MULTI-SCALE FLOW AND TURBULENCE IN COMPLEX TERRAIN UNDER WEAK SYNOPTIC CONDITIONS

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Abstract

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Mountain weather has a remarkable range of phenomena, which are signified by two flow types: modifications to the background (~100 km, synoptic-scale) flow driven by regional pressure gradients and thermal circulations (valley/slope flows, mesoscale) generated by local heating/cooling of the ground surface. Current state-of-the-art mesoscale models typically employ a grid resolution of ~1 km, with the assumption that over such scales there is sufficient spatial homogeneity to represent heat, mass and momentum transport within a grid cell. Nevertheless, because of topographic heterogeneities, differential heating/cooling and flow interactions, the flow within a mesoscale model grid element can be highly inhomogeneous, and this exemplifies the difficulty of mathematically representing (parameterizing) sub-grid processes of a slope-scale grid.

The aim of the MOUNTAIN TERRAIN ATMOSPHERIC MODELING AND OBSERVATIONS (MATERHORN) PROGRAM was to conduct fundamental research to
improve weather predictions for mountainous terrain, and improvements in understanding of sub-grid scale processes and their parameterizations were a significant part of it. To this end, field and laboratory studies were conducted and theoretical formulations were developed, which are described in this thesis. A suite of flow diagnostic techniques were used. At the core of the work are the subtopics of upslope flow separation in mountainous terrain, multi-scale interactions of slope and valley flows and measurement of turbulence in katabatic flows.

Owing to the vast range of scales involved, new cutting edge techniques were developed and deployed for process identification. These include tower mounted three-dimensional hot-film combo probes, consisting of sonic anemometers co-located with hot-film anemometers. The combo probes follow mean winds using a feedback control loop and use a Neural Network to calibrate the hot-films in-situ. Once calibrated, these probes can measure from mesoscale flow down to the Kolmogorov scale. Also deployed were three scanning Doppler LiDARs in coordination to visualize the velocity structure and to obtain three-dimensional velocity virtual towers up to 300 m AGL. Turbulence in slope and valley flows, upslope flow separation, flow collisions and interactions between different types of flow are discussed in this thesis, with particular emphasis on quantitative results of consequence for numerical modeling.
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CHAPTER 1:
INTRODUCTION

1.1 Overview

Since the epochal review of Queney (1948) on flow over low and steep hills, mountain meteorology has seen profound progress (Scorer 1997; Taylor et al. 1987; Blumen 1990; Whiteman 2000; Fernando 2010; Zardi & Whiteman 2010). Notwithstanding, the gaps in knowledge as well as computational and measurement technologies continue to challenge predictions of weather in mountainous terrain, especially in areas of steep hills, high vegetation and rapid land use changes. In these areas, the spatial and temporal gradients of meteorological variables are steep, and flow and thermodynamics processes are intricate. Furthermore, most of the complex terrain research has been conducted in regions of low hills located in slowly varying terrain since human settlements are prevalent in such areas due to abundance in riparian and karstic water resources and attractive navigation corridors. Only a few urban centers have emerged in very complex terrain with high, steep hills because of poor accessibility and severe environmental conditions (e.g., Juneau, Alaska).

About 70% of the Earth surface, and more so the urban areas, is covered by complex terrain, and thus basin-scale processes characteristic of urban airsheds have
been studied extensively in the context of air quality and aviation (Blumen 1990; Whiteman 2000; Fernando 2010). Yet, as mentioned, the weather in the thick of mountainous terrain has received less attention. The more recent military engagements and build up of major airport hubs (e.g. Hong Kong) in mountainous areas, however, have called for a renewed interest on weather in complex topography.

War fighting in rugged terrain is historically referred to as ‘mountain (or Alpine) warfare’, which is one of the most dangerous types of combat (Winters 2001). A principle element of mountain warfare is meteorological support, given the extreme weather (e.g., lightning, wind gusts, Venturi wind effects between ridges, stagnation, cold air pooling, travelling and stationary waves, up/down drafts, snow/ice, convective clouds and lightning), steep terrain, incomplete environmental information, intricate logistics, greater uncertainties and high risks that the combatants have to endure. The prediction of flow, turbulence and dispersion of contaminants and obscurants in complex terrain is extremely difficult because of the wide range of physico-chemical processes involved covering a broad spectrum of space-time scales. Moreover, many of these processes are of small-scale, less than a few km and on the order of an hour or less; to name a few, eddy shedding, aerosols and secondary aerosol formation, mountain waves and phase changes. Thus, in the context of mesoscale modeling of mountainous terrain, the applicability of conventional sub-grid parameterizations is in question whilst the role of sub-grid processes is more crucial than elsewhere. These complexities, confounded by the lack of high resolution meteorological data, have been the bane of predicting mountain weather.
The most important civilian applications of complex terrain weather are the orographic convection developing over mountains that lead to thunderstorms, turbulence and wind shear near airports, air quality deterioration due to low wind speeds in complex-terrain urban areas, prediction of road conditions against the threats of ice formation in the winter and dust storms in the summer, prediction of visibility on foggy and poor air quality days and forecasting for freeze protection of agriculture. Military applications also abound. In a specialized workshop held in Tempe, Arizona during February 1-2, 2010, a cadre of defense practitioners identified illumination, visibility, wind speeds and moisture as the most important current issues for Alpine Warfare, while deemphasizing wind shear, icing and lightning. Illumination is reduced by shadowing effects, clouds and walls of dust that appear during high wind events. For example, the 2009 drought in Afghanistan created islands of low visibility under high winds, and enhanced erosion of soil has impacted on low level helicopter operations; there is a drastic reduction of visibility when the helicopters descend to altitudes of 20-30 feet above ground level. Moisture deleteriously interferes with unmanned aerial vehicle (UAV) hardware, making them viable only when the relative humidity is < 30%.

Mountain weather has a remarkable range of phenomena as depicted in Figure 1.1, which schematizes an area located in a valley surrounded by sloping terrain. In this scenario, two flow types dominate the local mesoscale circulation: modifications to the background (synoptic-scale, ~100 km) flow driven by regional pressure gradients; and thermal circulations due to differential heating/cooling of the ground surface (Blumen 1990). Thermal circulations comprise of downslope (katabatic) and downvalley winds.
during the night and upslope (anabatic) and upvalley winds during the day. The valley winds are driven by either horizontal pressure gradients due to temperature variation along the valley or between a valley and nearby plane or by along-valley buoyancy forces. The slope flows originate due to the temperature differential between the air layer adjoining the slope and the free atmosphere. Differential surface heating, if present, produces secondary flows, for example, sea, lake, or land breezes driven by the temperature differential between land and water bodies (Whiteman 2000).

Figure 1.1: Possible flow processes in complex terrain. The left panel shows nocturnal conditions, while daytime conditions are on the right. Fernando 2010.

At night, at the ridge level, stable stratified synoptically influenced winds (i.e., those driven by regional scale pressure gradients) may form lee waves, flow separation, rotors, stagnation, and recirculation (Grubišić et al. 2008). Meanwhile along the slope, the ground continues to cool rapidly, forming a dense layer of air above the surface and
downslope (katabatic) flow strengthens. As the katabatic flow reaches the lower levels near the valleys it merges with the valley nocturnal atmospheric boundary layer (ABL), possibly undergoing hydraulic adjustment or colliding with the valley winds. In terrain-confined valleys, along-valley flows are weak, and stably stratified air accumulates.

In the morning, developing thermal convection breaks up the cold pools formed overnight and the valley/slope flows reverse forming upvalley/slope flows in a process called the morning transition. If a moderate synoptic flow is present a convergence zone between the synoptic flow and the upslope (anabatic) flow may form in the lee of the topography (Banta 1986). If instead the synoptic flow is weak, the anabatic flow may separate from the slope forming plumes. In either case the rising warm humid air may trigger cumuli and hence cumulus congestus and cumulonimbus clouds.

Solar radiation subsides in the evening, weakening the upslope/valley flows, which eventually reverse to form downslope/valley flows (i.e., evening transition). Under low synoptic conditions, the flow becomes nearly stagnant during morning and evening transition periods. These periods are problematic in numerical modeling, given the rapid flow evolution, nonequilibrium turbulence (Pardyjak et al. 2009) and spatial variability of processes. The relationships among synoptic, mesoscale, and small-scale processes are not well understood and quantified, but are crucial in the modeling of weather prediction in mountainous terrain.

The mesoscale meteorology for mountain weather prediction is supplied by models such as NCAR’s Weather Research and Forecasting (WRF) model, the US NAVY’s Coupled Ocean/Atmosphere Mesoscale Prediction System (COAMPS), the Regional...
Atmospheric Modeling System (RAMS) and HadRM (Hadley Center Regional Model), and their typical resolution is \(~1\) km, with the assumption that over such scales there is sufficient spatial homogeneity to represent conditions within a grid cell. Nevertheless, because of heterogeneous topography and differential heating/cooling, flow can be highly inhomogeneous within mesoscale grid cells. This is illustrated in Figure 1.2, in which Granite Mountain, Utah (which will appear throughout this thesis, given its central role in field experiments) is overlain by a 1-km grid. Clearly, strong topographic heterogeneities exist within mountain grid cells, which exemplify the difficulty of mathematically representing (parameterizing) sub-grid processes of a slope-scale grid and point to the necessity of finer scale simulations in dealing with weather prediction in mountainous areas.

Figure 1.2: (A) Topographic map of Granite Mountain, Utah with 10 m contours and (B) the associated 1-km grid used for mesoscale models. Note the inability to resolve the topographic features. (Courtesy of Zachariah Silver, University of Notre Dame).
1.2 Thermal Circulation

As mentioned, katabatic (downslope) and anabatic (upslope) flows that become well defined in complex terrain in the absence significant synoptic influence are called the thermal circulation. In terms of atmospheric flows, thermal circulation is at or near the small-scale end of a spectrum (Martilli et al. 2002) while its contribution is dominant in situations of weak synoptic forcing (i.e., high-pressure conditions). Being mesoscale, thermally driven flows are unresolved, sub-grid features in weather (resolution $\sim$40–100 km) and climate ($\sim$100–500 km) prediction models that are highly sensitive to topographic heterogeneities and differential near surface heating. Thermal circulation has been discussed extensively (e.g., Atkinson 1981; Blumen 1990; Whiteman 1990, 2000; Fernando 2010; Zardi & Whiteman 2010), but most of the reported research is on downslope flows because of their air-quality implications and the misconception that convection in upslope flow is fairly well understood (Doran et al. 2002).

In the earliest work on slope flows, (Prandtl 1942) obtained a 2D solution for the along-slope velocity of a uniformly heated or cooled slopes (Figure 1.3) with fixed surface temperature, which can be easily modified for the case of fixed surface heat flux $Q_s$ (equivalent to a buoyancy flux, $q_0 = \alpha q Q_s/\rho_0 c_p$) as,

$$u(n) = 4^{1/4} q_0 \left[ \kappa N^2 \sin^2 \beta_s \right]^{1/4} e^{-n l} \sin(n/l),$$

where $l = \left( 4 \kappa N^2 \sin^2 \beta_s \right)^{1/4}$ is a length scale in the slope-normal ($n$) direction, $\nu$ the kinematic viscosity, $\beta_s$ the slope angle, $N$ the background Brunt-Väisälä buoyancy
frequency, $\kappa$ the thermal diffusivity, $g$ the gravitational acceleration, $\rho_0$ a reference density, $\alpha$ the thermal expansivity and $c_p$ the specific heat.

Although equation 1.1 is applicable to general slope flows, whether up or down, the eddy diffusivities and flow structure for the two cases are very different. Katabatic flows are stably stratified, devoid of significant turbulence levels or an effective turbulent heat/mass transport mechanism. Mixing between katabatic flow and the ambient atmosphere is weak due to the presence of a significant stable buoyancy jump at the interface. The entrainment of ambient fluid into the katabatic flow is therefore slow, resulting in a thin layer along the slope (~20–50 m). This contrasts with the case of upslope flow, in which turbulence is intense, vertical turbulent transport is strong and entrainment is rapid, forming a thick anabatic flow layer (~500 m to 2 km). The upslope flows also have a tendency to separate from the mountain, especially at larger slope angles and heat fluxes.

1.3 Downslope Flows

To minimize complexity, idealized downslope flow studies assume 2D flow (Figure 1.3), neglect the complications of slope variation, lateral inhomogeneities, ambient shear, edge effects, and flow separation. To describe downslope flows, Manins & Sawford (1979) used 2D layer-averaged momentum and buoyancy equations,

$$\frac{\partial U_h}{\partial t} + \frac{\partial U_h^2}{\partial s} = -\frac{\partial}{\partial s} \left( \frac{1}{s} S_1 \Delta h^2 \cos \beta_s \right) + S_2 \Delta h \sin \beta_s - C_D U^2 - \left( \overline{u'w'} \right) H, \quad (1.2)$$

and
\[
\frac{\partial}{\partial t} (S_2 \Delta h) + U h N^2 (\sin \beta_s - S_3 E \cos \beta_s) + \frac{\partial}{\partial s} (U \Delta h) = B_0 - (\overline{g' w'}) H ,
\]

(1.3)

where the characteristic averaged velocity \(U\), local buoyancy deficit \(\Delta\), and characteristic velocity thickness \(h\) are defined in terms of the shape parameters \(S_1, S_2, \ldots\) as follows:

\[
U h = \int_0^H u d n , \quad U^2 h = \int_0^H u^2 d n , \quad U \Delta h = \int_0^H u g' d n ,
\]

\[
S_1 \Delta h^2 = 2\int_0^H g' n d n , \quad S_2 \Delta h = \int_0^H g' d n \quad \text{and} \quad \int_0^H w d n = U h - S_3 w_H h ,
\]

(1.4)

where \(u\) is the downslope velocity that vanishes at \(n = H\), \(\overline{g' w'}\) is the buoyancy flux, \(-\overline{u' w'}\) is the Reynolds stress (subscripts denote the location), \(B_0\) is the surface buoyancy flux including the accumulation of radiation energy near the surface (radiation divergence), \(-\overline{(u' w')} = C_D U^2\) and \(C_D\) is a drag coefficient. By vertically integrating the continuity equation, the entrainment velocity at the sharp upper interface is \(w_H = -E U = -\partial U h / \partial s\), where \(E\) is the entrainment coefficient (Fernando 1991). If the surface buoyancy flux, \(B_0\) is neglected equations 1.2–1.4 are similar to those of gravity currents on inclined surfaces, and hence the results of katabatic flow and gravity current studies are interchangeable. Quantitative comparisons between the two flow types however should be done with caution due to the dynamical implications (e.g., suppression of turbulence) of the strong stable stratification within katabatic flows induced by \(B_0\).

For strongly stratified cases, retaining unsteady effects, equations 1.2 and 1.3 accept oscillatory solutions with frequency \(\omega_c = N \sin \beta_s\) (Princevac et al. 2008), which are the critical internal waves that degenerate into turbulence upon reflection from a
sloping surface (De Silva et al. 1997). However, other forms of internal waves also can be present in the katabatic flow, in particular, long internal wave modes in the thin stable boundary layer.

As evident from the laboratory study of Baines (2001) on 2D gravity currents, a wide variety of entrainment mechanisms are possible at the edges of katabatic flows (e.g., at the density interface), including direct encroachment of fluid aloft and mixing due to shear (Kelvin-Helmholtz (K-H) and Hölmböe) instabilities. The K-H instabilities take the form of billowing motions, whereas Hölmböe instabilities result from the interaction between a gravity wave on the density interface and a vorticity wave on the maximum vorticity gradient above it, when the two are displaced from each other; they appear as cusps travelling on the interface, while ejecting wisps of fluid into the adjacent flow (Strang & Fernando 2001). The process of fluid leaving the main current and entering the environment is termed detrainment, which is possible in the case of katabatic flows, especially when the ambient fluid is highly stably stratified. Baines (2001) has classified entrainment, detrainment, or a mix thereof based on the local values of a Richardson number $R_i$, the Reynolds number $Re$ and the dimensionless parameter $M = QN^3/g^2$ and $\beta_s$, where $Q$ is the volume flux in the flow, a parameter that is an extension of that used by Ellison & Turner (1959) who did not consider the ambient stratification in their analysis of gravity currents on slopes.

Where there are significant slope variations, katabatic flows undergo strong adjustments such as the generation of hydraulic jumps. Assuming shape factors of unity
in Equation 1.4, negligible entrainment, and \( N = 0 \), the equation for hydraulic adjustment becomes (e.g., Turner 1973),

\[
\frac{\partial h}{\partial s} \left( 1 - Fr_i^{-2} \right) = \left( \beta_s - C_D Fr_i^2 \right),
\]

(1.5)

where \( Fr_i = \sqrt{U^2 / \Delta h} \) is an internal Froude number. At the critical depth, \( Fr_i = 1 \), yielding a critical slope angle of \( \beta_s^{cr} = C_D \). Taking \( C_D \approx 10^{-3} \) (Princevac et al. 2005), \( \beta_s^{cr} \approx 0.05^\circ \). A transition from supercritical to subcritical flow or vice versa occurs through this critical slope. The differentiation of hydraulic jumps from other complex-terrain spatially varying flow phenomena (Figure 1.1) is difficult and may require special methodologies/diagnostics (Drobinski et al. 2001).

Figure 1.3: (A) An idealized 2D slope flow used in theoretical and numerical studies; from Fernando 2010. (B) A Doppler LiDAR image of a katabatic flow over an escarpment with a gridded surface showing the topography in Wager-up, Western Australia. The color represents the radial velocity along the LiDAR beam. (Courtesy of Dr. Charles Retallack)

In this thesis, the downslope flows are studied in the context of interpreting data taken along a special track on the slope where extensive measurements have been
taken via a series of towers, LiDARs and a fiber optic distributed temperature measurement system. Measurements of small scale turbulence have also been made at selected locations of the track, which will be analyzed to obtain information of stable stratified turbulence.

1.4 Upslope Flows

Above a critical slope angle (which can be small $1^\circ$) a turbulent anabatic flow is generated upon heating an incline (Princevac & Fernando 2007). Similar to the katabatic flow case, layer-averaged equations analogous to equations 1.2 and 1.3 can be used to provide a framework for anabatic flows (Schumann 1989). In steady state, neglecting along-slope variation, these equations call for an integral balance between the mean along-slope buoyancy forces and surface friction. As in katabatic flows, including the unsteady effects in the analysis produces oscillations of mean velocity and temperature fields with a frequency of $N \sin \beta_s$ (Fernando et al. 2013). In the case of constant diffusivities, the velocity is given by equation 1.1. Alternative velocity scales are possible depending on the assumed form of eddy diffusivity. If both heat and momentum diffusivities are $K \sim U h$, where $h$ is the upslope layer depth and $U$ is the velocity, then $U \sim (q_0/N)^{1/2}$ and, when $K = 0.05 w_* h$, where $w_* = (q_0 h)^{1/3}$ is the convection velocity (Wyngaard 1984), then $U \sim (q_0/N)^{1/2} (\sin \alpha)^{1/2}$. If instead the temperature gradients within and outside the anabatic flow are assumed identical with $U \sim u_*$, where $u_*$ is the friction velocity, then $U \sim (q_0/N)^{1/2} (\sin \beta_s)^{-1/4}$ based on large eddy simulations (Schumann 1989). It is also possible to arrive at the scaling of $U \sim$
by balancing the upslope buoyancy with the bottom frictional effects (Kondo et al. 1989).

A more comprehensive analysis by (Hunt et al. 2003) considered the detailed internal structure of anabatic flow along a long, gentle slope of angle \( \alpha \) with a stably stratified ambient atmosphere and constant heat flux in the presence of negligible synoptic winds. Flow adjustments were argued to occur via long internal gravity waves, followed by the establishment of a quasi-stationary flow consisting of four dynamical zones identified in Figure 1.4: a surface layer \([S]\) of thickness \( h_s \), where buoyancy and momentum fluxes are important; a middle layer \([M]\) in \( h_s < n < h - h_i \) dominated by shear stresses and inertia; and an inversion layer \([I]\) of thickness \( h_i \) that separates \([M]\) from the background \([E]\) and where entrainment as well as inertial, buoyancy, and shear stress terms are important. In \([S]\), below the Monin-Obukov (MO) length scale, 
\[
L_* = \left( \frac{u_*}{w_*} \right)^3 h, \quad \text{where } u_*^2 = \tau_s \quad \text{and } w_* = (g_0 h)^{1/3}, \quad \text{the shear is dominant \([S_s]\)}, \\
\text{and above } L_* \text{ convection dominates \([S_c]\)}. \]
Using simplified forms of governing equations for each zone, by parameterizing momentum and heat fluxes in \([S]\) using the level-terrain MO flux laws, assuming a self-similar velocity profile for \([M]\), and by parameterizing entrainment in \([I]\), Hunt et al. (2003) derived an expression for the velocity for small \( \alpha \) as 
\[
U_m = \Gamma \alpha^{1/3} w_* \quad \text{where } \Gamma \sim 10.
\]
Figure 1.4: (A) A schematic of dynamical zones in upslope flows proposed by Hunt et al. (2003). (B) A Doppler LiDAR scan of upslope flow on Granite Mountain, Utah Dugway Proving Grounds during the MATERHORN-X-1 field campaign. The color represents the radial velocity along the LiDAR beam with darker blue representing upslope flow. (Courtesy of Dr. Yansen Wang, U.S. Army Research Laboratory)

In this thesis, upslope flow is studied in different contexts, first in the development of a model for flow separation (to be discussed next) and then for interpreting data taken during daytime period.
1.5 Upslope Flow Separation

Flow separation from mountain slopes has been a topic of considerable interest because of its relevance to severe weather and precipitation in the vicinity of mountains. In particular, heat carried by the separated flow and associated small-scale features are crucial for predicting deep convection, precipitation, moisture distribution and unsteady phenomena surrounding orographic features (Banta 1984; Hanley et al. 2011). Excessive precipitation over steep and high mountains is a well-known problem in mesoscale and climate models, which has been attributed to the parameterization difficulties of separated flows in mountain slopes, especially the sub-grid heat ventilation by separated upslope flows (Chao 2012). Without ventilation the heat flux and slope flows are overestimated, which together with excessive moisture transport lead to erroneous precipitation predictions (Figure 1.5A, B). In urban context, the upslope flow during daytime transports urban pollution plume to mountainous rural surroundings (Ellis et al. 2000; Fernando et al. 2001; Lee and Fernando 2013). The fate of the pollutants is largely dependent on whether the upslope flow separates or not. When separation occurs, pollutants are lofted to regional flow or recirculated at higher levels (Lu & Turco 1994) whereas in attached flows, pollutants return back to the city at night as downslope flow (Fernando 2010). In aircraft operations, flow separation in the lee or windward side of the mountains is a key issues of safety (Politovich et al. 2011). In predictive models, the separation is sensitive to the details of turbulent models employed (Mason 1987; Moreira et al. 2012), and hence an understanding of mechanisms involved is vital for developing sub-grid parameterizations.
Figure 1.5: (A) A schematic of upslope flow separation and possible cloud formation. (B) Upslope flow separation and cloud formation at Granite Mountain, Utah Dugway Proving Grounds during the MATERHORN-X-2 field campaign. (B courtesy of L. Leo)

Most of the work on flow separation has been on cases with background mean flow approaching mountains, driven by large-scale pressure gradients, both for neutral (Belcher & Hunt 1998) and stratified cases (Baines & Manins 1989; Baines 1995). The pioneering 2D linear analysis of Jackson & Hunt (1975) considered low hills, and its validity is restricted to small slopes, say < 10° (Hewer 1998). The analysis has been extended to 3D cases (Mason & King 1985), notably to practically relevant situations (Hunt et al. 1988; Carruthers et al. 2012), and the results show that flow features of 2D and 3D cases may differ substantially (Doyle and Durran 2007). For steeper slopes, when the surface shear stress is negligible, flow separation and a recirculation bubble may appear in the lee slope, wherein flow is non-linear. Weakly (Xu et al. 1994) and full (Apsley & Castro 1977; Crook & Tucker 2005) non-linear models have been developed for idealized topographies, which have been used for flow over hills and mountains (600
m relief is considered as the demarcation between hills and mountains, but this
description is not exhaustive; Barry 1992).

A model for flow separation has been developed by Wood (1995), and the
lowest critical slope for separation has been obtained. A number of numerical, field
experimental and wind tunnel (Britter et al. 1981; Gong & Ibbetson 1989) studies on
flow separation in neutral flows have been reported (Mason & King 1985; Hewer 1998;
Hunt et al. 1991), and the absence of thermal forcing in such models limits their
applicability to environmental flows.

The addition of thermal forcing changes the character of flow, depending on the
completion between ensuing thermal circulation and external flow. A few theoretical
studies have been reported in this regard, for both two dimensional 2D (Raymond 1972;
Smith & Lin 1982; Tian & Parker 2003) and three dimensional 3D (Reisner &
Smolarkiewicz 1994) cases. Numerical studies with periodic 2D hills show that for low
hills the convective flow structures developed without external flow is similar to that on
flat terrain, but the presence of even a weak mean flow destroys this circulation.
Convective rolls developed in the latter case have both transverse and longitudinal rolls,
compared to longitudinal rolls without the mean flow (Dornback & Schuman 1993).
Using scaling arguments, Mason (1987) derived a form of Froude number that helps
delineate convectively versus external flow dominated cases, which has broad
consistency with the expression later proposed by Lewis et al. (2008).

According to the 2D simulations of Dornback & Schuman (1993), an external flow
impacts convective structures for $U/w_* > 2$, whereupon the influence of topographic
details wanes in determining turbulent statistics of overlying flow. They showed that, for steeper terrain, heating facilitates leeside flow separation and a critical heating rate is required for a given flow velocity and a valley width to cause separation. Heating reduces the affinity for lee-side flow separation over low hills of sufficient steepness, and the characteristic size and velocity of separation bubble are reduced when heating is present (Allen & Brown 2006). This is due to enhanced vertical momentum diffusion in the windward-side due to convection, thus reducing the vorticity of the leeward separation bubble. WRF simulations of Lehner & Whiteman (2012) confirm such separation features for idealized axi-symmetric topographies.

Interestingly, Crook & Tucker (2005) showed that flow past a heated topography with horizontally homogeneous heating can be approximated by flow over an obstacle with heating isolated to only the obstacle. They developed linear and non-linear models, and in the former flow past a heated low hill could be analyzed by superposing separate responses of thermal and orographic forcing.

Convection over hills has been studied in numerical experiments (Krettenauer & Shumann 1989) but no laboratory experiments exist to validate these models, especially the flow separation. On the other hand, a number of numerical and field studies exist on neutral and stratified flow past hills or complex terrain (see Taylor et al. 1987; Blumen 1990; Raupach & Finnigan 1997; Whiteman 2000; Price et al. 2011).

In this thesis, the flow separation in the absence of synoptic effects is revisited using a laboratory experiment. LiDAR observations of flow separation are also presented.
1.6 Transitional Flows

1.6.1 Morning Transition

The morning transition in complex terrain is a very energetic period involving both the breakup of the stably stratified boundary layer (cold pools) and the reversal of slope/valley flows. During this period solar heating warms the air of the cold moist cold pool, generating a CBL, rising to form upslope flow which eventually overtakes the decaying remnants of downslope flow. Depending on the solar inclination, the destruction of downslope flow can be initiated over the valley or slopes. For basin topography, several cold pool destruction mechanisms have been identified, including the three scenarios proposed by Whiteman (1982). The first, but rarely observed, is as in flat terrain where CBL growth is initially rapid with eddies penetrating the stable buoyancy jump generated at the top of the growing CBL in a process referred to as penetrative convection followed by erosion by turbulent eddies, first by shear and then by both shear and convection. In the second, the upslope flow carries heated fluid along the basin walls, causing the stable core to descend into the valley, accompanied by subsidence (compressional) heating. As stable air descends, it is entrained into the CBL. The third, and perhaps the most common, is a combination of the first two mechanisms. More recently a fourth mechanism has been identified by Princevac & Fernando (2008) in which dominant intrusions generated at approximately the mid-height of the valley slowly descend into the stable core and merge with the CBL, expediting the breakup.
Figure 1.6: (A) The cold pool of air trapped in a complex topography (basin) with an initial inversion height $h_0$, and inversion strength (temperature gradient) $\gamma$. (B) The development of the convective layer and upslope flow. (C-D) The development of the convective layer of depth $H(t)$ and the descent of the stable core and top of the inversion $h(t)$. (E) Complete destruction of the stable core. The (potential) temperature profile through the middle of the basin is shown on the left (based on Whiteman 1982). From - Princevac and Fernando (2008).
1.6.2 Evening Transition

During the evening transition, radiative cooling occurs rapidly at the ground, forming a surface inversion followed by the dissolution of a CBL. Hunt et al. (2003) modeled evening transition by assuming that cooling of anabatic flows occurs according to a linear decrease of temperature with time (Figure 1.7). The simplified equations of motion for small slopes accepted a frontal solution for mean velocity, indicating that the transition occurs via a stagnation front that forms at a distance $\hat{x} = x_f$ from the bottom of the slope. As an air parcel moves along the slope, it is gradually cooled, and at some point the parcel becomes sufficiently dense to overcome the inertia forces of upslope flow. Following the front formation, for $\hat{x} < x_f$, the flow continues upslope, whereas at $\hat{x} > x_f$ a downslope flow is initiated near the surface while the flow aloft is still upslope (Figure 1.7). The front propagates downslope, trailed by the katabatic current, and the entire downslope flow undercuts the prevailing (but dwindling) upslope flows, and this completes the transition. Because there is only moderate interaction between the two, this counterflow situation may last for hours (Brazel et al. 2005), with katabatic current near the ground and a weak upslope flow at greater heights (Fernando et al. 2013).
Figure 1.7: A schematic diagram of the cooling of an upslope flow as proposed by Hunt et al. (2003). As the fluid cools, becoming sufficiently dense to overcome the inertial forces, a stagnation front forms which slowly starts advancing downslope. From Hunt et al. (2003).

1.7 Interactions Between Slope and Valley Flows

Pressure gradients due to temperature variation along the valley or along-valley buoyancy forces drive valley flows, which are most likely to be encountered by intermittent, pulsating flow through canyons and gaps within the topography. Flows originating from thermal gradients due to land use variation, shadow propagation from surrounding terrain, differential solar insolation on slopes of different orientation can interact with each other, and may continuously modify the structure and dynamics of the thermal circulations (Kuwagata & Kimura 1997; Ruffieux et al. 1995; Fernando et al. 2001; Matzinger et al. 2003; Lee et al. 2003; Hoch & Whiteman 2010). As slope flows descend, interaction occurs between flows originating at steeper (higher elevation) and gentler (foothills) slopes, leading to hydraulic adjustments and enhanced mixing. Farther below near the valley floor, interaction of slope and valley flows and at times collisions between them produce a host of small scale phenomena, to name a few, collision
fronts, intense turbulent patches, regions, intrusions and instabilities which contribute vigorously to sub-grid heat and momentum transfer. Conversely, depending on the background conditions or when the valley winds get strong enough, horizontal and vertical shears can greatly modify the slope flows or completely obliterate them (Doran et al. 1990). Such processes are not properly accounted for in currently used mesoscale models, and eduction of such processes via observations and their analysis is part of the work reported in this theses.

Interactions between various types of thermally driven as well as thermally and synoptically driven flows lead to highly variable weather, and physical processes underlying these interactions are not well understood. Until the recent MATERHORN field campaigns that will be described in Section 2, no detailed observations of these flow interactions existed (Fernando and Pardyjak 2013). While in principle such interacting flows may be best studied using numerical simulations, mesoscale models do not incorporate sub-grid flux contributions of these processes, and hence tend to be flawed in predicting the mean flows, let alone the interactions between desperate mean flows. As such, some of the detailed field observations to be discussed in this thesis will help understand and parameterize collisions and interactions between different types of flow elements present in mountain terrain winds.

1.8 Turbulence in Complex Terrain

Turbulence is complex terrain arises due to a variety of mechanisms, to name a few, flow separation, shear layer formation, convection, flow intrusions and collisions...
between gravity currents. Only the first three mechanisms have received prior attention, and in this thesis we identify the rest of the mechanisms and study them in detail (Section 4). Given the emphasis of this thesis on low synoptic flow conditions, the first mechanism is not discussed in the brief introduction below, but aspects of convection, unstably-stratified upslope flow and stably stratified katabatic flow are reviewed.

Previous studies have delved into convection in complex topography, and a recent review in this regard is given by Dallman et al. (2013). Turbulence parameterizations for upslope flows where surface shear and unstable buoyancy flux is present are listed in Table 1.1. Here \( w_* \) is the convective velocity and \( u_* \) is the friction velocity.

### TABLE 1.1

TURBULENCE PARAMETERIZATIONS FOR THE MIXED CASE (DALLMAN ET AL. 2013)

<table>
<thead>
<tr>
<th>Authors</th>
<th>Parameterization</th>
<th>Conditions</th>
</tr>
</thead>
<tbody>
<tr>
<td>André et al. (1987)</td>
<td>( \sigma_A = (1.75u_<em>^2 + w_</em>^2)^{1/2} )</td>
<td>Convective</td>
</tr>
<tr>
<td>Deardorff (1983)</td>
<td>( \sigma_D = (w_<em>^3 + \eta^3 u_</em>^3)^{1/3}, \eta = 1.8 )</td>
<td>Convective</td>
</tr>
<tr>
<td>Clarke et al. (1971)</td>
<td>( \sigma_C = C_u(w_<em>^2 + u_</em>^3 \lambda_{max}/(\kappa z))^{1/3}, C_u \approx 0.4 - 0.6 )</td>
<td>Convective, ( \lambda_{max} ) is the energy containing wave length;</td>
</tr>
</tbody>
</table>
Dallman et al. (2013) conducted a detailed study on the validity of these parameterizations, and found that all of them work reasonable well within the measurement errors. Perhaps of most interest to the atmospheric boundary layer community is turbulence during the evolution of the stable boundary layer, which has been a topic of many of the previous studies; however it is still a phenomenon that remains elusive because of its complexity, not to mention the measurement difficulties. Katabatic flows are highly stratified with considerable shear, and hence offer an excellent platform to study stratified turbulence in fundamental sense. Any progress in understanding and modeling of stratified turbulence will be of utmost value, as one of the drawbacks of mesoscale models is inadequate representation of processes occurring in sub-grid scales. Small scale mixing processes, conditions of their occurrences and the amount of fluxes associated with such processes (expressed as equivalent eddy diffusivity) are quantities of importance for improving nocturnal boundary layer predictions in general, let alone predictions of complex terrain boundary layers.

One of the most important parameters that determines the stability of stratified shear flows, and hence the turbulence therein, is the gradient Richardson number, $Ri_g$, defined as

$$ Ri_g = \frac{N^2}{\left| \frac{\partial \tilde{V}}{\partial z} \right|^2} = \frac{N^2}{\left( \frac{\partial U}{\partial z} \right)^2 + \left( \frac{\partial V}{\partial z} \right)^2}, \quad (1.6) $$

where the mean velocity vector is expressed as $\tilde{V} = (U, V, W)$, $U$, $V$ being the horizontal ($x$, $y$) velocities and $W$ the (small) vertical ($z$) velocity. According to the
linear stability theory of Miles (1961) and Howard (1961), the flow can become unstable when $Ri_g = N^2/(dU/dz)^2$ drops below a critical values of $Ri_{gc} = 0.25$. This critical value can be much different from 0.25 for non-linear disturbances (Miles 1986).

According to the experiments of Strang and Fernando (2001A,B) conducted with a parallel stratified shear flow having hyperbolic tangent velocity and density profiles, strong mixing events can persist up to $Ri_g \approx 1$. For $Ri_g < 1$, the major mixing mechanism was the K-H instability. Non-linear waves tend to generate and resonate with K-H billows when $Ri_g \sim 1$, thus producing the most efficient turbulent mixing. At still larger values of $Ri_g$, the K-H billowing and wave motions subside, paving the way to less intense Hölmböe waves.

The closure of Reynolds averaged equations of motions are conducted through vertical eddy diffusivities, which can be defined in terms of vertical momentum $(\overline{u'w'})$ and buoyancy $(\overline{w'b'})$ or temperature $(\overline{w'T'})$ fluxes as (Lozovatsky and Fernando 2012),

$$K_u = -\overline{u'w'}/\partial U/\partial z$$
and

$$K_v = -\overline{v'w'}/\partial V/\partial z,$$

(1.7)

and

$$K_T = -\overline{T'w'}/\partial T/\partial z,$$

(1.8)

where $K_M \equiv K_u = K_v$ and $K_T$ are the eddy viscosity and diffusivity, respectively, and their ratio $Pr_{tr} = K_M/K_T$ is the turbulent Prandtl number (Monin & Yaglom 1975).

When temperature $T$ is the major contributor to density $\rho$, as in the dry atmosphere, the buoyancy fluctuations become $b' = -g (\rho - \rho_0)/\rho_0 = -g \rho'/\rho_0 \approx \alpha g T'$, and $K_T$
can be interpreted as the mass diffusivity. Here \( g \) is the gravity, \( \rho_0 \) the reference density, and \( \alpha \) the thermal expansion coefficient. The instantaneous velocities \((u, v, w)\) in \((x, y, z)\) directions are written in the usual forms \( u = U + u', v = V + v', w = W + w' \), the temperature \( T = \bar{T} + T' \), and the density \( \rho = \bar{\rho} + \rho' \), where \( U, V, W, \bar{T} \) and \( \bar{\rho} \) represent the mean and primes the fluctuations. Typically, these eddy diffusivities can be written as functions of the Richardson number \( Ri_g \).

In the analysis of typical natural stratified shear flows, the turbulent kinetic energy equation is used in its stationary form,

\[
0 = -u'u_j \frac{d\bar{U}_i}{dx_j} + b'w' - \varepsilon ,
\]

assuming that the energy flux divergence is negligible in volume averaged sense. Here \( b'w' \) is the buoyancy flux and \( \varepsilon \) the dissipation rate of turbulent kinetic energy. A key parameter used in discussions of stratified shear flows is the flux Richardson number,

\[
Ri_f = -\frac{-\bar{b}'w'}{-u'u_j \frac{d\bar{U}_i}{dx_j} - \bar{u}w \frac{d\bar{U}}{dz} } = -\frac{-\bar{b}'w'}{-u'u_j \frac{d\bar{U}_i}{dx_j} - \bar{u}w \frac{d\bar{U}}{dz} } ,
\]

and in the last term of Equation 1.10 it is assumed that the vertical shear prevails. In stably stratified flows, stirring of fluid by turbulence causes \( -\bar{b}'w' > 0 \), and the shear production \( -\bar{u}'w'(d\bar{U}/dZ) > 0 \). The term \( -\bar{b}'w' \) represents the correlation of buoyancy fluctuations and the movement of fluid parcels by turbulent motions. Since the vertical motions of fluid parcels in a stably stratified fluid give \( b' < 0 \), the buoyancy flux in averaged sense it is negative. A \( \bar{b}'w' < 0 \) leads to a decrease of turbulent kinetic
energy, and hence the stationarity of turbulence can only be maintained by the shear production.

In a stratified shear flow at high $Ri_g$, say with $Ri_g > 1$, turbulence production is weak because of the inability of instabilities to continuously generate turbulence. Locally the flow can become unstable due to intermittent drop of $Ri_g$, generating isolated turbulent patches (Woods 1968). Owing to the lack of a sustained energy source, turbulence within such patches demise, eventually leaving a field of slowly decaying internal waves.

Stratified turbulence is present in highly stratified katabatic flows, and the evolution of turbulence therein can be characterized using parameters listed in (1.6) to (1.10), and attempts to do so can be found in many publications (Monti et al. 2002; Pardyjak et al. 2002). However, none of the previous studies have measured the high frequency/wave number turbulence. This is attempted in this thesis using a sophisticated measurement platform, were it is attempted to measure the turbulence scales down to Kolmogorov scale and inferences are made on the suitability of conventional measurement platforms (Section 5).
CHAPTER 2:

MATERHORN PROJECT OVERVIEW AND FIELD CAMPAIGN

2.1 Introduction

The MOUNTAIN TERRAIN ATMOSPHERIC MODELING AND OBSERVATIONS (MATERHORN) PROGRAM is an Office of Naval Research (ONR) funded multidisciplinary research initiative (MURI) grant to lead multi-institutional efforts to improve meteorological modeling in mountain terrain. The program scientific objectives are:

- Identify and study the limitations of the current state-of-the-science mesoscale models at the 1 km grid scale for mountain terrain weather prediction. Determine the scientific and computational knowledge gaps and devise methodologies to overcome these barriers.

- Conduct carefully designed field studies to obtain benchmark data for the models and identify interacting physical phenomena.

- Study known unknowns; uncover hidden physical processes, delve into them using laboratory studies and high-resolution numerical simulations (Large Eddy Simulation (LES) as well as Direct Numerical Simulations (DNS)); and develop high fidelity parameterizations.

- Bridge the science gaps, identify processes across scales and delineate their contributions to basic momentum, heat and mass transfer processes.

- Evaluate models with new and existing observations. Identify the drawbacks and determine what the possible remedies are.
Develop new computational and technological tools to help realize leaps in predictability.

Transition the new knowledgebase to improve predictive models (Weather Research and Forecasting (WRF) and Coupled Ocean/Atmosphere Mesoscale Prediction System (COAMPS)).

To achieve the project goals, a multidisciplinary team was assembled including atmospheric scientists, geophysicists, numerical/theoretical analysts, engineers and applied mathematicians culled from five main academic institutions (Figure 2.1) including; University of Notre Dame (lead institution), University of Utah, Naval Post Graduate School, University of California, Berkeley and University of Virginia. A number of additional US and foreign institutions joined MATERHORN, leveraging alternative funding sources, mainly the Army Research Office and Air Force Weather Office. The Environmental Technology Laboratory of National Oceanic and Atmospheric Administration (NOAA) and the Environmental Observing Laboratory of the National Center of Atmospheric Research (NCAR) provided additional instrumentation and personnel support, which augmented resources from the regular participants and collaborators (www.nd.edu/~dynamics/materhorn).

Figure 2.1: MATEHRORN participants and program logo.
The MATERHORN project enmeshed four synergistic components that worked symbiotically toward its goal:

- The Modeling Component (MATERHORN-M) studied the predictability at mesoscale, in particular, the error growth (i.e., the sensitivity to initial conditions at various lead times), and developed meaningful measures of skill relative to appropriate conditional climatologies (i.e., the skill of capturing specific phenomena when they are supposed to appear; e.g., turbulence generation when a Richardson number criterion is satisfied). Sensitivities to input properties and boundary conditions were also investigated. Data assimilation studies are being conducted, and different techniques (e.g., 4DVAR, ensemble Kalman filtering, 3DVAR) are compared.

- The Technology Component (MATERHORN-T) developed cutting edge technologies to enable some needed, yet currently unattainable, meteorological measurements. These include an instrumented UAV, and remote sensors and samplers for moisture measurements.

- The Parameterization Component (MATERHORN-P) employed high resolution simulations with novel modeling and terrain-representation methodologies as well as imaginative laboratory studies to deduce and quantify processes intermingled (and hidden) in field observations. Salient processes are being analyzed theoretically and described using conceptual models. Insights so gained are used to develop sub-grid parameterizations with improved physics. The new parameterizations will be implemented in mesoscale models, and their efficacy will be evaluated using new and archived data taken under diverse mountain weather conditions.

- The Field Experiment Component (MATERHORN-X) conducted measurements with unprecedented spatio-temporal detail and a full suite of high-end instrumentation to support modeling efforts and process studies. This component is described in greater detail in the remainder of Section 2.

The laboratory measurements made and theoretical concepts developed as a part of MATERHORN-M and the field campaigns of MATERHORN-X are predominately the focus of this thesis. In this chapter, the field campaign is described in detail, and
given the limitation of the scope only the data taken during the first campaign was analyzed.

2.2 Location

The MATERHORN-X field campaigns were conducted at the Dugway Proving Grounds (DPG) located 137 km southwest of Salt Lake City, UT (Figure 2.2) consisting of 3700 km$^2$ of encroachment-free terrain. DPG is a secured facility, with controlled roads and air space by the Department of Defense (DOD), and a special cooperative agreement between the University of Notre Dame, University of Utah and DPG permitted the conduct of the experiment at DPG field range. Within the boundaries of DPG, unique topography is present in the form of an isolated, Granite Mountain (Figure 2.3). The closest terrain perturbation to Granite Mountain is a much smaller mountain to the southeast, Sapphire Mountain, and their separation forms a gap between the valleys to each side of Granite Mountain and the mountains farther to the southeast. Farther from Granite Mountain, approaching the DPG boundaries, the peaks climb to upwards of 2.2 km above the valley on three sides while to the north, salt flats extend for 145 km.
Figure 2.2: The orange colored region is the Dugway Proving Grounds (DPG) in relation to the State of Utah. The area encompasses 48 x 84 km of the West Desert, an area larger than the State of Rhode Island.

Besides the terrain heterogeneities, DPG also has variational land coverage. To the west of Granite Mountain, the terrain is extremely smooth with salt flats (playa) while the terrain to the east is sagebrush covered, leading to differential cooling and variations in roughness. This location is semi-arid terrain, resembling to some of the complex terrain airsheds with air quality problems such as Phoenix and Los Angeles and some remote areas of interest for national defense. The weather is modulated by land-surface contrasts, complex topography, strong inhomogeneities of terrain and land use and intrinsic synoptic variability. The DPG, and the Granite Mountain area in particular, observes much of the flow phenomena of interest including thermal circulation flows, terrain-forced flows and turbulence, frontal cyclones, convective systems and dust storms (Rife et al. 2002; Schultz et al. 2002; Schultz & Trapp 2003; Shafer & Steenburgh 2008; West & Steenburgh 2010; Jeglum et al. 2010).
Figure 2.3: Topography of DPG featuring an isolated, Granite Mountain surrounded by flat terrain on all sides in the immediate proximity.

2.3 Granite Mountain Atmospheric Sciences Testbed (GMAST) and Instrument Placement

Not only is Granite Mountain a unique topographical feature, it is also the site of GMAST (Figure 2.4A). GMAST is conceivably the most densely instrumented complex-terrain testbed available, consisting of a dense instrumentation network (Figure 2.10) of 31 - Surface Atmospheric Measurement Systems (SAMS), 51 - mini-SAMS (Figure 2.4B) and 51 - Portable Weather Information Display Systems (PWIDS) (Figure 2.4C). SAMS and mini-SAMS are 10 m towers with anemometers located at 2 and 10 m to measure wind speed/direction and a temperature/relative humidity probe (Figure 2.4F) located
at 2 m. The difference between the two is their anemometers; the SAMS feature 3D sonics (Figure 2.4D) while the mini-SAMS have Wind Monitors (Figure 2.4E). PWIDS are 2 m masts sitting atop tripods for portability and have a Wind Monitor and temperature/relative humidity probe located at 2 m. All data from SAMS/mini-SAMS/PWIDS is transmitted wirelessly to the DPG Meteorology Division in Ditto (Figure 2.4G) via a spread spectrum radio.

In addition to GMAST, three heavily instrumented Extended Flux Sites (EFS) (Figure 2.5, orange dots) were developed to investigate: (i) surface energy budgets and fluxes, (ii) internal waves and fine-scale turbulence, and (iii) SBL and CBL. The EFS exploit several contrasting features, including albedo, roughness, moisture availability and slope angle.

- **EFS-Playa** was located on the salt playa west of Granite Mountain. It is characterized by a high albedo, low roughness length and large seasonal variations in albedo and moisture.

- **EFS-Sagebrush** was located east of the Granite Mountain. Covered by sparse sagebrush-type vegetation, it is highly representative of the land cover found at DPG. This site was influenced by the nocturnal DPG basin-scale mesoscale drainage flow.

- **EFS-Slope** was located on the eastern slope of Granite Mountain. Local slope flows played an important role at this site.

Important physical parameters for the main land use types of GMAST and surroundings are given in Table 2.1. Important physical parameters of Granite Mountain are listed in Table 2.2.
Figure 2.4: Granite Mountain Atmospheric Science Testbed (GMAST): A facility for complex terrain airflow studies. (A) Granite Mountain. (B) Surface Atmospheric Measurement Systems (SAMS)/Mini-SAMS. (C) Portable Weather Information Display Systems (PWIDS). (D) 3D Sonic anemometer. (E) Wind Monitor. (F) HMP45 temperature/relative humidity probe in radiation shield. (G) Ditto, Dugway Proving Ground, the location of the Meteorology Division. (A courtesy of Dr. Stephan de Wekker, University of Virginia, G from www.dugway.army.mil, and others courtesy of Dr. Laura Leo, University of Notre Dame)
TABLE 2.1
PHYSICAL PARAMETERS FOR THE MAIN LAND USE TYPES OF GMAST AND SURROUNDINGS

<table>
<thead>
<tr>
<th>Site</th>
<th>Roughness Height* ((z_0) , (m))</th>
<th>Albedo*</th>
<th>Thermal Conductivity* (W/(m , K))</th>
</tr>
</thead>
<tbody>
<tr>
<td>EFS - Playa</td>
<td>(5.7 \times 10^{-4})</td>
<td>0.89</td>
<td>0.31</td>
</tr>
<tr>
<td>EFS - Sagebrush</td>
<td>0.24</td>
<td>0.58</td>
<td>0.27</td>
</tr>
<tr>
<td>EFS - Slope</td>
<td>0.10</td>
<td>0.44</td>
<td>0.23</td>
</tr>
</tbody>
</table>

* = Roughness Heights courtesy of Derek Jensen, University of Utah.
* = Albedo and Thermal Conductivity courtesy of Dr. Sebastian Hoch, University of Utah.

TABLE 2.2
PHYSICAL PARAMETERS OF GRANITE MOUNTAIN

<table>
<thead>
<tr>
<th>Property</th>
<th>Value (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Length</td>
<td>11.8</td>
</tr>
<tr>
<td>Largest Width</td>
<td>6.1</td>
</tr>
<tr>
<td>Peak Elevation*</td>
<td>0.84</td>
</tr>
</tbody>
</table>

* = Above valley elevation of 1.3 km.
Figure 2.5: MATERHORN-X experimental design. Orange dots are the location of the Extended Flux Sites (EFS).

Each EFS site consisted of an instrumented tower at least 20 m in height. Measurements included: temperature and relative humidity at 2, 5, 10, and 20 m; 20 Hz velocities and temperatures, momentum and sensible heat fluxes (3D Sonics and fine wire thermocouples at 2, 5, 10, and 20 m(Figure 2.6A)); CO$_2$ and water vapor concentrations (infrared gas analyzers for latent heat and CO$_2$ fluxes) (Figure 2.6B); fine-structure temperature profiles (~25 thermocouples up to 10 m, with enhanced vertical resolution near the ground (Figure 2.6C)); full radiation budget at 2 m (LW in- and outgoing, SW in- and outgoing) (Figure 2.6D); IR surface temperature, soil heat flux, soil moisture and subsurface temperature probes. Sonics were also placed 0.5 m above surface at the EFS sites to investigate the formation of skin flows, an observed
phenomenon (Manins & Sawford 1979a, Thompson 1986; Manins 1992; Mahrt et al. 2001; Soler et al. 2002; Clements et al. 2003) that is unresolved by both LES and mesoscale models and only could be adumbrated by previous observations.

Figure 2.6: EFS-Site specialized instrumentation. (A) Fine wire thermocouples coupled with 3D sonic anemometers for momentum and sensible heat fluxes. (B) Infrared gas analyzers. (C) Dense array of fine wire thermocouples with enhanced vertical resolution near the ground. (D) Net radiometer as the tower mounted component of the energy budget. (Courtesy of L. Leo)

In addition to the other EFS sites, the EFS-Slope site (Figure 2.7) had a line of five heavily instrumented towers with a total of 30 Sonics, a series of 17 HOBO® weather stations and 17 Local Energy-balance Measurements Stations (LEMS) (Figure 2.8A), Doppler Light Detection and Ranging (LiDAR) (Figure 2.8B) instruments with hemispherical scanners to investigate up/downslope flows as well as slope and valley flow interactions; a fiber optic Distributed Temperature Sensing (DTS) system (Figure 2.8C,D) using the principals of Raman spectra methods to measure temperature variation along a 2 km track of the slope; hot-film combo probes (Figure 2.8E) specifically developed to measure fine-scale turbulence down to Kolmogorov scales; and a FLIR® IR camera (Figure 2.8F), facing uphill to investigate the spatial and temporal
response of surface temperatures that are important for model validation, but not easily obtained with point sensors. The EFS-Playa site also featured unique instrumentation capable of measuring the finest scales of turbulence, employing a near surface flux Richardson number hot-wire probe (Figure 2.8 G), however unlike the EFS-Slope site combo probe, measurements were focused in the near surface layer vs. the surface layer.

Figure 2.7: The heavily instrumented EFS-Slope site flux towers. From left to right, descending down the slope are ES5, ES4, ES3, ES2 (with combo probes) and ES1. (Courtesy of L. Leo)

Besides the EFS sites, towers were strategically located to capture the flow on and around Granite Mountain, including the small and large gaps southeast of Granite Mountain, the East Slope canyons (Figure 2.9A,B) of Granite Mountain, West Slope of Granite Mountain, top of the Sapphire Mountain and multiple locations in the DPG.
basin. Figure 2.10 shows a detailed scaled map of the multi-institutional equipment deployment during the first MATERHORN-X-1 experiment.

Figure 2.8: Supplemental instrumentation at the EFS-Slope and EFS-Playa sites. (A) HOBO® weather stations. (B) LiDAR scanning towards Granite Mountain. (C,D) Distributed Temperature Sensing System (DTS). (E) 3D hot-film combo probe. (F) FLIR® IR camera facing upslope of Granite Mountain. (G) Flux Richardson number hot-wire probe. (E courtesy of L. Leo; others courtesy of Dr. Dragan Zajic, DPG)
Figure 2.9: Supplemental towers (A) East Slope canyons of Granite Mountain. (B) View from within the instrumented canyon. (Courtesy of L. Leo)

Figure 2.10: MATERHORN-X-1 Instrument placement. The main map is the DPG basin while the upper left panel is the EFS-Slope site and the bottom left panel is an overview of DPG. The domain of Section 4 is encompassed within the red box and Section 5 by the blue circle. (Courtesy: Dr. Silvana Di Sabatino, University of Notre Dame).
These ground based measurements were supported by frequent tethered balloon soundings (Figure 2.11A) at the EFS-Playa and EFS-Sagebrush sites as well as by at least 8 radiosonde launches (Figure 2.11B) per IOP at an upwind location. The soundings along with Microwave Radiometer Profilers (MWRP) (Figure 2.11C), a Sonic Detection And Ranging (SoDAR) / Radio Acoustic Sounding System (RASS) (Figure 2.11F), Ceilometers (Figure 2.11D), mini-SoDARs (Figure 2.7E), and a Frequency-Modulated Continuous-Wave Radar (FM-CW) RADAR (Figure 2.11G) probed the thermodynamic structure of the atmosphere and background meteorology. Moisture was measured at km scales via two newly developed RF Polarimetric crosshairs (surface) (Figure 2.11H; Pratt et al. 2013), microwave radiometers (vertical profiles), infrared gas analyzers (fluxes) (Figure 2.6B) and Krypton Hygrometers (Figure 2.11I; water vapor fluctuations).
Figure 2.11: Instruments used to define the thermodynamic structure of the atmosphere and background meteorology. (A) Tethered balloon soundings. (B) Launching of radiosondes. (C) Microwave Radiometer Profilers (MWRP). (D) Ceilometers. (E) Mini-SoDARs. (F) SoDAR/RASS. (G) Frequency-Modulated Continuous-Wave Radar (FM-CW) Radar. (H) RF Polarimetric crosshairs surface moisture probes. (I) Krypton Hygrometer. (A and B courtesy of D. Zajic, others courtesy of L. Leo)
To round out the full suite of instrumentation, aerial measurements were performed by the US Navy piloted Twin Otter Wind LiDAR (TODWL) (Figure 2.12A, B), DataHawk UAV (Figure 2.12C) and Flamingo UAV (Figure 2.12D). Manned TODWL flights crisscrossed the basin at 2400 m AGL, transecting the ridge of Granite Mountain, conically scanning the terrain with the onboard Doppler LiDAR to probe for various flow phenomena including mountain generated flow patterns (lee waves, rotors, separation) during synoptic influenced days and thermal circulations (slope and valley flows) during quiescent periods. Unmanned DataHawk flights flew circular Auto-Helix patterns, spiraling from the ground level to a height of 700 m AGL and then back down traversing the EFS-Slope tower line, able to characterize the flow the towers could not reach. Elaborate multiple smoke releases helped visualize the flow around Granite Mountain (Figure 2.12E, F) and provided information on the structure of the katabatic flow, dividing streamlines and flow streak lines (Figure 2.12G) associated with the complex mountain flow physics. To our knowledge, such a configuration and dense instrumentation deployment allowed sampling of Granite Mountain and the surrounding basin to the greatest detail of any complex terrain field experiment to date. Details for each instrument, variable measured and accuracy are provided in Table 2.3.
Figure 2.12: (A) Twin Otter Wind LiDAR (TODWL) preparing for flight near Salt Lake City, UT. (B) Onboard TODWL instrumentation. (C) DataHawk UAV during flight. (D) Flamingo UAV during flight testing at the Elkhart, IN airport. (E) Smoke release from the canyons above the EFS-Slope site in the evening during the initiation of the katabatic flow period. (F) Smoke release above EFS-Slope site when katabatic flow fully developed using laser illumination. (G) Dividing streamline smoke release located on the northwest side of Granite Mountain in the early morning period when sufficient stable stratification was present. (A and B courtesy of S. de Wekker, C from Dr. Ben Balsley MATERHORN 2nd Investigator meeting presentation, D courtesy of B. Johnston, E and F courtesy of Scott Coppersmith of University of Notre Dame and G courtesy of L. Leo)
### Table 2.3

**Instrument Details**

<table>
<thead>
<tr>
<th>Instrument</th>
<th>Qty Measured</th>
<th>Range(^x)</th>
<th>Accuracy</th>
</tr>
</thead>
<tbody>
<tr>
<td>Midrange-SoDAR</td>
<td>U, WD</td>
<td>50 to 400 m</td>
<td>0.5 m/s, ± 2°</td>
</tr>
<tr>
<td>SoDAR/RASS</td>
<td>U, W, WD, T</td>
<td>30 m to 1 km</td>
<td>&lt; 0.3 m/s, &lt; 0.1 m/s, ± 1.5°, 0.2° C</td>
</tr>
<tr>
<td>Ceilometer</td>
<td>Cloud height</td>
<td>0 to 7.6 km</td>
<td>± 5 m</td>
</tr>
<tr>
<td>Radiosonde</td>
<td>T, RH, P, U</td>
<td>Up to 30 km</td>
<td>0.5° C, 5%, 1hPa, 0.2 m/s</td>
</tr>
<tr>
<td>FM-CW RADAR</td>
<td>C(^2)_n</td>
<td>Usually set to 4 km</td>
<td></td>
</tr>
<tr>
<td>Wind Profiler 449</td>
<td>U, WD</td>
<td>0.075 - 16 km</td>
<td>±1 m/s, ± 10°</td>
</tr>
<tr>
<td>Wind Profiler 924</td>
<td>U, WD</td>
<td>0.075 - 5 km</td>
<td>&lt; 1 m/s, &lt; 10°</td>
</tr>
<tr>
<td>Microwave Radiometer Profiler*</td>
<td>T, H(^2)O vapor profile, Liquid H(^2)O profile, derived RH</td>
<td>0 - 10 km</td>
<td>~ 2° C, ~ 0.5 g/m(^3), 0.1 g/m(^3), 20%</td>
</tr>
<tr>
<td>3D Sonic Anemometer</td>
<td>U, V, W, T</td>
<td>NA</td>
<td>±0.05 m/s, ± 2° C</td>
</tr>
<tr>
<td>Wind Monitor Anemometer</td>
<td>U, WD</td>
<td>NA</td>
<td>±0.3 m/s, ± 3°</td>
</tr>
<tr>
<td>HMP45 Probe</td>
<td>T, RH</td>
<td>NA</td>
<td>± 0.6° C, ± 3%</td>
</tr>
<tr>
<td>Fine Wire Thermocouple</td>
<td>T</td>
<td>NA</td>
<td></td>
</tr>
<tr>
<td>Tethersonde</td>
<td>T, RH, P, U, WD</td>
<td>0 to 500 m</td>
<td>0.5° C, 5%, 1.5 hPa, 0.1 m/s, 1°</td>
</tr>
<tr>
<td>RF Polarimetric Crosshair</td>
<td>Soil Moisture</td>
<td>1 km grid scale</td>
<td></td>
</tr>
<tr>
<td>Krypton Hygrometer</td>
<td>H(^2)O vapor fluctuations</td>
<td>NA</td>
<td></td>
</tr>
<tr>
<td>Infrared Gas Analyzer</td>
<td>CO(_2) and H(^2)O density</td>
<td>NA</td>
<td>1%, 2%</td>
</tr>
<tr>
<td>Streamline Doppler LiDAR</td>
<td>U(_r), SNR</td>
<td>up to 10 km</td>
<td>&lt; 0.5 m/s</td>
</tr>
</tbody>
</table>
TABLE 2.3 (CONTINUED)

<table>
<thead>
<tr>
<th>Instrument</th>
<th>Qty Measured</th>
<th>Range&lt;sup&gt;x&lt;/sup&gt;</th>
<th>Accuracy</th>
</tr>
</thead>
<tbody>
<tr>
<td>100S Doppler LiDAR</td>
<td>$U_r, SNR$</td>
<td>up to 12 km</td>
<td>&lt; 0.5 m/s</td>
</tr>
<tr>
<td>HOBO® weather stations</td>
<td>$T, RH, U, WD$</td>
<td>NA</td>
<td>± 0.21°C, ± 3.5%, ±0.5 m/s, ± 5°</td>
</tr>
<tr>
<td>DTS system&lt;sup&gt;*&lt;/sup&gt;</td>
<td>$T$</td>
<td>2 km</td>
<td>± 0.1°C</td>
</tr>
<tr>
<td>Heat Flux Plates</td>
<td>Soil $Q$</td>
<td>NA</td>
<td>-15% to +5%</td>
</tr>
<tr>
<td>Net Radiometer</td>
<td>$SW_l, SW_o, LW_l, LW_o$</td>
<td>NA</td>
<td>± 10%</td>
</tr>
<tr>
<td>Hot-Film Combo Probe</td>
<td>$u', v', w', \epsilon$</td>
<td>NA</td>
<td>4% for 30% TI</td>
</tr>
<tr>
<td>Flux Richardson Probe</td>
<td>$\overline{u'w'}$ and $\overline{w'T'}$</td>
<td>NA</td>
<td></td>
</tr>
<tr>
<td>FLIR® IR camera</td>
<td>$T$</td>
<td>Max FOV = 63.2° x 52.4°</td>
<td>± 2°C or ± 2%</td>
</tr>
<tr>
<td>Twin Otter Wind LiDAR (TODWL)&lt;sup&gt;*&lt;/sup&gt;</td>
<td>$U, V, W, SNR, T$</td>
<td>0.3 - 21 km</td>
<td>&lt; 0.1 m/s</td>
</tr>
<tr>
<td>DataHawk UAV&lt;sup&gt;*&lt;/sup&gt;</td>
<td>$U, V, W, T, CT^2, RH, P, \epsilon$</td>
<td>to 3km AGL</td>
<td>0.1 m/s, 0.3 ° C, 1.0e-6, 0.01 %, 1.0 Pa, 1.0e-6</td>
</tr>
<tr>
<td>Flamingo UAV&lt;sup&gt;*&lt;/sup&gt;</td>
<td>$u', v', w', \epsilon, T, RH$</td>
<td>12.8 km</td>
<td>4% for 30% TI</td>
</tr>
</tbody>
</table>

<sup>*</sup> = Only present during the Fall campaign.
<sup>*</sup> = Only present during the Spring campaign.
<sup>x</sup> = Represents the maximum possible range. Results depend on atmospheric conditions.

2.4 Intense Operational Periods (IOPS)

Each MATERHORN-X campaign included ten Intensive Observational Periods (IOPs) where all instruments operated in coordination. IOP days were selected based on guidance via weather briefings provided by DPG featuring DPG’s high-resolution weather modeling system using WRF and data assimilation, cycling 8 times a day at 1.1 km resolution and an ensemble of numerical modeling including the North American Mesoscale (NAM) and Global Forecast System (GFS) models as well as satellite products.
During the campaign, weather briefings were held by the DPG forecasters, called-in meteorologists from the MATERHORN group and the participants, and the go/no-go decision for the following day was made after careful deliberation of the data needs, equipment functionality and resource availability for that day (Figure 2.13). The IOP wind speed classification is presented in Table 2.4.

Figure 2.13: DPG weather briefing at the Meteorology Division, Ditto. DPG forecasters and MATERHORN participants present.
TABLE 2.4
INTENSE OPERATIONAL PERIODS (IOPS) WIND SPEED CLASSIFICATION

| IOP Classification | Definition
|--------------------|----------------
|                    | 700 mb wind speed |
| Quiescent          | < 5m/s           |
| Moderate           | 5 m/s - 10 m/s   |
| Transitional       | Variable, could be > 10m/s associated w/front Passage |

The fall, MATERHORN-X-1 experiments (September 25 to October 31, 2012) focused on quiescent, dry, fair weather (wind speeds < 5 m/s) wherein diurnal heating/cooling provided the main forcing. The fall study included six IOPs with quiescent fair weather, the rest with moderate and strong winds. Table 2.5 contains the details of the fall IOPs including the run date, times and IOP type.
TABLE 2.5
FALL, MATERHORN-X-1 IOPS

<table>
<thead>
<tr>
<th>IOP</th>
<th>Run Date and Times (MDT)</th>
<th>Type</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>9/25/2012 14:00 - 9/26/2012 14:00</td>
<td>Quiescent</td>
</tr>
<tr>
<td>1</td>
<td>9/28/2012 14:00 - 9/29/2012 14:00</td>
<td>Quiescent</td>
</tr>
<tr>
<td>2</td>
<td>10/1/2012 14:00 - 10/2/2012 14:00</td>
<td>Quiescent</td>
</tr>
<tr>
<td>3</td>
<td>10/3/2012 2:00 - 10/4/2012 2:00</td>
<td>Transitional</td>
</tr>
<tr>
<td>4</td>
<td>10/6/2012 14:00 - 10/7/2012 14:00</td>
<td>Moderate</td>
</tr>
<tr>
<td>5</td>
<td>10/9/2012 14:00 - 10/10/2012 14:00</td>
<td>Transitional (Quiescent - Moderate)</td>
</tr>
<tr>
<td>6</td>
<td>10/14/2012 2:00 - 10/15/2012 2:00</td>
<td>Quiescent</td>
</tr>
<tr>
<td>7</td>
<td>10/17/2012 12:00 - 10/17/2012 20:00</td>
<td>Transitional (Quiescent - Moderate)</td>
</tr>
<tr>
<td>8</td>
<td>10/18/2012 5:00 - 10/19/2012 12:00</td>
<td>Quiescent</td>
</tr>
<tr>
<td>9</td>
<td>10/20/2012 14:00 - 10/21/2012 14:00</td>
<td>Moderate</td>
</tr>
</tbody>
</table>

The spring, MATERHORN-X-2 study (May 1 - May 31, 2013) dealt with larger, synoptic flow effects, moister surface conditions, and covered mostly moderate (5 to 10 m/s) and strong (> 10 m/s) wind periods. Table 2.6 contains the details of the spring IOPs.
TABLE 2.6
SPRING, MATERHORN-X-2 IOPS

<table>
<thead>
<tr>
<th>IOP</th>
<th>Run Date and Times (MDT)</th>
<th>Type</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>5/1/2013 14:00 - 5/2/2013 14:00</td>
<td>Transitional (Moderate-Quiescent)</td>
</tr>
<tr>
<td>2</td>
<td>5/4/2013 14:00 - 5/5/2013 14:00</td>
<td>Moderate</td>
</tr>
<tr>
<td>3</td>
<td>5/7/2013 5:00 - 5/7/2013 17:00</td>
<td>Moderate</td>
</tr>
<tr>
<td>4</td>
<td>5/11/2013 14:00- 5/12/2013 14:00</td>
<td>Quiescent</td>
</tr>
<tr>
<td>5</td>
<td>5/13/2013 12:00 - 5/14/2013 12:00</td>
<td>Transitional (Moderate-Quiescent)</td>
</tr>
<tr>
<td>6</td>
<td>5/16/2013 12:00 - 5/17/2013 12:00</td>
<td>Transitional (Moderate-Quiescent)</td>
</tr>
<tr>
<td>7</td>
<td>5/20/2013 17:15 - 5/21/2013 14:00</td>
<td>Quiescent</td>
</tr>
<tr>
<td>8</td>
<td>5/22/2013 14:00 - 5/23/2013 14:00</td>
<td>Moderate</td>
</tr>
<tr>
<td>9</td>
<td>5/25/2013 10:00 - 5/26/2013 10:00</td>
<td>Moderate</td>
</tr>
<tr>
<td>10</td>
<td>5/30/2013 14:00 - 5/31/2013 10:00</td>
<td>Moderate</td>
</tr>
</tbody>
</table>
CHAPTER 3:
UPSLOPE FLOW SEPARATION IN MOUNTAINOUS TERRAIN

3.1 Introduction

A component of the research in this thesis is directed at understanding of complex-terrain flow physics using laboratory experiments and theoretical analysis, which is a part of MATERHORN-M. Mountain meteorology encompasses a host of flow phenomena, and addressing all of them is obviously insurmountable. Therefore, a process that has not received adequate attention in fundamental realm, yet believed to be of critical importance was selected for detailed study, with the hope that the results will help improve prediction of flow separation from mountains.

Flow separation from mountain slopes has become a topic of considerable interest recently because of its relevance to improving mesoscale models. In particular, heat carried by the separated flow and small-scale features associated with flow separation are crucial for predicting deep convection, precipitation, moisture distribution and unsteady phenomena surrounding orographic features (Banta 1984; Hanley et al. 2011). Excessive precipitation over steep and high mountains in mesoscale and climate models is a well-known problem, which has been attributed to the difficulties of parameterizing flow separation over slopes, especially the sub-grid heat
ventilation by separated upslope flows (Chao 2012). Without proper ventilation, the heat flux and slope flows are overestimated, which together with excessive moisture transport lead to erroneous precipitation predictions. In the urban context, more than 70% of world’s cities are in valleys surrounded by complex topography, and upslope (or anabatic) flows occurring during the daytime transports urban pollution plume to mountainous rural surroundings (Ellis et al. 2000). The fate of the pollutants is largely dependent on whether the upslope flow separates or not. When the separation occurs, pollutants are lofted to regional flow or form elevated recirculating flows (Lu & Turco 1994), whereas in attached flows there is a propensity for pollutants return back to the urban area with the downslope (katabatic) flow at night (Fernando et al. 2001). In aircraft operations, the flow separation in the lee or windward side of the mountains is a key issue of safety (Politovich et al. 2011).

In predictive models, flow separation is sensitive to the details of turbulent models employed (Mason 1987; Moreira et al. 2012), and hence an understanding of flow separation is vital for developing turbulence closure. While a substantial amount of theoretical, numerical and laboratory studies have been conducted on flow over mountains under stable and neutral conditions (Taylor et al. 1987; Baines 1995; Raupach & Finnigan 1997; Belcher & Hunt 1998; Whiteman 2000), fundamental studies on flow separation during convective conditions are meager. Heating a slope generates an upslope flow, and the vorticity of this flow is countered by the baroclinic generation of vorticity that facilitates flow separation (Crook and Tucker 2005). Thus the flow separation is governed by two opposing influences, and in this paper a simple model is
described to estimate the separation distance, which is validated against the results of a laboratory experiment.

3.2 Model

Consider a two dimensional, steady upslope flow along a uniformly heated incline of length $L_0$ and of angle $\beta_s$ with along ($s$) and normal ($n$) coordinates (Figure 3.1), where the flow develops at the base of the incline. The flow is assumed to be turbulent and Reynolds number independent, as observed to be the case so only inertial and buoyancy forces are considered important. We also assume no stratification is present and are using a Boussinesq approximation in scaling estimates. As the flow proceeds along the slope, a baroclinic vorticity develops that facilitates flow separation, viz.

$$u \frac{\hat{c} \omega}{\hat{s}} = \nabla \times (b \hat{k}),$$

(3.1)

where $\hat{k}$ is the vertical unit vector, in $z$ direction. The development of separation vorticity (lift off component) can be estimated as,

$$u \left( \frac{\hat{c} \omega}{\hat{s}} \right) \sim \left( \nabla \times b \hat{k} \right),$$

(3.2)

where $\left( \nabla \times b \hat{k} \right)_y = - \partial / \partial s \left( b \cos \beta_s \right)$ assuming vertically mixed and $\omega_y \approx \partial u / \partial n$. Introducing the scaling $\partial / \partial n \sim 1 / \delta$ and $\partial / \partial s \sim 1 / L_s$, where $\delta$ is the upslope layer thickness and $L_s$ is the separation length, we arrive at the vorticity at separation,

$$\omega_s \sim \frac{\Delta b_s \cos \beta_s}{u_s},$$

(3.3)
which is to be just balanced by the upslope flow vorticity \( u_s / \delta \) acting to keep the flow attached. Here \( u_s \) is the upslope velocity at separation. Thus we obtain,

\[
\begin{align*}
\frac{u_s^2}{\Delta b} \sim \Delta b \cos \beta_s \delta_s.
\end{align*}
\]

(3.4)

Now using the buoyancy conservation \( \partial (\Delta b u \delta) / \partial s \sim q_0 \), where \( q_0 = g \propto Q / \rho_0 C_p \), with \( \alpha \) the thermal expansivity, \( Q \) the heat flux per unit area, \( g \) the gravity, \( C_p \) the specific heat and \( \rho_0 \) the reference density, for upslope flow, this becomes,

\[
\Delta b u_s \delta_s \sim q_0 L_s,
\]

(3.5)

![Figure 3.1: Schematic representation of upslope flow.](image)

and from (3.4) and (3.5), we obtain the upslope velocity at separation,

\[
\begin{align*}
u_s = \gamma (q_0 L_s \cos \beta_s)^{1/3},
\end{align*}
\]

(3.6)

where \( \gamma \) is constant. Based on Hunt et. al. (2003), the general upslope velocity can also be written as \( u_s \sim (q_0 \delta_s \sin \beta_s)^{1/3} \), which is valid just before the flow separation.

Equating the two estimates, we find \( L_s / \delta_s \sim \tan \beta_s \).
During the separation, field observations show that the flow rises as a plume formed in the region between the separation of the flow and the apex of the slope, fed by the anabatic flow as shown in Figure 3.2.

![Figure 3.2: At separation the flow rises as a plume that is fed by the anabatic flow.](image)

The buoyancy flux fed into the plume is \( q_0 L_t \), where \( L_t = L_0 - L_s \), and the buoyancy flux over the equivalent surface horizontal planar is,

\[
q^*_0 = \frac{q_0 L_t}{L_t \cos \beta_s}.
\]  
(3.7)

The vertical plume velocity now becomes \( w \sim (q^*_0 L_t \cos \beta_s)^{1/3} \), where \( L_t \cos \beta_s \) is the plume width (Colomer et al. 1999). After substitution, the vertical plume velocity scales as,

\[
w \sim (q_0 L_t)^{1/3}.
\]  
(3.8)

The volume flux of the plume \( \sim (q_0 L_t)^{1/3} L_t \cos \beta_s \) needs to be balanced by that of the anabatic flow \( \sim u_s \delta \) feeding the plume. Equating the two at the point of separation, \( u_s \delta \sim (q_0 L_t)^{1/3} L_t \cos \beta_s \) and using (3.6) and \( L_s + L_t = L_0 \) an expression for \( L_s / L_0 \) is reached,
\[
\frac{L_s}{L_0} = \left[ 1 + \Pi \frac{\sin^{1/4} 2\beta_s}{\sin \beta_s} \right]^{-1},
\]  
(3.9)

where \(\Pi\) is a proportionality constant arising from scaling. Note that this is a more definitive form of the general expression obtained from the straight forward dimensional analysis, where \(L_s = f(q_0, L_0, \beta_s, \nu)\), where \(\nu\) is the kinematic viscosity.

The resulting form \(L_s/L_0 = f_1(H/L_0, Re) \rightarrow f_2(H/L_0)\) when \(Re = (q_0 H)^{1/3} H/\nu \rightarrow\) large obeying Reynolds number similarity. It is also evident that \(L_s/L_0\) has only a weak dependency on \(\beta_s\).

3.3 Experimental Procedure

The experiments were carried out in a rectangular glass tank (cross section of 125x35 cm) filled with deionized water to the depth of 30 cm. To reduce heat exchanged with the ambient air to a negligible minimum, experiments were performed with water at room temperature and thermal insulation on all sides was provided by 5 cm thick foam sheets. A removable window in the insulation allowed optical access to the flow. A unit serving as a 15.6x35 cm heated slope was custom manufactured and consisted of an array of densely packed heating wires embedded in silicon rubber. The bottom of the pad was fully insulated, hence heat flux from it to could be estimated as \(Q = IV/A\), which was independently verified using the standard calibration techniques describes in Hunt et al. (2003). The unit was heated by applying prescribed DC current, with the maximum possible heat flux of 7.1 kW/m\(^2\). The heat flux was regulated using a constant current DC power supply.
Two experimental configurations were considered, as shown in Figure 3.3. In the first, the slope was placed abutting the wall (Figure 3.3A) and by neglecting the no slip condition, the flow was assumed similar to the case of symmetric triangular-shaped mountain (Figure 3.3B). Obviously the no-slip assumption is invalid close to the vertical wall, but for cases of much larger thickness of (vertical) plume compared to the boundary-layer thickness at the wall, our assumption is plausible, in that the flow separation is unaffected by the vertical wall. A set of experiments were conducted, using the two configurations, under otherwise identical conditions, and the separation points for both cases were within ± 7 %. Given that configuration in Figure 3.3A allows larger flow development times before end walls become important, most of the experiments were conducted using that configuration.

Figure 3.3: Schematics of experimental configurations used. (A) Slope placed abutting the wall. (B) Case of symmetric triangular-shaped mountain.

In the experiments, the water tank was allowed to settle for two hours for residual motion to decay, whence a prescribed current was applied to heating elements. Video recordings of the field of view (FOV) were made of the initial transients, the quasi-steady current feeding the upslope flow with separation, and the experiment was terminated when substantial velocities were detected at the distant end wall. Post
processing of data allowed identification of the quasi-steady period. Temperature on the slope was obtained using a number of k-type thermocouples, and approximate uniformity of temperature along the slope was confirmed. Ten prescribed slope angles $\beta_s = 5^\circ, 10^\circ, 15^\circ, 18^\circ, 20^\circ, 25^\circ, 30^\circ, 35^\circ, 40^\circ$ and $45^\circ$ were used, and the heat flux was varied between 9.5 W/m$^2$ and 2.45 kW/m$^2$.

Various velocity parameters were obtained by Particle Tracking Velocimetry (PTV). The flow was seeded with polyethylene fluorescent particles 70 $\mu$m in diameter with density $\rho \approx 1$ g/cm$^3$. A 2 mm thick laser sheet, introduced through a slit in the insulation, illuminated the flow at the middle of the tank cross section. To obtain optimal particles track detection, special attention was paid to attain sufficient water clarity and particle seeding density. A 752x480 pixel CCD camera which was used to record videos of particle motion in the tank at 5 fps was positioned in front of the observation window. The field of view (FOV) of the camera was set to capture the flow at the slope and the immediate vicinity at 35 to 54 pix/cm spatial resolution. Recorded videos were split into individual frames and processed by in-house developed PTV software. Image enhancement to improve contrast and spatial filtering was first applied and then individual particles were detected in each frame, allowing reconstruction of 2D particle motion in the FOV. Velocity components were calculated using track information in later stages of data processing. Particle tracks and velocity components allowed detection of the distance along the slope at which the flow separates, $L_s$ and the mean local velocity at this point, $u_s$. 
The distance $L_s$ was detected by examining a map of particle tracks and with the help of Feature Tracking Visualizations (FTV). FTV reconstructed particle paths throughout the FOV, color mapping them according to the track length and/or velocity, hence obtaining a good representation of the time history of the flow. A combination of static paths maps and FTV animations allowed confident detection of $L_s$ with acceptable accuracy, which is discussed further in the results section. Values of $u_s$ were then obtained by ensemble averaging velocities at the vicinity of the separation point.

3.4 Results

3.4.1 Observations

After a transient period, the flow achieves a quasi-steady period, followed by the development of a circulation cell. It was the quasi-steady period that was of interest, in particular the flow separation. Figure 3.4B shows the PTV pathlines evaluated in plan view by illuminating a horizontal cross section by the sheet of laser light, at a height of 4 cm from the bottom, and a corresponding vertical section is given in Figure 3.4C. Note that the flow tends to be two dimensional, except near the tank walls where small deviations can be seen in Figure 3.4B.
Figure 3.4: Particle tracks at an illuminated cross section 4cm above the tank bottom. (A) Schematic of cross section. (B) PTV pathlines evaluated in plan view. (C) The corresponding vertical cross section in the middle ($\beta_s = 20^\circ$).

Given the 2D nature, most of the particles could be tracked for long times in FOV, thus providing long pathlines. Of course, the two dimensionality is more pronounced near the slope, complicated by the upslope flow, separation and plume formation. Many vertical and horizontal cross sections were evaluated (not shown), and the horizontal flow arriving at the slope in far field and their modification in the near field was noted, especially in the central region of the tank. The quasi-steady flow
established after a minimum of 200 s from the start of heating, but it varied with the experimental parameters. Another 100 s of quasi-steady data were possible. Repeatability of the results as well as similarity between flow configurations of Figures 3.4B-C away from the vertical wall were also investigated using velocity measurements.

One of the very distinct observations was the change of the nature of slope flow with the slope angle. At larger $\beta_s$ (> 20° or so), the flow consists of upslope flow at low elevations, flow separation at a particular location, beyond which it appears as if there is a rising plume fed by the separated upslope flow (Figure 3.5A). Upon formation, it also entrains fluid from the ambient fluid that approaches fluid horizontally. Given that the net vorticity of the shear flow and baroclinic term therein are small, the rapid rise of the plume is expected to be due to the heat flux in the separated region.

As the slope angle decreases, the separation point recedes, finally arriving at a scenario where the upslope is not developed (and the theoretical arguments in Section 3.2 are invalid). This appears to occur at angles $\beta_s < 20^\circ$ (Figure 3.5B). In this case, the plume flow rate cannot be fully supplied by the upslope flow, as assumed in section 3.2, and hence at the level of plume formation, strong entrainment of outside flow into the plume core is possible. Here the flow tends to be three dimensional. A further reduction of $\beta_s$ reduces the upslope development, finally causing the flow to be similar to that into a plume of finite source dimensions, as studied by Colomer et al. (1999); see Figure 3.5C. Now the separation point is determined by the balance between vertical inertial forces of entrainment flow and the buoyancy forces. This aspect is further discussed in Section 3.4.2.
Figure 3.5: A track map extracted from feature tracking visualization (FTV) animation. It illustrates the long-exposure path lines of fluid parcels. (A) $\beta_s = 25^\circ$. (B) $\beta_s = 15^\circ$. (C) $\beta_s = 5^\circ$. Note that in (A) few outside parcels penetrate the plume core causing upslope flow to feed the plume whereas in (B) and (C) there is substantial penetration.
3.4.2 Separation Length Observations

To obtain quasi-steady flow with minimum end wall influence a particular range of buoyancy fluxes $q_0$ was necessary, and the focus was on such cases. At $q_0$ below this range, the flow either developed numerous eddies that rolled up the slope (similar to the onset of laminar convection on slopes; Chen et al. 1991) and circulated in a counterclockwise direction as a single eddy of size approximately equal to the depth in the tank. When conditions for quasi-steady upslope flow prevailed, the mean flow was slope parallel, with entrainment at the edges, and it separated at a distance $L_s$ from the bottom (Figure 3.5A). Small fluctuations of the separating flow paths were detected, as observed by releasing small amounts of Fluorescent dye on the slope and by the FTV animations.

The distance to the separation point of the flow, $L_s$ was detected for each case examined and scaled by the corresponding slope length $L_0$. The detection was performed by manually examining particle tracks obtained from PTV and selecting the point of the first visible track separation from the slope. The detected separation point location was then verified, and corrected if necessary, by examination of FTV animations. A combination of both the static map of particle tracks and the animated FTV allowed a high level of certainty of the separation location detection, estimated to be less than the distance traveled by the particle during three consecutive frames. Keeping in mind the image acquisition frame rate of 5 fps a position error was estimated as $\Delta L_s = \pm 3\Delta t u_s$, where $\Delta t$ is the time delay between the frames. A method used to obtain values of $u_s$ is elaborated further in text.
Figure 3.6 shows the dimensionless distance to separation $L_s/L_0$ as a function of $Re$. The estimated errors of $L_s/L_0$ were around 12% for lower $\beta_s$ values and less than 4% for higher $\beta_s$, and are presented as error bars. The separation distance increased with the slope angle, from 0.16 $L_0$ for $5^\circ$ and up to 0.7 $L_0$ for the largest examined angle of $45^\circ$. A higher buoyancy flux $q_0$ was required to initiate an upslope flow with the decrease of slope angle. The obtained results show that at each angle the separation distance is virtually constant and therefore $Re$ independent at all range of $Re$ used. Also, for slope angles $\beta_s<10^\circ$, the normalized separation length remained approximately constant.

Next a slope of double the original length (31.2 cm) was used. Measurements were performed at slope angles of 20° and 30° resulting in comparable (deviation within 8%) $L_s/L_0$ values to those presented in Figure 3.6; a few data points are presented in Figure 3.6. The double length slope however increased the rising flux of the buoyant fluid above the slope causing the rapid formation of a thick, warm layer which violated the assumption of constant background density. Therefore results from the double length slope were limited to a few cases.
Figure 3.6: Dimensionless flow separation distance $L_s/L_0$, with error bars calculated based on Section 3.4.2. O shows data with a slope of double the original slope length.

The measured separation lengths are shown in Figure 3.7 as a function of $\beta_s$, together with the prediction (3.9). The model and data agree well for $\beta_s > 20^\circ$, with $\Pi = 0.35$ but below the critical $\beta_{sc} = 20^\circ$ the model over predicted the separation length. The data shows a sudden drop of $L_s/L_0$, and an approximately constant $L_s/L_0 \approx 0.2$. Based on the observations discussed in Section 3.4.1, this can be attributed to a change of flow structure from an upslope flow fed separated plume to one that resembles a plume of finite size of diameter $\sim L_0$. In the latter, the contribution of upslope flow is unimportant and the plume rises almost entirely from the slope. The volume of plume fluid is now supplied by the entrainment flow penetrating deep into the plume core, whence the model assumptions are invalid.
Figure 3.7: Ensemble averaged non-dimensional separation length $L_s/L_0$ distribution as a function of the slope angle $\beta_s$, $\Pi = 0.35$.

The flow situation at $\beta_s < \beta_{sc}$ indeed resembles to that of a buoyant plume of finite size, described by Colomer et al. (1999). It has a receding region at the edge of the source, giving the appearance of flow separation, but this is due to the influence of the entrainment flow into the plume that overshadow the rising thermals at the plume edge. The penetration distance of the entrainment flow before lift off can be estimated by the balance of buoyancy forces acting on the fluid parcels entering the plume via lateral entrainment flow and the vertical inertia forces on the fluid parcels, viz.

$$b \sim \frac{Dw}{Dt} \sim u \frac{\partial w}{\partial x} \sim \frac{u_e w}{L_s} \sim \frac{\alpha w^2}{L_s},$$

where $L_s$ is the separation distance from the edge, $w$ is the plume velocity outside the thermal boundary layer and the entrainment velocity $u_e = \alpha w$, $\alpha$ being the
entrainment coefficient. For the 2D case, the Colomer et al. (1999) scaling for axisymmetric plumes becomes,

\[ b \sim \left( \frac{q_0 L_0}{L_0} \right)^{2/3}, \quad w \sim \left( q_0 L_0 \right)^{1/3} \quad \text{and} \quad u_e \sim \left( q_0 L_0 \right)^{1/3}. \]  

(3.11)

Together, (3.10) and (3.11) gives \( L_s \sim \alpha L_0 \), which is in good agreement with \( \beta_s < \beta_{sc} \) observations of Figure 3.10. To further corroborate the velocity scales, the measured \( u_e \) and \( w \) are shown in Figure 3.8 at \( z = 0.3 L_t \cos \beta_s \) above the slope apex. When \( \beta_s = 5^\circ \), the plume is fed purely by entrainment with \( \alpha = 0.41 \), while \( \beta_s > 5^\circ \) shows a transition from a pure plume towards upslope flow with reduced entrainment from the ambient fluid.
Figure 3.8: Measured $u_e$ and $w$ of the plume at $z = 0.3L_T \cos \beta_s$ above the slope apex, with error bars calculated based on Section 3.4.3. $w$ is the plume velocity outside the thermal boundary layer and the entrainment velocity $u_e = aw$, $a$ being the entrainment coefficient. The 2D case for axi-symmetric plumes, (Colomer et al. 1999) is also shown as O, with $\alpha = 0.59$. The $\beta_s = 5^\circ$ case corroborates the velocity scale with $\alpha = 0.41$, while $\beta_s > 5^\circ$ shows a transition from a pure plume towards upslope flow with reduced entrainment from the ambient fluid.

3.4.3 Velocity Magnitude

Values of the local mean velocity magnitude at the point of separation, $u_s$ were obtained next. A small rectangular region of interest, ± 0.4 cm around the detected point of separation was chosen. Particle velocities were calculated in this rectangular region by dividing track lengths between two consecutive frames by the frame rate value (5 fps). A representative histogram of calculated velocity magnitudes is shown in Figure 3.9.
To ensure the full inclusion of the near slope boundary layer in the region of interest the rectangle was chosen to include a portion of the slope itself, resulting in partial combination of the raw velocity data with some falsely detected values. The main ensemble of velocity magnitude values were distributed in a well pronounced bell shape, close to a normal distribution, as seen in the histogram. Numerous values distributed close to zero are sourced from false particle detection at the slope surface caused by the laser sheet reflection. Also, a small number of velocity magnitude values from outside the boundary layer can be present which are much higher than the main body of data. False values were filtered out from the ensemble by carefully examining histograms for each case and the local mean velocity magnitude values were then obtained by ensemble averaging the filtered data. This method allowed a high level of certainty of the velocity, based on the particle position detection accuracy of ± 0.5 pixels, spatial resolution of 35 to 54 pix/cm and a frame rate of 5 fps, errors are estimated to be less than 6.8% to 10%.
Figure 3.9: Velocity at separation point magnitude distribution histogram. Conditions and the small frame are the same as in Figure 3.5(A). Small rectangular region of interest, ± 0.4 cm around the detected point of separation was chosen.

In the analysis of the data, quite interestingly, the along slope velocity data at the separation point $u_s$ were found to fall into three distinct regimes. All data for $\beta_s > \beta_{sc}$ (20°), followed (3.6), as shown in Figure 3.10 as a plot of $u_s$ versus $(q_0L_s\cos\beta_s)^{1/3}$.

Also evident from Figure 3.10 is the clustering of data for $\beta_s \approx 15°$ and 18°, which can be attributed to the observed deviation of flow from the assumptions made in deriving (3.6). At these angles, the separated plume could not be supplied via the upslope flow only, and additional entrainment flow was required for the separated region. The results for 5° and 10°, slopes also showed a distinct cluster, arguably representing the case of “pure” plume emanating from a source, commensurate with Figure 3.5C. These data are
essentially similar to $u_\phi$ data shown in Figure 3.8, where the separation point is now determined by the entrainment flow, and $u_s \sim (q_0 L_s \cos \beta_s)^{1/3} \sim (q_0 L_0)^{1/3}$ for small $\beta_s$.

![Graph](image)

Figure 3.10: Local mean upslope velocity, $u_s$ at point of flow separation. Dashed line - power fit (3.6) with $\gamma = 3.7 \times 10^{-2}$. Fit excludes $q_0$ below the minimum required for establishment of quasi-steady upslope flow.

3.5 Field Measurements

During the MATERHORN-X-2 field campaign, a portion of Granite Mountain, south of the EFS-Slope site was selected as a potential location where upslope flow separation might occur, since frequent observations showed the presence of a solitary cumulus cloud which formed directly above this portion of the slope (Figure 3.11). Upon further investigation of the site, it was found that the peak was nearly symmetric (Figure 3.12), the slope was nominally two dimensional over horizontal distances on the order of 1.1 km and was situated in a way that both sides of the peak were exposed to the sun.
throughout much of the day, thus satisfying the model assumptions in Section 3.2 to a reasonable degree.

![Figure 3.11: Granite Mountain, south of the main EFS-Slope site. Note the solitary cumulus cloud formed over the peak. (Courtesy: L. Leo).](image1)

The LiDAR was positioned at the base of the slope (Figure 3.13) and scanned continuously, sweeping from the base towards the peak using a program in which the azimuth angle was maintained while the elevation angle was varied with each scan.
taking 30 seconds to fully complete. The slope was scanned for a 24 hour period with the emphasis of capturing the morning transition, the evolution of the upslope flow and potential separation as well as the evening transition to downslope flow. During this period the background meteorology conditions were quiescent, in which a synoptic scale flow was not present and thermal circulations (slope/valley flows) were the dominant flow pattern.

![Image](image.jpg)

Figure 3.13: LiDAR positioned at the base of Granite Mountain scanning the slope. The slope angle, $\beta_s \approx 23^\circ$.

The LiDAR measurement beam was aligned perpendicular to the mountain, along the slope in the mean upslope wind direction. Using arguments similar to those proposed by Banta et al. (2013), this allows the projection of the wind components $u$ and $w$ onto the measured radial Doppler wind velocity, $u_r = u \cos \theta + w \sin \theta$, where $\theta$ is the elevation angle (Figure 3.14). Along the slope surface in the anabatic layer, $w$ can be assumed small such that along the slope $u_s \sim u_r / \cos(\theta - \beta_s)$. Outside of the anabatic layer, the upslope velocity $u_s$ goes to zero and $w \sim u_r / \sin \theta$. A LiDAR scan
during the upslope period is presented in Figure 3.15 in which the radial velocity is colored while vectors are overlaid to show the direction of the flow. No vectors are plotted in the region where neither $u$ nor $w$ can be assumed small.

![Figure 3.14: Schematic of LiDAR scan configuration where $\theta$ is the LiDAR beam elevation angle and $\beta_s$ is the slope angle.](image)

From the LiDAR results, it is clear that at low elevations an upslope flow travels along the slope (A). As the upslope flow continues to develop and gain heat along the slope, the separation vorticity develops which is balanced by the upslope flow vorticity $u_s/\delta$ acting to keep the flow attached. Unlike a typical upslope flow, in this case the flow ceases along the slope after some distance due to separation from the slope. As the flow separates at the estimated location (B), the anabatic layer thickens as the plumes begin to rise vertically, developing a strong vertical velocity. Beyond this point a plume is formed (C), fed by the upslope flow.
As evident from Figure 3.15, because of the measurement difficulties near the surface, estimation of the separation point is somewhat difficult, but it is possible to approximately determine the separation point as where the flow develops a strong vertical velocity indicating it is deviating from the surface. The observed dimensionless distance to separation (B) is $L_s/L_0 = 0.45 \pm 0.025$ (±1 LiDAR range gate) where $L_0 = 712$ m, while for $\beta_s = 23^\circ$, the model (3.9) predicts a value of $L_s/L_0 = 0.55$ with uncertainty of ±0.02. The upper limit of observed separation distance is within 39 m of the model prediction. The agreement is reasonable, considering the multitude of factors.
that potentially influence the separation distance, causing the flow to separate sooner than in the model.

Also note that in the field the angle of the lower-elevation slope is more gradual than the upper portion, increasing from $\beta_s = 16.5^\circ$ to $23^\circ$. Although a discontinuous slope has not been studied in the lab, as the slope angle decreases, the separation distance decreases (Figure 3.7) and it can be presumed the lower portion of the slope has an effect on the separation distance. Another factor influencing the separation distance is the added roughness of the slope compared to the laboratory slope which was smooth. Added roughness causes the boundary layer thickness, $\delta$ to increase, which decreases the upslope flow vorticity, $u_s/\delta$ acting to keep the flow attached allowing the separation vorticity to overcome the upslope vorticity at a lower elevation along the slope.

3.6 Conclusions

Laboratory experiments were performed to investigate the flow separation from isolated 2-dimensional mountains, specifically focusing on steeper slopes. The interest was on flow structure, separation distance, and velocity field. The experiments were performed using a (125x35x30 cm) tank, with mountain slope consisting of a specially designed heating pad located at one end. Particle tracking Velocimetry (PTV) was implemented in the regions of interest (slope and vicinity). PTV results were also used to produce Feature Tracking Visualization animations, color mapped according to track length and/or velocity. Measurements were performed on numerous slope angles (from
5° up to 45° from the horizontal) and buoyancy flux settings, collecting substantial sets of experimental data. The separation distance along the slope was determined by a combination of examining static maps of particle tracks and FTV animations. Antecedent experiments revealed that the tank side walls are unimportant for a period of at least 100 s between initial transients and formation of long recirculation cells. Flow was confirmed to be two dimensional during this period for steep slopes.

Three different flow regimes were identified. When the slope angle $\beta_s \geq \beta_{sc} \approx 20°$, the separated flow was found to be fed entirely by the upslope flow that precedes separation. A model that assumes the separation point is determined by the opposing vorticities due to baroclinic torque (separation tendency) and shear flow (attachment) produced a good prediction for flow separation in this case. The velocity scale at the separation point was also well captured.

When $\beta_s < \beta_{sc}$, the plume emanating at the separation point could not be fully provided by the upslope flow, and needed to be partly supplied by the ambient fluid as enhanced entrainment flow. The separation distance drastically reduced under these conditions, and at $\beta_s < 10°$ the volume spread over most of the slope, much similar to the case of plumes rising from area sources (finite source diameter). The entrainment flow penetrated about $L_s \approx 0.2L_o$, before being overpowered by rising plumes, which could be considered as separation length.

Mountain terrain weather is among the most difficult to predict, and flow separation that lead to strong updrafts and deep convection is a topic of great interest that is investigated in the ongoing MATERHORN (2011-2016) program. It is hoped that
the present work will contribute towards interpreting field studies of the MATERHORN program.
CHAPTER 4:
MULTI-SCALE INTERACTIONS OF SLOPE AND VALLEY FLOWS

4.1 Introduction

A component of research in this thesis that is of particular interest to MATERHORN-X-1 is directed at understanding complex multi-scale interactions between meso-α and β scale (up/down) valley flows and meso-γ scale thermally driven (up/down) slope flows. These interactions between various types of thermally driven as well as thermally and synoptically driven flows lead to highly variable weather, and are of key importance to mountain meteorology. Furthermore, the physical processes underlying these interactions are not well understood nor have they been discussed in detail previously. Until the recent MATERHORN field campaigns, no detailed observations of these flow interactions existed (Fernando and Pardyjak 2013). It is hoped that present research will improve the understanding of the nature of these interactions and quantifying them for the inclusion of mesoscale atmospheric models.

Valley flows, driven by pressure gradients due to temperature variation along the valley or along-valley buoyancy forces often encounter intermittent, pulsating flow through canyons and gaps within the complex topography. Slope flows descending into the valley floor interact with the valley flows and at times collide with them, producing a
host of small scale phenomena including; collision fronts, intense turbulent patches, intrusions and instabilities which contribute vigorously to sub-grid heat and momentum transfer. Conversely, depending on the background conditions or when the valley winds get strong enough, horizontal and vertical shears can greatly modify the slope flows or completely obliterate them (Doran et al. 1990). Such processes are not properly accounted for in currently used mesoscale models, and eduction of such processes via observations and reporting of their flux contributions are a part of the work reported in this section.

While in principle interacting thermally driven flows or thermal and synoptic flows may be best studied using numerical simulations, mesoscale models do not incorporate sub-grid flux contributions of these processes, and hence tend to be flawed in predicting the mean flows, let alone the interactions between desperate mean flows. Detailed field observations to be discussed in this section will help understand and parameterize collisions and interactions between different types of flow elements present in mountain terrain winds.

Our approach is to analyze MATERHORN-X-1 observations to identify possible collision events, measure short-term averaged fluxes, obtain averaged flux over a particular event and parameterize it using pre-collision and in-situ variables.

4.2 Experimental Domain and Instrumentation

Ideal meteorological conditions to properly investigate the interactions between slope and valley flows are quiescent, high pressure background flow in the absence of a
synoptic flow, wherein thermal circulation is the dominant flow. In addition, clear skies are needed to ensure maximum radiative cooling at the surface to establish strong thermally driven flow. Further requirements are that all of the needed instrumentation be in place and working at optimum conditions for detailed measurements. In the MATERHORN-X-1 campaign, IOP 2 (Table 2.5) met all of these requirements and was selected as the experimental period for the slope and valley flow interaction study.

The domain of the study was selected to include the rich topography of the East Slope of Granite Mountain (Figure 4.1) and the heavily instrumented EFS-Slope site and the Dugway basin. Embedded within this region is the dense instrumentation of GMAST (Figure 2.4A) and the EFS-Sagebrush site. Also of topographic interest within the domain are the small and big gaps on either side of Sapphire Mountain to the southeast of Granite Mountain. These gaps play an important role in channeling convergent flow and the setup of basin stratification in the nighttime hours. The domain of the slope and valley flow interaction study (encompassed within the red box) and detailed scaled map of the multi-institutional equipment deployment during the first MATERHORN-X-1 campaign is shown in Figure 2.10.

As previously discussed in Section 2.3, the EFS-Slope site consisted of a line of five flux towers (Figure 2.7) at least 20 m in height, with measurements of temperature, relative humidity, velocities, momentum and sensible heat fluxes and full radiation budget. This site also had Doppler LiDARs (Figure 2.8B) with hemispherical scanner on either side of the tower line, 600 m apart, capable of coordinated scanning of the incoming valley flow as well as the descending slope flow. Further instrumentation
included the DTS fiber optic temperature system (Figure 2.8C,D) sensing the near surface temperature of a 2 km track of the slope and a FLIR® IR camera (Figure 2.8F), facing uphill to investigate the spatial and temporal response of surface temperatures.

Figure 4.1: View of the EFS-Slope site and valley below from the Peak of Granite Mountain.

In the valley below, the dense GMAST instrumentation (Figure 2.10) (SAMS, mini-SAMS, and PWIDS) measured the temperature, relative humidity, wind speed and direction of the passing valley flow in great detail. At the EFS-Sagebrush site a 20 m tower measured temperature, relative humidity, velocities, momentum and sensible heat fluxes and full radiation budget. All of these measurements in the domain were supported by frequent tethered balloon soundings (Figure 2.11A) at the EFS-Sagebrush sites as well as by at least 8 radiosonde launches (Figure 2.11B) during the IOP at an upwind location.

Outside the Dugway basin, towers were strategically located within the small and big gaps to either side of Sapphire Mountain as well as on the mountain peak to
capture the flow into the Dugway basin within the surface layer, while a mini-SoDAR (Figure 2.11E) measured the flow aloft of the big gap.

4.3 Initiation of the Flow

As the sun passes to the west of Granite Mountain, shortly before sunset, radiative cooling occurs rapidly along the East Slope surface, forming a surface inversion followed by the dissolution of a CBL. According to the theory of Hunt et al (2003), a stagnation front forms at a distance \( \hat{x} = x_f \) from the bottom of the slope. As an air parcel moves along the slope, it is gradually cooled, and at some point the parcel becomes sufficiently dense to overcome the inertia forces of upslope flow. Following the front formation, for \( \hat{x} < x_f \), the flow continues upslope, whereas at \( \hat{x} > x_f \) a downslope flow is initiated near the surface while the flow aloft is still upslope (Figure 1.7). The front propagates downslope, trailed by the katabatic current, and the entire downslope flow undercuts the prevailing (but dwindling) upslope flows, completing the transition. Meanwhile at lower elevations in the Dugway basin, the evening transition in the valley flow is initiated a few minutes to approximately one hour before the sunset and is completed within minutes, accompanied by a drop in temperature, abrupt wind shifts, an increase of specific humidity, and low-level jets.

After the evening transition has occurred, the downslope flow quickly develops with velocities reaching up to 3 m/s, draining beyond the base of the slope out into the valley as shown in Figures 4.2 and 4.3. As the basin fills, the flow flows through the big gap leaving the basin (Figure 4.2A, 3:30 UTC (21:30 MDT)). This period is short lived as a
competing flow from the valley to the southwest reaches the gap and overtakes the exiting flow (Figure 4.2B, 3:45 UTC (21:45 MDT)) shifting the direction of flow. Slightly delayed, this also occurs at the small gap as flow overtops the passageway, merging with the downslope flow from the southern flank of Granite Mountain (Figure 4.3A, 4:00 UTC (22:00 MDT)). Upon the merger of these flows, convergence takes place (Figure 4.3B, 4:15 UTC (22:15 MDT)), accelerating the merged flows and big gap flow toward the valley flow. As the gap flows intersect the valley flow (Figure 4.4), $\partial v/\partial x > 0$ and $\partial u/\partial y < 0$ resulting in the development of positive vorticity, $\omega_z > \partial v/\partial x - \partial u/\partial y$ acting to deflect the valley flow towards the eastern slope of Granite Mountain.
Figure 4.2: DPG basin tower temperature and velocity data. $T$ and $U$ is measured at 2 - 5 m depending on the tower configuration. Vectors are colored by $\Delta T$ compared to the previous 15 minutes. (A) 3:30 UTC (21:30 MDT), flow drains out of the Dugway Basin through the small and big gap. (B) 3:45 UTC (21:45 MDT), flow through the big gap changes direction as now the valley flow from the southwest is pushing through.
Figure 4.3: DPG basin tower temperature and velocity data. $T'$ and $U$ is measured at 2 - 5 m depending on the tower configuration. Vectors are colored by $\Delta T'$ compared to the previous 15 minutes. (A) 4:00 UTC (22:00 MDT), downslope flow continues to drain into the valley below while the direction of flow has shifted at the small gap. (B) 4:15 UTC (22:15 MDT), valley flow flows past Granite Mountain, cutting off downslope flow. Note the flow converging at the small gap.
Figure 4.4: 4:15 UTC (22:15 MDT). As the gap flows intersect the valley flow, $\partial v / \partial x > 0$ and $\partial u / \partial y < 0$ resulting in the development of positive vorticity, $\omega_z > \partial v / \partial x - \partial u / \partial y$ acting to deflect the valley flow towards the eastern slope of Granite Mountain.

The descending downslope flow (Figure 4.5A) is much warmer at 22°C than the underlying valley flow, 12°C (Figure 4.5D). This strong temperature gradient is driven by differential cooling of the contrasting land surface cover of the mountain vs. the dry sparsely vegetated soil of the valley. The rocky surface of Granite Mountain absorbs and stores an immense amount of radiation during the day, analogous to an urban heat island (UHI). This differential cooling and mountain heat island (MHI) affects the set up of stable stratification within the Dugway basin, which strengthens throughout the evening. Stratification strengthens in the basin with the cold pooling valley flow forming the base layer followed by the flow through the big gap (Figure 4.5C), then the mixed
flow emanating from the small gap (Figure 4.5B) and topped by the downslope flow of Granite Mountain (Figure 4.5A).

Figure 4.5: Temperature in the Dugway basin during the development of the stable stratification. (A) ES2 tower as part of EFS-Slope site. (B) Small gap tower predominantly influenced by downslope flow off of the southern flank of Granite Mountain. (C) Big gap tower exposed to the valley to the southwest of the Dugway basin. (D) SAMS 29 tower, the closest SAMS site to the EFS-Slope site within the valley.
4.4 Initial Interaction of Flows

As the deflected valley flow continues to pour into the basin, it builds up in depth, pushing towards the slope (Figure 4.6, 4:30 UTC (22:30 MDT)). When the valley flow meets the downslope flow, a collision occurs, disrupting the latter (Figure 4.7A, 4:45 UTC (22:45 MDT)) and sending a pressure signal through the valley. This signal helps the flow to readjust, deflecting to the north. After the initial collision, the valley flow continues to push forward farther upslope, thus obliterating the downslope flow (Figure 4.7B, 5:00 UTC (23:00 MDT)). This upslope movement is a result of the along-slope flow momentum of the valley flow.

Figure 4.6: DPG basin tower temperature and velocity data. $T$ and $U$ is measured at 2 - 5 m depending on the tower configuration. Vectors are colored by $\Delta T$ compared to the previous 15 minutes. 4:30 UTC (22:30 MDT), valley flow has penetrated the lower extent of the slope flow, beginning to undercut the slope flow. Change in direction seen as well as a reduction in temperature as colder valley flow rushes in.
Figure 4.7: DPG basin tower temperature and velocity data. \( T \) and \( U \) is measured at 2 - 5 m depending on the tower configuration. Vectors are colored by \( \Delta T \) compared to the previous 15 minutes. (A) 4:45 UTC (22:45 MDT), collision has occurred. (B) 5:00 UTC (23:00 MDT), valley flow propagates farther upslope.

The collision event in Figure 4.7A is captured by the Doppler LiDAR at the EFS-Slope site, as shown in Figure 4.8. The first image is 4:41 UTC (22:41 MDT), while each
successive image is 15 min later. The early stage of the collision is marked by a thickening of the katabatic flow layer as the much denser valley flow collides with the downslope flow at the base of the slope (Figure 4.8A, 4:41 UTC (22:41 MDT)). A short time later the valley flow pushes up the slope, undercutting the downslope flow (Figure 4.8B, 4:51 UTC (22:51 MDT)). At the interface of the two colliding flows, the flow overturns rapidly, with enhanced mixing and turbulence. Still carrying a component of southeasterly momentum, the valley flow continues up the slope as a front as the downslope flow builds up in depth, flowing overtop (Figure 4.8C, 5:11 UTC (23:11 MDT)).
Figure 4.8: Time series of UU LiDAR data located at the EFS-Slope site near ES2. LiDAR scans oriented in azimuth of 270° for the left panel and 100° for the right panel. This event is the initial collision between the downslope (red) and valley flow (blue). (A) 4:41 UTC (22:41 MDT), the downslope layer thickens as the flows meet. (B) 4:56 UTC (22:56 MDT), the valley flow undercuts the downslope flow pushing up the slope. (C) 5:11 UTC (23:11 MDT), the bulk of the valley flow continues up the slope, undercutting the downslope flow while near the surface a thin layer of valley flow beings to descend the slope.

ES2 tower data near the merger of the slope and valley provides more detailed data about the characteristics of the collision and is presented in Figure 4.9 and 4.10.

When the valley flow pushes past the ES2 tower, a sharp gradient in temperature is recorded throughout all of the tower levels, (Figure 4.9A) dropping 5.9°C in the 16.8
minutes the collision lasts. During this time, within the near-surface layer measured by
the tower, the downslope flow formerly flowing at 3.2 m/s in a direction of 257° is
obliterated, as the valley flow overtakes the slope flow, first approaching from the east
at 1.8 m/s and then from the south (Figure 4.9B,C). At the instant of the collision, a
strong vertical velocity is generated (Figure 4.9D) as the downslope flow is violently
lifted, overturned, and subjected to turbulent mixing as indicated by spikes in TKE
(Figure 4.9E). Here, the TKE is defined as,

\[ TKE = \sigma^2 = \frac{1}{2} \left( u'^2 + v'^2 + w'^2 \right). \]  (4.1)

The collision contributes vigorously to the fluxes of momentum (Figure 4.10A) and
temperature (Figure 4.10B), creating turbulent drag force at the interface as the
currents move past each. Occurring simultaneously is the vertical transport of warm
downslope flow as a result of advection and turbulent mixing.
Figure 4.9: Characteristics of the slope and valley flow collision. Legend displayed in upper left of the figures. One minute was used for the averaging period. (A) Thermocouple temperature. (B) Sonic wind direction. (C) Sonic velocity transformed in a rotated coordinate system aligned with the mean flow. (D) Vertical velocity. (E) Turbulent kinetic energy.
Figure 4.10: Characteristics of the slope and valley flow collision. One minute was used for the averaging period. (A) Temperature flux. (B) Momentum flux.

4.5 Collision Events

After the initial collision, the valley current continues to push up the slope until its along-slope momentum is lost due to buoyancy effects, whence it slides back down the slope toward the valley. The secondary collisions are formed (Figure 4.11 A-C) as the descending flow meets the newly approaching valley current moving up the slope and lingering downslope flow. These secondary collisions can be quite strong, nearing the impact of the primary event as in the example shown; here the TKE levels (Figure 4.14, identified by the arrow) rival the primary collision and are higher than any other observed collisions. Other occurrences of secondary collisions can be identified as continual spikes in the vertical velocity (Figure 4.12), high TKE and fluxes (Figure 4.15 and 4.16) and as ripples in the temperature (Figure 4.13). The secondary events thereafter create sloshing or seiching within the Dugway basin, which have been observed with the IR camera as cyclic fluctuations in the surface temperature.
Figure 4.11: Time series of UU LiDAR data located at the EFS-Slope site. LiDAR scans oriented in azimuth of 270° for the left panel and 100° for the right panel. This event is a secondary collision between the downslope (red) and valley flow (blue). (A) 5:41 UTC (23:41 MDT), a very weak thin layer of downslope flow is present. (B) 5:56 UTC (23:56 MDT), shortly after the moment of the secondary collision the surge of turbulence is observed in Figure 4.14. (C) 6:11 UTC (0:11 MDT), piled up downslope flow, flows over the undercutting valley current.
Figure 4.12: ES2 vertical velocity. After the initial collision secondary collisions occur, identified by strong velocity fluctuations as the downslope layer is lifted by the undercutting valley flow. Primary and secondary collisions are labeled as P and S.

Throughout the evening the pattern repeats itself with the secondary collisions, which wane eventually, followed by the reemergence of downslope flow. Shortly thereafter another valley current collides with the downslope flow, forming a primary collision, and the process is repeated. During the IOP 2 nighttime period, three episodic periods were identified by an initial sharp drop in temperature as the valley current pushes past the ES2 tower followed by an increase in vertical velocity as the downslope flow is rapidly lifted by the incoming valley flow, spikes in TKE up to ten times the background levels, increased drag as the layers flow past each other and vigorous fluxes of temperature as heat is vertically transported. Secondary collisions and the reemergence of downslope flow follow. The primary collisions and collision periods are shown in Figure 4.13-4.17 as red dashed lines.
Figure 4.13: ES2 temperature. Three separate collision periods ($P_1$, $P_2$, $P_3$) are identified by an initial sharp drop in temperature as the valley current pushes past the tower followed by secondary collisions and the reemergence of downslope flow (periods of warming).

Figure 4.14: ES2 turbulent kinetic energy showing discernible spikes when collisions occur. The arrow points to the time of the secondary collision presented in the Figure 4.11B LiDAR scan.
Figure 4.15: ES2 momentum flux showing discernible spikes when collisions occur. The arrow points to the time of the secondary collision presented in the Figure 4.11B LiDAR scan.

Figure 4.16: ES2 temperature flux showing discernible spikes when collisions occur. The arrow points to the time of the secondary collision presented in the Figure 4.11B LiDAR scan.
One of the most important parameters that determines the stability of stratified shear flows and the turbulence therein is the gradient Richardson number, $Ri_g$, defined as,

$$Ri_g = \frac{N^2}{\left(\frac{\partial U}{\partial z}\right)^2 + \left(\frac{\partial V}{\partial z}\right)^2}$$  \hspace{1cm} (4.2)

where $U$ and $V$ are the horizontal velocities, $N^2 = g\alpha d\theta/dz$ is the background Brûnt-Väisälä buoyancy frequency and $\theta$ is the potential temperature. According to the linear stability theory of Miles (1961) and Howard (1961), when $Ri_g$ drops below the critical value of $Ri_{gc} = 0.25$, the flow can become unstable. For non-linear disturbances, according to the experiments of Strang and Fernando (2001A,B), strong mixing events can persist up to $Ri_g \approx 1$. For $Ri_g < 1$, the major mixing mechanism was the Kelvin-Helmholtz instability. Non-linear waves tend to generate and resonate with K-H billows when $Ri_g \sim 1$, thus producing the most efficient turbulent mixing. At still larger values of $Ri_g$, the K-H billowing and wave motions subside, paving the way to less intense Hölmböe waves. In the case of two colliding flows, $Ri_g$ however is not well correlated to the collision events as shown in Figure 4.17, which may be due to the dense valley flow undercutting the warmer downslope flow that results in strong stability, yielding $Ri_g > 1$. Nevertheless, collisions lead to intense mixing due to strong vertical and slanted shear, a topic that has not been subjected to theoretical or laboratory investigations.
Figure 4.17: ES2 tower gradient Richardson, \( Ri_g \) number calculated using the layer between 4 and 28 m with a red solid line indicating the critical value of 0.25. The non-linear theory value of 1 is also shown, Strang and Fernando (2001A,B). The times for the collision events are also indicated.

As \( Ri_g \) number is not well correlated and not providing much information about the collisions, a new approach was used to analyze the collision events. The events were first separated into primary and secondary events, and the angle between colliding flows was considered as a variable. Three possible collision types were identified:

1. Perpendicular to the slope collision (Figure 4.18A).
2. Along slope collision (Figure 4.18B).
3. Merging collision (Figure 4.18C).
Figure 4.18: Collision types. (A) Perpendicular to the slope collision where the angle between the downslope and valley flow is $\beta$. Each flow is transformed into components where $U$ is parallel, where $\theta = (180^\circ - \beta) / 2$. (B) Along slope collision. Again the angle between the downslope and valley flow is $\beta$ and $\theta = (180^\circ - \beta) / 2$. (C) Merging collision in which the downslope and valley flow merge together.
The results of the analysis are shown in Figure 4.19, in which $w'T'$ of each event is vertically averaged, except the measurements at 0.5 m. The onset and end of the collision were identified by the time at which the averaged $w'T'$ reached 10% of the maximum.

Figure 4.19: ES2 vertically averaged $w'T'$ of the collisions separated based on primary or secondary and collision type. The onset and the end of each collision were identified by the time the vertically averaged $w'T'$ reached 10% of the maximum.

Of particular interest were the collisions in which the colliding flows were nearly parallel to one another and the collision occurred perpendicular to the slope (Figure 4.18 A), as this resembles the collision of two gravity currents. A schematic representing the colliding flows is shown in Figure 4.20 where $U_1$, $\rho_1$, and $h_1$ are the downslope velocity, density and depth and $U_2$, $\rho_2$, and $h_2$ are the valley flow velocity, density and
depth, where $U_1$ and $U_2$ are transformed into the collision coordinate system where $U$ of each flow is parallel (see Figure 4.18 A,B).

Dimensional analysis reveals that the total buoyancy flux for each such event is given by, $\overline{\rho w'} = f(\Delta b, \Delta U, \Delta h^*)$, where $\overline{\rho w'} = \frac{1}{2} \int_0^t \overline{\rho w'} dt$, $\Delta b = g(\rho_2 - \rho_1)/\rho_0$ is the buoyancy jump, $h^* = (h_1 + h_2)/2$ and $\Delta U = (U_1 + U_2)$. The resulting form is,

$$\frac{\overline{\rho w'}}{\Delta b \Delta U} = f\left(\frac{\Delta h^*}{\Delta U^2}\right),$$

where the collision Richardson number is defined as,

$$Ri_c = \frac{\Delta h^*}{\Delta U^2}. \quad (4.4)$$

![Schematic representation of colliding flows. $U_1, \rho_1$ and $h_1$ are the velocity, density and depth of the downslope flow while $U_2, \rho_2$ and $h_2$ are the velocity, density and depth, respectively, of the valley flow.]

Table 4.1 contains details of the IOP 2 collision events. As mentioned previously, the onset and the end of the collision was identified by the time the vertically averaged $\overline{w'T'}$ reached 10% of the maximum. Doppler LiDAR was used to estimate the depth of
the currents. The ES2 tower sonics at 4 and 10 m were vertically averaged to determine the direction and velocity of the downslope flow while thermocouples were used for the temperature. The SAMS 29 tower (Figure 4.7A) sonics at 2 and 10 m located within the valley to the northeast of the EFS-Slope site were vertically averaged to determine the velocity and direction of the valley flow. In all cases the valley flow temperature was determined with HMP45 temperature/humidity probes located at 2 m on the SAMS 29 tower. The collision Richardson number (4.4) was calculated in two methods; $Ri_c^U$ considered only $\Delta U^2$ while $Ri_c^{UV}$ considered $\Delta U^2 + \Delta V^2$. The collision events in which the flows merged together are listed in the table; however $h^*$, $\Delta U$, $\Delta V$ and the collision Richardson numbers were not obtained because the underlying physics of these cases is different.
### TABLE 4.1

#### COLLISION EVENTS

<table>
<thead>
<tr>
<th>Event</th>
<th>$t_i$ (UTC)</th>
<th>$\Delta t$ (min)</th>
<th>$h^*$ (m)</th>
<th>$\Delta b$ (m/s²)</th>
<th>$\Delta U$ (m/s)</th>
<th>$\Delta V$ (m/s)</th>
<th>$R_{i_c}^u$</th>
<th>$R_{i_c}^{uv}$</th>
<th>$\bar{b}'w'$ (m²/s³)</th>
</tr>
</thead>
<tbody>
<tr>
<td>$P_1^p$</td>
<td>4:40</td>
<td>12</td>
<td>22.0</td>
<td>0.307</td>
<td>4.61</td>
<td>1.73</td>
<td>0.32</td>
<td>0.28</td>
<td>5.27e-5</td>
</tr>
<tr>
<td>$P_2^p$</td>
<td>9:19</td>
<td>9</td>
<td>21.4</td>
<td>0.367</td>
<td>2.45</td>
<td>0.80</td>
<td>1.26</td>
<td>1.15</td>
<td>2.57e-5</td>
</tr>
<tr>
<td>$P_3^a$</td>
<td>10:53</td>
<td>11</td>
<td>24.0</td>
<td>0.356</td>
<td>1.76</td>
<td>0.67</td>
<td>2.76</td>
<td>2.41</td>
<td>3.64e-5</td>
</tr>
<tr>
<td>$S_1^m$</td>
<td>5:49</td>
<td>7</td>
<td></td>
<td>0.325</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>4.64e-5</td>
</tr>
<tr>
<td>$S_2^m$</td>
<td>6:00</td>
<td>10</td>
<td></td>
<td>0.286</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1.55e-5</td>
</tr>
<tr>
<td>$S_3^m$</td>
<td>6:32</td>
<td>9</td>
<td></td>
<td>0.190</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1.54e-5</td>
</tr>
<tr>
<td>$S_4^m$</td>
<td>9:54</td>
<td>7</td>
<td></td>
<td>0.210</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1.21e-5</td>
</tr>
<tr>
<td>$S_5^p$</td>
<td>11:45</td>
<td>3</td>
<td>22.0</td>
<td>0.375</td>
<td>2.71</td>
<td>0.76</td>
<td>1.12</td>
<td>1.04</td>
<td>1.44e-4</td>
</tr>
</tbody>
</table>

$^p$ = Perpendicular to slope collision.

$a$ = Along slope collision.

$m$ = Merging collision.

$u$ = Calculated using only the colliding velocity component $\Delta U^2$.

$uv$ = Calculated using $\Delta U^2$ and $\Delta V^2$ velocity.

The results of Equation 4.3 are plotted in Figure 4.21 for collision events with nearly parallel colliding flows (merging collisions are not considered). Interestingly, the dimensionless buoyancy flux, $\left|\bar{b}'w'\right|/\Delta b\Delta U$ is nearly constant for primary collisions occurring perpendicular to the slope. The flux for along slope collisions is slightly higher, however, and this case differs from the others in terms of the angle of the collision.

Note the high flux of the secondary event. More quiescent IOPs need to be studied to determine if $\left|\bar{b}'w'\right|/\Delta b\Delta U$ is constant for the primary events as $R_{i_c}$ increases and if the secondary collisions generate more flux or this particular case is an outlying event.
Figure 4.21: ES2 vertically averaged, absolute value of the normalized buoyancy fluxes vs. the collision Richardson number \((Ri_c)\), (4.4), for the collisions in which the colliding flows were nearly parallel. \(^1 = Ri_c^U\) considered only \(\Delta U^2\), \(^2 = Ri_c^{UV}\) considered \(\Delta U^2 + \Delta V^2\).

Also of interest was the collision decay time,

\[
t \sim \left(\frac{\Delta b}{h^*}\right)^{-1/2},
\]

which is plotted in Figure 4.22. When taking into account the change in buoyancy and the averaged heights of the colliding flows, the primary collision decay time, the time for the flux to be negligibly small, \((\Delta b / h^*)^{-1/2}\) is nearly constant. Only one of the secondary events met the non-merging collision criteria so it is not possible to say whether this is also true for the secondary events.
Figure 4.22: ES2 vertically averaged normalized buoyancy fluxes vs. the normalized decay time, $(\Delta b/h^*)^{1/2} t$ for the non-merging collisions.

4.6 Adjustments in the Valley

During the collision process, both the slope and valley flows adjust, and this adjustment is of primary importance to inhabitants in the valley. In the pre-collision period (Figure 4.23A), the downslope flow drains into the basin, while the valley flow is unaffected due to its inertia or inadequate development of flow. Shortly thereafter, the gap flows intersect the valley flow and vorticity develops which deflects the valley flow towards the slope (Figure 4.23B). Upon collision of the currents, the valley flow undercuts the downslope flow, pushing up the slope and flowing to either side of the slope. The impact of the collision sends pressure waves, flushing the basin, allowing a thin layer of dense valley flow remnant to begin to drain out into the valley while the rest continues to push upslope (Figure 4.23C). The valley flow remnant that pushed up
the slope then loses momentum due to buoyancy effects and slides down the slope with the warmer mountain downslope flow, flowing over top, both draining out into the valley. Meanwhile the valley flow recovers from the initial event returning to the original flow direction while the downslope flow continues to flow out into the basin. Vorticity again develops as the gap flows intersect the valley flow and the valley flow deflects towards the slope, forming a secondary collision, completing the cycle (Figure 4.23 D). Figure 4.24 shows the wind direction and velocity at the EFS-sagebrush tower, which is embedded in the valley flow. Dashed lines are correlated to the collision cycle described in Figure 4.23 A-D, showing how the valley flow adjusts during the first cycle.
Figure 4.23: Schematic of flow adjustments in the valley. (A) The valley flow flows past Granite Mountain to the northwest, allowing the downslope flow to fully develop and flow out into the basin. (B) As the gap flow intersects the valley flow, vorticity develops, deflecting the valley flow towards the slope, initiating the collision. As contact is made, the valley flow undercuts the downslope flow, flowing upslope while the downslope flow adjusts and flows overtop. (C) After the collision the valley flow is flushed out of the basin. Some of the valley flow continues upslope while the rest begins to drain downslope. Meanwhile the mountain downslope flow flows overtop. (D) After the valley flow reestablishes, vorticity again develops and the valley flow deflects towards the mountain, colliding with the downslope flow, forming a secondary collision. After this, the pattern repeats.
Throughout the night of IOP 2, three collision periods existed in which the valley flow continues to adjust, flushing cold dense air out of the basin. The description above is simplified to give a general description. In some situations, it has been identified that the direction of basin flushing is shifted by 180° to what is shown in Figure 4.23C wherein the flow exits the big gap. Figure 4.25 is a plot of the wind direction and velocity at the big gap tower, showing flow reversal though the gap during the second collision period between 8:00 and 11:00 UTC (2:00 – 5:00 MDT). Shortly thereafter the flow reverses and the third collision period characteristics are more similar to the first collision period.
Figure 4.25: Big gap tower data showing flow reversal through the big gap after the first collision period. At the onset of the third collision period the flow reverses again flowing into the basin. Wind direction is plotted as blue circles and the velocity is the green line.

4.7 Conclusions

In the planning stages of the MATERHORN project, it was thought that due to the isolation of Granite Mountain and flat topography in the immediate vicinity of the mountain, the thermal circulations (slope flows) would be the dominant flow pattern, extending far out into the Dugway basin during the nighttime period. It is now clear that Granite Mountain does not act alone and that the downslope flows do not simply drain into the basin unmodified.

The flow in the Dugway basin is extremely complex, with valley flows to either side of Granite Mountain competing to flow through the Sapphire Mountain gaps. There is a competition driven by differential cooling due to land surface contrast and
surrounding mountain drainage that determines the direction of flow through the gaps. Eventually the flow leaving the basin through the Sapphire gaps is overtaken by incoming flow, temporarily closing off the exit while to the north, rapidly cooling valley flow pools into the basin, forming a barrier slowly moving to the northwest. During this time period, the flow continues to descend the slopes of Granite Mountain. The downslope flow, however, is much warmer due to mountain heat island (MHI) effects, which, together with the undercutting valley flow sets up a strong stable stratification in the basin.

Eventually the gap flows intersect the valley flow and vorticity develops which deflects the valley flow towards the slope, thus creating a violent collision of two currents marked by overturning, increased levels of turbulence, rapid temperature drop along the slope and enhanced mixing and spikes in buoyancy and momentum fluxes. A series of secondary collisions then follow as the once valley flow begins to descend the slope after the upslope momentum is lost. This impacts the newly emerging valley flow that pushes up the slope with less dense remnants of downslope flow riding atop characterized by rapid fluctuations of velocity and TKE with overlapping events. The secondary impacts send waves of disturbance throughout the basin, as if the basin pool is seiching while allowing the downslope flow to reemerge.

During the evening hours of IOP 2 of MATERORN-X-1, three main collision periods were identified, each with a primary collision followed by a series of weaker secondary collisions. Eventually, the secondary collisions wane and downslope flow
reeemerges for a short period to again meet a valley flow pushing up the slope, repeating the cycle. Of these collisions, three possible collision types were identified:

1. Perpendicular to the slope collision (Figure 4.18A).

2. Along slope collision (Figure 4.18B).

3. Merging collision (Figure 4.18C).

After distinguishing the collisions between primary and secondary, the collision events were further separated based on the collision type. For each type, $\overline{w' T'}$ of each event was vertically averaged, excluding the 0.5 m flux measurement, to determine the onset and end of collisions based on the time at which the averaged $\overline{w' T'}$ reached 10% of the maximum. The absolute values of the collision fluxes (omitting the merging collisions) were then integrated to determine the total buoyancy flux for each event.

An analysis showed that the normalized flux can be given as $\overline{b' w'}/\Delta b \Delta U = f(\Delta b h^*/\Delta U^2)$ where $R_i = \Delta b h^*/\Delta U^2$ is defined as the collision Richardson number. Over the examined range of collision Richardson numbers, the dimensionless buoyancy flux, $\overline{b' w'}/\Delta b \Delta U$ is nearly constant for the primary collisions occurring perpendicular to the slope. The along slope collision flux is slightly higher, however, and this case differs from the other in terms of the angle of the collision.

Also of interest was the collision decay time, $t \sim (\Delta b / h^*)^{-1/2}$, which takes into account the change in buoyancy and the averaged heights of the colliding flows. Applied to the primary collisions, the decay time, $(\Delta b / h^*)^{-1/2}$ is nearly constant for primary collisions.
These interactions within thermal circulation generate an intriguing set of small scale processes that contribute vigorously to sub-grid heat and momentum transfer. These processes include the collision of gravity currents, formation of intense turbulent regions, intrusions and instabilities. WRF and other mesoscale models do not account for such processes, and hence their incorporation is crucial in modeling mountain terrain winds.
CHAPTER 5:
MEASUREMENT OF TURBULENCE IN KATABATIC FLOWS

5.1 Introduction

Fine-scale turbulence in complex terrain can be originated via different mechanisms. When the thermal circulation is significant, turbulence is mainly generated by the interaction of mean winds with the topography, where the main areas of turbulence generation are the shear layers at the top and the sides of the mountains, at the boundaries of katabatic flows and in-between interleaving gravity currents. Other mechanisms are flow collisions and convection (in the case of unstable stratification). In the case of mean flow past mountains, Kármán vortex shedding is prevalent, which may modulate the turbulence generation. Flow over two-dimensional hills has been studied by Jackson & Hunt (1975), where they used the linearized theory to study the distortion of turbulence surrounding the hill. When the stratification is present, a part of the flow occurs above the mountain while the rest around the mountain, the separation between the two types occurring at the dividing streamlines (Sheppard 1956). Turbulence generation in this case has not been studied in detail, especially in the area of dividing streamline where directional shear is pronounced due to horizontal and vertical flows, below and above the dividing streamlines, respectively.
In deducing turbulence, in particular the TKE dissipation, in katabatic flows and other natural shear layers, it has been customary to use sonic anemometers to measure turbulent velocity and temperature fluctuations and then use the Kolmogorov inertial sub-range theory (Kaimal and Finnigan 1994; Grachev et al. 2013). However, the Kolmogorov theory is valid only for flows in equilibrium, where energy in the large scales has cascaded down to the small scales without accumulating in the intermediate scales. As a precursor to future stratified turbulence studies within katabatic flows as well as to obtain insights to the past studies on turbulence in katabatic flows, we have developed and deployed a novel turbulence measurement system that can capture the entire range of scales of turbulence in atmospheric flows - from energy containing scales to dissipating Kolmogorov scales. Using this system, the nature of turbulence in developing and developed katabatic flows could be determined, and the applicability of Kolmogorov theory to the two types of flows could be inferred. This chapter is dedicated to describe the development of the measurement system; the calibration as well as application to measure turbulence in katabatic flows. It is concluded that inference of turbulent energy budgets using sonic anemometers should be done with caution and that if used they may lead to misleading results with regard to turbulent energy dissipation.

While the said measurement system was particularly developed to study stratified turbulence in katabatic flows, fine scale turbulence measurements showed that there is a critical necessity for understanding the energetics of developing and developed katabatic flows, especially in view of the limitations of sonic anemometers in
the studies of TKE dissipation. Therefore, the stratified turbulence work was left for future studies, while the present work dedicated for identifying dynamical limitations of measuring energetics of katabatic flows using low resolution measurements such as sonic anemometers and LiDARs.

5.2 Combo Probe Development and Neural Network Explanation

Although the integral quantities of atmospheric turbulence are conveniently and commonly measured using sonic anemometers (sonic) (Figure 5.1A), their low space-time resolution [spatial (~10 cm) and temporal (~16 Hz)] limits the obtainable fine-scale turbulence information. The main frequency response limitation is imposed by the spatial separation between the paths, which introduces distortions at higher frequencies (Kaimal and Finnigen, 1994). The highest frequency that can be measured without distortion is,

\[ f_{\text{max}} = \frac{U}{2\pi L}, \]  

where \( L \) is the distance between two ultrasonic transducers and \( U \) the mean velocity.

On the other end of the anemometry spectrum is hot-film/wires (Figure 5.1 B, C), which have excellent spatial and temporal resolution and are perfectly capable of obtaining relevant fine-scale variables such as the kinetic energy dissipation. This method, however, remains a challenge outside of the laboratory due to two main difficulties. The first is the mean wind variability, which causes violation of the requirement that the mean winds have a specific alignment with the hot-film/wire
probe. To circumvent this problem, a combo of co-located sonic and hot-film anemometers, with the former measuring the mean winds and aligning the latter in the appropriate wind direction via an automated platform, was successfully designed and implemented.

![Figure 5.1: Anemometers. (A) RM Young 81000 Sonic anemometer. (B) X hot-film able to measure two orthogonal velocity components. (C) Three dimensional hot-film. (A courtesy of L. Leo and B, C from TSI.com)](image)

The second difficulty is the necessity of frequent and laborious calibrations akin to hot-film anemometry, leading to logistical difficulties during outdoor (field) measurements. Thus a reliable calibration procedure is necessary to convert the voltage to velocity. Such a procedure becomes more complicated for two- or three-dimensional probes wherein the wires are inclined to the flow direction. This is addressed by employing sonic measurements to calibrate the hot-films in the same combo, with the output (velocity) / input (voltage) transfer function for the hot-film is derived using a Neural Network (NN) model. The NN is trained using low pass filtered hot-film and sonic data taken in-situ.
5.2.1 Low Pass Filtering

Each data set consists of hot-film voltages for each wire and \((u, v, w, T)\) measured by the co-located sonic. All data is simultaneously recorded at 2 kHz for a duration of five minutes, resulting in 600,000 data points. Based on the maximum sonic sampling frequency of 32 Hz, according to (5.1), the highest frequency measurable by the sonic without distortion is 2.1 Hz where \(L = 0.15\) m, assuming a mean wind velocity of \(U = 2\) m/s. Clearly, the narrow bandwidth of the sonic can be perceived as a low-pass filter. To keep the correspondence between the spectral composition of sonic and hot-film signals, the hot-film signal is also filtered using the sonic cutoff frequency applying a block low-pass filter averaging every 600 samples of data to produce a single data sample. 600,000 samples were reduced to 1000 samples, which by the Nyquist theorem mean the set contains data up to a maximum frequency of:

\[
f_{\text{cutoff}} = \frac{1000}{2 \cdot 300s} = 1.67 \text{ Hz.}
\]  

Although the resolutions of the two instruments are significantly different, at low frequencies, the velocity field sensed by both probes is the same, and hence both ought to yield the same velocity field upon low-pass filtering.

5.2.2 Neural Network Method for Interpolation/Extrapolation

A NN is a computational device that produces appropriate outputs from inputs, based on a selected architecture and subsequent training to perform intended tasks. If the general relationships are mapped as functions from \(R^n\) to \(R^m\), then a NN with
smooth activation functions can approximate continuous functions with compact support (i.e. all continuous functions whose domains are closed and bounded in $R^n$).

This property is a result of the Stone-Weierstrass theorem, which stipulates that all continuous functions with compact support can be approximated to any degree of accuracy by a neural network of one hidden layer with a sigmoid or hyperbolic tangent activation function (Nguyen et al. 2003; Haykin 1998). NN of this class (Figure 5.2) are often referred to as Multilayer Perceptrons (MLPs) and have the following characteristics (Haykin, 1998):

1. Each neuron in the internal ("hidden") layers of the network includes a smooth nonlinearity, commonly sigmoidal nonlinearity as hyperbolic tangent function or logistic function of the form $y = 1/1 + \exp(-x)$ called the activation function.

2. The network contains one or more layers of hidden neurons which enable the network to learn, i.e. to adjust to represent complex mappings.

3. The network exhibits a high degree of connectivity.

The input layer units are passive, simply receiving a single value input, and duplicating the value to their multiple outputs creating a fully connected structure which contrasts the active hidden layer and output units which modify the data. The input values entering a hidden unit are multiplied by weights and then added to produce a single number. Before leaving the unit, this number is passed through a sigmoidal function limiting the unit’s output to asymptotical values (0 and 1 for logistic function, -1 and 1 for hyperbolic tangential). The following description was adopted from (Smith, 1997). The output from $n$-th hidden layer unit is given by,
\[ y_n = f_n^{\text{hidden}}(w_{(x1)n}x_1 + w_{(x2)n}x_2 + \ldots + w_{(xk)n}x_k + w_{(k+1)n} \cdot 1), \]  

(5.3)

where \( f_n^{\text{hidden}} \) is a sigmoidal function of choice, \( w_{(x1)n} \) is the weight factor of the \( i \)-th input of \( k \) and \( w_{(k+1)n} \) is the weight of the bias term of the unit.

![Diagram of the generated MLP neural network](image)

Figure 5.2: The structure of the generated MLP neural network with one hidden layer containing ten hidden units, four input layers and three outputs. From Kit et al. 2010.

In the output layer it is common to use linear units for regression tasks. Their output is similar to the one described in the hidden layer, except the activation function of the unit is linear sum,

\[ z_n = w_{(y1)m}y_1 + w_{(y2)m}y_2 + \ldots + w_{(ym)m}y_m + w_{(m+1)m} \cdot 1, \]  

(5.4)

where \( w_{(y1)m} \) is the weight factor of the \( i \)-th hidden layer unit output of \( m \) and \( w_{(m+1)m} \) is the bias term and \( z_n \) is the \( n \)-th network output.
In current work a three-layer MLP is employed with one hidden layer of hyperbolic tangential functions and an output layer of linear functions.

5.2.3 Training the Neural Network

The low pass filtered data is used as a training set to train the NN (Haykin 1998) for use for hot-film calibration. Training the NN means adjusting the weights and biases of the units of the network so that the network will reflect a desired mapping between its inputs and outputs. Since regression is the goal, the network should accurately reproduce the function it tries to reflect with minimal error, so that once the network is fed with an input sample, it will produce a result close to the output sample. A popular training method for MLP NNs is the back propagation algorithm.

Consider a feed-forward network with \( n \) input and \( m \) output units and a training set \( \{(x_1, t_1), \ldots, (x_p, t_p)\} \) consisting of \( p \) ordered pairs of \( n \) and \( m \) dimensional vectors, which are called the input and output samples. Since the activation functions are sigmoidal, they are differentiable. The initial weights and biases of the units are set at random. When the input pattern \( x_i \) from the training set is presented to this network, it produces an output \( z_i \), which is generally different from the target \( t_i \). The purpose is to make \( z_i \) and \( t_i \) (for \( i = 1, \ldots, p \)) as close as possible to each other. Specifically, the error can be minimized,

\[
E = \frac{1}{2} \sum_{i=1}^{p} (z_i - t_i)^2,
\]

by adjusting the network weights.
After minimizing this function for the training set, new unknown input patterns are presented to the network and we expect the network to interpolate.

The back propagation algorithm is used to find a local minimum of the error function. The network is initialized with randomly chosen weights. The gradient of the error function is computed and used to correct the initial weights. The weights in the network are the only parameters that can be modified to make the quadratic error \( E \) as low as possible. Because \( E \) is calculated by the extended network exclusively through composition of the activation functions (which are either sigmoidal or linear), it is a continuous and differentiable function of the \( l \) weights \( w_1, w_2, \ldots, w_l \) in the network. Thus \( E \) can be minimized by using an iterative process of gradient descent, for which the gradient needs to be calculated,

\[
\nabla E = \left( \frac{\partial E}{\partial w_1}, \frac{\partial E}{\partial w_2}, \ldots, \frac{\partial E}{\partial w_l} \right),
\]

(5.6)

each weight is updated using the increment,

\[
\Delta w_k = -\gamma \frac{\partial E}{\partial w_k} \quad \text{for } k = 1, \ldots, l,
\]

(5.7)

where \( \gamma \) represents a learning constant, i.e., a proportionality parameter which defines the step length of each iteration in the negative gradient direction.

The whole training problem now reduces to calculating the gradient of a network function with respect to its weights. Once a method to compute this gradient is established, the network weights can be iteratively adjusted. Using this method it is possible to minimize the error function, where \( \nabla E = 0 \).
5.2.4 NN Implementation

The Matlab Neural Network package is used for implementation of NNs. It contains the required data structures to host the network, randomization mechanism for initial network weights, the back propagation algorithm with an optimization for computing the derivatives and means to automatically simulate the Neural Network with arbitrary input. To reduce the dependence of final results on the randomized initial values of the network weights, several networks are trained and the averaged result among the nets is used.

5.3 Experimental Domain and Combo Probe Deployment

Of primary interest to the combo probe study were the stable stratified nocturnal downslope and down valley flows. For this reason the domain of the study was selected to include the East Slope of Granite Mountain featuring the heavily instrumented EFS-Slope site (Figure 2.7, 2.8), the portion of the GMAST instrumentation (Figure 2.4A) in the immediate vicinity and Doppler LiDARs (Figure 2.8B). The larger scale measurements (GMAST and LiDARs) were used to build up the mesoscale flow patterns during the periods of interest to aid in interpretation of the fine-scale combo probe results. A detailed scaled map of the multi-institutional equipment deployment during the first MATERHORN-X-1 campaign is shown in Figure 2.10. The domain of the combo probe study is encompassed by the upper left panel where the combo probes were located on the ES2 tower as indicated by a blue circle.
In the initial phase of the campaign, combo probes were embedded in the ABL surface layer at heights of 2 and 6 m above ground (Figure 5.3A) to capture the flow variability with height and the potential development of katabatic flows. In the field, the direction of atmospheric flows is variable and the mean flow direction must be followed. To do so, a specialized mount (Figure 5.3B) was constructed which rigidly held the sonic and stepper motor attached to the hot-film holder. The stepper motor could then rotate the hot-film probes via a feedback signal derived from the three velocity components measured by the sonic, which were interrogated at every 60 s interval. In so doing, the signals were averaged over the last 5 s of each 60 s record to allow sufficient time to filter out smaller scale fluctuations caused by the variability of wind direction. This allowed each combo to sweep through 120° of measurement between the sonic ears (Figure 5.4A). Due to the variability of the mean flow direction, in the second half of the campaign the 2 m combo probe was moved to the 6 m level and was rotated 180°, thus increasing the flow coverage to 240° (Figure 5.4B). A sophisticated LabVIEW program acquired the data at a sampling frequency of 2 kHz and provided the necessary feedback for probe rotation.
Figure 5.3: (A) ES2 tower with combo probes positioned at 2 and 6 m. Trailer positioned downwind out of the measurement range near the tower due to 25 m cable length limitations from the hot-film probes to the constant temperature anemometer (CTA) bridges. (B) Rigid combo mount including stepper motor and motor driver able to rotate the hot-film holder 120° between the sonic ears. (Approved by DPG)
Figure 5.4: Combo probe Setup. (A) During the initial setup period, two combo probes placed in the same orientation were able to measure 120°. The levels were 2 and 6 m. (B) During the second half of the campaign, the 2 m combo probe was moved to the 6 m level and was rotated 180° to increase the flow coverage to 240°.
The combo probe operation required a full suite of electronics (Figure 5.5) including power supplies for the sonics and stepper motors, constant temperature anemometer (CTA) bridges for the hot-films and A/D boards with PCs running LabVIEW to convert the incoming analog signals to digital signals for data acquisition and motor position feedback. All of these electronics were designed for laboratory use, making it impossible to store them in the outside field environment. Therefore, these instruments and processing units were housed in a trailer, which was positioned downwind outside of the combo probe measurement range but within 25 m of the hot-films due to cable length requirements between the probes and CTA bridges (Figure 5.3A).

Figure 5.5: Trailer contains the electronics needed for combo operations, including power supplies for sonic and motors, hot-film CTA bridges and A/D boards, with PCs running LabVIEW for data acquisition and motor feedback.

The field campaign gave us the unique opportunity to test the feasibility of multiple combo probe configurations. The previous work and underlying foundation of the combo probe method by Kit et al. (2010) used two orthogonally displaced X hot-film probes. During the deployment, the 2 m combo (Figure 5.6) featured a three-
dimensional hot-film probe collocated with two X hot-film probes, which allowed comparison of the performance of the two configurations. The potential benefits of the three-dimensional film include improved spatial resolution as well as a reduction of the needed CTA bridge channels.

Figure 5.6: The 2 m level combo probe featured two X hot-films and a three-dimensional hot-film to compare performance of the two combo probe configurations.

5.4 Data processing

Once collected, data was methodically processed to ensure proper training set selection and NN construction for calibration. The process included multiple steps as outlined below.

1. Selection of good data: Data is discarded if the hot-film probe was outside of the 120° of measurement. Each remaining data is analyzed. If the directional variability between the beginning and the end of a 60 s
record is less than 10° difference, the record is selected for data processing. This warrants the off-plane component to be small enough.

2. **Data conditioning:** A narrow band-cut-off filter is constructed for the hot-films based on user input, which filters out 60 Hz noise and its aliases. The good data meeting the criteria in step 1 is written to a separate file. Sonic data is low-pass filtered down to 16 Hz with a goal to reduce the noise at frequencies higher than Nyquist frequency and resampled at 2 kHz using the inverse Fourier transform. The narrow band-cut-off filter is applied to the hot-film data. At this stage the spectra of filtered sonic data, sampled at 32 Hz is created and plotted. In addition, statistics of velocity and angles for each unique minute in the data is compiled.

3. **Selection of data for NN training:** The compiled sonic statistics from step 2 are examined and graphics are generated allowing the selection of the desired five minutes for NN training and the time period of hot-film conversion from voltages to velocities.

4. **Sort data for NN construction and evaluation:** The data selected during step 3 is sorted, creating files containing the training set and the data minutes for conversion.

5. **NN construction and hot-film data conversion:** The sorted data in step 4 is used to construct the NN. To reduce the dependency on initial conditions, several networks with randomly selected initial conditions are trained and the average result among the nets is used. The constructed network is used to convert the hot-film voltages to velocities. Hot-film velocity statistics are compiled and the spectra and reconstructed hot-film velocities are plotted.

6. **Skewness of velocity derivatives:** As a means to check the quality of data, the skewness of the velocity derivative in the mean velocity direction is computed for each minute where,

$$
Sk = \left( \frac{\partial u}{\partial x} \right)^3 \left\{ \left( \frac{\partial u}{\partial x} \right)^2 \right\}^{3/2}.
$$

If $0.2 < Sk < 0.8$ the data record for this minute is selected for further processing and is used to calculate dissipation in the next step. If the skewness is outside of this range there is a high probability the signal is primarily noise. To convert the combo measurements to time instead of space, Taylor’s frozen turbulence hypothesis is used,
\[
\frac{\partial u}{\partial t} + U \frac{\partial u}{\partial x} = 0, 
\]

where it is assumed the turbulent patches are frozen as they pass the probe, carried with mean velocity \( U \) past the probe.

7. **Calculate dissipation**: If the data record for a specific minute passes the skewness criteria, the record is used to calculate the dissipation rate \( \varepsilon \) based on two expressions, which are equal for isotropic turbulence:

i. \[ \varepsilon = 15v \left( \frac{\partial u}{\partial x} \right)^2. \] (5.10)

ii. \[ \varepsilon = \frac{15}{2} v \left( \frac{\partial w}{\partial x} \right)^2. \] (5.11)

5.5 Measurements

In contrast to controlled laboratory conditions, not only the wind direction and speed vary significantly but also the diurnal thermal forcing that produces a convective boundary layer (CBL) during the day and stable boundary layer (SBL) at night. The former consists of turbulent convection that produces rising and falling thermals over the entire boundary layer and shear-produced (mechanical) turbulence near the ground. The latter, SBL, inhibits turbulence and confines it to intermittent patchy regions or maintains weak turbulence over the entire boundary layer (Monti et al. 2002; Pardyjak et al. 2002).

Of particular interested to the combo probe study were stable stratified downslope flows in which the ideal meteorological conditions are quiescent, high pressure background conditions that prevail in the absence of a synoptic flow. In addition, clear skies void are needed to ensure maximum radiative cooling at the surface.
to establish strong katabatic flow. In the MATERHORN-X-1 campaign, IOP 8 (Table 2.5) met all of these requirements, and was selected as the experimental period for the fine scale stratified turbulence study.

5.5.1 EFS-Slope Site Background Meteorology

During the evening of IOP 8, strong katabatic flow developed on the East Slope of Granite Mountain. Shortly after 4:00 UTC (22:00 MDT), strong horizontal shear developed as downslope velocities (Figure 5.7A) reached in excess of 3 m/s in the lower levels and as much as 6 m/s in the upper portion of the flow above 10 m AGL. After the katabatic flow developed for the period 4:00 - 5:30 UTC (22:00 – 23:30 MDT), the temperature was nearly constant (Figure 5.9A). The wind direction during this period was nearly constant at ~230° (Figure 5.7C), which was slightly to the south compared to the observed period in IOP 2. As the katabatic flow developed, the TKE slowly increased with time (Figure 5.8A) as another indicator of the presence of mixing and overturning as the katabatic layer depth increased. During the developing stages, the buoyancy flux was slightly negative as stability set in and was nearly constant (Figure 5.8B). Momentum flux increased as the katabatic flow developed, thus increasing the drag on the flow (Figure 5.8C). As the flow developed the horizontal shear increased and temperature stratification decreased, hence the stability of the flow decreased. The gradient Richardson number captures this transition to turbulence, as just prior to 4:00 UTC (22:00 MDT) the value drops below the non-linear theory critical value of 1 and then shortly thereafter, below the linear theory value of 0.25 (Figure 5.9B). In Figures
5.7-5.9, (a-d) correspond to 3:23 UTC (21:23 MDT), 3:30 UTC (21:30 MDT), 3:40 UTC (21:40 MDT) and 4:09 UTC (22:09 MDT) and are the starting times of combo probe measurements.

Figure 5.7: ES2 Tower, IOP 8, 3:00 - 6:00 UTC (19:00 - 24:00 MDT). One minute was used for the averaging period. (A) Sonic velocity transformed in a rotated coordinate system aligned with the mean flow. (B) Sonic vertical velocity. (C) Sonic wind direction. (a-d) Correspond to 3:23 UTC (21:23 MDT), 3:30 UTC (21:30 MDT), 3:40 UTC (21:40 MDT) and 4:09 UTC (22:09 MDT) and are the starting times of combo probe measurements.
Figure 5.8: ES2 Tower, IOP 8, 3:00 - 6:00 UTC (19:00 - 24:00 MDT). One minute was used for the averaging period. (A) Turbulent Kinetic Energy. (B) Buoyancy flux. (C) Momentum flux. (a-d) Correspond to 3:23 UTC (21:23 MDT), 3:30 UTC (21:30 MDT), 3:40 UTC (21:40 MDT) and 4:09 UTC (22:09 MDT), and are the starting times of combo probe measurements.
Figure 5.9: ES2 Tower, IOP 8, 3:00 - 6:00 UTC (19:00 - 24:00 MDT). One minute was used for the averaging period. (A) Temperature. (B) Gradient Richardson, $Ri_g$ number (5.1) calculated using the layer between 4 and 28 m with a solid red line indicating the critical value of 0.25. The non-linear theory value of 1 is also shown as a dashed line, Strang and Fernando (2001A,B). (a-d) Correspond to 3:23 UTC (21:23 MDT), 3:30 UTC (21:30 MDT), 3:40 UTC (21:40 MDT) and 4:09 UTC (22:09 MDT), and are the starting times of combo probe measurements.

5.5.2 Combo Probe

The time series of combo probe sonic velocity and rms velocity are shown in Figure 5.10. Note that the $v$ component in the upper panel is near zero, indicating the relatively low variability of flow. This data set represents 1400 continuous minutes starting at time, 18:57 UTC (12:57 MDT) on 10/19/2012 during IOP 8. Short segments of this data set will be analyzed with the combo probe corresponding to the times; 3:23 UTC (21:23 MDT), 3:30 UTC (21:30 MDT), 3:40 UTC (21:40 MDT) and 4:09 UTC (22:09 MDT) labeled in Figures 5.7-5.9 as (a-d).
Figure 5.1: Time series of the combo probe sonic data, representing 1400 continuous minutes starting at the time 18:57 UTC (12:57 MDT) on 10/19/2012 during IOP 8. The upper panel contains the instantaneous velocity and the lower panel contains the RMS velocity.

After completing steps 1-5 of the data processing, the hot-film velocities were obtained via the Neural Network calibration procedure. A 1-s time series of the reconstructed hot-film velocities ($u_{HF}$, $v_{HF}$, $w_{HF}$) compared to the sonic velocities ($u_{S}$, $v_{S}$, $w_{S}$) is presented in Figure 5.11 for the final time period of study (4:09 - 4:22 UTC (22:09 - 22:22 MDT)). As expected, the hot-film data contains more velocity fluctuations about the mean velocity due to improved temporal and spatial resolution compared to the sonic. This result also shows that the NN calibration method is working very well.
Figure 5.1: 1-s zoomed-in time series to show how well the NN is calibrating the hot-film probes. Hot-film velocities ($uHF, vHF, wHF$) compared to the sonic velocities ($uS, vS, wS$).

During the developing stages of katabatic flow ($Re = 5.2 \times 10^5$), the power spectra consists primarily of intermittent large scale eddies containing very low energy compared to the next time period, as evident from the spectra of velocity during this time period from 3:23 - 3:29 UTC (21:23 - 21:29 MDT) (Figure 5.12). Though the noisy spectra somewhat resembles Kolmogorov spectra at -$5/3$ (actually the slope is smaller), there is an inexplicable region ($k^{-1}$) between the low frequency and dissipative range where the bottleneck effect is evident as theoretically predicted by Tatsumi et al. (1978). This developing region previously could not be detected. Beyond this region, the spectra is mostly the result of noise as there is very little energy in this range and the SNR value is very low. This conclusion is also confirmed by the hot-film and sonic
statistics in which only one hot-film minute out of seven and zero sonic minutes had a skewness (5.8) in the range of $0.2 < Sk < 0.8$. The skewness, mean and RMS velocity and kinetic energy dissipation (5.10, 5.11) averaged over the number of minutes in the acceptable skewness range are presented in Table 5.1. The sonic spectra also show that the signals are comprised of very low frequencies.

Figure 5.12: Spectra for 3:23 - 3:29 UTC (21:23 - 21:29 MDT), developing katabatic flow ($Re = 5.2 \times 10^5$).
TABLE 5.1


<table>
<thead>
<tr>
<th></th>
<th>SK</th>
<th>U (m/s)</th>
<th>V (m/s)</th>
<th>W (m/s)</th>
<th>$\sigma_u$ (m/s)</th>
<th>$\sigma_v$ (m/s)</th>
<th>$\sigma_w$ (m/s)</th>
<th>$\varepsilon_u$ (m$^2$/s$^3$)</th>
<th>$\varepsilon_w$ (m$^2$/s$^3$)</th>
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<tbody>
<tr>
<td>HF</td>
<td>0.245</td>
<td>3.707</td>
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<td>0.052</td>
<td>0.095</td>
<td>0.153</td>
<td>0.049</td>
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<td>6.1e-4</td>
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<tr>
<td>Sonic</td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Good Minutes: HF = 1/7, Sonic = 0/7

“Good minute” means that it passed the skewness (5.8) test ($0.2 < Sk < 0.8$).

Figure 5.13: Spectra for 3:30 - 3:37 UTC (21:30 - 21:37 MDT), developing katabatic flow ($Re = 1.8 \times 10^5$).

When the katabatic flow further develops ($Re = 1.8 \times 10^5$), the TKE of spectral components increases and the energy continues to cascade towards smaller scales (Figure 5.13), nevertheless, an inertial Kolmogorov range for the velocity components is
still not very evident. It is possible to speculate that some large eddy structures develop but they are far from a developed three-dimensional turbulent spectrum. This is again confirmed by the computed statistics indicating that none of the minutes have a skewness in the acceptable range.

At the next measurement (Re = 4.2x10^5), the energy of spectral components continuously increased (Figure 5.14). There is still not a very evident inertial Kolmogorov range for the velocity components and the bottleneck region is present. The number of minutes that have a skewness in the acceptable range has increased as shown in Table 5.2, however the turbulence is far from local isotropy and the dissipation computed using the two equations are significantly different indicating the turbulence is still structured.

Figure 5.14: Spectra for 3:40 - 3:53 UTC (21:40 - 21:53 MDT), developing katabatic flow (Re = 4.2x10^5).
TABLE 5.2

STATISTICS FOR 3:40 - 3:53 UTC (21:40 - 21:53 MDT)

<table>
<thead>
<tr>
<th></th>
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<th>U (m/s)</th>
<th>V (m/s)</th>
<th>W (m/s)</th>
<th>$\sigma_u$ (m/s)</th>
<th>$\sigma_v$ (m/s)</th>
<th>$\sigma_w$ (m/s)</th>
<th>$\varepsilon_u$ (m$^2$/s$^3$)</th>
<th>$\varepsilon_w$ (m$^2$/s$^3$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>HF</td>
<td>0.48</td>
<td>4.06</td>
<td>0.091</td>
<td>0.03</td>
<td>0.44</td>
<td>0.37</td>
<td>0.25</td>
<td>4.4e-1</td>
<td>7.7e-2</td>
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<tr>
<td>Sonic</td>
<td>0.21</td>
<td>4.72</td>
<td>0.66</td>
<td>0.035</td>
<td>0.45</td>
<td>0.41</td>
<td>0.29</td>
<td>1.9e-4</td>
<td>1.0e-4</td>
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</tbody>
</table>

Good Minutes: HF = 8/14, Sonic = 1/14

By 4:09 UTC (22:09 MDT), the turbulence in the katabatic flow (Re = 1.4x10$^6$) is fully developed and the spectra of all three components exhibits the inertial Kolmogorov range, showing decay rates similar to the well known -5/3 slope at frequencies below 100 Hz and decay faster at higher frequencies (Figure 5.1). The bottleneck effect is no longer present as the energy has cascaded all the way down to Kolmogorov eddies and is now dissipating. The hot-film statistics shown in Table 5.3 show the skewness was within the desirable range for 10/14 minutes and the average value is 0.36; this compares favorably to previous wind tunnel, grid-generated turbulence studies in the laboratory. Batchelor and Townsend (1947) suggested Sk has an average value of 0.39, independent of the Reynolds number; Arora and Azad (1980) reported values of 0.38 $\leq Sl \leq 0.5$ depending on the axial location downstream of a pipe exit within a diffuser; Tsinober et al. (1992) used a 12 wire hot-wire probe (three arrays of four wires) and found 0.41 $\leq Sl \leq 0.5$ at various locations downstream the grid; and Tatsumi et al. (1978) theoretically obtained a value of 0.3 $\leq Sl \leq 0.65$ for isotropic turbulence. Of further interest is that the dissipation rate calculated with Equations 5.10 and 5.11 are
identical, indicating that even though the entire flow is clearly not isotropic (the RMS values of various components differ significantly) the very small scales of inertial range display local isotropy.

![Spectra for 4:09 - 4:22 UTC (22:09 - 22:22 MDT), fully developed katabatic flow (Re = 1.4x10^6).](image)

**Figure 5.15**: Spectra for 4:09 - 4:22 UTC (22:09 - 22:22 MDT), fully developed katabatic flow (Re = 1.4x10^6).

**TABLE 5.3**

<table>
<thead>
<tr>
<th></th>
<th>SK</th>
<th>U (m/s)</th>
<th>V (m/s)</th>
<th>W (m/s)</th>
<th>(\sigma_u) (m/s)</th>
<th>(\sigma_v) (m/s)</th>
<th>(\sigma_w) (m/s)</th>
<th>(\varepsilon_u) (m^2/s^3)</th>
<th>(\varepsilon_w) (m^2/s^3)</th>
</tr>
</thead>
<tbody>
<tr>
<td>HF</td>
<td>0.36</td>
<td>4.36</td>
<td>-0.006</td>
<td>0.044</td>
<td>0.43</td>
<td>0.35</td>
<td>0.25</td>
<td>4.2e-3</td>
<td>4.2e-3</td>
</tr>
<tr>
<td>Sonic</td>
<td>0.30</td>
<td>4.45</td>
<td>0.14</td>
<td>0.067</td>
<td>0.49</td>
<td>0.38</td>
<td>0.26</td>
<td>2.1e-4</td>
<td>1.2e-4</td>
</tr>
</tbody>
</table>

Good Minutes: HF = 10/14, Sonic = 4/14
5.6 Conclusions

A hot-film combo system consisting of hot-film probes and sonic anemometers (sonic) was developed, with the capability of sonics calibrating the hot-films using a Neural Network (NN). In this system, the hot-film probes are co-located with a sonic and the data from both are acquired and processed simultaneously. Note that the NN model is used here for obtaining an in-situ velocity/voltage transfer function for calibration rather than for any predictive purpose, which is the typical application of NNs. Sonics are commonly used in atmospheric research, but due to their low space-time resolution, the utility in obtaining information on fine-scale turbulence is severely limited. Hot-film/wire probes can provide such information, but they require frequent and laborious calibrations as well as specific alignments of the probes with the flow. The latter problem was circumvented by designing a platform where the hot-film probes automatically adjust to the wind direction, prompted by in-situ information provided by the co-located sonic anemometer. To address the former problem, a NN for in-situ calibration of hot-films was used.

For homogeneous and isotropic turbulence (Hinze 1975), the one-dimensional spectra of the lateral and longitudinal velocity components are related through a following relation:

$$E_2(k_1) = E_3(k_1) = \frac{1}{2} \left[ E_1(k_1) - k_1 \frac{\partial E_1(k_1)}{\partial k_1} \right].$$

(5.12)

The above expression explains the difference in values of spectral components, depicted in Figure 5.12-5.15: the power density of the lateral components at small $k_1$ is
1/2 of their longitudinal counterpart while in the inertial range (large $k_1$) the lateral components become by $4/3$ greater than the longitudinal. It is worth noting here that these low frequencies represent the main part of the TKE measured by sonic anemometer.

Turbulent velocity fluctuations are often modeled as stochastic processes with certain characteristics. Usually velocity fluctuations are modeled as a normally distributed process with correlations in space and time. The spectra of velocity fluctuations collected in the field or by laboratory set-ups such as pipes, wind tunnels or jet facilities, such as in Figure 5.12-5.15, reveal that the spectrum of fluctuations in the mean velocity direction differs from the two lateral components, as expected for spatial one-dimensional spectra. These are the spectral components measured by the hot-film probe along the line coinciding with the mean velocity, wherein the frozen Taylor hypothesis is applied. However, the RMS values of the lateral velocity fluctuations components also differ from the notably larger RMS of the $u$ component, suggesting that flow characteristics are dissimilar in every direction, i.e. the measured flow is not isotropic. Actually, all real-world flows, especially shear flows, are not isotropic (i.e. with characteristics independent of the spatial direction). Yet, for the high Reynolds number ($Re = 1.4 \times 10^6$) measurements presented above, there exists an inertial sub-range where the flow is locally isotropic, but only when the shear flow (katabatic current) is fully developed. Therefore, the use of Kolmogorov theory for evaluating turbulence characteristics of developing katabatic flows is highly discouraged and to determine whether or not a katabatic flow is fully developed requires the high resolution
measurements of the combo probe.
CHAPTER 6: CONCLUSIONS

The aim of the MOUNTAIN TERRAIN ATMOSPHERIC MODELING AND OBSERVATIONS (MATERHORN) PROGRAM was to conduct fundamental research to improve weather predictions for mountainous terrain. A significant part of the program has been dedicated to the understanding of physical processes, via which improvements in understanding of sub-grid scale processes and their parameterizations are sought. To help achieve this goal, field and laboratory studies were conducted and theoretical formulations were developed in this thesis. A suite of state-of-the art flow diagnostic techniques were used and new instrumentation was developed. At the core of the work in this thesis are the subtopics of upslope flow separation in mountainous terrain, multi-scale interactions of slope and valley flows and the measurement of turbulence in katabatic flows.

To investigate the upslope flow separation in mountainous terrain, laboratory experiments were performed specifically focusing on steeper slopes. The interest was on flow structure, separation distance, and the velocity field. The experiments were performed using a water tank, with mountain slope consisting of a specially designed heating pad located at one end. Particle tracking Velocimetry (PTV) was implemented in the regions of interest (slope and vicinity) and Feature Tracking Visualization animations
were developed and color mapped according to track length and/or velocity. Measurements were performed for numerous slope angles (from $5^\circ$ up to $45^\circ$ from the horizontal) and buoyancy flux settings, collecting substantial sets of experimental data.

When the slope angle $\beta_s \geq \beta_{sc} \approx 20^\circ$, the separated flow was found to be fed entirely by the upslope flow that precedes separation. A model that assumes the separation point is determined by the opposing vorticities due to baroclinic torque (separation tendency) and shear flow (attachment) produced a prediction for flow separation in this case, and the predictions compared reasonably well with laboratory results and field results obtained during MATERHORN-X-2. The velocity scale at the separation point was also well captured.

Also of key importance to mountain meteorology and of interest to the MATERHORN program are the complex multi-scale interactions between meso-$\alpha$ and $\beta$ scale (up/down) valley flows and meso-$\gamma$ scale thermally driven (up/down) slope flows. Until the recent MATERHORN field campaigns, no detailed observations of these flow interactions existed (Fernando and Pardyjak 2013) nor had the physical processes underlying these interactions been identified in detail.

During the field campaign, it became clear that in the Dugway basin the flow is extremely complex and multi-scale in both space and time, with competing valley flows to either side of Granite Mountain driven by differential cooling due to land surface contrast and downslope flows draining into the basin. The downslope flow, however, is much warmer due to mountain heat island (MHI) effects, which, together with the undercutting valley flow sets up a strong stable stratification in the basin.
Eventually the gap flows intersect the valley flow and vorticity develops which deflects the valley flow towards the slope, thus creating a violent collision of two currents marked by overturning, increased levels of turbulence, rapid temperature drop along the slope and enhanced mixing and spikes in buoyancy and momentum fluxes. A series of secondary collisions then follow as the once valley flow begins to descend the slope after the upslope momentum is lost. This impacts the newly emerging valley flow that pushes up the slope with less dense remnants of downslope flow riding atop; rapid fluctuations of velocity and TKE with overlapping events characterized this flow situation. The secondary impacts send waves of disturbance throughout the basin, as if the basin pool is seiching, while allowing the downslope flow to reemerge.

During the evening hours of IOP 2 of MATERORN-X-1, three main collision periods were identified, each with a primary collision followed by a series of secondary collisions. Of these collisions, three possible collision types were identified, and primary collisions signified by nearly parallel colliding flows (Figure 4.18A) were selected for further analysis as they are analogous to two colliding gravity currents. Interestingly, over the examined range of collision Richardson numbers, the dimensionless buoyancy flux and the decay time are nearly constant for these collisions.

Above-mentioned interactions within thermal circulation generate an intriguing set of small scale processes including the collision of gravity currents, formation of intense turbulent regions, intrusions and instabilities. WRF and other mesoscale models do not account for such processes, and hence their incorporation is crucial in modeling mountain terrain winds due to their significant contributions to sub-grid heat and
momentum transfer. It should be emphasized that the identification of these processes was possible due to the extensive suite of remote sensors and in-situ measurements deployed in MATERHORN-X-1.

To measure the smallest scale processes, new cutting edge techniques were developed and deployed. This included tower mounted three-dimensional hot-film combo probes developed as a part of this thesis, which consist of sonic anemometers co-located with hot-film anemometers. The combo probes adjust to the wind direction using a feedback control loop and use a Neural Network to calibrate the hot-films in-situ. Once calibrated, these probes can measure from mesoscale flow down to the Kolmogorov scale, thus circumventing the issue of low space-time resolution of commonly used sonic anemometers.

The spectra of velocity fluctuations collected in the field by the combo probes during katabatic flow development reveal that the spectrum of fluctuations in the mean velocity direction differs from the two lateral components, as expected from spatial one-dimensional spectra for isotropic turbulence. These are the spectral components measured by the hot-film probe, along the line coinciding with the mean velocity, using the frozen Taylor hypothesis. The RMS values of the lateral velocity fluctuations components also differ from the notably larger RMS of the $u$ component, suggesting that flow characteristics are dissimilar in every direction, i.e. the measured flow is not isotropic. Yet, for the high Reynolds number measurements presented ($Re = 1.4 \times 10^6$), there exist an inertial sub-range where the flow is locally isotropic, and it existed only when the shear flows (katabatic current) is fully developed. Therefore, the use of
Kolmogorov theory for evaluating turbulence characteristics of developing katabatic flows is discouraged, although it has been a common practice. In determining whether or not a katabatic flow is fully developed requires the high resolution measurements of the combo probe.
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