channeling along the valley due to its geometry (Sakai et al., 2004), while for HF there is no such effects. All these aspects, different land cover /land use and small scale (~<2 km) large scale topography (>> 2km) are the reason for such differences in the dependence of the flux with mesoscale flow, but due to experimental limitations we are unable to point out which aspect is the most important. The only point we can make from this comparison is that, there is no universal relation of the regional and local turbulent
fluxes with mesoscale flow stability ($Ribr$). The dependence of the regional fluxes with $Ribr$ changes with the regional landscape type, and the local fluxes with the local landscape type.

### 5.3.2 Spatial distribution of fluxes

Spatial heterogeneity is reflected in observed flux across the landscape. Figures 5.14 and 5.15 shows that for the same background flow conditions ($Ribr$) the different stations have different momentum and heat exchanges. It also was found in Chapter 4, that spatial heterogeneity influences the turbulence activity. However, there we were not interested in determining which landscape type (surrounding the station) corresponded to most fluxes, but what kind of influence each surface characteristic (e.g. curvature, slope, $TF$, and land cover /land use) of the local landscape had on turbulence, and under what circumstances one or other is dominant. Here, it is clear that landscapes like the ones found around station 5, 10, 4 and 6 with rough convex surface, or flat surfaces with high exposure or moderate exposure, or relatively convex with high exposure tend to dominate the turbulent exchanges (Figures 5.14 and 5.15 and Tables 5.3 and 5.4) over the HVAMS network.
Figure 5.14: Normalized station relative contribution to the nightly network average momentum flux versus $Ribr$. Note the total length of the each bar is normalized by the maximum network average momentum flux (-0.26 kg m$^{-1}$ m$^{-2}$) found during the IOP. The widths indicate different stations (thickest = 1$^{st}$ dominant station, second thickest = 2$^{nd}$ dominant station, second thinnest = 3$^{rd}$ dominant station, and thinnest = all other stations) and the length the individual contributions of each station. Note only the dominant stations for the nights are shown, all others stations fall into the residual contribution “o”.
Figure 5.15: Normalized station relative contribution to the nightly network average sensible heat flux versus Ribr. Note the total length of the each bar is normalized by the maximum network average heat flux (-44.22 W m$^{-2}$) found during the IOP. The widths indicate different stations (thickest = 1$^{st}$ dominant station, second thickest = 2$^{nd}$ dominant station, second thinnest = 3$^{rd}$ dominant station, and thinnest = all other stations) and the length the individual contributions of each station. Note only the dominant stations for the nights are shown, all others stations fall into the residual contribution “o”.
<table>
<thead>
<tr>
<th>Flux</th>
<th>St. 1</th>
<th>St. 2</th>
<th>St. 3</th>
<th>St. 4</th>
<th>St. 5</th>
<th>St. 6</th>
<th>St. 7</th>
<th>St. 8</th>
<th>St. 9</th>
</tr>
</thead>
<tbody>
<tr>
<td>Momentum(%)</td>
<td>11.17</td>
<td>6.77</td>
<td>6.22</td>
<td>13.81</td>
<td>24.78</td>
<td>12.90</td>
<td>6.42</td>
<td>9.18</td>
<td>8.73</td>
</tr>
<tr>
<td>Heat (%)</td>
<td>8.53</td>
<td>7.17</td>
<td>9.71</td>
<td>14.48</td>
<td>18.32</td>
<td>12.51</td>
<td>12.05</td>
<td>8.28</td>
<td>8.94</td>
</tr>
</tbody>
</table>

Table 5.3: Average station relative contribution (%) during the IOP for momentum flux and sensible heat flux excluding station 10.

<table>
<thead>
<tr>
<th>Flux</th>
<th>St. 1</th>
<th>St. 2</th>
<th>St. 3</th>
<th>St. 4</th>
<th>St. 5</th>
<th>St. 6</th>
<th>St. 7</th>
<th>St. 8</th>
<th>St. 9</th>
<th>St. 10</th>
</tr>
</thead>
<tbody>
<tr>
<td>Momentum(%)</td>
<td>8.40</td>
<td>7.33</td>
<td>4.92</td>
<td>12.82</td>
<td>21.43</td>
<td>11.76</td>
<td>4.82</td>
<td>8.09</td>
<td>7.58</td>
<td>12.85</td>
</tr>
</tbody>
</table>

Table 5.4: Average station relative contribution (%) when station 10 was operating, for momentum flux and heat flux.

5.4 Spatial distribution of mixing and stability

In this section we present the results of the local Rib (Ribl) and local fluxes (Figure 5.16). For the stations 1-9, the Ribl was obtained from the temperature difference ($\Delta \theta$) between 10m AGL and the surface radiation temperature, the wind at 10m, and $\Delta z = 10m$, corresponding to the height of the measurements of wind and temperature. For station 10 from the tower wind and temperature profiles, corresponding to 12m AGL and 3m AGL (See Table A.1.3 on Appendix A.1). The fluxes used corresponded to periods of increased turbulence (sub-periods within an intermittent events, Chapter 3) and calm periods as well.

The fluxes showed a better relationship with Ribl during periods of increased turbulence (red dots) than during calm periods (black dots) probably because turbulence is more intense and stationary. Even though, the site exposures of the HVAMS stations were not ideal for practically all the stations, most of them behaved in a classical way, of well-exposed station with homogenous surrounding terrain with of continuous and
“strong” turbulence. The local fluxes tend to vanish when $Ribl > \sim Ricr$. For station 7 it is not true, there is still significant heat flux for $Ribl > Ricr$ range. Perhaps, because this is the most sheltered (lowest $TF$) station amongst all the flux stations, the locally observed turbulence is not entirely related to the local wind and temperature profiles. It might represent part of the turbulence that was generated at the surrounding nearby windbreaks and was advected to the site. The other possibility is that part of the observed turbulence was generated by shear in drainage flows, which were forced by presence of the Catskill Mountains to the west of this site. The fact that most of the periods identified as periods of increased turbulence lays at left side $Ricr$ shows that the approach developed in Chapter 3, to identify intermittent turbulent events that really cause momentum and heat exchanges between the surface and atmosphere, is adequate.

Figure 5.16 shows that the network of flux stations is capable of measuring turbulence, despite the fact that sites did not have a large enough fetch, but our point interest is to know which site is more likely to mix under given conditions of background flow. According to Figures 5.14 and 5.15, it is clear that there are preferential sites for occurrence of mixing. This proves observationally that there are “hot pots’ within a grid cell model for the occurrence mixing, as was supposed by Mahrt (1987) and observationally indirectly shown by AF2003. These preferential places (hot spots) are sites with negative concavity and well exposed (This result was also indirectly shown in AF2003). Perhaps, an empirical attempt to better adjust the amount of surface fluxes would to develop stability functions with adjusting coefficients dependent on the grid cell average $TF$ and relative elevation ($mz - z$). Probably $TF$ depends on the number and typical size of the clearings composing the area grid-cell, while the effects of concavity
could be more easily accounted by $m_z - z$ parameter than the Laplacian because $m_z - z$ is less sensitive to the area-size.

Figure 5.16: Four upper panels: momentum flux versus $Ribl$ for stations 5, 7, 8 and 10. X-axis is in logarithmic scale and y-axis is in linear scale. The vertical red-dotted line is $Ribl = \sqrt[3]{4}$. The four lower panels: the same but for sensible heat flux. The red squares (windy and clear skies), circles (windy and overcast), triangles (calm and clear) and “+” (calm and overcast) represents periods with increased turbulence activity (intermittent events), and black ones periods of weak or vanishing turbulence.

5.4 Summary

The stability of the regional flow was evaluated by developing a regional Richardson bulk number ($Ribr$), where $\Delta \theta$ was determined using $\theta_{top}$ at the SBL top and the minimum network $\theta_{min}$, $\Delta U$ using the wind at the top of the SBL $U_{sbl}$, with $\Delta z$ taken as the SBL thickness $h$. The network regional surface fluxes depend on the stability of the background flow. Nevertheless, this dependence was not clear for $Ribr < Ricr$, the fluxes did not necessarily increased with decreasing $Ribr$ in such range. For the range $Ribr > Ricr$ the fluxes decrease with increasing $Ribr$, however they never vanish even at very
high values of $Ribr$ ($Ribr \approx 50$). A previous work done by Mahrt (1987), using $Ribr$, obtained in a similar fashion from aircraft data, showed close results however for $\sigma_n^2$. He found no clear dependence of the $\sigma_n^2$ with $Ribr$ within the range $0 < Ribr < 1$, but above this range, $\sigma_n^2$ decreased with increasing $Ribr$.

The need for extra mixing above $Ricr$ in NWP models, first found empirically in the attempt to avoid a cold bias in minimum temperatures, is justified by the observational results presented, which show that there are still network-averaged fluxes in the regime $Ribr >> Ricr$.

For momentum fluxes the short-tail stability functions perform better for $Ribr < 2$, but they yield flux underestimates for $Ribr > 2$. The long-tailed distributions perform well in conditions of calm winds but overestimate the momentum flux in windy conditions. The heat flux signal is noisier. Both functions perform adequately for windy conditions but underestimate the heat flux in calm conditions.

Spatial heterogeneity of landscape forces mixing not to be uniform across the landscape. Places located at convex surface with high elevation, or well exposed or with rough surface have larger contribution to regional momentum and sensible heat exchanges. At these places the local stability is weak and turbulence tend to be higher. This result shows that within a model grid-cell these places can be “hot spots” for momentum and heat exchanges, which might dominate total area transfer (AF2003). This result is consistent with previous work done by Mahrt (1987), who found that the usage of area-averaged gradient may lead to serious reduction in the aerodynamic exchange coefficient, which would cause the surface flux to be underestimated.
6. Stable boundary layer heat budget

6.1 Heat budget equation

The relative contributions of radiative and turbulent cooling to the bulk cooling of the stable boundary layer (SBL) depend on specific situations (Sun et al. 2003; André and Mahrt, 1982; Drüe and Heinemann, 2007; Garratt and Brost, 1981). We estimate SBL bulk total cooling, and SBL bulk turbulent cooling, using Albany 00-UT and 12-UT soundings data, and network surface flux data. For the calculations of the SBL bulk radiative cooling rates, we use simulated radiation flux data. For these calculations, we use a range of SBL heights varying from 80m to 310m, above ground level (AGL) found during 19 nights of the IOP (See Table 6.1).

Absent appreciable evaporation or condensation, ignoring subsidence warming, and considering no horizontal temperature gradients, the bulk heat budget for the SBL simplifies to:

\[ \int_{t_1}^{t_2} \frac{\partial \theta}{\partial t} \, dt = -\int_0^h \frac{\partial}{\partial z} \left< \overline{w'\theta'} \right>_\text{spatial} \, dz - \int_0^h \frac{1}{\rho c_p} \frac{\partial R_z}{\partial z} \, dz \]

(1)

Where the 1st term is the bulk SBL cooling, 2nd term - bulk turbulent cooling, and 3rd term- bulk radiative cooling. If Leibnitz’ rule is applied to the first term and the SBL height \( h \) is assumed constant during the night, it leads to \( < \theta >_{\text{sbl}} = \frac{1}{h} \int_0^h \theta(z,t) \, dz' \). If \( < \overline{w'\theta'} >_{\text{spatial}} = 0 \) at \( z = h \). After performing the integration of the second and third terms of Eq. (1), it reads as:

\[ \frac{\partial < \theta >_{\text{sbl}}}{\partial t} = \frac{< \overline{w'\theta'} >_{\text{spatial}}(z = 0)}{h} - \frac{R_n(z = h)}{\rho c_p h} - \frac{R_n(z = 0)}{\rho c_p h} \]

(2)
6.2. Estimating terms of the heat budget

Following Fitzjarrald and Lala (1989) and André and Mahrt (1982), the first term was obtained by differencing the 12-UT and 00-UT NWS soundings vertical $\theta$ profiles, the latter being modified to carry the supposed potential temperature value of the afternoon convective boundary layer (CBL) to the surface (see Figure 6.1). When possible, these data were supplemented by aircraft soundings at local airports between Poughkeepsie and Albany made during the HVAMS IOP (See Figure 6.1). In the example (Figure 6.1), the area between the red line and the early morning sounding divided by the SBL height ($h$) is defined as the total cooling. The SBL thickness $h$ was defined as the height AGL just beneath the first significant temperature inversion, when turbulence creates a steep inversion, or the height above the temperature inversion (Figure 6.1), when the turbulence is weaker and no strong inversion above the surface layer is observed (André and Mahrt 1982). The second term in Eq. (2), turbulent sensible heat flux divergence, was obtained by network nocturnal averages or single station nocturnal averages of the heat flux measured at $z = 7m$ (AGL) (taken for this purpose as the surface) divided by $h$. The third term in Eq. (2), radiative flux divergence, was computed as the difference of the nocturnal spatially averaged net long-wave radiation between $z = h$ and $z= 1.4$ m (the lowest model level) divided by $h$. The upward and downward long-wave fluxes were simulated by the Meso-NH model, and were available at every 30 minutes for each station site. To obtain the radiative and turbulent cooling rates, we divide the radiative flux divergence and turbulent flux divergence by $\rho c_p$. 

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Figure 6.1: The top two plots are early evening and early morning Albany soundings during 10/06-07/2003 and 10/07-08/2003. The vertical red line shows the approximate potential temperature at early evening chosen for the calculation of the total bulk cooling. The dotted line indicates the SBL height (elevation) and the dashed line the site elevation. The bottom two plots use the same approach for aircraft soundings at early evening (21:44 UT, 21:49UT) and early morning (12:29 UT, 12:35UT) at Green Acres and Columbia County airports, respectively, during the night 10/30-31/2003.
6.3 Results and discussion

SBL heights in Table 6.1 were chosen by applying one of the SBL height definitions above to the potential temperature profile obtained from Albany 12-UT soundings (Figure 6.1). The total cooling, radiative cooling, and turbulent cooling in the SBL were calculated using these heights.

<table>
<thead>
<tr>
<th>Night</th>
<th>SBL elevation (m)</th>
<th>SBL total cooling rate at ALB (K/hr.)</th>
<th>Network average</th>
<th>Max.</th>
<th>Min.</th>
<th>Network average</th>
<th>Max.</th>
<th>Min.</th>
<th>Mean wind at SBL top (m/s) at ALB or KU</th>
<th>Cloud cover at ALB</th>
</tr>
</thead>
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<tr>
<td>259 – 260</td>
<td>389</td>
<td>-0.13</td>
<td>-0.01</td>
<td>-0.10</td>
<td>0</td>
<td>NA</td>
<td>3.06</td>
<td>0</td>
<td></td>
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</tr>
<tr>
<td>260 – 261</td>
<td>339</td>
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<td>2.38</td>
<td>0</td>
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<tr>
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<td>241</td>
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<td>-0.04</td>
<td>-0.11</td>
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<td>NA</td>
<td>6.5</td>
<td>0.08</td>
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<tr>
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<td>-0.07</td>
<td>-0.18</td>
<td>0.14</td>
<td>NA</td>
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<td>0.23</td>
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<td>0.05</td>
<td>-0.53</td>
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<tr>
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<td>-0.01</td>
<td>-0.03</td>
<td>0.02</td>
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<td>289 – 290</td>
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<td>-0.06</td>
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<td>NA</td>
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<td>-0.04</td>
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<td>-0.05</td>
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<td>-0.08</td>
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<td>300 – 301</td>
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<td>-0.18</td>
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<td>NA</td>
<td>2.39</td>
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<td>302 – 303</td>
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<td>-0.46</td>
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<td>NA</td>
<td>7.23</td>
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<tr>
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<td>-0.12</td>
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<td>NA</td>
<td>10.04</td>
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Table 6.1: SBL heat budget components and mesoscale background conditions for all HVAMS IOP nights with no frontal system approaching or overhead. It is presented as SBL elevation above the river rather than height AGL. To get the SBL local ALB site elevation (89m) was subtracted. *Note for night 303 – 304 aircraft soundings available during dusk and dawn, respectively, for days 303 and 304, were used to estimate the SBL height and total cooling. In this case the average site elevation was 79 m.

On calm nights (< 5 m/s at SBL top), turbulent heat flux divergence accounts for 10% of total bulk cooling. For winds ≥ 5 m/s, the turbulent cooling is higher (29%).
maximum turbulent cooling in the network, at individual sites with specific local surface characteristics, can get as high as 40%. At sites with minimal cooling rates it ranges near 5%. For the four nights simulated by the mesoscale model, radiative cooling dominates the total cooling during calm and somewhat windy conditions ($\leq 5\text{m/s}$). This indicates that the radiative cooling plays a major role in the total bulk cooling of the SBL for these wind conditions. For similar or somewhat higher wind conditions, Drüe and Heinemann (2007), using aircraft measurements and tower data over a homogeneous surface in Greenland, found that the radiative cooling is an important or even dominant term in the SBL heat budget in extremely light wind conditions. On the other hand Andre and Mahrt (1982) using data from two experiments at lower latitudes found that turbulent and radiative cooling were roughly equal contributors of the SBL bulk cooling. Sun et al. (2003) not only attributed the total cooling in a tower layer of $\sim 50\text{m}$ deep to radiative and turbulent cooling, but they found that each can contribute approximately the same cooling in the SBL over the night, but that radiative cooling tends to be more important in the early evening and the turbulent cooling afterwards.

Another important factor that affects the estimated cooling rate is the SBL depth selected. (Note that in Table 6.1 SBL elevations above the river are presented instead of height AGL because the study region is over complex terrain and experimental sites do not have necessarily the same elevation.) The turbulent cooling contribution is larger on windy nights with shallower SBL, probably because a larger portion of the SBL depth ($h$) is filled with turbulence or was filled with turbulence at certain periods of its evolution (See Figures 6.2 (a) and (b)). If the values of bulk Richardson number for the regime $Ribr \sim < \frac{1}{4}$, are used to infer the presence of turbulence (Andre and Mahrt 1982), there
are indications that on the night 10/05-06/2003 (278-279), with lower wind winds, the turbulent layer (Figure 6.2 (a)) was more intermittent than on the night 10/08-09/2003 (281-282) (Figure 6.2 (b)), with higher winds. As consequence the turbulent cooling was stronger during the night of 10/08-09/2003 than on the night of 10/05-06/2003 (fourth column on Table 6.1). Figure 6.2 (a) shows that during the night of 10/05-06/2003 turbulence was present at the levels 52m, 71m and 95m (AGL) during ~ the first four hours of the night (2100 UT - 0100 UT), and after was seen intermittently at 52 m (AGL). On the other hand, during the night 10/08-09/2003 the wind was higher and turbulence was continuously present at the heights of 42m and 66m (AGL) during at least the first 7 hours of the night (2130 UT – 0430UT) and at the 86m (AGL) for the first three hours (Figure 6.2 (b)).
Figure 6.2 (a): Bulk Richardson number calculated using tethered balloon data at Anchor (station 10) site during the night of 10/05-06/2003. The different line colors indicate the heights (AGL) for which \( \text{Rib} \) were calculated using the five-sounding levels of TAOS. The dotted horizontal line shows the critical Richardson number \( (Ricr = \frac{1}{4}) \).
Figure 6.2 (b): Bulk Richardson number calculated using tethered balloon data at Anchor (station 10) site during the night of 10/08-09/2003.

In Drüe and Heinemann (2007) the thickness of the turbulent layer was found to approach the inversion height \( h \) only when winds were > 10 m/s at \( z \sim 100 \) m AGL, but for moderate or weak wind < 10 m/s, turbulent layer was found always to be smaller than the inversion height \( h \).
Surprisingly, cloud cover showed no clear cooling influence on the bulk SBL cooling rates. It may be that the cloud cover fraction statistic does not carry information about cloud height and thermal structure, which together with the reflection of upwelling long-wave radiation would impact the downwelling long-wave radiation. A second possibility is that cold or warm advection affected the total SBL cooling. Our selection of IOP nights, with no frontal system approaching or overhead, might have failed for certain nights. A third possibility, is that, the water vapor column across the SBL was significantly different from night to night, in way that it caused differences in the radiative flux divergence. However, we did not perform integration of the specific humidity for the SBL to make any strong statement.

According to the UIB simulations, radiative cooling (Figure 6.3) in the early evening is largest near the surface and moves upward with the growth of the SBL. Hence, just before sunrise the maximum radiative flux divergence lies a couple of hundred meters above the surface near the top of the SBL. This happens because as the layer cools, a strong temperature gradient is establishes at boundary between the cooled layer and the warmer layer above, where radiative flux divergence is strongest. In Figure 6.3 the radiative cooling at the end of the 4 simulated nights 278-279 (10/05 -06), 279-280 (10/06-07), 280-281 (10/07-08), and 281-282 (10/08-09) was largest at few hundred meters AGL, and for the first three nights it corresponds roughly to the observed inversion heights ~ 189m, 259m, 316m AGL (- 89m site elevation) on the same nights. This result agrees with findings of Andre and Mahrt (1982), who concluded that the radiative cooling was largest at the top of the temperature inversion. For the windiest
nights, 276-277 and 304-303, which had moderate SBL depth, the turbulent cooling contribution were the highest ones, 40% on 276-277 and 50% on 303-304.

The fact that radiative cooling is important for the total cooling near the surface only during the early part of the night, and that the turbulent cooling tends to act in the lower part of the SBL (Andre and Mahrt, 1982) during any time of the night, suggests that the nocturnal minimum temperatures measured by surface weather stations at ~ 2m (AGL) are more likely to be affected by the occurrence of turbulence than by radiative loss. If this is the case, local site characteristics, such as sheltering and local topography, may dictate the occurrence of mixing, which, depending on the heat flux divergence within the layer of the temperature measurement, will prevent the near-surface temperature to reaching a minimum during the course of the night.
Figure 6.3: Four days of simulation of the evolution of the spatially averaged radiative cooling rate (color code at right) and net radiative flux at 1.4 m (black line; 30 minute time intervals).

For the four nights with calculated radiative flux divergences, we cannot close the layer heat balance, even though we selected nights with no frontal systems overhead or approaching, reducing the likelihood of large scale (> 100 km) temperature advection. In the night 278 (10/05) – 279 (10/06) 15 % (-0.06 K/h) remains unexplained, in 279 – 280 18% (-0.1 K/h), in 280 – 281 10% (+0.05 K/h), and in 281 – 282 9% (+0.07 K/h).

Perhaps, a smaller scale (<100 km) temperature advection due to land surface contrasts affected our heat budget balance.
6.4 Summary and conclusions

Radiative flux divergence at the surface and across the SBL overall nights accomplishes most of the total cooling of the SBL. In all cases, turbulent heat flux divergence contributes least to the total SBL cooling. However, its importance increases with increasing mesoscale wind. On windy nights (>5m/s) it can be responsible for 40% - 50% of the total cooling. For the simulated nights, close to calm wind conditions, radiative cooling was always a dominant term of the cooling >60%.

Although radiative cooling plays the major role in the SBL cooling, local mixing can still be an important mechanism affecting or even controlling local minimum temperatures at certain sites well exposed to the flow (See section 7.4 on Chapter 7), if other processes such as drainage flows are unimportant. Under this situation, in situ radiative cooling is most important at the beginning of the night and moves upward during the course of the night, while turbulent mixing episodes can happen at any time.

Some of our difficulties may come from the simplifying assumptions we used, but such difficulties in closing the energy budget observationally are not ours alone. Few researchers even attempt this calculation from data. Sun et al. (2003) used local measurements of radiative and turbulent flux in a 50 m tower layer but were also unable to close the heat budget locally. Horizontal and vertical advection effects were the main terms in the heat budget residual. It is interesting that both in our calculation and in the Sun et al. (2003) ones the residuals were positive and negative at different times in the evening.
The Garratt and Brost (1981) simulations of the nocturnal boundary layer showed that bulk cooling in the SBL for \( 0.1 \, h < z < 0.8 \, h \) is dominated by the turbulent cooling and that the radiative was cooling important only at the edges of the SBL, e.g. close to surface \( (z > 0.1 \, h) \) and near the top of the SBL \( (z < 0.8 \, h) \). Perhaps because they specified a 10 m/s wind at SBL top (~100m), they examined weakly stable, more turbulent layers than the ones we are concentrating on (see Table 6.1 columns 2 and 9).
7. Nocturnal turbulence behavior at a single site over seasons.

7.1 Introduction

In this section we analyze data from the HVAMS long-term operational phase (LTOP), considering data from the single flux tower (#10) that continued to operate, data from 7 weather stations, and regular radiosonde data from Albany, NY. From the five weather stations, 1-minute average temperature, wind and pressure data are used. From two stations that are part of the regular weather surface observation data acquisition (ASOS stations at Poughkeepsie and Albany NY), we obtained hourly cloud base data to calculate the nocturnal average cloud cover. The flux station provided us with 10 Hz turbulence data plus 1-minute averages of temperature, wind and surface pressure. The turbulence data covers a period of approximately 3 years. The sounding and weather station data approximately 2 years, but we use only two years to assure data overlap. (See Section 2.1.3 of Chapter 2 and Appendix A.1 for experimental array details, operational periods, and localization.)

As in Chapter 5, we use a regional Richardson number ($Ribr$) to evaluate the background flow state and relate it to the observed turbulence. We investigate how $Ribr$ varies seasonally. Ideally, we would need the same flux station network used in Chapter 5 for the IOP to obtain the region average fluxes. Sensor limitations restricted our flux data to station 10. Since station 10 was one of the most turbulent stations in the network during the IOP, and was well exposed (See Figure 2.5 for the station 10 TF). It is appropriate to represent the most active sites, which better sense the regional SBL state over the HVAMS region.
In Chapters 4 and 5, we could investigate not only background flow influences but also examine landscape influences on the local turbulence because we looked to different nights with a spatially distributed network of surface stations. However, the collection of nights was restricted to second half of September/2003 and entire October/2003. In this chapter, background flow influences on the local turbulence are studied for a much longer period. Though we are fixed to one landscape type, having years of data leads to adequate natural variability in $Ribr$. The aim here is to develop a climatology of the boundary layer for stable nights, to determine the conditions favorable for the occurrence of nocturnal mixing events at a site representative of the Hudson Valley open areas, and also to investigate the distribution of temperatures according to the surface curvature using 6 stations for a period of approximately 2 years.

### 7.2 Results – Variations in the LTOP turbulence and fluxes with mesoscale flow state

The technique used in Section 3.1 was applied to the LTOP of station 10 to identify the intermittent turbulent events and to obtain the nocturnal fluxes. In Figure 7.1 we can see dependence of the station 10 momentum and sensible heat fluxes on $Ribr$. Both fluxes become more negative with decreasing $Ribr$ and for $Ribr > Ricr$, there is still appreciable flux. This result reinforces the main result presented in Chapter 5. However, now the evidence is stronger because it was found using ~ 2 years of data.
The number of events per night at station 10 is smallest under weakly stable flow ($Ribr < Ribcr$), largest in the range $Ricr < Ribr < 5$, and intermediate for $Ribr > 5$ (Figure 7.2). As additional information, in Figure 7.3 is presented the intermittency index as a function of $Ribr$, which shows that in the range $Ribr < Ricr$ approximately 9/10 of the nighttime is turbulent, while for $Ribr > 5$ only 3/5 of the nighttime is
turbulent. The result presented in Figure 7.3 somewhat agrees with the one from Doran (2003), which showed that fraction of time with turbulence within 1-hour period increases with decreasing stability. However, in that work the author used a local Rib instead of regional one. Combining this information with the number of events per night, shows that the turbulence is more continuous at low Ribr than at high Ribr, showing that in the Ribr < Ricr regime turbulence tends to be strong and continuous during a few long-duration events (1.7 events / night), while for Ricr ~ < Ribr <~ 5, intermittent turbulence with several turbulence breakdowns (2.6 events / night) are characteristic. For Ribr > 5 there is also intermittent turbulence but fewer, short-lived events (2.1 events / night) were observed.
Figure 7.2: *Number of events per night* as function of *Ribr*. Each point represents 30-night average and the vertical line is the standard error.
Figure 7.3: Intermittency index as a function of Ribr. Each point represents 30-night average and the vertical line is the standard error.

7.3 Results - Seasonal changes in mesoscale flow state and in the site turbulence

The seasonal median number of intermittent events per night is minimum during spring and winter (~1 event / night), intermediate for fall (1.5 event/night), and maximum in summer (~2 events/night; Figure 7.4 (a)). Seasonal variation of the seasonal median number of events per night (1 to 2 events/night) is small, though within each season there
is a quite large spread. This is most evident for summer (1 to 5 events/night). The seasonal median intermittency index also shows variation across the seasons (∼ 0 to ∼ 0.1), though it is small compared to the nightly basis intermittency index (0 to 1; See Figure 7.4 (b)). (Note from Chapter 4 Section 4.3 that intermittency index = 0 means turbulence occurs during the entire night, and = 1 indicates no turbulence at all during a nighttime period). Both results combined show that turbulence tends to be less intermittent and more continuous during the winter and spring, and more intermittent and less continuous during fall and summer. There is practically no difference in the turbulence behavior for winter and spring. For summer and fall there are differences. Fall exhibited less intermittency (1.5 event/night) with the least turbulent time (∼ 0.22 continuous turbulence time fraction) while summer was more intermittent (2 events/night) with longer turbulent time (∼ 0.1 intermittency index).

Owing to its sensitivity to small changes in the denominator, Ribr shows large spread skewed toward large values (Ribr >> 1) in the nightly basis for all the seasons (See Figure 7.5 (c)). On the nightly basis, within the bins, Ribr can be as wide as near zero to over 6 in the fall, with some slightly unstable cases during winter. Such slightly negative Ribr values can be the result of cold air being advected over warmer surface. On the other hand, the seasonal median Ribr do not differ much amongst the seasons; they lie within the range ¼ < Ribr < 1. Which indicates that Ribr larger than critical (Ricr) is common over the seasons. This result with the previous two, which showed more turbulent time than none turbulent time for all seasons, reinforces the result found in Chapter 5, that turbulence does exist for Ribr > Ricr. However, it is important to recall
that this is a regional Rib and the wind and temperature profiles used to determine it are not local but rather represent a certain region.

In addition, if we compare the seasonal median turbulence indicators (number of events per night and intermittency index) with the seasonal median Ribr, we find that winter is the least stable and the most turbulent season followed immediately after by spring, with the second least stable and second most turbulent season. In the other extreme there is the summer and fall seasons, which are the most stable seasons, with more intermittency and less turbulent time.
Figure 7.4: a) *Seasonal median number of events per night* for the years of 2005 and 2006. b) *Season intermittency index* for the same period as in (a). c) *Seasonal median nocturnal cloud cover fraction* for the same period. d) *Seasonal median SBL height* ($h$) for the same period. The whiskers show the entire range, the upper and lower notch of the box the interquartile range, respectively ($3/4$ and $1/4$), and the solid horizontal segment the median.
Figure 7.5: a) Seasonal median wind at the SBL top ($U_{sbl}$) for the years of 2005 and 2006. b) Seasonal median potential temperature difference across the SBL ($\Delta \theta$) for the same period. c) Season median regional Richardson number ($Ribr$) for the same period. The horizontal dashed line shows the critical Richardson ($Ricr$) value ($\frac{1}{4}$).

We discuss separately results from wind speed at the SBL top ($U_{sbl}$), SBL height ($h$), and potential temperature difference across the SBL ($\Delta \theta$) because all this information is contained in the $Rib$ formulation, and it represents the effects of all them acting together on the flow stability. As noted earlier in the Introduction, $\Delta \theta$ and $h$ are controlled by the external parameters $U_{sbl}$, cloud cover fraction, and surface type. The former is
important when considering a heterogeneous region, which is not our emphasis in this section, focusing on a single station. It is interesting to speculate about the influence of these two parameters on $\Delta \theta$ and $h$ across the seasons. Recall that the wind at the SBL top is the result of a balance of a mesoscale pressure gradient, Coriolis force, momentum flux divergence, and momentum advection and therefore it is not a true external parameter. The horizontal pressure gradient should be considered as an external parameter but because, we do not have measurements of it above the surface, and because flow motion only exists because there is pressure gradient we assume $U_{sbl}$ as an external parameter to our discussion. Therefore, $U_{sbl}$ affects $Ribr$ explicitly and implicitly through $\Delta \theta$ and $h$.

The median winter $U_{sbl}$ (~12 m/s) is the second highest (See Figure 7.5 a), cloud cover fraction (~0.6) the highest (See Figure 7.4 (c)), and $\Delta \theta$ the smallest (~6 K) (See Figure 7.5 (b)) amongst the seasons. A small $\Delta \theta$ might be partially explained by the high seasonal median cloud cover fraction that reduced the long-wave radiation loss to the space. On the other hand, $h$ was the lowest for this season (~310 m) (See Figure 7.4 (d)), which seems to be contrary to the idea that higher winds cause more turbulence leading to a thicker SBL. Recall that in Chapter 5 we used the first significant temperature inversion top to define the SBL height. So, it is possible that the SBL thickness chosen refers rather to a turbulent layer and not to the top of a deep temperature inversion. In case of clear turbulent layer, we would see an inversion right near the surface, in the surface layer, and other narrow steep one (temperature jump) away from the surface at the top of the turbulence, which we use to define $h$.

One could argue that the high wind speeds close to the surface during the winter, might be a sign that there is large wind shear near the surface due to the SBL decoupling
from the surface. Nevertheless, the low number of intermittent events per night and low intermittency index indicate more turbulence during this season than the rest of the seasons that invalidates the argument above.

During the summer season $U_{sbl}$ was the second lowest ($\sim 9$ m/s) and cloud cover fraction was on the lowest level (0.5). Surprisingly, the summertime $\Delta \theta$ was the second lowest ($\sim 7$ K), not the largest. This result seems to contradict the assumption that with low cloud cover fraction there is high radiative loss and with weak winds there is less turbulence making the temperature inversion to be stronger, extending from the surface up to some height above the surface layer. This season tends to be moister than the others, and perhaps the possible extra moisture contributed to reduce the long-wave loss. $h$ is the second highest ($\sim 380$m) during this season. Such a thick SBL might refer to a deep weak inversion from the surface up to ~ 380 m, and not to a deep turbulent SBL marked by a strong narrow inversion away from the surface.

7.4 Results – Temperature distribution according to the landscape

Figure 7.6 shows that the nocturnal surface temperature distribution depends on features in the local landscape. In Chapter 4, it was not clear that the warmer stations were localized at the more convex surface, nor that the colder ones were located at the more concave surfaces (See Figure 4.14 (a) and (b) on Chapter 4). It was also not true that stations higher than immediately surrounding were necessarily the warmer ones in the network and that the ones lower were necessarily cooler (Figure 4.16). From Figure 7.6 we can see that the potential temperature spread factor ($SF = (T_{st} - T_{min})(T_{max} - T_{min})$, where $T_{st}$ = station’s temperature, $T_{min}$ = nocturnal median network minimum temperature, $T_{max}$ = nocturnal median network maximum temperature) is maximum at
stations higher than the immediately surrounding terrain (negative $mz-z$) and decreases for stations lower than the immediately surrounding terrain (positive $mz-z$). In other words, $SF$ decreases with increasing $mz-z$. Here, we used areas of ~ 2km x 2km to evaluate $mz-z$. Such a scale choice is based on the fact that most of the elevation variance is capture on this scale and perhaps can account for most of the terrain induced flows (See Appendix A2).

Under conditions of widespread turbulence, forced by strong wind and/or high cloud cover fraction ($Ribr < ¼$), it would be expected that the $SF$ would be more uniform because of reduced spatial temperature differences, however $SF$ does not change significantly with $Ribr$ for the majority of the stations (e.g. station 6, 4, S, and 12 (H2)). Only station 10 and 15 (H5) seem to have significantly higher $SF$ values under these conditions. The high $TF$ at station 10 suggests that such increase in $SF$ could have been caused primarily due to an increase of the turbulence locally, but at station 15, which has intermediate $TF$ it is not clear (See Figure 2.5 on Chapter 2 for station’s 10 and H5 $TF$s). On the other hand, stations 4, S (H3) and 12 (H2) did not have their $SF$s changed significantly for the different classes of background flow ($Ribr$). Opposite to station 10, the reduced exposure of station S might have impeded any turbulence increase. This station is located inside a patch of woods on the west side riverbank approximately 150m from it.

Perhaps, we could not verify a reduction of the $mz-z$ effects on the temperature distribution, for low weakly stable SBLs ($Ribr < ¼$), because the differences in local sheltering did not allow the turbulence level to homogenize across the different sites.
Ideally, to investigate the \( mz-z \) influences on \( SF \) we would need sites with similar exposure.

Figure 7.6: SF versus \( mz – z \) using a network composed by 5 weather stations (all HOBOs) plus station 10. Note that station number or symbols are given in the top of the plot. (Table A.1.1 in appendix A.1). The whiskers show the entire range, the upper and lower notch of the box the interquartile range, respectively (3/4 and ¼), and the solid horizontal segment the median. The symbols, red-square, green-circle, blue-triangle, and light blue –plus, represent the four different background flow stability conditions.
7.5 Summary

The turbulence (momentum and sensible heat fluxes and intermittency) depends on $\text{Rib}_r$. At low positive $\text{Rib}_r (0 < \text{Rib}_r < \text{Rib}_{cr})$ turbulence is most continuous and momentum and sensible heat fluxes are relatively large downward fluxes. As $\text{Rib}_r$ increases ($\text{Rib}_{cr} < \text{Rib}_r < 5$), turbulence becomes intermittent and the total nighttime period filled with turbulence is shortened, while the fluxes become less negative. In this regime there is the highest occurrence of intermittent events but the individual event duration is shortened. At high positive values ($\text{Rib}_r > 5$), turbulence is still intermittent, but the events happen less often and are short lived. At this extreme the fluxes are the least negative.

Seasonal medians of turbulence behavior do not show great variability compared to night-to-night variability. The seasonal medians of number of events per night and intermittency index do depend on the seasonally varying $\text{Rib}_r$, and this dependence is close to what is observed on the night basis (Figure 7.2 and 7.3).

The range of $\text{Rib}_r$ values found within the seasons is close to that found over the entire year. Thus, it suggests that the $\text{Rib}_r$ values found during the IOP might be comparable to the ones found during the LTOP. Figures 7.7 (a) and (b) show that even though in the LTOP there are more cases than in the IOP, most of the cases in both LTOP and IOP occur in the range $0.1 < \text{Rib} < 100$, however the LTOP distribution is wider, including very weak stable cases and very strong stable cases, respectively. In conclusion, the IOP period had the most expected values of $\text{Rib}_r$ but did not cover all the range found in the two years of the LTOP, and therefore the range of stabilities of the
background flows found in the IOP is not entirely representative of the ones in the LTOP. Nevertheless, it does not invalidate the common agreement between the results found in both periods (LTOP and IOP), that there are still significant turbulence (fluxes) for $Rib_r > Ric_r$.

Depending on the background flow state, which is controlled by the wind at the SBL top and cloud cover fraction, turbulence may or may not be forced regionally, while the different types of landscape makes the local flow to respond differently to the regional flow state. If there were no spatial inhomogeneities, the landscape would not cause any influence, and any change in the flow would be entirely caused by changes in the state of the background flow. Now, if the background flow were kept fixed and the landscape was heterogeneous, any spatial difference would be uniquely caused by spatial changes in landscape. In this sense, analyzing the long-term record focuses primarily on changes in the background flow (Figures 7.1, 7.2, and 7.3), while IOP changes in the background flow (Figures 5.11 and 5.12) as well when considering the network average fluxes, but also on changes of landscape when looking to the individual stations (Figures 5.14 and 5.15) with different local site topography and sheltering. When comparing Figure 7.7 (a) with 7.7 (d), it is clear that the range of variability of both, regional $Rib$ and local $Rib$ ($Rib_l$), are compatible, showing that what was achieved temporally in terms of regional flow state in ~ 2 years is comparable to what is achieved spatially in terms of local flow in just 1.5 months. Here, it is worth noting that because $Rib_r$ and $Rib_l$ were obtained in a different manner ($Rib_r$ using bulk $\Delta \theta$, $\Delta U$, and $h$, while $Rib_l$ using local temperature and wind profiles with a much smaller fixed height) the $Rib_r$ distribution is skewed towards larger values of $Rib$, while $Rib_l$ towards to lower values. In addition,
comparing Figure 7.7 (c) with (d), we can see that the local flow states found at a single
stations under the influences of wide range of background flow states is not as diverse in
terms of stability as the ones found over ten different stations but under the influences of
narrow range of background flow states. This shows that under a wide range of
background flow conditions local landscape can limit the range of local flow stabilities
achievable by a station.

Figure 7.7: a)- Probability distribution of $Ribr$ in the LTOP. b)- As in (a) but for the IOP.
c) – Probability distribution of $Ribl$ in the LTOP. d) – As in (c) but for the IOP.
The temperature distribution analysis showed that in mesoscale areas with heterogeneous topography, such as the Hudson Valley, sites higher than the immediately surroundings are warmer, and lower sites are cooler. The assumption that during weak stability conditions spatial temperature should be diminished was not clear using the network of stations during LTOP. Perhaps, because there was no selection in terms of topography, sheltering, and land cover/land use prior to the installation of such array, the topographic-terrain induced effects on the temperature distributions could not be isolated. Site sheltering might have masked such influences, which on the other hand might have influenced the investigation of sheltering effects on the temperature distribution done in Chapter 4 (See Figure 4.14 on Chapter 4).
8. Summary and conclusions

In this section we return to the scientific questions outlined at the beginning of the thesis.

8.1 Stable Boundary Layer heat budget

Chapters 4, 5, and 6 addressed the main question: *How does the atmospheric stable boundary layer (SBL) form over realistic heterogeneous terrain?* Chapter 6: *The specific question—Which agent (turbulent or radiative cooling) is responsible for the bulk SBL cooling?*

Under very stable conditions (e.g., wind speed at ~ 200m AGL < 5m/s and cloud cover fraction < 0.5) in complex terrain topography with patchy land cover, radiative cooling is the principal cause of the establishment of the SBL, while turbulent mixing within the colder surface plays a relatively minor role. Radiative cooling dominates near the surface for the first 2-3 hours after sunset. Abrupt onset of near surface turbulent mixing usually leads to sudden warming, as heat from the potentially warmer air above encounters the colder surfaces below, but subsequent radiative cooling maintains steep vertical temperature gradients. If the mixing event occurs internally to the SBL, potential temperature within the mixed region tends to be constant, depending on the vigor of the mixing. This leads to a steep potential temperature gradient at the top of the SBL. This internal mixed layer will be immediately affected by the action of radiative flux divergence in the mixed layer such that it seeks a more stable isothermal equilibrium. For
weakly stable conditions, the relative contribution of turbulent cooling contribution is increased.

Other process such as drainage flows and mesoscale advection forced by land cover contrasts may also play a role in the development of the SBL. However, these influences have a smaller effect than does radiative cooling and turbulent cooling, under any circumstances observed here. Processes such as microscale advection try to horizontally homogenize the SBL (AF2003 and Mahrt et al., 2001).

For most of the heat budget case studies (see Table 6.1) the thickness of the layer in which the budget was performed refers to the observed inversion height rather than the actual height of a turbulent layer. There is evidence from tethered balloon (TAOS) profiles (e.g., Figure 3.3 (a) and Figure 6.2 (a)) that the surface layer was disconnected from the upper SBL at times. Under this circumstance turbulent cooling did not contribute directly to the cooling in the upper SBL, but only to the cooling of the surface layer if when there was mixing at the surface. Note that the observed surface layer temperature can rise if mixing brings warmer air from above. However, the regional average SBL will have net flux divergence (cooling) rather than convergence (warming) only if mixing touches the cooler surface. Other possibility is that turbulence might exist aloft for most of the time but reach the surface only intermittently, characterizing a condition of upside down SBL, as described by Mahrt (1999). When there is no mixing at the surface, there would be no cooling or warming but rather a redistribution of heat. Once intermittent mixing happens then there would be net transfer of from the SBL to the surface. But in any circumstance the heat flux at the surface would be an upper limit for the heat transfer between the SBL and the surface.
Other possibilities are that the flux stations were not located at the most active sites in the region or that the modeled radiative cooling data was overestimated.

The study presented here is novel. It differs from previous work (e.g. Garratt and Brost, 1981; Sun et al., 2003; and André and Mahrt, 1982) because we attempted to close the SBL heat budget observationally for a heterogeneous landscape. In Garratt and Brost (1981) the modeled heat budget closure referred to higher wind conditions than those addressed here, and included assumptions of a homogeneous surface with a stationary geostrophic wind. In our case we referenced the observed wind ~ 200m AGL and the surface was clearly not homogeneous. Under conditions of stronger background flow their work showed that the turbulent mixing accomplishes most of the cooling of the SBL and that the radiative cooling was important only near the surface and near SBL top, where the temperature gradient was strongest. Sun et al. (2003), using only experimental data to estimate all the terms in the budget, showed that turbulent and radiative cooling were comparable. Near the surface, radiative effects were important only during the first part of the night, while turbulent effects were important during the entire night. Even though Sun et al. (2003) did use experimental data, their measurements were limited to single site, not necessarily representative of a wider area, such as the HVAMS region. André and Mahrt (1982), using indirect flux measurements and modeled radiative cooling, found that for the bulk SBL layer depth, which corresponded to the inversion layer, turbulent and radiative cooling are equally important. In contrast to André and Mahrt (1982), the present work used actual network flux averages, while in previous work turbulent sensible heat flux was estimated indirectly based on the surface energy budget or temperature and wind profiles. They found that in the lower SBL turbulent
effects are more important, while in the upper SBL clear air radiative cooling might be the only agent acting. Their work is most similar to the analysis presented here. The current work indeed suggests that turbulent cooling might act at the surface and radiative cooling at the upper parts of the SBL, but our results indicate that the for the bulk of the SBL, radiative cooling was dominant.

The SBL heat budget presented here would have benefited from better estimates of the radiative cooling based on local temperature and specific humidity profiles, or even better, if real radiative flux profiles were available. We did not have regular vertical soundings at all the surface station locations in the Hudson valley nor real radiation profiles at any location, but only one point measurements of net radiation. The main source of uncertainty in the heat budget in the current work comes from the radiative cooling estimates, which were based on modeled data, and the total SBL cooling, which was for the majority of the cases based on regular weather soundings at a single location.

8.2 Regional Rib – (Mesoscale influences)

Chapters 5 and 7 addressed the scientific question: What do the background mesoscale conditions (wind and cloud cover) do to the regional turbulence activity?

We found that the stability of the background flow, controlled by the horizontal mesoscale pressure gradient and regional cloud cover, is responsible for forcing the regional turbulence activity. A critical $R_i$ that serves as a turbulence cut off does not seem to exist. This work has demonstrated that mixing during regionally supercritical conditions is an observational consequence of spatial averaging, which reflects the
influence of a few “hot spots” of mixing (AF2003), that might dominate the regional momentum and heat transfers. Turbulence seems not to vanish for any value of $Ribr$, although it decreases with increasing $Ribr$. It can be imagined that even for a condition with no average mesoscale pressure gradient, turbulence can still be generated locally by drainage flows, or any microscale circulation induced by topography or land cover contrasts. In this way, turbulence would entirely be controlled by the stability of the mesoscale flow only if the surface were flat and homogeneous. Otherwise, features of the local landscape will influence the local turbulence activity. This result is consistent with previously work (Mahrt, 1987) suggesting that spatial heterogeneity leads the local $Rib$ to differ from the $Rib$ determined from the area-averaged wind and temperature profiles.

The area-averaged $Rib$ tends to be smaller than the profile averaged-$Rib$ causing the flux in models, which use the latter approach, to be underestimated and therefore extra mixing has to be allowed in the models. Mahrt’s work, however, was based on a hypothetical distribution of $Rib$ within a grid size model, while the current results are based on observed spatially averaged network fluxes, locally observed $Rib$, and regional determined $Rib$, based on upper SBL wind and temperature and network surface minimum temperature.

The current work shows that local mixing (momentum and sensible heat fluxes, the number of intermittent events and the intermittency index) responds to the mesoscale background flow. However, the landscape nearby imposes limits for the range of stability achievable at a single site. It was shown that even during the Long Term Observational Period (LTOP), when a more diverse set of background flow conditions was encountered than during the IOP, the range of local flow stabilities at station 10 were not as diverse as
those found locally during IOP. Another interesting result is that intermittency is not only affected by terrain curvature (see Chapter 5), but it also responds to the background flow stability. The nighttime filled with turbulence, quantified by the intermittency index, seems to decrease with increasing $Ribr$, but the number of intermittent events per night shows a more complex behavior. It increases with $Ribr$ in the range $0 \sim < Ribr \sim < 5$, and decreases thereafter. If we make an analogy of local concavity with $Ribr$, turbulence tends to be more continuous at low $Ribr$ ($Ribr \sim < \frac{1}{4}$) (analogous –most convex sites), with few intermittent events that may last more than $\frac{1}{2}$ of the nighttime while at high $Ribr$ (most concave sites) turbulence is more intermittent with a high number of intermittent events that last $< \frac{1}{2}$ nighttime period. But the difference between the effects of $Ribr$ and concavity is that for $Ribr$ the number of intermittent events does not always increase with increasing $Ribr$, there is a value where the number of events peaks, while for concavity this is not the case. At $Ribr \sim 1$, the number of events is maximum and for larger values decreases. This regime reflects the effects of strong stability.

These results support earlier findings but extend them by providing observational support for the necessity of mixing at the high $Ribr$, first found empirically by modelers. For future work, it would be desirable to have a denser network of stations covering wider range of landscapes. Deeper vertical wind and temperature profiles taken at different locations would provide a better estimate of $Ribr$ that would be close to the model.
8.3 Landscape influences

Chapters 4 and 7 addressed the scientific question: *What is the role of topography and landscape structure in determining the turbulence activity and representativeness of local measurements?*

We can draw some qualitative conclusions about the relative importance of the different landscape influences on nocturnal turbulent exchange. Developing more precise quantitative relationships for turbulent fluxes, the *number of intermittent events*, *intermittency index* and surface temperature distributions in terms of local terrain curvature, slope, *TF*, elevation, *mz - z* and land cover/land use must await deployment of a more extensive surface station network. The network analyzed here, consisting of 10 flux-measuring stations and 5 additional conventional weather stations, was necessarily limited. Site selection was determined primarily by identifying the locations that were technically feasible that spanned the region. With the benefit of the analyses presented here we now have a better idea of how to select specifically for a representative sampling of landscape features before locating stations. These limitations meant that the individual influences of each parameter, for example, absolute elevation (*z*), *mz - z*, curvature, *TF*, and slope could not be completely isolated. However, the importance of each parameter (except for slope) was quantified at the extremes of its range of variability. Despite these limitations we can reach certain conclusions about the relative importance of local terrain curvature, *TF*, and elevation under given conditions of background flow and cloud cover fraction in nocturnal conditions.
Very local (e.g. typical length scale ~ 0.3km) obstructions and terrain curvature directly control turbulence, and indirectly influence local surface temperature. For conditions of strong background flow (e.g. > 5m/s), site obstruction is more important, while under weak background flow and clear skies (e.g. < 5m/s and cloud cover < 0.5), concavity becomes more important. With strong winds momentum and heat exchanges greatly differ for winds from obstructed and open directions. During calm conditions overall exchanges are smaller as are the differences between the different directions. Local concavity, on the other hand, is of greater importance in calm conditions. Under this condition the local flow stability is stronger, and is consequently more heterogeneous across the landscape, with the most stable flows at concave sites and less stable at convex ones. Turbulent fluxes in this situation increase (magnitude) with decreasing concavity. However, when the winds are stronger and the sky overcast (e.g. > 5 m/s and cloud cover > 0.5) heterogeneity, perhaps the result of local flow stabilities, is reduced and turbulence depends less on concavity.

The effect of the main valley central depression, represented by the absolute elevation ($z$), was demonstrated to have a control on the surface temperature distribution. For calm winds and clear skies, a cold air pool is established at bottom of the valley. From the case studies of all calm and clear skies nights observed during the IOP, the cold pool was about 200 m thick from river level (which is sea level). Moreover, isentropic surfaces near the cold pool top and above were not completely horizontal, but were bent up parallel to the valley sidewalls slopes. Another possibility (Fitzjarrald et al, 2008) is that the cold pool is tilted across the valley and cross-valley pressure gradient is established to balance the Coriolis force. Nevertheless, we did not investigate this
hypothesis. For stations within the cold pool, $z$ is apparently not relevant; the relative elevation $mz - z$ is better related to temperature distribution. This result differs from that discussed by LeMone et al. (2003), who found that potential temperature varied linearly with absolute elevation. In HVAMS, the surface potential temperature follows a more complex variation with elevation. This may be because in the previous work the landscape structure was less complex than in the present work. Therefore, the effects of site sheltering, local curvature, and other local topographic features of the terrain affected the temperatures less. According to our results, the Transmission Factor ($TF$) and curvature affects turbulent sensible heat flux, which ultimate affects temperature, while nearby topographic features such as surrounding slopes and landscapes contrasts can cause local circulations causing horizontal temperature advection.

The surface temperature distribution follows neither the very local terrain concavity nor $TF$, nor slope. A possible explanation for this might be that the scales of all the processes controlling surface temperature are not comparable. Turbulence appears to have spatial length scale on the order of a few hundred meters or less, while drainage flows or other types of microscale horizontal cold or warm advection, and/or in situ radiative cooling, may have larger scales. The former assertion was not directly shown in this work but we hypothesize it to be true based on the results of surface temperature and $mz - z$ (Figure 7.6). When comparing $SF$ with $mz - z$ (with length scale on the order of ~2 km), it was found that potential temperature increases with decreasing $mz - z$. To summarize, in a region with complex terrain, places higher than the 2km x 2km surrounding areas tend to be warmer and those in hollows are cooler than their immediately surroundings (Figure 7.6). For the HVAMS study region, the terrain
elevation variance converges to nearly constant values for horizontal length scales at the order of a couple kilometers. The length scale for which elevation variance converges should be comparable to the typical length scale of the hill/dips composing the area.

Quantitatively relating turbulence intermittency to landscape properties constitutes a new result. We have shown that the number of turbulent breakdowns (number of intermittent events) increase with local terrain curvature, in such a way that they are maximum at the most extreme convex places and minimum at the most extreme concave places. An index of turbulent continuous fraction was developed, and it showed that the nighttime exhibits a longer period of turbulence at convex sites than at concave places. Therefore, one should find that stations placed near hill tops would tend to have a few long-lasting turbulent events, which characterizes a more continuous turbulent regime, while the ones placed on dips would tend to have an elevated number of short-lived turbulent events, more intermittent turbulence. For the range of small terrain curvature or for near flat surfaces, these two indices did not show any pattern probably because changes on other landscape features, such as local slope, and TF, overwhelmed changes on concavity.

The results found in the current work of surface temperature with mz - z, fluxes and mz - z, and fluxes and TF agree with the earlier findings of AF2003. They showed a similar tendency (increasing fluxes (magnitude) with decreasing mz - z, and increasing TF, and increasing surface temperatures with decreasing mz - z) for a denser network of stations in a smaller area (Figures 4.9 (a) and 4.9 (b), A3.1, and A3.2). In terms of intermittency or more specifically number of intermittent events per night, the situation was different. We found an increase of the number of intermittent events with increasing
$mz - z$, while AF2003 found the reverse. In our case such behavior was not only valid for $mz - z$ but also for the least squares fitted surface concavity. Perhaps, because in AF2003, lacking direct turbulence measurements, weaker events could not be detected by using their method of identification of turbulence breakdowns, which was based on periods of simultaneous increase of temperature and wind speed. However, the results founded here, specifically for turbulent fluxes and the number of intermittent events events/night, are strong evidence that fluxes become more negative or stronger downward with decreasing $mz - z$ or increasing $TF$, while the number of intermittent events/night increase with increasing $mz - z$.

The parameter $mz - z$ was important because it was the only factor strongly related to turbulence and surface temperature. As was mentioned before, it does not isolate any landscape features effect but rather combines effects of relative local elevation, slope and curvature together. This parameter may be suitable for the formulation of a relation of turbulent fluxes and surface temperatures with the landscape. For example, one could take a region like the one in the Hudson Valley and draw a grid with area size of ~ 2km x 2km and weight the temperatures and fluxes by the values of $mz - z$ if mathematical expressions were available.
8.4 What are the consequences of understanding landscape influences on turbulent mixing, on temperature distribution, and on efforts to model the stable BL?

Contemplating the present works and speculating on the way forward.

Understanding topographic and land cover influences is important for interpolating climate data to areas for which surface climate data is unavailable. This is particularly important for complex landscapes (Daly, 2006). Many interpolation schemes are currently used. The simplest ones only use the distance from the near stations to weight their influences, while the most sophisticated not only use the distance but take into account the similarity of various aspect of the terrain such as elevation, slope, slope aspect, proximity to coastal areas, moisture availability, and topographic positions (e.g. top of a ridge middle of a valley; Daly, 2006) of the nearest stations with the unknown point to predict its temperature. These have been used largely to estimate surface temperature patterns. In the most sophisticated approaches, effects such as drainage flows, pooling of cold air, and presence of persistent temperature inversions are considered but in a statistical way.

It is also very important to assess the representativeness of local measurements (surface temperature) when using a long record temperature data from specific site to infer temperature trends. In recent years assessing the suitability of long time series from surface climate station temperature records for climate change studies has assumed increasing importance (Klotzbach et al., 2009).

The procedures just outlined are essentially empirical. To date, no one seems to take into account the effects of turbulence, and perhaps the present work aiming to link
the temperature distribution, turbulent mixing occurrence with landscape features could help to improve these schemes. The approach presented in this thesis, relating how turbulent mixing is distributed across the landscape, can assist in the improvement of NWP/GCM models. It could provide a systematic methodology first to understand spatial variability in climate variables and then to parameterize them in model and climate studies. This is especially true for momentum and heat exchanges between the surface and lower atmosphere, and for describing the diffusion of pollutants. This new knowledge would also aid forecasting minimum temperatures, though the effects of larger scale phenomena such as local circulations and drainage flows still must be more clearly understood.

The approach and results of the work described in this thesis can pave the way for future studies in which a great number of flux and weather stations would be deployed. We showed that carefully considering topography and land cover before siting stations is extremely important. In this way one can assure that the network as a whole spans the range of each topographic parameter (local curvature, elevation, and slope) or land cover (TF, and roughness, and other properties of the surface such as heat capacity, soil thermal conductivity, and emission/absorption) that characterize the landscape in question.

For example, it is now feasible to construct a network of a hundred flux-measuring stations, (Oncley et al., 2010 – CentNet). Careful study of the microtopography of a proposed study region can identify important elements, such as small hills and hollows that the curvature analysis presented in Chapter 4 points toward. Second, a proposed region can be divided in terms of land cover types (i.e., forest, wetland, and urban) and appropriate canopy heights and roughness measures associated
with each. A key aspect of future work should be to relate the sheltering measure used here, the *transmission factor*, to the conventional notions of roughness length and displacement height. With more stations available, one would seek to ‘span the parameter space’ for the variables considered in this thesis. For example, were one to deploy surface stations in the HVAMS study region again, it would be desirable to perform a more elaborate study of local curvature, to identify sites with similar elevation and obstruction, sites with similar concavities and sheltering, and for a particular elevation range, sites with similar sheltering and concavity. It is known that sheltering can be estimated indirectly before any deployment if one knows that distance from near obstacles and their heights (see FW1982).
9. Appendix

### A.1 – HVAMS sites and instrumentation during the IOP

<table>
<thead>
<tr>
<th>Site name</th>
<th>Symbol</th>
<th>Station type / equipment</th>
<th>Elevation (m)</th>
<th>Longitude (deg)</th>
<th>Latitude (deg)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Alexander Farm</td>
<td>1</td>
<td>Integrated Surface Flux Facility or Portable Automated Mesonet (&quot;PAM&quot;)</td>
<td>156</td>
<td>-73.61623</td>
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<td>Black Horse Farms</td>
<td>2</td>
<td>-</td>
<td>47</td>
<td>-73.8508</td>
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<tr>
<td>Southlands Farm</td>
<td>3</td>
<td>-</td>
<td>45</td>
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<td>Green Acres</td>
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<td>-</td>
<td>94</td>
<td>-73.7522</td>
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<tr>
<td>Fix Bros</td>
<td>5</td>
<td>-</td>
<td>108</td>
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<td>42.180</td>
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<tr>
<td>Van Orden</td>
<td>6</td>
<td>-</td>
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<td>Zena Cornfield</td>
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<tr>
<td>Pertgen</td>
<td>9</td>
<td>-</td>
<td>76</td>
<td>-73.75</td>
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<tr>
<td>Greig – Anchor</td>
<td>G</td>
<td>20 m micrometeorological tower and the NCAR Tethered Atmospheric Observation System (TAOS)</td>
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<td>-73.8642</td>
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<tr>
<td>Albany ASOS</td>
<td>ALB</td>
<td>Standard surface weather station</td>
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<tr>
<td>Poughkeepsie ASOS</td>
<td>POU</td>
<td>The same</td>
<td>50.3</td>
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<td>41.6266</td>
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<td>SchProfile</td>
<td>L</td>
<td>Details on table</td>
<td>115.2</td>
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<td>42.8525</td>
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<td>Kingston Ulster</td>
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<td>Details below</td>
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<tr>
<td>Lake Taghkanic SP</td>
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Table A.1.1 - Names, symbols and coordinates of all HVAMS sites during the IOP.
### Observations, sensors, equipment and their abbreviations

<table>
<thead>
<tr>
<th>Sensor type</th>
<th>Abbreviation</th>
<th>Sensor type</th>
<th>Abbreviation</th>
<th>Sensor type</th>
<th>Abbreviation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Temperature (°C)</td>
<td>T</td>
<td>Upward short-wave Radiation (W/m²)</td>
<td>S&lt;sub&gt;up&lt;/sub&gt;</td>
<td>Propeller anemometer (m/s and deg)</td>
<td>PA-vane</td>
</tr>
<tr>
<td>Relative Humidity (%)</td>
<td>RH</td>
<td>Downward long-wave radiation (W/m²)</td>
<td>L&lt;sub&gt;dw&lt;/sub&gt;</td>
<td>Cup anemometer (m/s and deg)</td>
<td>CA-vane</td>
</tr>
<tr>
<td>Pressure (mbar)</td>
<td>P</td>
<td>Upward long-wave radiation (W/m²)</td>
<td>L&lt;sub&gt;up&lt;/sub&gt;</td>
<td>Sodar wind profiler (m/s and deg)</td>
<td>Sodar</td>
</tr>
<tr>
<td>Microbarograph pressure (μbar)</td>
<td>ubar</td>
<td>Net radiation (W/m²)</td>
<td>R&lt;sub&gt;net&lt;/sub&gt;</td>
<td>915 MHz wind profiler (m/s and deg)</td>
<td>915 MHz</td>
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<tr>
<td>CO₂ (ppm)</td>
<td>Same</td>
<td>Photosynthetically active radiation (μmol/m²/sec)</td>
<td>par</td>
<td>Soil temperature (°C)</td>
<td>ST</td>
</tr>
<tr>
<td>O₃ (ppb)</td>
<td>NO</td>
<td>3D-sonic anemometer (m/s and C)</td>
<td>3D-SA</td>
<td>Soil moisture (m³ of H₂O / m³ of soil)</td>
<td>SM</td>
</tr>
<tr>
<td>Downward short-wave radiation (W/m²)</td>
<td>S&lt;sub&gt;dw&lt;/sub&gt;</td>
<td>2D-sonic anemometer (m/s and C)</td>
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<td>Soil heat flux (W/m²)</td>
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<td>Hygrometer (KH₂O)</td>
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<td>Infrared surface temperature (°C)</td>
<td>IST</td>
<td>Rain Gauge (mm)</td>
<td>RG</td>
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Table A.1.2 - List of equipment and/or sensors, measurements type, and abbreviations.
## Sensors and equipment

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<thead>
<tr>
<th>Site</th>
<th>Equipments, sensors and installation height or depth</th>
</tr>
</thead>
<tbody>
<tr>
<td>PAM 1</td>
<td>3D-SA (7m), PA-vane (10m), KH2O (10m), T/RH (2m), $S_{dw}$/$S_{up}$/L$<em>{dw}$/L$</em>{up}$/R$_{net}$ (2m), ST (at 1 and 4 cm depth), SM (at 5 cm depth), SHF (at 5 cm depth), P, IST, and RG</td>
</tr>
<tr>
<td>PAM 2</td>
<td>3D-SA (7m), PA-vane (10m), KH2O (10m), T/RH (2m), $S_{dw}$/$S_{up}$/L$<em>{dw}$/L$</em>{up}$/R$_{net}$ (2m), ST (at 1 and 4 cm depth), SM (at 5 cm depth), SHF (at 5 cm depth), P, IST, RG, ubar, CO$_2$ (10m), and O$_3$.</td>
</tr>
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<td>PAM 3</td>
<td>3D-SA (7m), PA-vane (10m), KH2O (10m), T/RH (2m), $S_{dw}$/$S_{up}$/L$<em>{dw}$/L$</em>{up}$/R$_{net}$ (2m), ST (at 1 and 4 cm depth), SM (at 5 cm depth), SHF (at 5 cm depth), P, IST, RG, ubar, CO$_2$ (10m), and O$_3$.</td>
</tr>
<tr>
<td>PAM 4</td>
<td>3D-SA (7m), PA-vane (10m), KH2O (10m), T/RH (2m), $S_{dw}$/$S_{up}$/L$<em>{dw}$/L$</em>{up}$/R$_{net}$ (2m), ST (at 1 and 4 cm depth), SM (at 5 cm depth), SHF (at 5 cm depth), P, IST, RG, ubar.</td>
</tr>
<tr>
<td>PAM 5</td>
<td>3D-SA (7m), PA-vane (10m), KH2O (10m), T/RH (2m), $S_{dw}$/R$_{net}$ (2m), ST (at 1 and 4 cm depth), SM (at 5 cm depth), SHF (at 5 cm depth), P, IST, RG.</td>
</tr>
<tr>
<td>PAM 6</td>
<td>3D-SA (7m), PA-vane (10m), KH2O (10m), T/RH (2m), $S_{dw}$/R$_{net}$ (2m), ST (at 1 and 4 cm depth), SM (at 5 cm depth), SHF (at 5 cm depth), P, IST, RG, ubar, and CO$_2$.</td>
</tr>
<tr>
<td>PAM 7</td>
<td>3D-SA (7m), PA-vane (10m), KH2O (10m), T/RH (2m), $S_{dw}$/R$_{net}$ (2m), ST (at 1 and 4 cm depth), SM (at 5 cm depth), SHF (at 5 cm depth), P, IST, RG.</td>
</tr>
<tr>
<td>PAM 8</td>
<td>3D-SA (7m), PA-vane (10m), KH2O (10m), T/RH (2m), $S_{dw}$/R$_{net}$ (2m), ST (at 1 and 4 cm depth), SM (at 5 cm depth), SHF (at 5 cm depth), P, IST, RG, and ubar.</td>
</tr>
<tr>
<td>PAM 9</td>
<td>3D-SA (7m), PA-vane (10m), KH2O (10m), T/RH (2m), $S_{dw}$/R$_{net}$ (2m), ST (at 1 and 4 cm depth), SM (at 5 cm depth), SHF (at 5 cm depth), P, IST, RG, ubar, and CO$_2$.</td>
</tr>
<tr>
<td>G (20 m tower) or station 10$^1$</td>
<td>3D-SA (15.7m), 2D-SA (3 heights 12.13, 6.2, 2.9 m), KH2O (15.4m), CO2 (5 inlet heights 15, 12.13, 9.14, 6.2, 2.9 m), $S_{dw}$/S$<em>{up}$/L$</em>{dw}$/L$<em>{up}$/R$</em>{net}$ (16m), ST (at 2.5, 5 and 10 cm depth), SM (at 5 cm depth), SHF (at 5 cm depth), P, O3 (1 inlet height 3m), RG, and GPS.</td>
</tr>
<tr>
<td>HOBO 1</td>
<td>CU-vane (~3m), $S_{dw}$ (~3m)/par (~3m), RG, T/RH (~2m), ST/SM/SHF (at 5 cm depth), and P.</td>
</tr>
<tr>
<td>HOBO 2</td>
<td>The same</td>
</tr>
<tr>
<td>HOBO 3</td>
<td>The same</td>
</tr>
<tr>
<td>HOBO 4</td>
<td>The same</td>
</tr>
<tr>
<td>HOBO 5</td>
<td>The same</td>
</tr>
<tr>
<td>TAOS</td>
<td>PA-vane/T/RH/P (5 levels)</td>
</tr>
<tr>
<td>Schodack</td>
<td>Sodar</td>
</tr>
</tbody>
</table>

Table A.1.3 - Sensors and equipment operated during the IOP and LTOP
Table A.1.4 (below) indicates the operational periods of every equipment and/or site during the IOP, and as well after it, during the long term observational period (LTOP). In addition, this table also cites which sites were removed, relocated and installed after IOP.

*Instrumental details of some sites and/or observational unit*

- Greig – Anchor is micrometeorological 20 m tower equipped with an eddy correlation system (at the top of the tower capable of measure momentum, heat, water vapor, and CO$_2$), wind, temperature, humidity, and CO$_2$ profiles.
- Albany Asos is a standard surface weather station, which provides standard weather data at Albany international airport.
- Poughkeepsie ASOS also a standard surface weather station, however at Poughkeepsie airport.
- SchProfile is the NOAA 915 wind profiler at the Schenectady airport.
- Kingston is the University of Alabama-Huntsville Mobile Integrated Profiling System station equipped with a surface weather station, 915 wind profiler, sodar profile, ceilometer, and Microwave passive radiometer located at the Kingston airport.
- Schodack-IS this site is located at the Schodack Island State park and had two set ups: during the IOP there was only a wind profiler (Sodar) and after the end of IOP it was replaced by the standard weather station that was located at
- UWKA refer to the vertical profiles done by University of Wyoming Aircraft during the transition times of the PBL (dawn and dusk) at the five locations: Alexander Farm, Green Acres, South Albany, Columbia County, and Kingston Ulster. This plane was equipped with micrometeorological sensor (to measure momentum, heat, water vapor, and CO₂ turbulent fluxes), O₃ sensor, pressure sensor, temperature sensor, RH sensor, effective surface emissivity temperature, and other navigation sensors as well.
### Operational periods during IOP and LTOP

<table>
<thead>
<tr>
<th>Timeline</th>
<th>Intense Observational Period</th>
<th>Long Term Observational Period</th>
</tr>
</thead>
<tbody>
<tr>
<td>PAMs</td>
<td>15…20…25…30…5…10…15…20…25…30…31</td>
<td>Stopped running</td>
</tr>
<tr>
<td>TAOS</td>
<td>3 5…10 13.14 16.18 20 24.25 30.31</td>
<td>Stopped running</td>
</tr>
<tr>
<td>Schodack IS</td>
<td>15…31</td>
<td>Ran from 11/01 to 12/03/03</td>
</tr>
<tr>
<td>UWKA</td>
<td>3 6…11 13.14 16.18 24.25 28 30.31</td>
<td>Stopped running</td>
</tr>
<tr>
<td>G-Anchor</td>
<td>16…31</td>
<td>Still running</td>
</tr>
<tr>
<td>HOBO1</td>
<td>18…31</td>
<td>Ran from 11/01 to 11/24/03, but was relocated to Green Acres on 11/24/03, and still running</td>
</tr>
<tr>
<td>HOBO2</td>
<td>16…31</td>
<td>Still running</td>
</tr>
<tr>
<td>HOBO3</td>
<td>18…31</td>
<td>Ran from 11/01 to 11/20/03, but was relocated to Schodack-IS on 07/01/04, and stopped running in Sep/08</td>
</tr>
<tr>
<td>HOBO4</td>
<td>16…31</td>
<td>Continued running from 11/01 to 11/25/03, but was relocated to Van Orden on 11/25/03, and stopped running in Sep/08</td>
</tr>
<tr>
<td>HOBO5</td>
<td>15…31</td>
<td>Still running till Mar/08</td>
</tr>
<tr>
<td>MIPS Profiler</td>
<td>26…24</td>
<td>Stopped running</td>
</tr>
<tr>
<td>Sodar</td>
<td>25…31</td>
<td>Stopped running</td>
</tr>
<tr>
<td>Weather/st ceilometer</td>
<td>25…31</td>
<td>Stopped running</td>
</tr>
</tbody>
</table>

Table A.1.4 - Observational period of all HVAMS sites during the IOP and LTOP’s. Two numbers with points in indicates the period of operation. The number at the left side means the beginning day, and at the right side the ending day.
A.2 – *Landscape topographic description*

Quadratic surface were adjusted by least-square fit to the local topography of all the stations to define the local surface curvature ($V^2z(x,y)$), and central point (tower) slope ($\lvert \nabla z \rvert$)

Figures A.2.1 (a) and (b), (c) and (d), and (e) and (f) show, respectively, local topography around station 5 and the adjusted quadratic surface by least-squares method to the respective area for three different area size. The coefficient of determination ($r^2$) for (b) $r^2 = 0.41$, (d) $r^2 = 0.59$, and (f) $r^2 = 0.98$. 
The quality of the fitting ($R^2$) increased with decreasing area size for most of the stations as shown in figure A2.2. For the central Hudson Valley region the area was fixed (~ 41km x 66km) and $R^2$ was 0.44.

**Length scale and least-squares fitting coefficient**

![Graph showing $R^2$ versus length scale.](image)

Figure A.2.2: The coefficient of determination ($r^2$) versus length scale.

The station relative elevation compared to the immediately surrounding did not show unique asymptotic limits as shown in figure A2.3. It is also shown in Figure
A2.4 that there was no unique length scale for which the elevations variance converged.

Every station had a different value.

Figure A.2.3: Station relative elevation (mz-z) versus length scale of area size. The vertical redline indicates the chosen scale.
Figure A.2.4: Elevation variance versus length scale ((area size)$^{1/2}$).
A.3 Relations of topography parameters $mz - z$ and slope with turbulent fluxes, intermittency and surface temperatures

**Figure A.3.1**: Momentum flux as function of relative elevation ($mz - z$) evaluated for an area of 2km x 2km around each station. Number indicates the stations and colors differentiate background conditions, black – *average* (including all conditions), green – *windy and clear skies*, blue – *windy and overcast skies*, light blue – *calm and clear skies*, and pink – *calm and overcast*.
Figure A.3.2: Sensible heat flux as function of relative elevation (mz-z) evaluated for an area of 2km x 2km around each station. Number indicate the stations and colors different background conditions, black – average (including all conditions), green – windy and clear skies, blue -windy and overcast skies, light blue – calm and clear skies, and pink - calm and overcast.

10. References


Intermittency and the exchange of scalars in the nocturnal surface layer, Boundary-
Layer Meteorology, 119, 41-55.


Exploring the Possible Role of Small-Scale Terrain Drag on Stable Boundary Layers

Sun, J., Lenschow, D. H., Burns, S., Banta, R. M., Newsom, R. K., Coulter, R., Frasier,
S., Ince, T., Nappo, C., Balsley, B. B., Jensen, M., Mahrt, L., Miller, D., and
Skelley, B.: 2004, Atmospheric disturbances that generate intermittent turbulence in

Sun, J., Bunrs, S. P., Lenschow, D. H., Banta, R., Newsom, R Coulter, R., Frasier, S.,
turbulence associated with density currents passage in the stable boundary layer,


Soc.*, 125, 2401 – 2426.

scale flows and winds within a valley, *Journal of Applied Meteorology*, 32, 1669 –
1682.

