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Numerical Study of Canopy Flows in Complex Terrain

By

Xiyan Xu

A dissertation submitted to the Graduate Faculty in Earth and Environmental Sciences in partial fulfillment of the requirements for the degree of Doctor of Philosophy,
The City University of New York

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Abstract

Numerical study of canopy flows in complex terrain

By

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Canopy flow plays a substantial role in regulating atmosphere-biosphere exchanges of mass and energy. The worldwide FLUXNET has been developed to quantify the net ecosystem exchange of mass and energy through fluid dynamics in and above vegetation canopy using tower-based eddy covariance (EC) technique. However, EC measurements are subject to advection errors in complex terrain, particularly during nights when atmospheric stability is strong. Because EC measurements are one-dimensional (1D), three-dimensional (3D) air movement, CO₂ transport, and temperature variation around the instrumented tower are unknown. We employ a Computational Fluid Dynamics (CFD) model to investigate the impact on CO₂ transport of 2D and 3D characteristics of canopy flow resulting from interactions between large-scale synoptic flows and local topography, vegetation and thermal conditions.

Under neutral conditions, flow distortion over a forested hill is asymmetric, with recirculation on the lee slope. The presence of vegetation and steepening slope intensifies recirculation depth and extension. The recirculation regions are responsible for CO₂ build-up behind the hills. The contribution of advection to the CO₂ budget is significant and topography-dependent. Gentle slopes can cause larger advection error than steep slopes. However, the relative importance of advection to CO₂ budget is slope-independent. Under calm and stable conditions, canopy flow is thermally stratified: super-stable layers above and in the deep canopy and an unstable layer inbetween. Vertical exchanges of mass, momentum, and energy are limited
by the stabilities of these layers. The pattern of two drainage flows are significantly modified by the interaction between thermal stratification and slope, and are better understood with the distribution of vortices, and the sources and sinks of turbulent kinetic energy. The numerical method is applied to the alpine forest at Renon, Italy to investigate how thermo-topographic and synoptic flows interact to govern canopy flow dynamics and CO$_2$ transport. We found that recirculation with high CO$_2$ concentration developed only when local slope wind is enhanced by synoptic wind. There’s no recirculation formed as synoptic wind direction is opposite to the local wind direction and CO$_2$ is quite well mixed. No recirculation appears without synoptic condition under which CO$_2$ builds up mainly at downwind locations.
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Chapter 1 Introduction and Background

1.1 Research Motivation, Goal and Questions

One of the key issues concerning climate change is the climate-carbon cycle feedback. Attention on the global carbon cycle over more than 30 years has focused on the imbalance in the carbon budget. Of the total amount of CO₂ released to the atmosphere from known sources (fossil fuel emission, land use change), about 40% contributes to an increase of CO₂ in the atmosphere and about 30% is absorbed by the ocean. The remaining 30% is not clearly identified and is referred to as the missing carbon sink (Figure 1.1). The missing carbon can be estimated in the northern mid-latitudes by satellite and forest inventory data and global atmospheric transport models (Tans et al., 1990; Gurney et al., 2002, Myneni et al., 2001), and the uptake of carbon by northern forests is increasing (Liski et al., 2003, Qian et al., 2010) in response to the warming climate. In contrast, Stephens et al. (2007) discovered that tropical ecosystems play a larger role in sinks of CO₂. However, the carbon sink in tropics may decreases due to the warming climate (Wang et al., 2013).

The mystery missing sink and uncertainties in contributions of CO₂ from terrestrial ecosystems to the global carbon cycle have motivated numerous studies to quantify the exchange of CO₂ between atmosphere and terrestrial ecosystems. A global network (FLUXNET) of micrometeorological flux measurement sites has been established to measure the exchange of carbon dioxide, water vapor and energy between the biosphere and atmosphere (Baldocchi et al., 2001). The tower-based measurements in FLUXNET account for the turbulent flux above the canopy and CO₂ storage changes below the maximum height of measurements above the canopy. It is suspected that the tower measurements underestimate the nighttime CO₂ flux from the
ecosystem because sensors above the canopy on a single tower cannot detect the flows beneath the canopy that transport CO$_2$ laterally from its sources, the so-called advection fluxes (Goulden et al., 1996; Aubinet et al., 2000). The underestimation related to neglect of advection fluxes is called advection issue. The advection issue has been reported to be very common during calm nights in ecosystems with complex terrains, which is correlated to mechanisms on nocturnal canopy flow, e.g. turbulent ramps, gravity waves, small-scale turbulence, intermittent turbulence, land, sea or lake breezes and drainage flows (Aubinet, 2008).

A state-of-the-art approach to dealing with the nocturnal advection problem in the FLUXNET community is to judge the quality of data with the turbulent mixing indicator u*. Data with low-u* are discarded and replaced with regressions between high-u* nocturnal NEE and temperature. This is called a u*-filter approach (biological model). Nevertheless, the u* correction is subject to several drawbacks. First, the u* filter threshold is arbitrary and varies from site to site. Second, we may use bad data to fill bad data. The philosophy of u* filtering is to discount NEE measurements at low u* condition and treat them as bad data, and fill up these bad data gaps with high u* data. Third, biological modeling is based on climatic responses of organisms, and has no physical connection to aerodynamic processes, while the advection correction is based on aerodynamic mechanisms and is independent of biological activity (Yi et al., 2008).

Multi-tower and multi-level measurement systems are applied to make direct measurement of CO$_2$ advection at FLUXNET sites, such as three European FLUXNET forest sites with different topography (Renon/Ritten, Italian Alps, Italy; Wetzstein, Thuringia, Germany; Norunda, Uppland, Sweden) and AmeriFlux Niwot Ridge site. During the advection measurement campaign, CO$_2$ concentration and wind field are measured in three-dimensional
(3D) multi-tower cube setup. The direct advection measurements indicate that all the experimental sites experience advection problems. Horizontal advection and vertical advection are the main CO$_2$ transport processes at night (Aubinet et al., 2005; Yi et al., 2008; Feigenwinter et al., 2008). The advection contribution is closely correlated to local and synoptic meteorological conditions. Local orographic flow mostly occurs within the canopy, while synoptic is dominant above the canopy. However, synoptic flows can penetrate into the open canopy and interact with orographic flow (Sun et al., 2007). The interaction mechanism is that the synoptic winds alter the direction and enhance or attenuate orographic wind, depending on the direction and strength of prevailing synoptic winds (Feigenwinter et al., 2010a). Accordingly, the modified orographic flows have direct influence on CO$_2$ pooling or mixing.

Although the direct advection measurement provides insights in the wind fields and CO$_2$ transport at the research sites, the conclusions drawn from the experimental sites may not be applicable to other FLUXNET sites subject to local terrain and vegetation conditions and large-scale synoptic conditions. In addition, how representative the multiple-tower measurements are is very sensitive to the multi-tower setup and methodology to derive the fluxes from measurements (Aubinet et al., 2010a; b). How can we take advantage of the single tower measurements on the set-up of most FLUXNET sites to understand ‘site-specific’ CO$_2$ transport processes? This study aims to apply a Computational Fluid Dynamics (CFD) model to provide insight into thermal stability, sub-canopy small scale structures and related CO$_2$ transport by addressing these questions: How does the canopy layer modify the thermal condition in the complex terrain? What’s the response of sub-canopy flow to slope variation, canopy structure and synoptic forcing? What are the roles of the sub-canopy flow on CO$_2$ transport?
The organization of this thesis is as follows. In chapter 1, we describe the circulations in forested complex terrain under different stabilities and how the circulation influences the CO$_2$ transport. We highlight the current studies of canopy flows in complex terrains. The numerical method and CO$_2$ source parameterization are described in chapter 2. Simulations of neutrally stratified canopy flow and its related CO$_2$ transport on two-dimensional hilly terrains are presented in chapter 3. Simulation of canopy flow structure under stable condition on a two-dimensional single hill is described in chapter 4. In chapter 5, the numerical method is applied to a three-dimensional case of alpine forests to investigate interactions between local thermotopographic flow and large-scale synoptic flow and related CO$_2$ transport. Lastly, in chapter 6, we summarize our research and discuss the prospects for future research.
**Figure 1.1** Carbon flux showing sources and sinks between 1850-2000. Source: Woods Hole Research Center.
1.2 Circulation in Complex Terrain

A large part (70%) of the earth’s land surface is covered with mountains and hills. Thermally driven circulation, such as slope flows and valley winds are a common phenomenon observed in mountainous regions throughout the world (Whiteman, 1990; 2000). These circulations exert a powerful control on local weather and climate, mass and energy cycles.

1.2.1 Mountain-Valley Flow

Typically, over a mountain slope under quiescent synoptic conditions, winds blow upslope during the day (Figure 1.2a) and downslope (Figure 1.2b) at night, known as ‘thermal circulation’ (Monti et al., 2002). The primary driving force of the slope winds is the vertical buoyancy force with additional horizontal pressure gradient. During the day, the slope is heated by solar shortwave radiation. Air on the slope is warmer than the air above, resulting in a vertical buoyancy force. In addition, the air on the slope is warmer than the ambient air at the same altitude which is further away from the heated surface, resulting in a horizontal pressure gradient force. At night, due to the radiative longwave cooling induced temperature inversion, the air parcel in the vicinity of the cooling slope is negatively buoyant and the horizontal pressure gradient is reversed. The air parcel tends to accelerate down the slope. When the terrain is undulated with ridges and valleys, the two-dimensional slope winds can evolve to a three-dimensional circulation, known as valley wind (Rampanelli et al., 2004). The valley winds result from horizontal pressure gradients that are driven by temperature differences between the air in the valley and air at the same elevation over adjacent plains or larger valleys. Valley winds blow along the valley axis, up-valley during the day and down-valley at night. Slope winds usually interact with valley winds in a local mountain-valley wind system.
Characteristics of mountain-valley winds are dependent on variations of slope angle, land cover, and atmospheric stability. As down-slope flow is initialized by sufficient surface inversion (Sturman, 1987), wind speed is reversely proportional to temperature stratification (McNider and Pielke, 1984). The typical speed of the slope flows ranges between 1-5 ms\(^{-1}\) during periods of strong temperature gradient. During the early morning and late afternoon transition period, slope winds are usually weak (< 0.5 ms\(^{-1}\)) (Papadopoulos and Helmis, 1999). On slopes with snow and ice cover, the katabatic flow could be faster than 10ms\(^{-1}\) over long slopes (Monti et al., 2002; Pettré and André, 1991). Slope flows usually confined to slope walls are relatively shallow, especially the nocturnal katabatic flow. The depth of the katabatic flow layer increases with distance downslope, wind speed and surface inversion (Dickerson and Gudiksen, 1983; Clements and Nappo, 1983;). On a simple slope, the flow depth increases with downslope distance and slope angle (Briggs, 1981). Large eddy simulations by (Smith and Skyllingstad, 2005) showed that the relationship between katabatic layer depth with downslope distance and flow maximum velocity located just above the surface is near linear.

As cold air flows down-slope into the valley, warmer air from aloft will replace the cold air by advection. The warm air is subsequently cooled and accumulates in the valley with strong temperature inversion (Daly et al., 2009), which is called the cold-air pool. Due to continuing cooling, some of the valleys and basins have recorded extremely low temperatures, large daily temperature ranges and very strong temperature inversions. Peter Sinks basin in Utah experienced -56.3°C on a clear, dry September night (Pope and Brough, 1996). A 30°C diurnal temperature range and 22°C temperature inversion at 100m depth were observed in the summer (Clement et al., 2002). A -52.6°C was recorded in the Gruenloch sinkhole in the eastern Alps, Lower Austria (Whiteman et al., 2004a). The diurnal temperature range as high as 28.1°C and a
20°C temperature inversion over a depth of 70m were observed in October (Whiteman et al., 2004b). The depth of the cold air pool evolves with radiation and ambient winds. The cold pool is particularly deep when along-valley winds are weak (Clements et al., 2003). Barr and Orgill (1989) showed that under ideal conditions of radiative cooling and light ambient wind, the depth of cold air in the valley can reach the ridge top but linearly decreases with wind at the ridge top. In addition, clouds and humidity caused less cooling and turbulent mixing reduces the depth of the valley wind. Whiteman (1982) observed three patterns of inversion break-ups. The first is growth of a convective boundary layer due to surface heating, which usually occurs on a summer day with wide valleys. When the valley is vastly covered by snow in wintertime, the surface heating is not enough for inversion break-up (Vrhovec and Hrabar, 1996). Thus, the second break-up pattern is caused by sinking of the warmer top of the inversion layer through advection and turbulent mixing. In most situations, inversion in valleys is destroyed by a combination of the previous two patterns.

1.2.2 Flow Separation

Flow separation and recirculation are very common phenomenon in mountain-valley systems. Under ideal clear-sky and calm condition, the symmetric double circulations are expected on perfect symmetric topography. Air rises along lateral slopes and subsides in the valley center during the day (Figure 1.3b), with reversed flow circulation at night (Figure 1.3a). In reality, the topography is much more complex and the diurnal heating on the slopes is heterogeneous. The scale associated with flow separation over and around mountain ranges is defined as Froude number $Fr = U / NH$, where $N$ is the Brunt-Väisälä frequency, a measure of the stratification and $U$ and $H$ are typical velocity and height scales, respectively. In strongly stratified flows ($Fr \ll 1$), air flows around, rather than over, the hill. These flows are associated
with vortices development downwind, which are induced by tilting of horizontal vorticity into vertical (Smolarkiewicz and Rotunno, 1989; 1990; Lin and Jao, 1995). This horizontal vorticity is generated baroclinically and tilted on the lee of obstacles. In the numerical model, the vortex stretching and advection terms are demonstrated to be important to initialize vortex and enhance the vorticity to the lee, and Coriolis force influences the propagation of the vortex (Sun and Chern, 1993). In addition to leeward vortices, reversed flow is developed on the upwind side as the aspect ratio (across-stream length/along-stream length) of obstacles exceeds a critical value (Smolarkiewicz and Rotunno, 1990; Smith, 1989). The reversed westerly flow was predicted on the windward side of the island of Hawaii, accompanied with convergence and upward motion (Chen and Feng, 2001).

If there is weaker stability and light winds (Fr ~ 1), flows at higher altitude can pass the top of the hill, while flows at lower altitude tend to be blocked at the windward slope of the orography and become stagnant (Baines and Hoinka, 1985; Pierrehumbert and Wyman, 1985). With very strong wind and near neutral stability (Fr >> 1), thermal conditions have minimum influence on aerodynamics. Air flows over the hill with very little lateral displacement. Local circulation in hilly terrain can only be driven by strong synoptic wind or the upstream basic flow (Belcher and Hunt, 1998). The longitudinal momentum governing the equation of steady, neutral flow is:

$$ u(\partial u/\partial x) + w(\partial u/\partial z) = -\partial P/\partial x + \partial \tau/\partial z, $$

where $u$ and $w$ are the time and spatially averaged velocity component, $\tau$ is the turbulent shear stress, $P$ is the pressure per unit density. Across the hill, airflow is dominated by the pressure gradient. On the windward slope, $-\partial P/\partial x$ is positive, which drives the wind to accelerate up the slope, reaching maximum wind speed at the crest. On the leeward slope, $-\partial P/\partial x$ is negative. This adverse pressure gradient decelerates wind down the slope, leading to flow separation and recirculation formation on the leeside of
hills (Figure 1.3c). The recirculation zone is characterized by high intensities of the fluctuations of the streamwise velocity and wall-normal velocity. The size of the recirculation region and the velocity in the recirculation zone are associated with slope angles (Poggi and Katul, 2007; Atkinson and Shahub, 1994) and vegetation structures (Poggi and Katul, 2007). Froude number is also used as a diagnostic scale for flow regimes in the valley. In a sloped valley, Holden et al. (2000) found that when \( Fr \leq 2 \), flow in the valley is dynamically stable and decouples from the flow aloft, while \( Fr \geq 2 \), flow remains attached to orography with production of turbulence.

Recirculation in mountain-valley systems plays an important role in local weather and climate and redistribution of mass and energy. Reversed flow on the windward side and vortices on the leeside of a mountain can cause redistribution of moisture and associated precipitation events (orographic precipitation), which are studied in the relative large scale (100s kilometers, Carbone et al., 1998; Yang et al., 2008) and small-scale (10s meters in height and 100s of meters in length, Arazi et al., 1997) topographies. Cool pools resulting from strong surface cooling can cause fog, stratus clouds and air pollutant build up within the pool (Savov et al., 2002) for reduced vertical mixing and ventilation. Air pollutants can be trapped in the pool for many days with diurnal cycles causing severe air quality problems. High respiratory disease deaths were reported in cities in the basin or valley topographies, e.g. Utah Valley (Pope et al., 1992) and Kathmandu Valley (Shrestha, 2012). Stagnation induced by the recirculation region can enhance strong horizontal gradients of scalar quantities (Finnigan and Belcher, 2004). Katul et al. (2006) incorporated \( CO_2 \) sources in the two dimensional hilly terrain, and numerically demonstrated the presence of recirculation zones as evidence of high accumulation of \( CO_2 \).
Figure 1.2 Thermally produced slope winds: a) anabatic; b) katabatic.
Figure 1.3 The recirculation forms along slopes (a) cool surface (Fleagle, 1950); (b) warm surface (Atkinson and Shahub, 1994); (c) Neutral condition with background wind (Finnigan and Belcher, 2004; Poggi and Katul, 2007).
1.3 Canopy Flows and CO₂ Transfer

Canopy flow is fluid dynamics occurring within and immediately above vegetation canopies, which has a significant influence upon various biological and physical processes. The canopy flow has drawn more attention in FLUXNET community since uncertainties were recognized in the measurement of net ecosystem exchange (NEE).

1.3.1 Canopy Flows

Canopy elements change the nature of mean flow and its turbulent characteristics in the lower boundary layer. The wind profile of canopy flow is uneven, governed by the surface roughness and plant structure. The mean velocity profile is characterized by strong inflection near the top of the canopy with a maximum shear. The velocity inflection leads to mixing-layer instability, which is responsible for the formation of coherent eddies in fully-developed turbulent flow (Raupach et al., 1996). Coherent eddies are demonstrated to be responsible for upward and downward transfer of momentum and are scalar through low frequency ‘sweeps’ and ‘ejections’ (Thomas and Foken, 2007). With the effects of canopy morphology, wind profile within the canopy is characterized by an S-shaped wind profile, with a second wind maximum that is often observed within the trunk space of forests and a minimum wind speed in the region of maximum foliage density, described as a ‘bulge’ by Ni (1997). The S-shaped wind profiles are recorded in field studies (Shaw, 1977; Turnipseed et al., 2003; Yi et al., 2005). Yi (2008) theoretically described the S-shaped profile by leaf area density and local drag coefficient profiles. In the case of drainage flow, the second wind maximum is likely due to the combined effects of local drainage flows that reach relatively high speeds in the lower region of the canopy (Yi et al., 2005), and resistance to the mean wind flow that can reduce wind speeds in the region of the canopy with high leaf area density (Massman, 1997; Mohan and Tiwari, 2004).
In hilly terrain, the presence of a canopy layer modifies the flow patterns on the slope and valley. In the near-neutral condition, the canopy reduces the acceleration of the flow over the hill (Brown et al., 2001, Allen and Brown, 2002), but increases the asymmetry of the flow above the crest of the hill. The canopy adds an effective drag on the hill, which increases the deceleration on the leeside slope and the tendency for the flow to separate (Finnigan and Belcher, 2004), therefore promoting the onset of recirculation behind the hill (Patton and Katul, 2009). Patton and Katul (2009) simulated the recirculation with dense and sparse canopy, showing a deeper recirculation region with dense canopy. Poggi and Katul (2007) studied recirculation in a flume experiment. It shows that the recirculation region is not displayed as a rotor but intermittent reversed flow when $H/L = 0.1$, where $H$ is hill height and $L$ is the half-length scale of the hill.

Thermal stratification within a plant canopy usually differs from that above a canopy due to the morphology of canopy and its related heat transfer. During the daytime, upper canopy is well heated by incoming short-wave radiation. Thus, temperature in the upper and above canopy is unstable. However, the shaded deep canopy and ground surface beneath remains cold resulting in an inversion developing within the canopy layer. At night, due to loss of long-wave radiation from upper dense canopy layer, the air above-canopy becomes stably stratified, while in the deep canopy, air is unstably stratified because of heat from the soil (Jacobs et al., 1991; 1992). Currently, there have been very few studies quantifying thermal stability within canopy. Shaw et al. (1988) and Leclerc et al. (1990) applied the stability parameter $(z/L)$ to assess stability and its effects on turbulent properties within and above canopy, where $z$ is the height above ground, $L$ is Monin-Obukhov length. Thermal effects are most significant in the upper canopy where the canopy is densest and radiation load is highest. In very stable condition, turbulent activity is suppressed by buoyant production (Leclerc et al., 1990). Yi et al. (2005)
applied the gradient Richardson number (Ri) to analyze S-shaped wind profile associated
stability in the sub-canopy. A super stable layer is predicted at the height of maximum leaf area
density, which impedes vertical mixing between the sub-canopy and atmosphere aloft (Yi et al.
2005). Belcher et al. (2008) showed Ri profiles in a homogeneous canopy under moderate
stratification. Ri indicates turbulent flow above the canopy (Ri < 0.2), but Ri increases
dramatically down to deep canopy indicating suppressed turbulence within the canopy. Burns et
al. (2011) applied a bulk Richardson number Ri_b evaluating the wind and scalar fields. For strong
stable conditions (Ri_b > 1), the vertical scalar gradient is at its maximum and keeps constant with
increasing stability.

1.3.2 CO₂ Transport between Vegetation and Atmosphere

Fluid dynamics within and just above the vegetation canopy governs the energy and mass
exchange between the vegetation and the atmosphere above. The worldwide FLUXNET program
(Baldocchi, 2008, Baldocchi et al., 2001) has been established using eddy covariance methods,
monitoring long-term exchanges of energy, CO₂, and other scalar quantities between plant
canopies and the atmosphere.

The time averaged mass conservation for CO₂ states that biological sources and sinks in a
controled volume is balanced by the time rate of CO₂ mixing ratio, advection and flux
divergence in the vertical and horizontal directions:

\[
\frac{\partial \bar{c}}{\partial t} + \frac{\partial \bar{c}}{\partial x} + \frac{\partial \bar{c}}{\partial z} + \frac{\partial \bar{u}' \bar{c}'}{\partial x} + \frac{\partial \bar{w}' \bar{c}'}{\partial z} = \bar{S}_c
\]  

(1.1)

where \(c\) represents the mixing ratio of carbon dioxide, \(x\) is aligned with the horizontal mean
wind direction, \(z\) is perpendicular to the long-term average stream line, \(u\) and \(w\) are the wind
velocity along the \(x\) and \(z\) directions, respectively. The terms \(\bar{u}' \bar{c}'\) and \(\bar{w}' \bar{c}'\) are the time-
averaged turbulent fluxes of CO₂ in the x and z coordinate, respectively; the sum of CO₂ sources or sinks in a control volume, which is non-negligible only within canopy. The overbars denote the time average of 30 minutes and the primes represent the fluctuation from the mean values. Integrating the equation (1.1) from 0 (represent the ground level) to z\_r, z\_r is the canopy height or the measurement height), the net ecosystem exchange NEE is balanced by CO₂ stored in the control volume (F\_S), or exchanged by the turbulence (F\_T), vertical advection (F\_Av), horizontal advection (F\_Ah), or horizontal divergence of turbulent flux (F\_D) (Figure 1.4).

\[
\text{NEE} = \int_0^{z_r} \overline{S}_c \, dz + \left(\overline{w'c'}\right)_{z=0}
\]

\[
= \int_{F_d}^{z_r} \overline{\partial c / \partial t} \, dz + \left| \int_{F_s}^{z_r} u \overline{\partial c / \partial x} \, dz \right| + \int_{F_{Ah}}^{z_r} \overline{w \partial c / \partial z} \, dz + \int_{F_{Av}}^{z_r} \overline{\partial u'c' / \partial x} \, dz
\]

(1.2)

In equation (1.2), the divergence term of the horizontal turbulent flux (F\_D) could be ignored, provided that the length of the footprint of the turbulent flux measurement is much larger than z\_r (Yi et al., 2000). In the classical eddy covariance method, it is also assumed that the strong mixing and homogeneous conditions are fulfilled so that the terms F\_Av, and F\_Ah, can be considered as negligible due to very low CO₂ gradient. In these conditions, NEE can be estimated from only two measurements made at a single tower:

\[
\text{NEE} = F_S + F_T = \int_0^{z_r} \overline{\partial c / \partial t} \, dz + \left| \overline{w'c'} \right|_{z_r}
\]

(1.3)

The storage term (F\_S) is obtained from a vertical CO₂ concentration profile and the turbulent flux (F\_T) is measured directly using an eddy covariance system at z\_r. With regard to equation (1.3), it is clear that the night flux error that affects such a system results either from an incorrect evaluation of F\_S and F\_T or from the fact that the advection terms and horizontal
turbulent flux become significant. The bias in NEE estimation without advection terms is called advection error.

The distribution of CO$_2$ presents significant spatial and temporal variation in forests. CO$_2$ concentration was measured at six levels on a 447m tall television transmitter in the Chequamegon National Forest in northern Wisconsin. During the day, photosynthetic uptake of CO$_2$ dominates and the boundary layer is convectively mixed, CO$_2$ mixing ratio is vertically uniform. At night, the boundary layer is stably stratified and shallow (<150m) and respiration provides sources for CO$_2$. CO$_2$ builds up very high underneath temperature inversions and decreases dramatically with height, up to about 200m. The difference in CO$_2$ mixing ratio between 30m and 396m is as high as 13mol (Yi et al., 2001). In complex terrain, the spatial variability of CO$_2$ in the canopy can be significantly influenced by thermally driven gravity flows. Cold air has been observed draining down the hill at night (Goulden et al., 2006). Large amounts of CO$_2$ are carried down the slope from the plateau and stored in the valley during the stable night, and is not flushed until mid-morning (Araujo et al., 2008). CO$_2$ concentration on the plateau is high at 0.5m but decreases with height at any time. Nighttime CO$_2$ concentration on the slope and valley is uniformly high in the vertical direction but decreases with height in the morning due to better mixing at the upper levels as sun rises, resulting in a large horizontal and vertical CO$_2$ gradient.

An orographic wind-related CO$_2$ gradient is responsible for significant advection. Sun et al. (2007) showed that CO$_2$ advection dominated the turbulent and storage terms (Equation (1.2)) in nighttime NEE estimates. Horizontal advection was significant even in the early morning and evening transition periods with weak turbulent mixing in the canopy. As the local slope wind system dominated at the Renon site in Italy, Feigenwinter et al. (2008) reported that due to
down-slope winds and higher concentrations of \( \text{CO}_2 \) on the slope surface, horizontal and vertical advection are positive during night. The highest horizontal advection occurred in a relatively shallow part of the lower canopy, close to the ground surface. An uncommon condition is a small negative horizontal advection observed during the day, which requires further investigation. The averaged non-turbulent advection terms had the same order as turbulent flux. The net contribution of \( \text{CO}_2 \) advection to NEE estimate is complicated because of the uncertainties in the magnitude and sign of vertical and horizontal advection, which is largely determined by local slope-valley wind system and land use changes. Aubinet (2008) classified advection into three categories according to the vertical velocity (Figure 1.5). Negative vertical \( \text{CO}_2 \) gradients in the night results in (1) positive vertical advection with convergence flow, while horizontal advection is (i) negative with decreasing NEE, (ii) zero with constant NEE and (iii) positive with increasing NEE; (2) negative vertical advection and positive horizontal advection with divergence flow; and (3) zero vertical advection and positive horizontal advection with constant mass flows.
Figure 1.4 The different fluxes between a control volume and the atmosphere.
Figure 1.5 Classification of the different advection patterns in relation to source intensity distribution and mass flow characteristics. Abbreviations are: HA, horizontal advection; VA, vertical advection; NEE, biological source/sink; B, Belgium; D, Germany; CND, Canada; I, Italy; and F, France. In the source intensity column, X represents the downwind position. Source: Aubinet, 2008.
1.4 Numerical Methods in Canopy Flow

In the numerical studies of canopy flow, the canopy is usually considered horizontally homogeneous and isotropic, and structural diversity, flexibility and elasticity of plant part are neglected. These simplified assumptions make it possible to characterize the canopy by the leaf area density (LAD) which is only dependent on vertical distribution. A drag force is introduced for relating the flow resistance caused by canopy elements (Shaw and Schumann, 1992),

\[ F_{dh} = -C_d a |U| \bar{u} \]  

(1.4)

The drag force term has been implemented in numerical models. In addition, the canopy elements generate turbulent wake by converting the mean kinetic energy (MKE) into turbulent kinetic energy (TKE) at the canopy-elements scale, resulting in high turbulence within canopy. The energy converted by wake production from MKE to TKE is equal to the work done by the flow against drag force in the wake of canopy elements.

\[ P_w = |U| F_{dh} = -C_d a | \bar{u} | U^2 \]  

(1.5)

The first-order closure models have been used in canopy flow assuming that fluxes within a plant canopy are governed by local diffusion (Wilson et al., 1998; Pinard and Wilson, 2001; Ross and Vosper, 2005; Katul et al., 2006).

\[ F_s = -\rho K_s \frac{\partial \bar{s}}{\partial z} \]  

(1.6)

where \( F_s \) is the vertical flux density of a property with mean concentration \( s \) per unit mass, \( K_s \) is turbulent diffusivity. However, the mean fluxes are reported against the corresponding mean gradient (Denmead and Bradley, 1985). Raupach (1987) proposed that canopy flow is controlled by a diffusive contribution, representing flux from sources at distances exceeding the local Lagrangian integral length-scale, and a non-diffusive contribution from nearby sources. The non-diffusive contribution from coherent eddies with length-scale of the order of the canopy height.
leads to counter-gradient fluxes (Raupach, 1996), which is theoretically unable to be simulated by first order closure models. When mixing length scheme is applied to first-order closure models,

$$K_x = l_m^2 \frac{\partial \overline{u}}{\partial z}$$  \hspace{1cm} (1.7)

where $l_m$ is the mixing length, which is assumed constant in the canopy (Ross and Vosper, 2005; Katul et al., 2006) and increases linearly above the canopy. Mixing length must satisfy von Karman’s rule, $l_m = \kappa \left| \frac{d \overline{u}}{dz} / \frac{d^2 \overline{u}}{dz^2} \right|$, where $\kappa$ is Von Karman’s constant. The mixing length is minimum at the extreme values of the wind profiles (second maximum wind speed at trunk region and minimum wind speed at the level around maximum foliage density), where $d\overline{u}/dz = 0$ and $d^2 \overline{u}/dz^2 \neq 0$ and reaches a maximum at the inflection point of the wind profile, where $d\overline{u}/dz \neq 0$ and $d^2 \overline{u}/dz^2 = 0$. The varying mixing ratio within canopy with S-shaped wind profile has been predicted by an analytic model (Finnigan and Belcher, 2004), which is contradicted by the constant mixing length assumption.

As an alternative, the higher-order closure is capable of simulating non-local, non-diffusive second order moments in canopy flow and eliminates eddy viscosity in modeling Reynolds stress. Wilson and Shaw (1977) first proposed the second-order closure scheme for canopy flow. Three terms, the pressure-strain term, momentum flux-transport term, and turbulent kinetic energy dissipation rate term in the second-order closure model require closure parameterization with prescribed length scale or time scale. The length scales are introduced in approximation for the second-order moments with many adjustable and empirical constants. Wilson and Shaw (1977) used an empirical length scale varying with height within the canopy, while (Wilson, 1988) proposed a relaxation timescale defined by the ratio of turbulent kinetic
energy and mean kinetic energy dissipation rate. The justification of these two parameterizations is controversial (Katul and Chang, 1999). Wilson and Shaw’s model had a better prediction of mean velocity near the forest-atmosphere interface while Wilson’s model better reproduced the longitudinal velocity standard deviation. Both failed to model the third moments for flux gradient approximation in the flux-transport term, which is another deficiency of the second-order closure models. Triple-velocity correlation (flux-transport term) in the second-order closure models is parameterized by flux-gradient approximation resulting in unrealistic momentum flux transport profiles near the top of the canopy for canopy flows (Katul et al., 1998). Meyers and Paw U (1986) applied third-order closure to improve the triple-velocity correlation terms, in which the fourth moment is related to second moments by a zero-fourth cumulant expansion. However, the third order closure model didn’t show improved prediction over the flux-gradient approximation, especially the prediction of velocity variance profiles (Katul et al., 1998).

Since the pioneering works of Smagorinsky (1963), Lilly (1967), Deardorff (1970), and Leonard (1974), large-eddy simulation (LES) has been widely applied in studies of atmospheric boundary layer. The advantage of LES over high-order closure models is the ability to resolve large scale eddy flow, rather than relying on parameterizations of turbulence fluxes. LES models the smallest scale eddies rather than solving them in direct numerical simulation. Shaw and Schumann (1992) first applied Large-eddy simulation (LES) for canopy flow. Brown et al. (2001) first applied LES to canopy flow over hills. The results of LES are in better agreement with wind tunnel experimental data than the results of a first order mixing length model. LES have been confirmed to be able to reproduce many turbulent characteristics of canopy flow and improved our understanding of canopy flow with complex conditions, such as canopy flow at the forest edge (Yang et al., 2006a; b; Dupont and Brunet, 2008) and over complex terrain (Dupont et al.,
The main problem for LES has been the computer power, for the requirements of very high resolution on the rough surface. LES with inadequate resolution could produce worse results than the mixing-length closure with even coarser resolution (Hobson et al., 1999). In addition, the application of LES is limited to stably stratified regions near the surface where turbulence scales became small, which cannot be explicitly resolved in the sub-grid scale modeling (Mason, 1994).

With parameterization of canopy effects, numerical simulations have been capable of modeling the mean and turbulent properties and reproducing the small scale structures of canopy flow, such as recirculation zones occurring behind hills reproduced by numerical modeling and coherent structure, which is demonstrated to be responsible for upward and downward transfer of momentum and scalar through low frequency ‘sweeps’ and ‘ejections’ (Thomas and Foken, 2007). Currently, these numerically reproduced canopy flows are confined to ideal conditions: (1) Neutral (Ross and Vosper, 2005; Dupont et al., 2008; Ross, 2008; Patton and Katul, 2009; Katul et al., 2006) or weakly unstable (Wang, 2010) atmospheric stability. During neutral and unstable periods, the boundary layer is better mixed due to turbulent and convective mixing. The effects of canopy flow on mass and energy transport are relatively small. While the boundary layer is stably stratified, some of the processes are identified to be responsible for the uncertainties in nocturnal measurement of CO₂ and energy balance, such as, turbulent ramps (Cava et al., 2004), small-scale turbulence (Mahrt and Vickers, 2005) and intermittent turbulence (Mahrt and Vickers, 2002; Doran, 2004; Nakamura and Mahrt, 2005); (2) Flat terrain with homogeneous and extensive canopy (Huang et al., 2009; Dupont et al., 2010). In practice, most of the FLUXNET sites are on undulating and sloping terrain. Some of the uncertainties associated with
complex terrain have been reported, such as advection problem and energy imbalance, which are identified as a result of thermal driving slope flows (Sun et al., 2007; Feiginwinter et al., 2010).
Chapter 2 Methods

2.1 Conservation of Mass and Momentum

Computational fluid dynamics (CFD) uses numerical methods and algorithms to solve and analyze problems that involve fluid flows. The Navier–Stokes (N-S) equations are the basic governing equations for almost all CFD problems. Three dimensional time dependent N-S equations can be expressed in a Cartesian coordinate system.

The momentum and mass balance equations in the canopy sub-layer can be written as:

\[ \frac{\partial \rho}{\partial t} + \frac{\partial \bar{u}_j}{\partial x_j} = 0 \]  \hspace{1cm} (2.1)

\[ \frac{\partial \bar{u}_i}{\partial t} + \bar{u}_j \frac{\partial \bar{u}_i}{\partial x_j} = -\frac{1}{\rho} \frac{\partial P}{\partial x_i} + \nu \frac{\partial^2 \bar{u}_i}{\partial x_i \partial x_j} - \frac{\partial}{\partial x_j} \left( \bar{u}_i \bar{u}_j' \right) - g \beta (\bar{\theta} - \theta_\infty) - F_{Di} \]  \hspace{1cm} (2.2)

\[ \frac{\partial \bar{\theta}}{\partial t} + \bar{u}_j \frac{\partial \bar{\theta}}{\partial x_j} = \Gamma \frac{\partial^2 \bar{\theta}}{\partial x_i \partial x_j} - \frac{\partial}{\partial x_j} \left( \bar{\theta} u_j' \right) + Q_{\text{source}} \]  \hspace{1cm} (2.3)

where \( \bar{u}_i \) and \( \bar{u}_j \) are the mean velocity components along the \( x_i \) and \( x_j \) direction, respectively. \( \bar{\theta} \) is the mean potential temperature, \( u_i' , u_j' \) and \( \theta' \) are the fluctuations from their mean value \( \bar{u}_i \) and \( \bar{u}_j \) and \( \bar{\theta} \), \( \rho \) is the air density. \( \nu \) is kinematic viscosity of air, \( P \) is the deviation of pressure from its reference value, and \( \beta \) is the thermal expansion coefficient of air. \( \theta_\infty \) is the reference temperature, \( g_i \) is the gravity acceleration in \( i \) direction, and \( \Gamma = \nu / P' \) is turbulent viscous diffusion coefficient. \( P' \) is turbulent Prandtl number, \( Q_{\text{source}} \) is the energy source and \( F_{Di} \) is the drag force exerted by the canopy elements:

\[ F_{Di} = \frac{1}{2} K_r u_i |U|, \]  \hspace{1cm} (2.4)
where $K_r$ is the resistance coefficient, which is derived from an empirical relationship given by Hoener (1965),

$$K_r = \frac{1}{2} \left[ \frac{3}{2\phi} - 1 \right]^{2},$$

(2.5)

where $\phi$ is porosity of the canopy layer, which can be obtained from leaf area density (Gross, 1993),

$$\phi = \frac{\sqrt{1 + 4\alpha} - 1}{2\alpha},$$

(2.6)

$F_D$ is zero above the canopy.

### 2.2 Conservation of Scalar Quantities

The conservation for scalar CO$_2$ with mean molar mixing ratio $c$ is given by

$$\frac{\partial \bar{c}}{\partial t} + \bar{u}_j \frac{\partial \bar{c}}{\partial x_j} = D \frac{\partial^2 \bar{c}}{\partial x_i \partial x_j} - \frac{\partial}{\partial x_j} \left( \bar{c}' u'_j \right) + S_c,$$

(2.7)

where $\bar{u}_j$ is solved by Eq. 2 and 3, $c'$ is the fluctuation from its mean value $\bar{c}$, $D$ is the molecular diffusivity of CO$_2$, $S_c$ is the source of CO$_2$. The sources of CO$_2$ from ecosystem are associated with heterotrophic respiration ($R_h$) from soil and autotrophic respiration ($R_a$) from plant leaves/stems.

In WFIS model, Sun et al., (2006) ignored the autotrophic respiration because leaf and stem respiration contributes much less than soil respiration. In this research, CO$_2$ source term $\bar{S}_c$ is parameterized with total ecosystem respiration (TER) of soil respiration ($R_S$) and above ground respiration ($R_L$), where $R_S = \sigma \cdot R_L$. Here $\sigma$ is proportional coefficient fulfilling $TER = R_S + R_L$. The net contribution of CO$_2$ from the soil respiration is quantified as the Q10 exponential relation
with temperature (van t’Hoff, 1884). The empirical formula obtained at the Renon site to quantify soil respiration is:

\[
R_S = R_{\text{ref}} \cdot Q_{10}^{(T - T_{\text{ref}})/10}
\]

where \( R_{\text{ref}} = 3.69 \text{ mol m}^{-2} \text{ s}^{-1} \) is the respiration rate at reference temperate \( T_{\text{ref}} = 10^\circ \text{C} \), \( Q_{10} = 3.64 \) is the factor by which \( R_S \) increases for an increase in soil temperature of \( a = 10^\circ \text{C} \) (Montagnani et al., 2009).

Foliage respiration is assumed to be only exponentially related to air temperature (Figure 2.1, Law et al., 2001; Okhubo et al., 2007; Urban et al., 2007), however, the measurements indicated that for a single tower at Renon site, the vertical temperature difference within the canopy is about 2-3°C at the same moment. It is assumed that at this temperature difference, the temperature dependence of respiration is negligible and the respiration above ground depends only on the leaf area density. As suggested by Janssens et al. (2001), we assume soil respiration accounts for about 63% of forest ecosystem respiration. Therefore, total respiration above ground is \( R_L = (37/63) \cdot R_S \). For each vertical layer \( z_k \), above ground respiration is

\[
R_L(z_k) = R_L \left\{ \frac{1}{LAI} \int_{z_k/2}^{z_k+1/2} a(z)dz \right\}
\]
Figure 2.1 The exponential dependency of foliage respiration on temperature.

Source: Law et al., 2006.
2.3 RNG k-\(\epsilon\) Model

The RNG model was developed using Re-Normalization Group (RNG) methods by Yakhot and Orszag (1986a; b) to renormalize the Navier-Stokes equations, to account for the effects of smaller scales of motion. The RNG \(k - \epsilon\) turbulence model does not involve any experimentally adjustable parameters and does not use mixing-length theory (Smith and Reynolds, 1992). In the standard k-\(\epsilon\) model, the eddy viscosity is determined from a single turbulence length scale (e.g., mixing-length), so the calculated turbulent diffusion is that occurs only at the specified scale, whereas in reality all scales of motion will contribute to the turbulent diffusion. The RNG approach, which is a mathematical technique, can be used to derive a turbulence model similar to the k-epsilon, resulting in a modified form of the epsilon equation which attempts to account for the different scales of motion through changes to the production term. Speziale and Thangam (1992) compared the modeling results from RNG k-\(\epsilon\) turbulent model with standard k-\(\epsilon\) model and Chen’s k-\(\epsilon\) model (Chen and Kim, 1987). It is acknowledged that k-\(\epsilon\) model did a better simulation when applied to two-dimensional valley flow (Maurizi, 2000). Kim and Patel (2000) suggested using RNG k-\(\epsilon\) model to predict the pollutant transport under neutral conditions.

In RNG k-\(\epsilon\) model, the Reynolds stress in equation (2.2), turbulent heat flux in equation (2.3) and turbulent CO\(_2\) flux in equation (2.7) are solved by turbulent viscosity, respectively, as

\[
-u'_i u'_j = \mu_t \left( \frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i} \right) - \frac{2}{3} \delta_{ij} k ,
\]

\[
-\theta' u' = \mu_\theta \frac{\partial \theta}{\partial x_j} ,
\]

\[
-c u'_j = \mu_c \frac{\partial \bar{c}}{\partial x_j}.
\]
where $\mu_t, \mu_\theta = \mu_t / P_t$ and $\mu_c = \mu_t / S_c$ are the turbulent viscosities of momentum, heat and CO2, respectively, $\delta_i$ is Kronecker delta, and $k$ is the turbulent kinetic energy. The turbulent Schmidt number $S_c$ is 0.6 as suggested by Flesch et al. (2002). The Turbulent Prandtl number $P_r$ is 0.5.

RNG $k-\varepsilon$ model assumes that turbulence viscosity in Equation (2.10) is related to turbulence kinetic energy $k$ and dissipation $\varepsilon$

$$\mu_t = \rho C_\mu \frac{k^2}{\varepsilon},$$  \hspace{1cm} (2.13)

where $k$ and $\varepsilon$ are determined from the coordinate-invariant semi-empirical transport equations; $C_\mu$ is a dimensionless constant.

The prognostic equation of turbulent kinetic energy and its dissipation are written as:

$$\frac{\partial k}{\partial t} = - \frac{\partial}{\partial x_j} (\bar{u}_i k) - \frac{\partial}{\partial x_i} \left( - \frac{\mu_t}{\sigma_k} \frac{\partial k}{\partial x_i} \right) + P_s + P_b + P_w + T_p - \varepsilon,$$  \hspace{1cm} (2.14)

$$\frac{\partial \varepsilon}{\partial t} = - \frac{\partial}{\partial x_j} (\bar{u}_j \varepsilon) - \frac{\partial}{\partial x_i} \left( - \frac{\mu_t}{\sigma_\varepsilon} \frac{\partial \varepsilon}{\partial x_i} \right) + C_{\varepsilon 1} \frac{\varepsilon}{k} P_k - C_{\varepsilon 2} \frac{\varepsilon^2}{k} - S,$$  \hspace{1cm} (2.15)

where $P_s$ is shear production given by

$$P_s = \mu_t \frac{\partial \bar{u}_i}{\partial x_j} \left( \frac{\partial \bar{u}_i}{\partial x_j} + \frac{\partial \bar{u}_j}{\partial x_i} \right),$$  \hspace{1cm} (2.16)

$P_b$ is buoyancy production, given by

$$P_b = - \mu_t g_i \beta \left( \frac{\partial \bar{\theta}_j}{\partial x_i} \right)$$  \hspace{1cm} (2.17)

$P_w$ is wake production caused by canopy elements

$$P_w = \bar{u}_i F_{\text{wake}} = C_d a |U| \bar{u}_i^2$$  \hspace{1cm} (2.18)
$T_p$ is pressure collection term, it is calculated as residual of other TKE components.

$S$ is a volumetric source term given by

$$S = \frac{C_\mu \eta^3 \left(1 - \frac{\eta}{\eta_0}\right)\varepsilon^2}{(1 + \beta_0 \eta^3)k},$$

(2.19)

$$\eta = \frac{k}{\varepsilon} \left[\frac{P_k}{\mu_t}\right]^{1/2},$$

(2.20)

The empirical constants $C_\mu$, $\sigma_k$, $\sigma_\varepsilon$, $C_{\varepsilon_1}$, $C_{\varepsilon_2}$, $\beta_0$, and $\eta_0$ are 0.0845, 0.7194, 0.7194, 1.42, 1.68, 0.012, and 4.38 (Yakhot and Orszag, 1986a; b).
Chapter 3 Neutrally Stratified Canopy Flow

Abstract

Flow distortion over a forested hill is asymmetric, forming a recirculation region on the lee slope that increases the complexity in understanding atmosphere-biosphere interaction. To understand the complexity, we examine the effect of the geometry of forested hills on recirculation formation, structure, and related CO$_2$ transport by performing numerical simulations over double-forested hills. The ratio ($= 0.8$) of hill height ($H$) to half length ($L$) is a threshold value of flow patterns in the recirculation region: below 0.8, sporadic reversed flow occurs; at 0.8 one vortex is formed; and above 0.8 a pair of counter-rotating vortices are formed. The depth of recirculation increases with increasing $H/L$. The contribution of advection to the CO$_2$ budget is non-negligible and topographic-dependent. Vertical advection is opposite in sign to horizontal advection but cannot exactly offset in magnitude. Height-integrated advection shows significant variation in fluxes across hills. Gentle slopes can cause larger advection error. However, the relative importance of advection to CO$_2$ budget is slope-independent.

Keywords: Advection CO$_2$ flux, Complex terrain, Forested hill, Recirculation
3.1 Introduction

A large part (70%) of the earth’s land surface is covered with mountains and hills. These rugged surfaces, particularly those covered with forests, distort airflows near the ground and create complexity in understanding land-atmosphere exchanges of mass and energy (Whiteman, 2000; Kaimal and Finnigan, 1994). One of the most important features in the distorted flows is recirculation formed behind forested mountains or hills. These recirculation bubbles (regions) operate through different mechanisms to influence mass and energy exchange between forests and atmosphere, such as wind direction alteration and scalar redistribution (Katul et al., 2006; Ross, 2011). Because of its importance, studies of recirculation over a forested hill have recently received more attention in experiments, numerical simulations and analytical models. Almeida et al. (1993) found in their water tunnel experiments that the length of recirculation formed behind a single hill is longer than that between multiple hills. Katul’s research group at Duke University has investigated the influence of canopy density on recirculation formation by flume experiments (Poggi and Katul, 2007). Recirculation formation behind forested hills has been studied by many numerical models (e.g. Dupont et al., 2008; Ross, 2008). These numerical simulations have focused on the turbulent characteristic of recirculation. Although Yi (2009) pointed out that a terrain slope is one of the most important control factors to terrain-induced canopy flows, the impact of terrain slopes on recirculation development is still poorly understood.

The goal of this paper is to explore the dependence of recirculation development on terrain geometry by using CFD (Computational Fluid Dynamics) method. We employed the renormalization group (RNG) k-e model developed by Yakhot and Orszag (1986) to simulate airflows in double-forested hills of different shapes to analyze formation conditions and the structure of recirculation. In order to understand how the scalar transport processes are affected
by recirculation development and its complexity, we studied the distribution of advective CO$_2$ fluxes over forested hills with different hill shapes. The numerical model designs of the simulation are described in section 3.2. Recirculation development is discussed in section 3.3. CO$_2$ transport and advections are discussed in section 3.4, and concluding remarks are in section 3.5.

### 3.2 Numerical Implementation

The computational domain extends over 1400×200 m in a Cartesian grid, corresponding to 700×151 grid intervals in the $x$ and $y$ directions. Double hills covered with 15 m tall forest canopy occupy the middle 200 m domain horizontally. The horizontal resolution at forested hills is 1 m and at the bare flat ground is stretched with a power law, starting with a horizontal grid spacing of 1 m at the edge of the forest. The meshes are stretched in the vertical using a power law, starting with a vertical grid spacing of 0.8 at the surface. The stretching powers in horizontal and vertical are 1.15 and 1.1, respectively. Ground surface roughness height is set to be 0.01. Wind velocity at west boundary is set to be a constant, $u = 3$ m s$^{-1}$. The upstream of the velocity field is fully developed by the bare ground to be logarithmic velocity profile before reaching the double hills. Pressure is fixed at the top boundary and east boundary, where the pressure is close to 0.0 Pa, relative to the external pressure.

In this study, the topography is specified with double sinusoidal hills (Figure 3.1) to include both valley and ridges. The shape function of the hill is defined as

$$H(x) = \frac{H}{2} \cos(\frac{\pi x}{2L} + \pi), \quad (3.1)$$
where $H$ is the hill height, $L$ is the half length scale, $x$ is longitudinal distance with $x = 0$ at the left trough of the first hill. Variation of the slope ($H/L$) is achieved by changing $H$ with a constant $L = 25$ m.

The porous canopy layer is designed to be horizontally homogeneous along the slope and vertically uniform. Leaf area density ($a$) is specified as mean values from observation (Yi et al., 2005). The hill surface is assumed to be a source of atmospheric CO$_2$. The efflux rate is 4 µmol m$^{-2}$ s$^{-1}$. 
**Figure 3.1** Schematic diagram of double forested hills. $H$ and $L$ are the height and half-length of hills, respectively. The total length of the double hills is $8L$. $h$ denotes the height of canopy. The symbols a-i indicate the locations of nine hypothetical sites across double hills. The horizontal distance between two locations is $L$.

**Table 3.1** Parameters of hill and canopy properties in the model

<table>
<thead>
<tr>
<th>Parameters</th>
<th>Values</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Hill</strong></td>
<td></td>
</tr>
<tr>
<td>$L$ (m)</td>
<td>25</td>
</tr>
<tr>
<td>$H/L$</td>
<td>0.2, 0.4, 0.6, 0.8, 1.0,</td>
</tr>
<tr>
<td><strong>Canopy</strong></td>
<td></td>
</tr>
<tr>
<td>$h$ (m)</td>
<td>15</td>
</tr>
<tr>
<td>$LAI$ (m$^2$ m$^{-2}$)</td>
<td>3.2</td>
</tr>
<tr>
<td>$a$ (m$^{-1}$), $\Phi$</td>
<td>0.22, 0.85</td>
</tr>
</tbody>
</table>
3.3 Recirculation in Forested Hill

3.3.1 Flow Distribution

Figure 3.2 shows the distributions of streamwise velocity, vertical velocity, and CO$_2$ concentration in forested hills with different slopes ($H/L$). It is not surprising that airflow accelerates up the windward slopes and reaches a maximum at crests, and then decelerates on the leeward slopes to the feet of hills, resulting in flow stagnation behind hills, as predicted in many previous studies (Finnigan and Belcher, 2004; Belcher et al., 2008). However, our results indicate that the position of the maximum velocity at crests shifts from above canopy into canopy as slope increases. The shapes of wind profiles on leeward slope and in the valley are substantially different: (1) being logarithmic with gentle slope ($H/L < 0.8$), and (2) exponential as $H/L > 0.8$ (Figure 3.3). The transition of the shape of the wind profile behind hills from gentle to steep hill (Figure 3.3d, e, h, i) occurs because stagnation becomes stronger and reversed flows appear as slope increases.

The behavior of vertical velocity is more complicated. The most important feature is that the profile of vertical velocity is inflected at the top of the canopy (Figure 3.4), i. e. the continuity of vertical velocity gradient ($\partial \bar{v}/\partial z$) is broken at the top of canopy. The vertical velocity gradient is important in calculating vertical advection CO$_2$ flux. Lee (1998) proposed a method of calculating vertical CO$_2$ flux by assuming that vertical velocity gradient is constant, in which vertical velocity is assumed to linearly increase from ground to the sensor height of an eddy-flux tower. Lee’s method has been widely used in eddy flux communities. However, our simulations indicate that Lee’s assumption may be challenged in complex terrain. The vertical velocity is largely inflected at the top of the canopy near forest edges (Figure 3.4a, i). The assumption of constant vertical velocity gradient is valid only for above canopy or within the
canopy, but the magnitude and sign of the vertical velocity gradient are different between above and below the top of canopy (Figure 3.4a, i). Inflection of vertical velocity around the top of canopy is a common characteristic for all locations across forested hills with low slopes ($H/L \leq 0.4$, Figure 3.4). Increasing of slopes enhances vertical velocity significantly. The behavior of the vertical velocity profile becomes complicated on leeward slopes and in the valley (Figure 3.4d, e, f, h) for different slopes, and is closely related to recirculation formation and structure (see discussions in next section).
Figure 3.2 Mean field distribution of streamwise velocity ($u$, left panel), vertical velocity ($w$, middle panel) and CO$_2$ mixing ratio ($c-c_0$, right panel) for $H/L = 0.4, 0.8, 1.2$. Streamwise velocity and vertical velocity are normalized by the velocity on the top of canopy for each $H/L$. The CO$_2$ mixing ratio (in ppm) is represented by the difference from the mean atmospheric CO$_2$ mixing ratio $c_0$. White solid lines indicate the top of canopy.
**Figure 3.3** Profiles of streamwise velocity \( (u, \text{ m s}^{-1}) \) at the hypothetical locations (a-i, shown in Figure 3.1) across double hills for \( H/L = 0.2, 0.4, 0.6, 0.8, 1.0, 1.2, \) and 1.6.
**Figure 3.4** Profiles of vertical velocity ($w$, m s$^{-1}$) at the hypothetical locations (a-i, shown in Figure 3.1) across double hills for $H/L = 0.2, 0.4, 0.6, 0.8, 1.0, 1.2, \text{ and } 1.6$. Note that $w$ on the slope surface (location b, d, f, and h) is not zero due to the centers of bottom grid cells in the numerical calculation are not exactly at the surface.
3.3.2 Recirculation Development

Recirculation is characterized by reversed flow behind hills. It is a region where the flow is decoupled from above mean airflow. Our numerical simulation results indicate that the flow patterns in the recirculation are closely related to hill geometry, which is shown in Figure 3.5. $H/L = 0.8$ is a threshold value for flow-pattern formation in recirculation. Our numerical results with gentle hills ($H/L < 0.8$) show that the recirculation is characterized by flows with sporadic negative streamwise velocities at low levels of the canopy layer (Figure 3.5a). When $H/L$ is about 0.8, one clockwise vortex is formed in the valley, inclining toward the leeward slope of the hills (Figure 3.5b). The inclination of the recirculation vortex behind the second hill is stronger than that of the first hill. The top of a single recirculation vortex can reach the middle canopy layer. When $H/L$ is greater than 0.8, the reversed flow is in the middle to upper canopy layer with a counter-rotating vortex pair (Figure 3.5c), which is a common phenomenon on the leeside of mountain barriers (Smolarkiewicz and Rotunno, 1989; Bauer et al., 2000; Epifanio, 2003). The upper vortex is clockwise and the lower vortex is anti-clockwise. The height of the upper vortex increases with increasing slope and can reach $2.5h$ at $H/L = 1.6$. In comparison with the recirculation formed behind the second hill that extends farther away from the second leeward slope, the recirculation formed in the valley is deeper because flow in the valley is deflected upward by the second hill.

The recirculation formation in the valley or leeward side can be understood by the relative contribution of three major forces within the canopy: adverse pressure gradient (Figure 3.6), Reynolds stress gradient, and drag caused by canopy elements. In the upper canopy layer, although adverse pressure gradient and canopy drag act to decelerate mean flow on the leeward slope, the Reynolds stress gradient is large enough to maintain positive streamwise velocity. In
the lower part of the canopy, airflows are dominated by adverse pressure gradient because the observed Reynolds stress profile is exponential (Yi, 2008) (Figure 3.6) and hence its gradient is very weak. Canopy drag is also weak in the lower part of the canopy because wind speed is low. Thus, reversed flow in the lower part of the canopy (Figure 3.5a, b) results mainly from adverse pressure gradient. As slope increases, adverse pressure gradient becomes larger because it is theoretically proportional to the height of hill (Finnigan and Belcher, 2004). Consequently, adverse pressure gradient becomes the dominant control even in the upper canopy layer for deep slope hills, which causes the upper vortex formation in upper and just above the canopy layer (Figure 3.5c). The lower vortex is a result of dynamic response to the upper vortex.

We define the recirculation depth $H_R$ as the mean height of the flows with streamwise velocity $u = 0$. The depth of recirculation in the valley increases quasi-linearly with slope for both forested and bared hills ($R^2 > 0.98$, Figure 3.7a). However, recirculation is deepened by vegetation canopy. The relative contribution of vegetation to the depth of recirculation decreases with $H/L$ (Figure 3.7b). This implies that the effects of canopy layer on recirculation are diminished as slope increases.
Figure 3.5 The streamlines of recirculation patterns for different $H/L$: (a) $H/L = 0.4$ (top panel), sporadic reversed flow; (b) $H/L = 0.8$ (middle panel), one vortex; and (c) $H/L = 1.2$ (bottom panel), counter-rotating vortices pair. $H/L = 0.8$ as a threshold value above which two vortices exist, below which no vortex appears, and equal to which only one vortex exists, is valid under condition that the ambient wind satisfies, $1 \leq \bar{u} \leq 5$ (m s$^{-1}$). The colored background indicates mean wind speed (m s$^{-1}$). Grey solid lines indicate the top of canopy. White solid lines are streamlines.
Figure 3.6 Deviation of pressure (Pa) from its reference value in the hilly terrain for $H/L = 0.8$ (upper panel) and a schematic diagram depicting the adverse pressure gradient ($-PG$) around a single hill which is positive on windward slope and negative on the leeward slope (lower panel). $\tau_h$ and $\tau_0$ are Reynolds stress on the top of canopy and at the ground surface, respectively. The exponential Reynolds stress profile is predicted by a canopy momentum transfer (CMT) model developed by Yi (2008). The dash lines indicate the top of canopy.
Figure 3.7 (a) Depth of recirculation ($H_R$) in the valley with slope ($H/L$) for forested hill and bare hill. $H_R$ is the mean height at which level streamwise velocity $u$ is zero in the valley. (b) Relative contribution of canopy to $H_R$ versus slope, $H/L$. Here, 

$$\beta = \left( \frac{H_R' - H_R^b}{H_R^b} \right) \times 100\%.$$  

$H_R'$ is the depth of recirculation with canopy and $H_R^b$ the depth of recirculation without canopy.
3.4 CO$_2$ Transport with Recirculation

3.4.1 CO$_2$ Distribution

The distribution of CO$_2$ in double-forested hills (Figure 3.2, right panel) is primarily dictated by flow patterns of recirculation. When $H/L$ is smaller than 0.8, ejection from upper canopy occurs behind hillcrests. The ejection has little effect on CO$_2$ transfer because CO$_2$ released from slope surfaces is confined to a very shallow surface layer. CO$_2$ concentration in the recirculation region behind hills could be 4 - 5 ppm higher than on windward slopes and crests. When $H/L$ is 0.8, the ejection from the lower canopy behind hillcrests can carry CO$_2$ to the atmosphere above-canopy in the valley. Large amounts of CO$_2$ from slope surface circulate in the single recirculation vortex. Because the top of the single vortex is within the middle canopy level, it prevents CO$_2$ from venting out of the canopy layer, causing as much as a 13 ppm difference in CO$_2$ concentration between the recirculation vortex on the leeside slope and above-canopy. When $H/L$ is greater than 0.8, CO$_2$ is partly ejected from the canopy layer to the atmosphere above the canopy in the valley. Transfer of CO$_2$ into the recirculation region is split into two streams behind hillcrest. One moves down-slope, mixing with respired CO$_2$ from the deep valley, and causing CO$_2$ to be trapped in the lower recirculation vortex. The other moves downstream then diverts downward upon hitting the windward slope of the second hill, resulting in CO$_2$ being stored in the upper recirculation vortex. The upper vortex that is located across the top of the canopy is responsible for CO$_2$ transport from within the canopy to the atmosphere above. Thus, CO$_2$ concentration in the lower vortex is much higher than in the higher vortex. A double vortex system makes it possible for CO$_2$ to ventilate from the deep valley to above the canopy, leading to a smaller CO$_2$ gradient in the canopy layer in comparison with the non-vortex and single vortex recirculation.
3.4.2 Advective Fluxes

In this study, we explore the CO$_2$ transfer characteristic in the steady-state neutral atmosphere. Thus, the storage term of CO$_2$ is not taken into consideration. In steady state, NEE equation (1.3) is expressed as

$$\text{NEE} = \int_0^{z_r} \frac{\partial (w'c')}{\partial z} dz + \int_0^{z_r} \left( \frac{\partial \overline{w^2}}{\partial x} + \frac{\partial \overline{w}}{\partial z} \right) dz + \int_0^{z_r} \frac{\partial (u'c')}{\partial x} dz$$

(3.2)

where $z_r$ is a reference height above canopy, I is integrated turbulent flux, II is integrated advection fluxes and III is integrated horizontal divergence. The eddy covariance technique (EC) is the most widely used method for quantifying ecosystem carbon flux (Baldocchi et al., 2001; Baldocchi, 2003). The method was developed to work on perfectly flat topography and homogeneous land-cover. But it generates significant errors when used to estimate net ecosystem-atmosphere exchange due to neglect of non-turbulent advection fluxes (Equation 3.2, II) and horizontal divergence (Equation 3.2, III) in complex terrain with heterogeneous vegetation (Aubinet, 2008; Sun et al., 2007).

In order to evaluate the influence of recirculation on the CO$_2$ fluxes, we first evaluate the role of terms I, II and III in NEE. $z_r$ is set to be triple the canopy height ($3h$) and terms I, II and III are averaged along slope to indicate the average contribution of I, II and III to NEE (Table 3.2). Height-integrated horizontal divergence III is small (III $\ll$ I) in comparison with height-integrated turbulent flux (I) as the slope is gentle ($H/L \leq 0.6$). As recirculation vortices appear ($H/L \geq 0.8$), the ratio of height-integrated horizontal divergence (III) to height-integrated turbulent flux (I) increases. The magnitude of III increases to about 37% of II as $H/L = 1.6$ due to greater horizontal divergence in the valley. The height-integrated advection flux (II) is comparable with height-integrated turbulent flux (I) in magnitude but opposite in sign. The value
of height-integrated advection flux (II) is over 60% of height-integrated turbulent flux (I) except at $H/L=1.6$, where it is 30% of height-integrated turbulent flux (I). The ratio of height-integrated advection flux (II) to height-integrated turbulent flux (I) is independent of slope. The neglect of advection fluxes can cause significant errors in NEE estimations by EC methods in the complex terrain. To further understand the role of advection fluxes in total NEE estimations for different terrain geometry and different positions on the terrain, we use nine hypothetical sites (Figure 3.1a-i) to illustrate the spatial variation of advection fluxes across double hills.

Figure 3.8 shows the distribution of horizontal advective flux $F_h = \bar{u} \partial \bar{c} / \partial x$ (left panel), vertical advective flux $F_v = \bar{w} \partial \bar{c} / \partial z$ (middle panel) and total advective flux $F_T = F_h + F_v$ (right panel) across double hills. Distribution of simulated advective fluxes is strongly dependent on hill geometry. When $H/L$ is smaller than 0.8, advective fluxes are important in a very shallow layer within canopy. Horizontal CO$_2$ flux is positive on the windward slope and negative on the leeward slope. However, the influence of recirculation, which is weak as $H/L < 0.8$, on horizontal advective fluxes is demonstrated by a very shallow layer with positive value in the lower part of the leeside slope. The sign of $F_v$ across the forested hills is mainly determined by the sign of vertical velocity because vertical CO$_2$ gradient is always negative. The relatively smaller magnitude of advective fluxes observed over the first hill as compared to the second can be explained by the fact that there is better air mixing on the first hill (stronger wind) than the second hill.

When $H/L$ is greater than 0.8, recirculation vortices cause CO$_2$ to vent out of the deep canopy layer and result in larger advective fluxes scattered and skewed in the valley and in a shallow layer on the slopes. When $H/L = 1.6$, the advective fluxes can become significant even at the level of three-canopy height in the valley. Our simulations demonstrate that $F_h$ and $F_v$ are
generally opposite in sign at the same location. However, \( F_h \) and \( F_v \) cannot exactly offset each other in magnitude as demonstrated in field experiments (Aubinet et al., 2003; Feigenwinter et al., 2008; Yi et al., 2008).

Figure 3.9 shows the height-integrated advective fluxes at nine hypothetical sites. The sign and magnitude of height-integrated advective fluxes depend on site-locations and slopes. When \( H/L < 0.8 \), height-integrated \( F_h \) and \( F_v \) are smaller in troughs (a, e, i) and over crests (c, g) but larger on slopes (b, d, f, h). This is not surprising because the maximum wind velocity over crests can cause better mixing of CO\(_2\), thus very low CO\(_2\) gradient. Although CO\(_2\) gradient is high in the troughs, the stagnated flow leads to small advective fluxes. The location of minimum magnitude of height-integrated total advective flux \( F_T \) varies with slopes, which is at h as \( H/L = 0.2 \), at g as \( H/L = 0.4 \) and at f as \( H/L = 0.6 \) (Figure 3.9).

Recirculation vortices cause large variations in height-integrated \( F_h \) and \( F_v \) across double hills when \( H/L \geq 0.8 \). Although recirculation pattern for \( H/L = 1.0, 1.2 \) and 1.6 are similar, variation in height-integrated \( F_h \) and \( F_v \) is significantly different, especially in the valley (d, e, f). When \( H/L = 1.2 \), the location of minimum magnitude of height-integrated \( F_T \) is at h. Even though the individual integrated \( F_h \) and \( F_v \) are significant in the valley as \( H/L = 1.0 \) and 1.6, height-integrated \( F_T \) are approximate to zero. The different locations of minimum height-integrated \( F_T \) indicate that advective error is strongly dependent on hill geometry and flux-tower location in hills.

In order to clarify the overall influence of hill geometry on total advective flux, \( F_T \) is integrated through the double-hill domain (Figure 3.10). We find that domain-integrated \( F_T \) decreases with increasing slope as \( H/L \leq 1.0 \). Although the relative importance of advective fluxes is slope-independent, the magnitude of advective error could be greater on gentle
topographies than steeper ones. We speculate that the decrease of domain-integrated $F_T$ with increasing slope is caused by increasing depth of recirculation that leads to better CO$_2$ mixing between vegetation and atmosphere.
Table 3.2 The ratio of Height-integrated advection flux (II) and horizontal divergence (III) to turbulent flux (I) in Equation 3.2

<table>
<thead>
<tr>
<th>$H/L$</th>
<th>$II/I$</th>
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<td>1.6</td>
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<td>-0.37</td>
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Figure 3.8 Spatial variation of advection components (in $\mu$mol m$^{-3}$ s$^{-1}$): horizontal advection ($\bar{u} \frac{\partial \bar{C}}{\partial x}$, left panel), vertical advection ($\bar{w} \frac{\partial \bar{C}}{\partial z}$, middle panel), and total advection ($\bar{u} \frac{\partial \bar{C}}{\partial x} + \bar{w} \frac{\partial \bar{C}}{\partial z}$, right panel) for $H/L = 0.4, 0.8, 1.2$. Dark solid lines indicate the top of the canopy.
Figure 3.9 Height-integrated advective fluxes (in µmol m$^{-2}$ s$^{-1}$) at hypothetical locations (a-i, shown in Figure 3.1). The advective fluxes are integrated from hill surface to triple the canopy top h. $\int F_h$, $\int F_v$ and $\int (F_h + F_v)$ are integration for horizontal advective flux, vertical advective flux and total advective flux, respectively.
Figure 3.10 Domain-integrated total advective flux in double-forested-hill. Total advective flux is integrated in the double-hill domain shown in Figure 3.8 to include advective fluxes below and above canopy. The Domain-integrated total advective flux is normalized by the domain-integrated total advective flux at $H/L = 0.2$. 
3.5 Conclusions

The impacts of hill geometry on flow and CO₂ transfer are explored over double-forested hills by numerical approach. We used a double-hill simulation setup because it can demonstrate flow characteristics not only through hills but also through a valley (i.e. between two hills). Flow recirculation is a typical phenomenon in complex terrain and plays a key role in vegetation-atmosphere exchanges of mass and energy. Our CFD predictions indicate that the complexity and structure of recirculation strongly depend on slope. For gentle forested hills ($H/L < 0.8$), the recirculation structure is simply characterized by reverse flows without vortex, which are limited in the lower part of the canopy layer on leeward sides. The near-surface reverse flows greatly alter CO₂ distribution near the ground rather than enhance CO₂ exchange in vertical. For steep forested hills ($H/L > 0.8$), recirculation bubbles become larger and even deeper than vegetation height with one or multiple vortices, enhancing mixing of CO₂ and energy between vegetation and atmosphere. Consequently, steep slopes cause less advective CO₂ fluxes. The conclusions from numerical experiments provide insights into the issues caused by complex terrain and canopy structure in eddy-flux measurements. However, the flow patterns found under neutral condition in this paper may be different from that under stably stratification. These numerical results also need to be justified by intensive field experiments in the future.
Chapter 4 Stably Stratified Canopy Flow

Abstract

Difficulties in measuring atmosphere-ecosystem exchanges of mass and energy have been attributed to the complexity of canopy flows under stably stratified conditions in hilly terrains. The complex characteristics of stably stratified canopy flows in complex terrain are investigated by employing the Renormalized Group (RNG) $k$-$\varepsilon$ turbulence model. Three stratified layers are found that: (1) a primary super-stable layer with temperature inversion is above the canopy; (2) a secondary super-stable layer with temperature inversion is in the deep canopy; and (3) an unstably stratified layer with negative temperature gradient is between the two super-stable layers. The vertical exchanges of mass, momentum and energy are limited by the stabilities of three stratified layers. The motion patterns of two drainage flows associated with the two super-stable layers are significantly modified by the interaction between thermal stratification and terrain slope, and can be better understood with the distribution of vortices, and the sources and sinks of turbulent kinetic energy.

**Keywords:** Canopy flow, complex terrain, stable stratification, Richardson number, turbulent kinetic energy, CFD, RNG
4.1 Introduction

Canopy flow occurring within and immediately above vegetation canopies plays a substantial role in regulating atmosphere-biosphere interaction. The canopy layer is an interface between land and atmosphere, in which most natural resources humans need are produced by tremendous biochemical reactions. Canopy flow influences those biochemical processes through the control of gas exchange between the vegetation and the atmosphere (e.g., influencing reaction rates by changing gas concentrations), heat exchanges (e.g., influencing reaction conditions by changing temperature), and momentum exchanges (e.g., changing turbulent mixing conditions). Better understanding of canopy flow behavior has many practical implications in accurately determining, for instance, terrestrial carbon sinks and sources (Sun et al., 2007), the fate of ozone within and above forested environments (Wolf et al., 2011), forest fire spread rate (Cruz et al., 2005), bark beetle management (Edenburg et al., 2010), and others.

The typical patterns of canopy turbulent flows are characterized by an S-shaped wind profile with an exponential Reynolds stress profile rather than the widely used logarithmic wind profile and constant Reynolds stress observed over bare ground (Yi, 2008). S-shaped wind profiles have been observed within forest canopies in numerous studies (Baldocchi and Meyers, 1988; Bergen, 1971; Fons, 1940; Lalic and Mihailovic, 2002; Landsberg and James, 1971; Lemon et al., 1970; Meyers and Paw U, 1986; Oliver, 1971; Shaw, 1977; Turnipseed et al., 2003; Yi et al., 2005; Queck and Bernhofer, 2010; Sypka and Starzak, 2013). The S-shaped profile refers to a secondary wind maximum that is often observed within the trunk space of forests and a secondary minimum wind speed in the region of greatest foliage density. The features of S-shaped wind profiles imply that K-theory and mixing-length theory break down within a forest canopy layer (Denmead and Bradley, 1985; Yi, 2008). Particularly, the
assumption of a constant mixing-length within a canopy is not consistent with the original mixing-length theory. This is because a mixing-length \( (l_m) \) must satisfy Von Karman’s rule (Von Karman, 1930; Schlichting, 1960; Tennekes and Lumley, 1972), which indicates that a mixing length is a function of velocity distribution (Schlichting, 1960), as:

\[
l_m = \kappa \left| \frac{dU/dz}{d^2U/dz^2} \right|, \tag{4.1}
\]

where \( \kappa \) is von Karman’s constant, \( U \) is wind speed, and \( z \) is height within the canopy. The mixing length of the S-shaped velocity distribution is not constant, being minimum at the local extreme values of the wind profile \( (dU/dz = 0, d^2U/dz^2 \neq 0) \) and maximum at the inflection point of the wind profile \( (dU/dz \neq 0, d^2U/dz^2 = 0) \) (Wang and Yi, 2012). A mixing-length that varies with height within canopy has been demonstrated by large-eddy simulations (Coceal et al., 2006; Ross, 2008) and by water tank experiments (Poggi and Katul, 2007).

The features of S-shaped wind profiles also dictate the existence of super-stable layers near levels where wind speed is maximum (or minimum) and temperature inversion (temperature increasing with height) exists, leading the Richardson number to be extremely large or infinity (Yi et al., 2005). A super-stable layer acts as a ‘lid’ or ‘barrier’ that separates fluid into two uncorrelated layers: (1) the lower layer between the ground and the super-stable layer, and (2) the upper layer above the super-stable layer. This canopy flow separation was verified by SF\(_6\) diffusion observations (Yi et al., 2005) and carbon isotope experiments (Schaeffer et al., 2008). The lower layer is sometimes called a ‘decoupled layer’ (Alekseychik et al., 2013) that is shallow, usually within the trunk space of a forest. Because the super-stable layer prohibits vertical exchanges, the decoupled layer channels air in the horizontal direction. The characteristics of the channeled air are highly dependent on soil conditions, containing a high
concentration of soil respired CO$_2$ and soil evaporated water vapor, and consisting of colder air cooled by radiative cooling at the ground surface (Schaeffer et al., 2008). The channeled air is sometimes termed ‘drainage flow’, and is a common phenomenon in hilly terrains under stable atmospheric conditions, such as on calm and clear nights (Yi et al., 2005; Alekseychik et al., 2013). The drainage flow limits the accuracy of tower-based estimates of ecosystem-atmosphere exchanges of carbon, water, and energy. Sensors on the tower above the canopy cannot measure the fluxes conducted by drainage flow because the layer above the canopy is decoupled from the drainage flow by the isolating super-stable layer. This advection problem is a well-known issue that has not yet been solved using eddy-flux measurements (Goulden et al., 1996; Aubinet et al., 2003; Staebler and Fitzjarrald, 2004; Sun et al., 2007; Yi et al., 2008; Montagnani et al., 2009; Feigenwinter et al., 2010a; 2010b; Aubinet and Feigenwinter, 2010; Queck and Bernhofer, 2010; Tóta et al., 2012; Siebicke et al., 2012).

The concept of a super-stable layer is useful in interpreting data associated with stratified canopy air (Schaeffer et al., 2008). However, stratified canopy flows over complex terrain are far too complex to be able to understand considering only a super-stable layer. Canopy structure (leaf area density profile), terrain slope, and thermal stratification are three key parameters in understanding the details of stratified canopy flows over complex terrain. The thermal stratification plays a leading role in the development of pure sub-canopy drainage flows (Chen and Yi, 2012): strong thermal stratification favors drainage flow development on gentle slopes, while weak or near-neutral stratification favors drainage flow development on steep slopes. Interaction between thermal stratification and terrain slopes may result in multiple super-stable layers and complicated flow patterns, causing difficulties in understanding the mechanisms and rates of exchange of mass and energy between the terrestrial biosphere and the atmosphere.
In this paper, we attempt to use a computational fluid dynamics (CFD) technique to examine the micro-structure of stratified canopy flows to provide insight into the role of physical processes that govern drainage motion and its turbulent characteristics within canopy in complex terrain. There are many challenges to face when pursuing this goal. First, the mixing-length theory and K-theory that are widely used as closure approaches to momentum equations (Wilson et al., 1998; Pinard and Wilson, 2001; Ross and Vosper, 2005; Katul et al., 2006) have been shown to have questionable validity within a forest canopy layer both theoretically (Yi, 2008) and observationally (Denmead and Bradley, 1985). Second, even though CFD models have been used to simulate flow within and above the canopy in numerous published studies, most numerically reproduced canopy flow is confined to idealized cases: either neutral (Ross and Vosper, 2005; Dupont et al., 2008; Ross, 2008; Katul et al., 2006) or weakly unstable (Wang, 2010) atmospheric conditions; or flat terrain with a homogeneous and extensive canopy (Huang et al., 2009; Dupont et al., 2010). Simulations of stratified canopy flow have received little consideration. This might be attributed to difficulties in numerical simulations arising from small scales of motion due to stratification (Basu et al., 2006), and complex interaction between wind and canopy drag elements (Graham and Meneveau, 2012). Large eddy simulation has been quite successful in producing turbulent flow in unstable cases (Shen and Leclerc, 1997; Wang, 2010). However, under stable conditions, due to flow stratification, the characteristic size of eddies becomes increasingly small with increasing atmospheric stability, which eventually imposes an additional burden on the LES-SGS models (Basu et al., 2010). If enough resolution can be employed, any turbulent flow can be simulated accurately by LES. In fact, given sufficiently fine resolution, LES becomes Direct Numerical Simulation (DNS), demanding very fine spatial and
temporal resolution (Galperin and Orszag, 1993), which is currently beyond the reach of available computational power.

In this paper, we employ the renormalized group (RNG) $k-\epsilon$ turbulence model rather than the standard $k-\epsilon$ turbulence model. In the standard $k-\epsilon$ model the eddy viscosity is determined from a single turbulence length scale, so the calculated turbulent diffusion is that which occurs only at the specified scale, usually assuming a constant mixing length for canopy flows. The constant mixing-length assumption and the mixing-length theory are invalid within canopy as discussed above (Yi, 2008; Ross, 2008; Wang and Yi, 2012). The RNG $k-\epsilon$ turbulence model was developed by Yakhot and Orszag (1986a) using the renormalized group methods that can be used to account for the different scales of motion through changes to the production term. The RNG $k-\epsilon$ turbulence model does not involve any experimentally adjustable parameters and does not use mixing-length theory (Yakhot and Orszag, 1986b; Smith and Reynolds, 1992). The initial successes in applying the RNG $k-\epsilon$ turbulence model to generate canopy flows have been demonstrated by Graham and Meneveau (2012).

4.2 Numerical Implementation

The two dimensional computational domain extends over 1400m×130m in a Cartesian coordinate system, corresponding to 1200×157 grid intervals in the x and y directions. A single hill covered with a 15m tall homogeneous forest canopy occupies the middle 100m domain in horizontal. The mesh spacing in both horizontal and vertical at the forested hill is 0.5m and is stretched with a power law, starting with a grid spacing of 0.5m throughout the canopy, with a larger grid spacing stretching outwards from the edge of the forest and the top of the canopy on
the hill crest. The stretch power in both horizontal and vertical is 1.15. Ground surface roughness height is set to be 0.01.

In this study, the topography is specified with a sinusoidal hill. The shape function of the hill is defined as

\[
H(x) = \frac{H}{2} \cos\left(\frac{\pi x}{2L}\right) + \frac{H}{2} \tag{4.2}
\]

where \(H\) is the hill height, \(L\) is the half length scale (half of the hill width at mid-slope height), \(x\) is longitudinal distance with \(x = 0\) at the center of the single hill. The variation of the slope \((H/L)\) is specified by changing \(H\) with a constant \(L = 25\) m.

The porous canopy layer is designed horizontally homogeneous along the slope. Leaf area density \(a\) is specified as values from observation of an actual forest (Yi et al., 2005). The ambient temperature is \(\theta_0(z) = \theta_{00} + \gamma z\), where \(\theta_{00} = 288\)K, is the potential temperature at \(z = 0\), \(\gamma\) is ambient lapse rate, set to -6°C km\(^{-1}\). Upward radiative heat flux is zero in the lower canopy layer (0-8m) and then linearly increased to -8 Wm\(^{-2}\) at the top of canopy. Heat flux at the ground surface is -15Wm\(^{-2}\). Since we are most interested in calm night-time conditions, no wind in the domain is initially specified.

### 4.3 Stably Stratified Sub-Canopy Flow

After a quasi-equilibrium condition is approached, all the solved fields in the studied cases are developed to be near symmetric horizontally (in the x-direction) with respect to the center of the modeled hill at \(x = 0\) due to the homogeneous boundary conditions and initial settings. We restrict our discussion to the right half of the hill. Our results show (Figure 4.1) that wind structure is differentiated into down-sweep \((H/L \leq 0.6)\) and up-draft \((H/L \geq 0.8)\) within
canopy. The temperature, wind and turbulence characteristics on representative gentle \((H/L = 0.6)\) and steep \((H/L = 1.0)\) hills are illustrated (see Figure 4.1) to explore the thermal and mechanical processes that govern the airflow structures.

### 4.3.1 Thermal Analysis

In the model, strong stratification develops with distinct thermal distribution on the slope, subject to heat loss on the slope surface and the upper canopy layer. The heterogeneous distribution of heat within the canopy causes a ‘fish-head’-shaped temperature distribution on the slope, with the upper jaw in the upper canopy layer and the lower jaw attaching to the slope surface. The jaws consist of cold air while the open mouth shows relatively warmer air (Figure 4.2). In comparison with the upper jaw which is confined to the middle and lower slope, the lower jaw extends up to the crest of the hill. As the slope intensity is reduced, the fish-head effect’s upper jaw is diminished. For a very gentle slope (i.e., \(H/L << 1\)), the model produces a horizontal isotherm pattern with cold air at the bottom of the slope and warm air upslope, as would be expected in real-world conditions. A significant difference in temperature distribution among varied slopes results in a different angle of orientation of the fish-head temperature profile. Isotherms are inclined parallel to the slope surface because they tend to follow the shape of the slope and the top-canopy layer since the cooling along the slope surface is uniform. The temperature distribution on a gentle hill is shown as an angled fish-head shape, while the fish-head is tilted by the slope on the steep hill, which is shown by the isotherms on the lower jaws. The different fish-head profile’s angle can explain specific flow structures in the canopy (see section 4.3.2). In accordance with the fish-head temperature distribution, temperature profiles are shown in three layers (Figure 4.3a-d). A strong inversion layer is developed across the lower jaw, above which temperature slightly decreases with height in a thermal transition zone and a weak
inversion layer is formed across the upper jaw. The temperature gradient and the depth of the lower inversion layer increases, since cold air flowing down the slope results in a cool pool on the lower slope where a single inversion layer extends above the canopy (Figure 4.3e, f). The temperature difference from the hill surface to the top of the canopy at the hillcrest is about 0.8°C and 0.4°C for gentle and steep hills, respectively, while the difference increases to around 3.2°C in the canopy layer at the feet of both hills. The inversion strength near the surface is larger than in the upper canopy, which is due to the stronger radiative cooling effect on the surface. The temperature gradient and inversion on the steep hill are predicted weaker than on the gentle hill, because at the same horizontal x/L location, the canopy layer is at a higher elevation on the steep hill. Regardless of the horizontal location x/L, we find that inversions both near the surface and in the upper canopy are stronger on the steep hill than on the gentle hill at the same elevation, which benefits the development of stronger drainage flow on the steep slope.

Richardson number (Ri) is the ratio of the relative importance of buoyant suppression to shear production of turbulence, which is used to indicate dynamic stability and formation of turbulence. Ri is calculated based on mean profiles of wind and temperature. For different purposes and data availability, gradient Richardson number (Ri_g) and bulk Richardson number (Ri_b) are used to predict the stability within canopy. Yi et al. (2005) found that the gradient Richardson number,

$$ Ri_g = \frac{(g/\bar{\theta}) (\partial \bar{\theta}/\partial z)}{(\partial \bar{U}/\partial z)^2} $$

(4.3)

with \(\partial \bar{U}/\partial z = 0\) and \(\partial \bar{\theta}/\partial z \neq 0\) at the inflection points of the S-shaped wind profile resulted in an infinite Ri_g, which describes the super-stable layer. In a forest, wind and temperature are typically only measured in a few levels, making \(\partial \bar{U}/\partial z\) and \(\partial \bar{\theta}/\partial z\) impossible to directly
calculate. Therefore, $Ri_b$ is commonly used to quantify stability between two levels ($z_1$ and $z_2$) using the measured temperature and wind speed (Zhang et al., 2010; Burns et al., 2011; Alekseychik et al., 2013),

$$Ri_b = \frac{g}{\theta} \frac{\theta(z_2) - \theta(z_1)}{(U(z_2) - U(z_1))^3} (z_2 - z_1). \quad (4.4)$$

In our modeling setting, the gridding space in vertical is $\Delta z = z_2 - z_1$, which is 0.5m in the canopy layer. We define a local Richardson number to evaluate stability around the forested hill and examine the local stability in response to the heterogeneous distribution of heat. The local Richardson number in grid $(m, n)$ is calculated as,

$$Ri_l = \frac{g}{\theta_{m,n}} \frac{(\theta_{m,n} - \theta_{m,n-1})}{\left(u_{m,n} - u_{m,n-1}\right)^2 + \left(w_{m,n} - w_{m,n-1}\right)^2} \left(z_{m,n} - z_{m,n-1}\right). \quad (4.5)$$

Local Richardson number indicates that, within the canopy, flow is stably stratified except for an unstable region penetrating from the hill summit into the middle slope within the thermal transition regime (Figure 4.1). $Ri_l$ is found to be extremely large ($\sim 10^5$) just above the canopy on the upper to middle slope (Figure 4.4a-d) indicating a thin primary super stable layer just above the top of canopy. The primary super stable layer is elevated and deepened on the lower slope (Figure 4.4e, f), extended from the height of 1.3-1.4h to about the height of 2h. The deep primary super stable layer is caused by the strong cooling and temperature inversion at the base of the hill, regardless of slope intensity. Within canopy, a secondary super stable layer with extremely high $Ri_l$ is developed below 0.5h. On the lower slope, the depth of the secondary super stable layer extends from the slope surface up to 0.5h due to strong temperature inversion and wind stagnation. The absence of a secondary super stable layer on the summit could be explained by stronger entrainment of warmer air from above-canopy. Air in the transition region with
negative temperature gradient is unstably stratified. The transition region is developed by the downwelling of cool air from the upper canopy with relatively warmer air upwelling from the lower canopy. The results show that for a sufficiently steep slope, the effects of the hill dominate the atmospheric profile, while for more gentle slopes the effects of the canopy dominates the resultant atmospheric profile.

The nocturnal stable canopy layer could be used to clarify the occurrence of within- and above-canopy flows decoupling observed in prior studies. Gorsel et al. (2011) reported a very stable nighttime canopy layer \((Ri_b >1)\) using the bulk Richardson number, indicating that the canopy layer is decoupled from air aloft. Decoupling at the top of the canopy is more likely to occur as the buoyancy is more dominant and air at the top of the canopy is strongly stable. The canopy top decoupling weakens vertical exchange of mass and heat between the vegetation and the atmosphere aloft. The measurement data show large temperature and CO\(_2\) gradients (Burns et al., 2011) as decoupling occur in strongly stabilized atmosphere. Decoupling at the top of the canopy produced stronger carbon dioxide and temperature gradients than within canopy decoupling (Alekseychik et al., 2013). The primary super stable layer in our study is shown as a lid located at the top and above canopy, which could terminate the vertical exchange between the canopy and the air above. During nighttime, soil respiration contributes about 60-70\% (Janssens et al., 2001) of the total CO\(_2\) emission from terrestrial ecosystem. The soil respired CO\(_2\) could be blocked by the secondary super stable layer forming a very shallow pool on the slope surface.
Figure 4.1 Simulated streamlines in the forested hill: (a) $H/L = 0.6$; (b) $H/L = 1.0$. The translucent yellow masks indicate the regimes with instability within canopy. The black ‘WV’ marks the region of wake vortices next to the edge of canopy. The white ‘DS’ in (a) and ‘UD’ in (b) indicate the region of down-sweep wind and up-draft wind on the gentle and steep slopes, respectively. The background color indicates the temperature distribution with the blue of warmer air and yellow at the bottom indicating cold air.
Figure 4.2 Contours of temperature (K) along the right slope: (a) $H/L = 0.6$; (b) $H/L = 1.0$. The temperature difference between isotherms is 0.25 K. The numbers on isotherms indicate the temperature. The x-axis is normalized by the half length scale of the hill $L$ and y-axis is normalized by the height of the canopy $h$. White dashed lines indicate the top of canopy and the isotherms marked with hot pink dashed lines highlight the ‘fish-head’ temperature distribution.
Figure 4.3 Temperature (K) profiles on the slope for $H/L = 0.6$ (blue) and $H/L = 1.0$ (red). The locations of the six sections are labeled as a-f, and their locations with respect to the hill are presented, which is normalized by the half length scale $L$. The green curves indicate the thermal transition zone with negative temperature gradient.
Figure 4.4 Locations of super stable layers for H/L=0.6 and H/L=1.0 (left panel). The primary super stable layers are marked by dash-dotted lines with yellow solid circles and secondary super stable layers are marked by dash-dotted lines with red solid circles. The Ri numbers at locations indicated by the yellow and red solid circles are extremely large, which are illustrated on the right panel for the locations (b) and (e).
4.3.2 Wind Structures

Figure 4.1 shows that air above the canopy sinks and converges towards the hill and undergoes direction shift within canopy. Flow converges to the hill from all sides, and is then inflected near the top of the canopy, following the shape of the slope as drainage flow within the canopy. The height of inflection points increases as the air flows down the slope. The inflection points are approximately at the bottom of the primary super stable layer. As a result of the abrupt convergence in the top the canopy at the base of the hill, wake vortices are developed near the forest edge, after the wind leaves the hillside within the primary super stable layer. The wake vortices can extend to about 2.6L in horizontal and 1.3h in vertical. According to the flowing location within the canopy, we identify the drainage flow as two streams: the majority air mass within the upper-canopy inversion layer is called the upper-canopy drainage flow (UDF) layer; and the majority air mass within the inversion layer in the lower-canopy is called the lower-canopy drainage flow (LDF) layer. The UDF is developed as the air above the canopy sinks from the lateral sides towards the hill. However, the sinking motion is diverted following the shape of the top-canopy layer as it reaches the top of the canopy (Figure 4.1). The UDF accelerates down the slope between the top of the unstable layer and the bottom of the primary super stable layer, reaching its maximum wind speed of 0.25 meters per second (ms\(^{-1}\)) at location (Figure 4.5d) on the gentle slope and 0.35 ms\(^{-1}\) at location (Figure 4.5e) on the steep slope, and then decelerates down to the feet of the hills. The air sinking over the crest can directly reach the surface of the crest and flow along the slope to form the LDF. The maximum wind speed of the LDF is at location (Figure 4.5d) for a gentle slope (0.18 ms\(^{-1}\)) and at location (Figure 4.5c) for a steep slope (0.29 ms\(^{-1}\)). The maximum wind speed in LDF occurs on the slope surface, below the secondary super stable layer. Deceleration of the flow towards the base of the hill should occur for a
number of reasons. The pool of cool, dense air at the base of the hill resists incoming flow. Also, the drag force acting against the wind is dependent on the speed of the air flow squared.

UDF and LDF show different patterns within canopy for different slopes, which essentially determines the shift direction within canopy (Figure 4.1). On the gentle slope (H/L = 0.6), UDF is much thicker compared with LDF. Air in UDF accelerates within the regime of the upper inversion layer reaching its maximum at the top of thermal transition region and then decelerates to a minimum (u = 0 and w = 0, Figure 4.5) at the top of the slope surface inversion layer. Then, UDF sweeps horizontally to join the shallow LDF on the slope surface, which is shown as negative streamwise velocity and near-zero vertical velocity in Figure 4.5 (down-sweep). When the slope is steep (H/L = 1.0), UDF is much shallower than LDF. Air in LDF accelerates on the upper slope (Figure 4.5a-c), followed by deceleration and stagnation. The stagnated flow jumps perpendicularly from deep canopy layer to join the shallow UDF in the upper canopy layer (The up-draft, with u>0 and w>0, is visible in Figure 4.1 and 4.5). The shifting winds on both gentle and steep slopes are parallel to the isotherms in the warm ‘fish mouth’ region of the profile. Rotational vortices are formed below the shifting winds.

The generation and direction of the shifting-wind structure are primarily driven by the slope and stratification. Under calm and stably stratified conditions, the dominant driving force of sinking drainage flow on the slope is the hydrostatic buoyancy force which is given as: 

\[ F_{hs} = g(\Delta \theta / \theta_0) \sin \alpha \]

where \( \alpha \) is the slope angle, \( \Delta \theta \) is the potential temperature difference between the ambient air and the colder slope flow, \( \theta_0 \) is the ambient potential temperature. The drainage flow on both the gentle and steep slopes is initiated by the dominant \( F_{hs} \) as the air is calm and stably stratified (Froude number <<1, Belcher et al., 2008). The magnitude of \( F_{hs} \) increases with slope angle \( \alpha \) so that \( F_{hs} \) is much larger on a steep slope than a gentle slope, leading to a stronger
sinking motion above the crest. The sinking air penetrates to the lower part of the canopy at the hilltop. Thus, the LDF layer is deeper than the layer of UDF for a steep slope. However, the sinking motion above the crest on the gentle slope is diverted to follow the shape of the slope in the upper canopy due to smaller $F_{hs}$, which is not strong enough to completely penetrate the canopy. As a result, UDF is deeper than the LDF on gentle slopes, in contrast to that on steep slopes. The heterogeneous cooling in the canopy layer causes two baroclinic zones consistent with the UDF and LDF: the upper canopy layer and slope surface layer. The strong baroclinicity on the steep slope surface causes the deep LDF wind to rotate counter-clockwise (i.e., turning upwards on the lower slope, perpendicular to the hill slope). However, the rotated wind is forced to shift down when hitting the top-canopy UDF. The wind at the baroclinic zone with a deep UDF on a gentle slope rotates clockwise, but shifts downslope when hitting the layer of the LDF.
Figure 4.5 Profiles of streamwise velocity (m s\(^{-1}\), top panel) and vertical velocity (m s\(^{-1}\), bottom panel) for \(H/L=0.6\) (blue) and \(H/L = 1.0\) (red). The locations of the six sections are labeled as a-f, and their locations with respect to the hill are presented in Figure 4.3.
4.4 Turbulent Properties of Stratified Sub-Canopy Flow

4.4.1 Turbulent Fluxes of Momentum and Heat

Figure 4.6 shows profiles of shear stress $\overline{-u'w'}$. Shear stress is most significant in the region near the top of the canopy where wind impinges on the canopy resulting in strong wind shear. Another region of large shear stress is in the lower canopy. This is related to the wind shifts which lead to strong wind shear. Shear stress is small on the upper slope but increases down the slope. The maximum shear stress at the top of the canopy is located at the wake region (Figure 4.6e, f), where the wake vortices are formed. Shear stress is positive above the canopy indicating the downward transfer of momentum which is different from the usually observed downward transport of momentum in the upper canopy. It could be explained by the strong stability above the top of canopy, because strong stability substantially reduces the downward transport of momentum (Mahrt et al., 2000). The momentum transfer is reversed to upward ($\overline{-u'w'} < 0$) when approaching the top of the canopy where airflow is diverted into canopy layer because of the UDF and shear-production of turbulence. Strong upward momentum transfer near the top of canopy on the lower slope is associated with the wake generation behind the hill. In the upper canopy at midslope and downslope, shear stress decays rapidly as $z$ decreases, because of the momentum absorption by the dense crown. The upward momentum ($\overline{-u'w'} < 0$) in the lower canopy indicates momentum sources in the LDF. The LDF was recognized as jet-like flow in lower canopy, which has important effects on momentum transfer within canopy (Mao et al., 2007). Upward momentum transport in the canopy is a very common, occurring in stable atmospheric conditions (Zhang et al., 2010).
The dominant positive turbulent heat flux, \(-\overline{u'\theta'}\) indicates downward heat transfer above and within the canopy (Figure 4.7). Heat transfer on the upper slope (Figure 4.7a, b) is weak because the temperature difference between the canopy and the atmosphere above is small. The downward heat transfer is much stronger on the lower slope, where the air is cooled as a ‘cool pool’ with the greatest temperature gradient. Turbulent heat flux increases towards the top of the canopy indicating increasing downward heat transfer \((\overline{-u'\theta'} > 0)\) but the downward heat transfer decreases in the upper canopy layer. The peak of turbulent heat flux near the top of the canopy is due to the strong radiative cooling in the upper canopy. Below that the near zero and slightly positive turbulent heat flux (Figure 4.7) is due to near neutral and negative temperature gradient in the thermal transition zone. As a result of the strong cooling in the ground surface, there are significant downward heat flux transfers in the lower canopy.
Figure 4.6 Profiles of shear stress, $-\overline{u W'}$ ($10^{-3}$ m$^2$ s$^{-2}$) on the slope for $H/L = 0.6$ (blue) and $H/L = 1.0$ (red). The locations of the six sections are labeled as a-f, and their locations with respect to the hill are presented in Figure 4.3.
Figure 4.7 Profiles of turbulent Heat Flux, $-\overline{u' \theta'}$ ($10^{-2}$ K m s$^{-1}$) on the slope for $H/L = 0.6$ (blue) and $H/L = 1.0$ (red). The locations of the six sections are labeled as a-f, and their locations with respect to the hill are presented in Figure 4.3.
4.4.2 Turbulent Kinetic Energy (TKE) Budget

In steady state, TKE budget Equation (2.14) can be written as:

$$0 = T_a + T_r + T_p + P_s + P_b + P_w - \varepsilon$$ (4.6)

where $T_a$ is the advection of TKE by the mean wind, $T_r$ represents the turbulent transport of TKE, $T_p$ represents the transport of TKE by pressure perturbation, $P_s$ is the shear production of TKE, $P_b$ is buoyancy production of TKE, $P_w$ is wake production of TKE and $\varepsilon$ is viscous dissipation of TKE. We calculate all the terms in the TKE budget equation individually except $T_p$ which is treated as the residual of other terms.

TKE is examined to show the intensity of turbulence along the slope (Figure 4.8). TKE is usually low within the canopy implying a low turbulence flow under strongly stable atmospheric conditions. TKE is available near the top of canopy on the midslope and downslope. The region with strongly shifting winds is on the lower slope where the wind shear is strong. The largest TKE is found in the region of wake vortices across the canopy edge. The TKE value is larger on the gentle slope than on the steep slope.

Contributions from transport and production terms of TKE are complicated. $P_b$ is a principal sink of TKE under stable conditions (Figures 4.9 and 4.10). $P_b$ exhibits negative values near the top of the canopy and slope surface, where flow is stably stratified, which suppresses the turbulence around the top of the canopy and within the deep canopy. In the thermal transition zone, the contribution of $P_b$ is minimal ($P_b \approx 0$ or slightly positive). Buoyancy production is neglected in some studies because $P_b$ is (1) unimportant compared with other terms in TKE budget (Lesnik, 1974) and (2) difficult to measure (Meyers and Baldocchi, 1991), restricting the modeling and measurement studies to near-neutral conditions. Shen et al. (1997) showed that
near the top of the canopy, the buoyancy production increases as instability increases, although it is smaller than 10% of shear production in unstable conditions. Leclerc et al. (1990) illustrated a strong positive correlation between buoyancy production and stability ($P_b < 0$) or instability ($P_b > 0$) both within and above the canopy, which is confirmed in our modeling results.

Wake production ($P_w$) is a principal source of TKE in the upper half of the canopy where the canopy is dense (i.e., for large values of $a$ and $K_r$) on both steep and gentle slopes. Although the magnitude of $P_w$ is very small on a steep slope, the relative contribution of $P_w$ is very large in comparison with other TKE components. Even in the lower canopy layer on the upper slope, $P_w$ is a dominant source of TKE. This unusual phenomenon is induced by the deeper and stronger drainage flow on the slope surface (large $u_i$).

The positive shear production $P_s$ indicates the net transfer of kinetic energy from the mean flow to the turbulent component of the flow (Figure 4.9 and 4.10). $P_s$ is smaller than $P_w$ except near the top of the canopy, which is consistent with the observations in soybeans (Meyers and Paw U, 1987), deciduous forests (Shi et al., 1987; Meyers and Baldocchi, 1991) and an artificial canopy (Raupach et al., 1987). $P_s$ peaks at the top of the canopy, due to strong wind shear. Shear production is not as important as buoyancy and wake production in the canopy because of strong stability. Observational data also showed that shear production decreases with increasing stability in the lower two-thirds of the canopy (Leclerc et al., 1990).

Transport terms are the dominant source to maintain turbulent kinetic energy near the top of the canopy where strong buoyancy suppression occurs (Figure 4.9 and 4.10). TKE is weakly transported by turbulence upward near the canopy top ($T_i < 0$) and downward ($T_i > 0$) in the canopy, because turbulence is limited by strong stability above the canopy. TKE transport by advection and turbulence is unimportant at all levels and all slopes in comparison to pressure
transport. The field measurement of pressure transport $T_p$ is difficult and the behavior of $T_p$ in the TKE budget is uncertain (Raupach et al., 1996; Finnigan, 2000). Maitani and Seo (1985), Shaw et al. (1990) and Shaw and Zhang (1992) have confirmed that $T_p$ is not small enough to be neglected according to the surface pressure measurements. Pressure diffusion is recognized as an important sink of TKE in the upper canopy and source of TKE below (Dwyer et al., 1997) under unstable conditions. Our results show that the contribution of pressure transport to the overall TKE budget is significant when it is identified as a residual of other TKE components. $T_p$, which is of the same order as the production terms, supplies TKE in areas where the buoyancy suppression is very strong and extracts TKE where wake production is dominant. On gentle slopes, $T_p$ is important to compensate the TKE loss by buoyancy near the top of the canopy and in the lower part of the canopy, and compensate TKE gain by wake motion in the upper half of the canopy (Figure 4.9). On steep slopes, $T_p$ on the lower half of the slope plays the same role as on gentle slopes to compensate the TKE loss by buoyancy and gain by wake (Figure 4.10d-f), but the relative significance of wake production becomes more prominent. On the upper slope (Figure 4.10a-c), pressure transport is important in the whole canopy to work against wake production. Our results suggest that the pressure perturbation is stronger compared with other terms on steep slopes. In addition, thermal effects on the upper steep slope are diminished and the canopy effect is magnified since the air is warm and the temperature gradient is small on the elevated topography.
Figure 4.8 Contours of turbulent kinetic energy (m$^2$s$^{-2}$): (a) $H/L = 0.6$ and (b) $H/L = 1.0$. The black dashed lines indicate the top of canopy.
Figure 4.9 Profiles of TKE components \((10^{-3} \text{ m}^2 \text{ s}^{-3})\) for H/L = 0.6. \(T_a\) is the advection of TKE by the mean wind, \(T_t\) represents the turbulent transport of TKE, \(T_p\) represents the transport of TKE by pressure perturbation, \(P_s\) is the shear production of TKE, \(P_b\) is buoyancy production of TKE, \(P_w\) is wake production of TKE and \(\varepsilon\) is viscous dissipation of TKE. The locations of the six sections are labeled as a-f, and their locations with respect to the hill are presented in Figure 4.3.
Figure 4.10 Profiles of TKE components ($10^{-3}$ m$^2$ s$^{-3}$) for H/L = 1.0. $T_a$ is the advection of TKE by the mean wind, $T_t$ represents the turbulent transport of TKE, $T_p$ represents the transport of TKE by pressure perturbation, $P_s$ is the shear production of TKE, $P_b$ is buoyancy production of TKE, $P_w$ is wake production of TKE and $\varepsilon$ is viscous dissipation of TKE. The locations of the six sections are labeled as a-f, and their locations with respect to the hill are presented in Figure 4.3.
4.5 Summary and Conclusions

The thermal stability and its influence on the canopy flow in complex terrain are explored in calm and stably stratified conditions. The thermal distribution and stability occurring within the canopy are substantially different from the ambient atmosphere. The stability around the canopy is characterized by stratification with a primary super stable layer above the top of the canopy, a secondary super stable layer in the lower canopy, and an unstable layer within the canopy.

The thermal stratification around the canopy primarily drives the airflow and turbulent characteristics in the canopy layer on the slope. Airflow converges to the hill from all sides and the crest above the canopy, and is then inflected near the top of the canopy, following the shape of the slope, becoming drainage flow within the canopy. The drainage flow within the canopy is separated into two streams: the majority air mass in the upper canopy inversion layer is the upper-canopy drainage flow (UDF) layer; and the majority air mass in the inversion layer in the lower canopy is the lower-canopy drainage flow (LDF) layer. On gentle slopes, air in UDF sweeps horizontally to join the shallow LDF on the slope surface, while on steep slopes, the stagnated flow in LDF jumps upwards, perpendicular to the slope, from the lower canopy layer to join the shallow UDF in the upper canopy layer. The generation and direction of the shifting-wind structure within the canopy are induced by the hydrostatic buoyancy force and baroclinic instability on the slope, which are functions of the inclination of the slope.

The turbulent properties of the stratified canopy flow are closely associated with thermal and dynamic conditions on the slope. The downward transport of momentum in the canopy is reduced due to strong stability. Upward transport of momentum occurs in the deep canopy. The heat flux is predominantly transported downward with the minimum heat flux in the thermal
transition zone in the middle of the canopy. TKE is available near the top of canopy on the midslope and downslope. The region experiencing strong wind shifts is on the lower slope where the wind shear is strong. The largest TKE values are found in the region of wake vortices across the canopy edge. Buoyancy production of TKE is a principal sink of TKE under stable conditions, which suppresses turbulence significantly near the top of the canopy and in the deep canopy. TKE is generated by shear production near the top of the canopy and by wake production in the canopy. The transport of TKE by pressure perturbation, which is of the same order as the production terms, supplies TKE where the buoyancy suppression is very strong and extracts TKE where wake production is dominant. Our results suggest that the relative importance of pressure perturbation is enlarged as the slope increases.

The findings in our numerical simulations have disclosed sub-canopy wind structure and turbulence characteristics on a single forested hill terrain in calm and stable conditions. The complicated slope-canopy flow has great influence on the energy and mass exchange between vegetation and the atmosphere aloft, which may not be adequately captured by considering only a simple drainage flow regime. More focused modeling and experimental studies are required – especially in complex terrain under relatively stable atmospheric conditions – in order to refine understanding of biosphere / atmosphere exchange of carbon dioxide, moisture and energy.
Chapter 5 Three-Dimensional Canopy Flow in a Forested Terrain

Abstract

Canopy flow resulting from interaction between thermo-topographic drainage flow and large-scale synoptic flow is extremely complicated and has been poorly understood. We apply a computational fluid dynamics approach to solve the three-dimensional variability of airflow and CO₂ transport based on measurements on one flux tower, but tested by multiple tower measurements conducted during the ADVEX campaign at the Renon site in the Italian Alps. The model is run with and without large-scale synoptic conditions to explore the interactions between local orographic flow and large-scale synoptic winds and related CO₂ processes. We found that the thermal condition in the canopy is directly related to the canopy morphology: dense canopy causes stronger cooling but limits vertical exchange of heat flux, resulting in weak temperature inversion in the deep canopy. Recirculation with high CO₂ concentration is developed only under the condition that local slope winds is enhanced by synoptic winds. There is no recirculation formed, as synoptic wind direction is opposite to the local wind direction and CO₂ is quite well mixed. No recirculation appears without synoptic conditions under which CO₂ builds up mainly at downwind locations. This numerical method brings to light a better understanding of the CO₂ closure problem with one-tower measurements in the FLUXNET community.

Keywords: Canopy flow, CO₂ transport, Complex terrain, Recirculation, Slope flow, Synoptic winds
5.1 Introduction

Accurate quantification of net ecosystem-atmosphere exchanges (NEE) of mass and energy is a fundamental and critical step in reducing the uncertainty from the potential effects of climate change on ecosystems as sources or sinks for atmospheric CO$_2$ (Yi et al., 2010; 2012). The eddy covariance (EC) technique has proven to be a useful approach to quantify net ecosystem carbon sequestration in the daytime when strong turbulent mixing occurs, while nocturnal flux measurements carry significant advection errors that can be of the same order as the eddy flux itself when flux sites are located in a complex terrain (Massman and Lee, 2002; Feigenwinter et al., 2004; 2008; 2010a; 2010b; Aubinet et al., 2003; 2008; 2010a; 2010b; Finnigan, 2008; Goulden et al., 2006; Montagnani et al., 2009). EC measurements are one-dimensional (1D) and particularly in complex terrain we do not know the three-dimensional (3D) details of air movement, CO$_2$ transport, and temperature variation around the instrumented tower. Massman and Lee (2002) stated “Clearly, a proper understanding of 2D and 3D flows and their role in micrometeorological flux observation is of importance to any site, but the problem of 2D and 3D flows is most difficult to treat at sites on non-flat topography.”

The advection issue is reportedly very common during calm nights in ecosystems with complex terrain, which is correlated to mechanisms of nocturnal canopy flow, e.g. turbulent ramps, gravity waves, small-scale turbulence, intermittent turbulence, land, sea or lake breezes and drainage flows (Aubinet, 2008). The gravity induced nocturnal drainage flows have been recognized as a dominant reason for advection issues, because drainage flow can lead to large spatial variability of CO$_2$ concentration on the slope. At night, the ecosystem behaves as a CO$_2$ source because of soil and above ground vegetation respiration. CO$_2$ tends to accumulate near the ground due to air stratification particularly in conditions of a super-stable layer within the
canopy (Yi et al., 2005), resulting in a strong negative CO$_2$ gradient (Yi et al., 2000; Araújo et al., 2008). The negative CO$_2$ gradient with subsiding background wind contributes to positive vertical advection. Along the slope, much higher CO$_2$ concentration is observed at lower altitude (slope and valley) than at higher altitude (plateau) (Araújo et al., 2008). The positive CO$_2$ gradient from high altitude to low altitude, along with drainage flow, results in a positive contribution to horizontal advection. Although positive vertical and horizontal advection is very common at night, some measurements confuse the advection issue, such as entrainment of poor CO$_2$ airflow from the top of the canopy to the surface flow contributing to a negative horizontal CO$_2$ gradient along the drainage flow direction and resulting in negative horizontal advection (Aubinet, 2003). Local terrain and vegetation effects cause positive vertical velocity at night, resulting in negative vertical advection (Turnipseed et al., 2003). Vertical gradient of CO$_2$ distribution is large on the upper slope but quite uniform on the lower slope (Reiners and Anderson, 1968; Araújo et al., 2008), implying the smaller vertical advection in the CO$_2$-pooled valley than on the upper slope. All these observations have demonstrated the complexity and 3D characteristics of the advection issues in eddy flux measurements.

The eddy flux community has gone to great efforts to use multi-tower/multi-level measurement systems to capture the 3D characteristics of wind fields and CO$_2$ movement to address the advection issues, such as the European ADVEX field campaigns (Feigenwinter et al., 2008) and advection measurements conducted at the AmeriFlux Niwot Ridge site (Sun et al., 2007; Yi et al., 2008). The measured advection fluxes (non-turbulent flux) are of similar magnitudes as the turbulent fluxes during calm nights, and vary from site to site (Feigenwinter et al., 2008; Yi et al., 2008). The important feature is that the advection contribution is closely correlated to local and synoptic meteorological conditions. Local orographic flow is most likely
to occur within the canopy, while synoptic wind is dominant above the canopy. However, synoptic flows can penetrate into the open canopy and interact with orographic flow (Sun et al., 2007). The process of interaction includes synoptic winds that alter the direction and enhance or attenuate orographic wind, depending on the direction and strength of prevailing synoptic winds (Feigenwinter et al., 2010). Accordingly, the modified orographic flows have direct influence on CO$_2$ pooling or mixing.

Although the direct advection measurement provides insights into the wind fields and CO$_2$ transport at the research sites, the conclusions draw may not be applicable to other FLUXNET sites subject to local terrain and vegetation conditions and large-scale synoptic conditions. In addition, the representativeness of the multiple-tower measurements is very sensitive to the multi-tower setup and methodology to derive the fluxes from measurements (Aubinet et al., 2010a; b; Montagnani et al., 2010). How can we take advantage of the single tower measurements found at most FLUXNET sites to understand ‘site-specific’ CO$_2$ transport processes? In this study, we aim to numerically resolve the 3D spatial variability of airflow and CO$_2$ transport based on one-tower measurements, but tested by multiple tower measurements conducted during the ADVEX campaign at the Renon (RE) site in the Italian Alps. The model is simulated with and without large-scale synoptic conditions to explore the interactions between local orographic flow and large-scale synoptic winds, and related CO$_2$ processes. We first describe the characteristics of terrain, vegetation and measurement set-up, then we present the numerical method, and finally we discuss the results of our simulations.

5.2 Site and Data

This numerical study is conducted based on the extensive measurements performed during the ADVEX campaign at the Renon-Selva Verde study site (RE, 46°25’ N, 11°17’ E). RE
is situated at about 1735 m above sea level (ASL) in the Italian Alps, 12.2 km North-Northeast of Bolzano. The Digital Elevation Model DEM of the 2×2 km area around the RE is shown in Figure 5.1. The topography is characterized by alpine conditions with a typical slope of about 11.0° dipping to the west in the north and south-east in the south.

The RE site is characterized by a coniferous forest with gaps between groups of older and younger trees. The forest species are dominated by *P. abies* (85%), *Pinus cembra* (12%) and *Larix decidua* (3%), with the mean leaf area index (LAI) of 5.1 and maximal canopy height of 29-30m in the 240 x 240 m research area (D2 in Figure 5.1). The vegetation structure is varied at towers. A field survey in October 2009 (Dr. Montagnani, personal communication) classified the vegetation in the D2 research area into three categories (Figure 5.2): (1) Sparse forest in grassland; (2) Forest edge (regrowth); and (3) Mature forest.

The meteorological conditions at RE are dominated by the ‘Tramontana’ winds from the north or northwest (northerlies), typically in winter and occasionally in summer. Winds from the south (southerlies) tend to come from a low-pressure system located over the Western Mediterranean area. Upslope winds during the day and down-slope winds during the night characterize the local slope wind system.

The extensive measurement dataset (half-hourly averaged) was collected from the ADVEX campaign carried out from 1 May to 15 September 2005. The ADVEX setup consisted of four external towers (A, B, C and D) and a permanent central tower (M) (Figure 5.3). Each external tower was equipped for measurements of CO₂, H₂O and wind vector at 1.5, 6, 12, and 30m above ground level (AGL). An additional wind velocity measurement was made at 41.5m at tower C. Measurements at tower M were at 1.5, 6, 12, and 32m. A more detailed description of the site and data is given in Feigenwinter et al. (2008) and Montagnani et al. (2009).
Figure 5.1 The topography around Renon site. A, B, C, D and M indicate the five tower locations in ADVEX campaign. D1 is the outer domain (2000×2000m) and D2 is inner domain (240×240m) in our numerical simulation.
Figure 5.2 Vertical leaf area density profiles of three vegetation categories (left panel) and their distribution in the 240×240m research area (right panel) at Renon site: (1) Sparse forest in grassland, (2) Forest edge, and (3) Mature forest.
Figure 5.3 The tower and instruments set-up during ADVEX campaign. Source: Feigenwinter et al., 2008.
5.3 Numerical Implementation

The main goal of this research is to understand the relationship among the three-dimensional wind field, temperature, and CO$_2$ concentration in this complex terrain with heterogeneous vegetation, which is solved in the steady state. The initial and boundary conditions are constrained with mean nocturnal time. We use 21:00-4:00 local standard time (LST) observations, because these periods show strong negative net radiation. The prevailing wind periods during the ADVEX campaign (Table 5. 2) based on the data availability and validity (Montagnani et al., 2009) are:

1. A 48-hour period characterized by synoptically driven northerlies, locally called ‘Tramontana’ from 0030 LST of 11th July (day 192) to 2400 LST of 12th July (day 193),

2. A 30-hour period characterized by southerlies: from 0030 LST of 25th July (day 206) to 0600 LST of 26th July (day 207),

3. A 108-hour period characterized by a local mountain-valley wind system in the area, resulting in below-canopy down-slope (northerly) winds at night, and upslope (southerly) winds during the day, from 1200 LST of 26th July (day 207) to 2400 LST of 30th July (day 211).

The model is first tested against the measurements at five towers (A, B, C, D and M) during the local slope wind period with changing grid spacing between 1 to 10m in both the horizontal and vertical. The reasons why we use the local slope wind data to test the model are that (1) these are the longest available measurement data during the local slope winds, (2) local slope winds are locally thermally driven with a minimum synoptic-scale disturbance which cannot be represented in our local scale model, and (3) the scale of air motion during calm nights should be smaller than nights with strong synoptic wind. The fact that the grid spacing fits the calm night means it is capable of resolving air flows during nights with stronger northerlies and
Because of the weak wind and high variation in wind direction, we use the 99.99% confidence level to plot the lower ($R_{\text{lower}}$) and upper ($R_{\text{upper}}$) bounds of the measured variable $R$ ($u, v, w, T$) at the measurement heights. We use the grid space setting based on when the modeling results best match the measurements. The matching standard is that about 70% of modeled variables are in the range of lower and upper bounds of the measurements (Figure 5.4).

The computational domain extends over $2 \times 2 \times 2$ km as our outer domain (D1), corresponding to $260 \times 260 \times 188$ grid intervals in the $x, y$ and $z$ directions. The $240 \times 240$ m inner domain (D2) is located in the middle of the outer domain with finer grids in the horizontal. The horizontal grid space at D2 is 4 m and at D1 it is stretched with a power law, starting with a horizontal grid spacing of 3 m at the edge of the forest and about 11 m at the lateral boundaries. The mesh is stretched in the vertical using a power law, starting with a vertical grid spacing of 3 m at the ground surface and 15 m at the top boundary. The stretching power in both horizontal and vertical is 1.3. The topography is extracted from the NASA Shuttle Radar Topographic Mission (SRTM) 90 m digital elevation data v4 (http://srtm.csi.cgiar.org/). The distribution of three canopy classifications covering D2 is shown in Figure 5.2. The outer region of D2 is mainly covered by grassland. But due to the lack of vegetation data, we specify the ground surface roughness height of 0.8.

The prescribed wind profile function is used for wind velocity at the north and south inflow boundaries of $2 \times 2$ km domain for northerlies and southerlies, respectively.

$$U_g(z) = \frac{u_\pi}{k} \ln \left( \frac{z + d}{z_0} \right)$$

(5.1)

where $d = 0.6h$ is the zero plane displacement height, $h$ is the height of canopy, $z_0 = 0.8$ is roughness length. The friction velocity $u_\pi$ is derived by the relation that $-u'w' = u_\pi^2$, where $-u'w'$. 


is measured by the sonic at level 4 (32m) on the permanent tower M, based on the assumption that Reynolds stress is constant above the canopy (Yi, 2008). The vertical velocity at the inflow boundary is specified to be zero. At the outflow boundaries, a zero gradient boundary condition is applied. The zero gradient boundary condition is specified for the top and four lateral boundaries of local slope winds.

The ambient temperature is $\theta_0(z) = \theta_{00} + \gamma z$, where $\theta_{00}$ is the potential temperature at $z = 0$, which is specified with soil temperature measured at the depth of 5 cm. $\gamma$ is ambient lapse rate, set to -6°C km$^{-1}$. The energy source $Q_{\text{source}}$ is specified as upward radiative heat flux. The upward radiative heat flux $Q_h$ at the top of the canopy is the sensible heat measured at the 32m sonic. Because heat flux peaks at the top of the canopy (Dupont and Patton, 2012), we specify exponential decrease of $Q_{\text{source}}$ from the top of canopy to zero at $z = 0$ correlated with accumulative leaf area density $L(z)$,

$$\frac{Q(z)}{Q_h} = a \exp\left(b \frac{L(z)}{LAI}\right) \tag{5.2}$$

where $a$ and $b$ are constants and $L(z) = \int_0^z a(z)dz'$. The heat flux at the top of the canopy and soil heat flux for each wind is shown in Table 5.2.
### Table 5.1 Typical wind systems and modeling periods

<table>
<thead>
<tr>
<th>Cases</th>
<th>Time Period (LST)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Northerlies</td>
<td>00:00, July 11 (day 192) 00:00, July 13 (day 194)</td>
</tr>
<tr>
<td>Southerlies</td>
<td>00:00, July 25 (day 206) 06:00, July 26 (day 207)</td>
</tr>
<tr>
<td>Slope wind</td>
<td>00:00, July 28 (day 209) 00:00, July 30 (day 211)</td>
</tr>
</tbody>
</table>

### Table 5.2 Parameter values used in the model

<table>
<thead>
<tr>
<th>Cases</th>
<th>H (Wm⁻²)</th>
<th>G (Wm⁻²)</th>
<th>Ta (°C)</th>
<th>Ts (°C)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Northerlies</td>
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<td>-0.39</td>
<td>9.81</td>
<td>9.04</td>
</tr>
<tr>
<td>Southerlies</td>
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<td>1.26</td>
<td>12.54</td>
<td>12.37</td>
</tr>
<tr>
<td>Slope wind</td>
<td>-10.53</td>
<td>1.81</td>
<td>15.94</td>
<td>14.44</td>
</tr>
</tbody>
</table>
Figure 5.4 The variables (u, v, w, T) profiles at 99.99% confidence level for local slope wind measured at A, B, C, D and M towers during the nighttime period (9pm-4am LST) from ADVEX campaign (light blue), showing the lower and upper values of the variables at the measurement height. The pink dots indicate the average values at the measurement height. The dark blue dotted lines denote the profiles by the numerical model.
5.4 Mean Wind Field

As our model is configured with detailed vegetation data in the D2 (240×240m) domain where the tower measurements are available, we confine discussion of results in that domain.

5.4.1 Local Slope Winds

The local slope winds are thermally-forced drainage flow \((w < 0)\) without any synoptic scale disturbance (Figure 5.4). Winds generally blow towards the southeast along the diagonal of D2, following the direction of slope. However, in the north, easterly drainage winds are dominant due to the slope falling to the east above about \(y = 150m\). In the south, winds shift from southwest to south because of the lowest elevation between \(x = 100\) and 200m where the slope is steepest. The drainage winds are only significant within the canopy, and the flow patterns remain similar at all levels through the canopy (Figure 5.5a). However the streamlines become inflected at the top of the canopy. The inflection is caused by canopy disturbance, which is notable when drainage winds become very weak. Above the canopy, air is sinking into the canopy, \(i.e.,\) only vertical velocity is important and the slope-following wind patterns are diminished due to reduced thermal-topographic effects.

The drainage flow is primarily driven by thermal-topographic effects, which can be described as hydrostatic buoyancy force given as:

\[
F_{hs} = g(\Delta \theta / \theta_0) \sin \alpha,
\]

where \(\alpha\) is the slope angle, \(\Delta \theta\) is the potential temperature difference between the ambient air and the colder slope flow, and \(\theta_0\) is the ambient potential temperature. However, the wind profiles within the canopy indicate the combined effects of topography and vegetation structure (Figure 5.6). The wind profiles at Tower A, B and M display similar profile shape with a
maximum wind speed around 15-20m. The maximum wind at Tower A is much smaller than that at Tower B and M, because hydrostatic buoyancy forces are small at Tower A where the slope is gentler than at Tower B and M. In addition, Tower A, B and M are surrounded by re-growth forest that is characterized by dense foliage in the lower canopy. The dense foliage exerts a large drag force on the drainage flow near the ground surface. Therefore, the level of maximum wind speed is elevated up to about mid-canopy, while most drainage flow occurs at lower trunk-level in open-trunk forests (Turnipseed et al., 2003; Yi et al., 2005; Belcher et al., 2008).

Mature forest in our model configuration is characterized by typical foliage distribution with open trunk and maximum leaf area density in the upper canopy. Wind profile at Tower C is determined by the canopy morphology of mature forest with a primary wind maximum at the trunk region and a secondary wind maximum near the top of the canopy. Although drag force at Tower C is significant, wind speed at Tower C remains strong compared with wind speed at towers on the upper slope. This is because the hydrostatic buoyancy force (equation 5.3) is large enough to maintain drainage flow on the steep lower slope. In addition, airflow experiences long slope acceleration before reaching Tower C. Although Tower D experiences small hydrostatic buoyancy force, it encounters the strongest drainage flow near the ground surface, which is attributed by the small drag force exerted by sparse trees in grassland. The fast attenuation of wind speed from ground surface to the top of the canopy at Tower D indicates that thermal forcing is only important near the ground surface for the sparse canopy, because of LAD-related heat flux setting.

5.4.2 Northerlies

During northerlies periods, wind direction and wind speed of the local slope winds are modified after interacting with synoptic northerlies (Figure 5.5b). However, the modification is
limited within the canopy, indicating the dominant local topographic effects. As the slope generally falls to the southwest, the wind is blowing in the south and southwest direction at the boundary of the research domain and then diverted to the west as the local drainage flow in the north, where there’s a slope falling to the west. These easterly winds only occur in the lower canopy (Figure 5.5b, 5m and 10m). Winds twist from southwest to south and southeast in the south, where the elevation is lowest and slope is steepest. The topographic effects decrease with height and become very weak at the top of the canopy, where northerlies prevail. Even so, we can still see southward winds in the southeast due to the steep slope. The initial prescribed wind speed of northerlies is about 7.4 ms\(^{-1}\) at the top of the canopy, according to equation (5.1). The strong northerlies enhance local slope winds both within and above canopy. The enhancement is much stronger in the upper canopy and above canopy than in deep canopy. The limited influences of northerlies on local slope within canopy winds demonstrate that deep canopy flow is governed by local radiative forcing of the topography and vegetation instead of downward momentum transport from synoptic northerlies above the canopy.

A remarkable flow pattern is the up-slope winds blowing to the northwest in the lower canopy between Tower A and M. This up-slope motion is associated with the recirculation region in the north-south (N-S) direction (Figure 5.8a, b) and updraft in the east-west (E-W) direction (Figure 5.8c, d). Instead of occurring on the lower level of the lee slope, recirculation occurs at the gentle upper slope in our research domain, which is attributed to the vegetation structure and distribution. Our simulation (2\times2\text{km}) is conducted for the clearing-forest-clearing configuration. As the wind blows from the northern clearing into the forest, it lifts after passing the sparse trees of the grassland that is located north of Tower A. Between Tower A and M, the dominant vegetation is re-growth forest characterized by dense foliage at trunk level. The dense
canopy exerts a large drag force, which retards flow in the deep canopy, resulting in a region of reversed flow, *i.e.* recirculation, underlying the lifted flow. The upward action was observed in the stable night at the Niwot Ridge Ameriflux site, which is also partly explained by vegetation distribution (Turnipseed et al., 2003). The explanation of flow convergence within canopy is also applicable to our results. Flow from north and east converges between Tower A and M, forcing a rising motion.

The recirculation bubbles extend through the re-growth forest in the N-S direction (Figure 5.8a, b). The reversed flow ends before reaching the edge between re-growth and mature forest in the west (cross sections at $x = 88, 104,$ and 120 in Figure 5.8a and b). In the east, the extension of reversed flow is further north than that in the west. The reversed flow extends across the edge between re-growth and mature forest, because of northward coverage of mature forest. At N-S cross section $x = 88$m (Figure 5.8a, b) where northerlies experience the longest expanse of re-growth forest, the recirculation breaks into two at about $y = 136$m. From the west to east, the average depth of the recirculation region (mean height at which $v = 0$) increases from 8m to about 10m, which confirms that increasing slope angle leads to increasing depth of recirculation (discussed in Chapter 3). The maximum depth of recirculation is at the mid-canopy, about 15-16m high.

The updraft motion in the west domain is due to temperature difference in the E-W direction (Figure 5.8c, d) which is mainly driven by vegetation distribution. In the east domain, re-growth forest and mature forest are dominant. LAI for both forests is around 5. The dense canopy radiates much more long-wave radiation than the grassland with LAI = 1.04, leading to stronger cooling in the east. As the colder air from the east advances into the warmer air in the west, the less dense warm air is lifted upward. In front of the lifted flow, there’s a region of
reversed flow (eastward) above a shallow layer of westward wind near the ground surface. The reversed flow occurs below the top of the canopy in re-growth forest (cross sections at \(y = 136\) and \(152\) in Figure 5.8c, d), but extends to above canopy in grassland in the south (cross sections at \(y = 88, 104, 120\) in Figure 5.8c, d).

### 5.4.3 Southerlies

The modification of southerlies to local slope winds is significant, although the southerlies are relatively weak (Figure 5.5c). Wind speed of local slope winds is strengthened, but with altered wind direction. The predicted wind speed is weaker than \(0.5 \text{ ms}^{-1}\) at the top of the canopy, which is much weaker than for the southerlies (about \(3 \text{ ms}^{-1}\)) but stronger than local slope winds. In the north, winds remain southerly with up-slope motion \((w > 0\) at Tower A and M). However, the wind shifts eastward \((u > 0)\) in the south. This westerly component is much stronger than the southerly component, leading to cross-slope flow instead of up-slope flow (Figure 5.5c). The altered flow pattern remains significant up to triple canopy height above which the southerlies become dominant.

Streamlines at the cross sections show the flow pattern in both the vertical and horizontal during northerly wind periods (Figure 5.9). Air sinking is obvious above the canopy during southerly wind periods, because southerlies are weak. In the N-S direction, air subsides strongly in the south between \(y = 80\) and \(100 \text{ m}\) (Figure 5.9a, b) where there is stronger radiative cooling from mature forest. The subsiding air diverges to the north and south near the ground surface. On the northern upper slope, the slope is gentle, so weak local slope winds are reversed by the prevailing southerlies. In the south, the steep slope results in strong drainage flow overwhelming the southerlies. Even so, the drainage flow is not strong enough to reverse southerlies, but both
flows interact to become westerly winds. The westerly winds are accompanied with sinking motion from above the canopy, due to radiative cooling.

**Figure 5.5** Wind streamlines at vertical levels ($z = 5, 10, 20$ and $30$ m) in $240 \times 240$ m domain for (a) Local slope winds, (b) Northerlies and (c) Southerlies winds. The colored isolines at $z = 0$ denote the elevation in the $240 \times 240$m domain.
Figure 5.6 Wind profiles at five towers during local slope wind period.
5.5 Thermal Analysis

Temperature inversion develops in all the vegetation canopies due to radiative cooling. However, the temperature distribution and vertical profiles are different under different prevailing wind systems and vegetation categories (Figure 5.7-5.10).

5.5.1 Local Slope Winds

During local slope wind periods, the apparent temperature difference between in- and above canopy is caused by radiative cooling in the vegetation canopy (Figure 5.7a, c). The cooling occurs all through the canopy, resulting in a temperature difference of 3-3.5°C between ground surface and the canopy top. This temperature difference is in the range of observed temperature differences in the 10 m high orchard (Dupont and Patton, 2012) and 21.5 m high aspen forest (Mahrt et al., 2000) under stable conditions with weak wind. The uneven distribution at the top of cooled air is caused by the leaf area density difference in the modeled three canopy shapes. Weak cool air pooling within canopy in the downwind direction is caused by cold air draining down the slope. A relatively warmer canopy layer at the domain boundaries is found because there is no extra canopy-depth cooling outside of the 240×240m domain where there is no vegetation.

Temperature profiles clarify that strong inversion occurs in the upper canopy because of strong radiative cooling, and temperature becomes uniform above the canopy for all the forest types because there is no cooling source (Figure 5.10a). However, the temperature profiles show different shapes in lower canopy, which is directly induced by the canopy structure-related radiative cooling. At Tower A, B and M, the dominant vegetation is re-growth forest. The re-growth forest is characterized by dense foliage below about 12m, above which temperature
inversion is developed. The inversion becomes very weak below 11m. At Tower M and B, negative temperature gradient is found below 6m, resulting in a ‘cold bulge’ at about 6m where leaf area density reaches its maximum. The maximum leaf area density in the lower canopy reduces the effects of radiative cooling above 6m, so the thermal condition below 6m is mainly determined by ground heat flux. The positive ground heat flux during local slope wind periods leads to weaker inversion and warmer air near the ground surface, implying the near-neutral or even unstable layer in dense canopy (see chapter 4 and Shaw et al., 1988; Jacobs et al., 1994; Dupont and Patton, 2012). In lower canopy, temperature at Tower M is much colder than temperature at Tower A and B, because Tower M is at downwind direction of slope wind, experiencing cooled drainage flow. In the upper canopy, air at tower B is colder than at Tower A and M, because Tower B is located at about 10m east to the mature forest. A stronger cooling in the upper dense canopy of mature forest can enhance cooling in the upper canopy of re-growth forest in the neighborhood.

Temperature profiles at Tower C depicts the cooling in the upper canopy of mature forest, with the coldest air and strongest temperature inversion above 17m high compared with temperature at other towers (Figure 5.10a). In lower canopy, the temperature inversion is reduced due to very week radiative cooling in the lower canopy and positive ground heat flux. In contrast, temperature inversion at Tower D is significant in lower canopy with warmer temperature above 8m due to the fact that Tower D is surrounded by sparse grassland with its maximum leaf area density at ground surface. Radiative cooling in sparse trees in grassland is stronger in lower canopy than at the same height in re-growth and mature forest. The strong cooling in lower canopy overwhelms surface positive heat flux, which results in strong inversion near the ground surface.
5.5.2 Northerlies

During northerly wind periods, air is warmer in the north but cooler in the region of recirculation and south (Figure 5.8a). The warmer north is attributed to the higher elevation and gentler slope. Wind in the recirculation region is weak and reversed, resulting in cool air stagnation. In the south, the downslope northerlies drain cooler air to the lower slope, especially in the southeast, where elevation is lowest and slope is steepest. The temperature gradient in the E-W direction is mainly attributed to the vegetation distribution. Most of the mature forest is distributed in the east where stronger radiative cooling occurs, leading to relatively cooler east and warmer west. The temperature difference is responsible for the updraft flow in the E-W direction (section 5.4.2).

During northerly wind periods, the maximum temperature difference from the top of the canopy to the ground surface is about 2.5 °C (Figure 5.10b). The temperature difference is much smaller and less variable compared with local slope wind periods, which can be explained by strong northerlies causing better mixing throughout the canopy. Temperatures at towers A, B and M surrounded by re-growth forest are very similar to temperatures during slope wind periods: warmest temperature is at tower A, which is at more elevated upwind direction; coldest temperature is at tower B, which is close to dense mature forest with strong radiative cooling; temperature at Tower M is intermediate between Tower A and B. The difference in profile shape is related to the flow pattern during northerly wind periods. Tower A and M are located in the region of recirculation. The temperature in the reversed flow is about 3°C cooler than air in the upwind direction on the upper slope (Figure 5.8a). Unlike temperature at tower B and M, temperature inversion is very strong in the lower canopy at tower A, because tower A is located at the boundary of reversed flow, experiencing strong temperature gradient in both the horizontal
and vertical. The temperature profile at Tower C shows the coldest air in the upper canopy and weak inversion in the lower canopy during local slope wind periods. Temperature at tower D indicates homogeneous cooling in grassland due to its canopy structure and strong wind mixing. The wind speed at tower D is stronger than at other towers because of less drag force exerted by canopy elements and acceleration in the down slope wind.

5.5.3 Southerlies

During southerly wind periods, the temperature difference is lower than 2°C in the research domain (Figure 5.9), which is induced by both very weak radiative cooling and cross-slope wind mixing. The weak cooling doesn’t have obvious effects on the temperature near the boundaries of the domain, even in the south where drainage flow is developed. The weak drainage flow cannot effectively drain the cold air down the slope.

The temperature difference is less than 1°C between ground surface and the top of the canopy at all the towers except tower D (Figure 5.10c). At tower A, B, C and M, temperature inversion only develops in the upper to above canopies. Temperature is almost constant in the lower canopy (below 10m high), due to less radiative cooling and the positive ground surface heat flux. Temperature at tower A is still warmer than at tower B and M, but with larger temperature difference throughout the canopy. The effects of mature forest cooling on temperature at tower B becomes trivial, showing a minor difference in temperature above 20m at tower B and M. The ‘cold bulge’ occurring during local slope wind periods is also present during southerly wind periods at tower M because of positive heat flux from the ground surface.

Although the cooling in the dense mature forest at tower C is stronger than at other towers, temperature at tower C is relatively warmer and inversion is weaker than at towers A, B, and M, which is contrary to the condition during local slope wind and northerly wind periods.
The warmer temperature at tower C can be partially explained by the weak radiative cooling during southerly wind periods. Although tower C is located on the slope of the drainage flow, the drainage flow isn’t strong enough to pool cool air in the lower slope. At tower D, temperature is warmest in the upper canopy and inversion is strongest near the ground surface, as during local slope wind periods. The highest foliage density-related radiative cooling in the lower canopy dominates the thermal condition near the ground surface.
Figure 5.7 Distribution of Temperature (°C) and CO₂ difference (ppm) at cross sections during Local slope wind period in N-S direction: (a) Temperature and (b) CO₂ difference; E-W direction: (c) Temperature and (d) CO₂ difference. The white solid lines with arrows denote the wind streams at the cross sections.
Figure 5.8 Distribution of Temperature (°C) and CO₂ difference (ppm) at cross sections during Northerly wind period in N-S direction: (a) Temperature and (b) CO₂ difference; E-W direction: (c) Temperature and (d) CO₂ difference. The white solid lines with arrows denote the wind streams at the cross sections. The yellow dashed lines in (a) and (b) denote the top of reversed flow.
Figure 5.9 Distribution of Temperature (°C) and CO₂ difference (ppm) at cross sections during Southerly wind period in N-S direction: (a) Temperature and (b) CO₂ difference; E-W direction: (c) Temperature and (d) CO₂ difference. The white solid lines with arrows denote the wind streams at the cross section. The dashed lines in (a) and (b) denote the boundary where the winds diverge to south and north.
Figure 5.10 Temperature Profiles at five Towers during prevailing wind periods (a) Local slope winds, (b) Northerlies and (c) Southerlies.
5.6 CO₂ Distribution

CO₂ emission rate is only determined by soil temperature in equation (2.8, 2.9). The total amount of CO₂ emission is very similar for different canopies experiencing the same wind period, because the temperature difference on the ground surface is very small. CO₂ emission rate is variable vertically levels dependent on leaf area density distribution in equation (2.9).

During local slope wind periods, CO₂ is built up in the south and east (Figure 5.7b and d). In the N-S direction, a maximum of 17 ppm higher CO₂ concentration is in the downwind of the local slope wind as compared to the upper slope. Particularly high CO₂ concentration accumulates in the southwest, where slope is steep and elevation is low. Relatively high concentration of CO₂ can extend north to the upper slope and fill in the whole canopy layer, which is related to the distribution of mature forest. In the E-W direction, CO₂ concentration is high in the east due to the westerly slope wind. We expect the CO₂ concentration in the west to be as high as in the east because of the westward slope winds. However, it shows good mixing in the west, probably due to sparse trees in grasslands in the west, which makes it easier for CO₂ to vent out of the canopy. Relatively stronger wind in the lower canopy and wind shear throughout the canopy at tower B and D, as compared to tower B, can be demonstrated by the wind profiles that (Figure 5.6).

During northerly wind periods, CO₂ is builds up in the recirculation region in the N-S direction, which coincides with cool air pooling on the mid-upper slope (Figure 5.8). The highest CO₂ concentration is in the west where the recirculation is prolonged to the south because of the decreased coverage of mature forest. Most of the CO₂ is restricted below about 16m, above which CO₂ is well mixed by strong prevailing northerlies. On the lower slope of the downwind direction, there is no obvious CO₂ enrichment because of elevated prevailing northerlies, as CO₂
is mainly emitted from the ground surface. In the E-W direction, CO$_2$ is confined to a very shallow layer on the ground surface, mostly in the west, which is caused by flow descending from east to west. The updraft flow has minor influence on CO$_2$ transport because the updraft flow is at a higher level.

During southerly wind periods, much better mixing of CO$_2$ is seen in comparison with northerly and local slope wind periods, although there is a very shallow layer below 6m with higher concentration of CO$_2$ (Figure 5.9). CO$_2$ distribution in the surface shallow layer is quite homogeneous along the slope. The concentration difference is mostly less than 10 ppm due to the accompanied sinking motion with the cross-slope flow. The cross-slope winds enhance the CO$_2$ mixing in the canopy layer. Even on the upper slope, the CO$_2$ difference is about 6ppm due to that descending winds flowing northward.

### 5.7 Overview of Model-Measurement Comparisons

As the prescribed boundary conditions are tested against the local slope wind, our model shows the best prediction for local down-slope drainage flow (Figure 5.3). About 70% of the model predictions at the measurement heights are in the lower and upper bounds of measurements within the 99.99% confidence level. Among the five towers, the wind speed tends to be under-predicted at most heights compared with the measured mean wind speed. Haiden and Whiteman (2005) indicated that the drainage flow accelerates down the long slope. However, our modeling domain (2×2km) is just a small part of the long north to south Alps slope, so that the acceleration on the upwind slope out of the modeling domain is missed.

During northerly wind periods, our model successfully predicts the down-slope wind measured at all the towers except at tower A, where the wind is blowing from north to south with upward motion (Figure 5.11). The strong upward motion (w is increasing with height to be as
large as 1 m s\(^{-1}\)) indicates flow convergence in the lower canopy with local vegetation distribution. During southerly wind periods, our model predicts southerlies throughout the canopy (Figure 5.12). However, just above and in the upper canopy, predicted southerlies are much weaker than the measured southerlies but stronger than pure local slope winds. This can reflect the interactions between southerlies and drainage flow. The predicted u component is opposite to the measurements during both northerlies and southerlies because the westerly component in the synoptic northerlies and easterly component in synoptic southerlies (Feigenwinter et al., 2010a) have not been taken into account in our model. It implies that wind direction is very sensitive to the large-scale wind regime. However, the large-scale wind direction cannot just be determined by the measurement near the top of the canopy, because of canopy flow characteristics near the canopy top (strong shear, inflection, etc.).

It is encouraging that our model shows a better prediction in the deep canopy than in the upper canopy. It is due to airflow in the deep canopy being dominated by local thermo-topographic forcing, while predictions in the upper canopy are the result of interactions between local slope winds and synoptic winds. However, the representations of synoptic influences in our local-scale model is very limited. Our model predicts the typical nocturnal temperature profiles in the canopy: temperature inversion in the upper canopy due to outgoing long-wave radiation and isothermal or inversion profile in the lower canopy (Baldocchi et al., 1983), and even negative wind gradients near the ground surface. However, the predicted temperature profiles are isolated from tower measurements. The mismatch between modeled and measured temperature is caused by the heat flux configuration such that the full energy balance equation is simplified to an outgoing radiative cooling. The temperature in the canopy is also very sensitive to water stress (Jensen et al., 1990).
The other uncertainties in solving the local scale topographic-canopy-flow arise from the rough topography data and vegetation classification. The SRTM 90m digital elevation data is very rough compared to 4m horizontal grid spacing in our model setting. It means a lot details of the topographic flow cannot be resolved on this 90m topography resolution. The real vegetation distribution is much more complicated than the three category classification. The 14.7% of the clearings cover in the 240×240m area (Montagnani et al., 2009) is not classified in our model configuration. In addition, the tower measurements we used from the ADVEX campaign was conducted in July 2005, while the field survey of vegetation classification was done in October 2009. The change in vegetation distribution can contribute to the mismatch between the modeled and measured data.
Figure 5.11 The variables (u, v, w, T) profiles at 99.99% confidence level for Northerlies measured at A, B, C, D and M towers during the nighttime period (9 pm-4am LST) from ADVEX campaign (light blue), showing the lower and upper values of the variables at the measurement height. The pink dots indicate the average values at the measurement height. The dark blue dotted lines denote the profiles by the numerical model.
Figure 5.12 The variables (u, v, w, T) profiles at 99.99% confidence level for Southerlies measured at A, B, C, D and M towers during the nighttime period (9 pm-4am LST) from ADVEX campaign (light blue), showing the lower and upper values of the variables at the measurement height. The pink dots indicate the mean values at the measurement height. The dark blue dotted lines denote the profiles by the numerical model.
5.8 Summary and Conclusions

We apply computational fluid dynamics (CFD) model to investigate nocturnal flow dynamics and its related thermal and CO$_2$ transport under different synoptic forcing in a forested complex terrain. In the absence of synoptic-scale forcing, thermal-driven topographic flow dominates, which is characterized by weak (a maximum wind speed of 0.22 ms$^{-1}$) local down-slope northerly winds. The down slope winds blow throughout the vegetation canopy and contribute to rich CO$_2$ accumulation within the canopy in the downwind direction. The modification of local slope winds by synoptic northerlies is limited to the deep canopy. However, the northerlies significantly intensify local slope winds. Recirculation is developed in the N-S direction when prevailing northerlies blow down-slope into regrowth forest which has high leaf concentration in the lower canopy, exerting large drag force on canopy flow. The wind direction of local slope winds can be modified by the weaker southerlies all through the canopy and up to triple canopy height, above which the local slope winds disappear. The interactions of local slope wind and southerlies result in reduced up-slope southerlies on the upper slope and dominant down-slope wind on the lower steep slope where the local down-slope winds exceed the southerlies.

Temperature distributions indicate that dense mature and re-growth forest canopies are subject to stronger cooling in the upper canopy. The upper canopy cooling has minor influence on the thermal conditions in the deep canopy because the layer with dense foliage density reduces vertical heat transfer, leading to very weak inversions in the deep canopy. In contrast, cooling in the canopy overwhelms the weak heat flux at the ground surface in the sparse grassland, resulting in strong inversions near the ground surface. Radiative cooling is the primary driving force of the within-canopy down-slope winds. The correlation between thermal
condition and wind microstructures in the canopy is obvious during northerly wind periods, which is expressed as the cooler recirculation in the N-S direction, and updraft occurring when winds blow from cooler regions to warmer regions in the E-W direction.

CO₂ transport is primarily determined by the wind. The local down-slope winds throughout the vegetation canopy contribute to rich CO₂ accumulation within canopy in the downwind direction. During northerly wind periods, recirculation in the N-S direction is characterized by high concentration of CO₂ up to the top of reversed flow, while CO₂ is built up in a very shallow layer in the E-W direction, where flow subsides from east to west. The updraft motion above the subsided flow does not affect CO₂ in the shallow surface layer, as the main CO₂ source is the ground surface. Compared with northerly wind and local slope wind periods, there is no extensive CO₂ accumulation during southerly wind periods, which is attributed to the cross-slope winds resulting in better mixing. However, a shallow layer with relatively high CO₂ concentration is found near the ground surface, which is attributed to the nocturnal air sinking.

Our CFD application in three-dimensional canopy represents a breakthrough in the modeling of interactions between local thermo-topographic flow and large-scale synoptic flow. However, this CFD model shows a weaker representation of large-scale synoptic influences. Downscaling techniques, such as coupled meteorology and CFD models, are necessary to improve the prediction of large scale forcing. In addition, high-resolution topography and vegetation data will help to solve the thermo-topographic airflow and its related ecosystem-atmosphere exchanges.
Chapter 6 Conclusions and Recommendations for Future Research

6.1 Conclusions

Canopy flow occurring within and immediately above vegetation canopies plays a substantial role in regulating atmosphere-biosphere interaction. However, it is poorly understood for its three-dimensional characteristics and variability relevant to canopy structure and distribution, synoptic weather condition and local topography. We apply the CFD model to investigate the three-dimensional canopy flow and its effects on CO₂ transport in forested terrain.

The CFD model is first applied to two-dimensional hilly terrain to explore the canopy flow with recirculation development. We find that flow recirculation is a typical phenomenon in complex terrain and plays a key role in vegetation-atmosphere exchanges of mass and energy. The complexity and structure of recirculation strongly depend on slope. For gentle forested hills ($H/L < 0.8$), the recirculation structure is simply characterized by reverse flows without vortex, which are limited in the lower part of the canopy layer on leeward sides. The near-surface reverse flows greatly alter CO₂ distribution near the ground rather than enhance CO₂ exchange in vertical. For steep forested hills ($H/L > 0.8$), recirculation bubbles become larger and even deeper than vegetation height with one or multiple vortices, enhancing mixing of CO₂ and energy between vegetation and atmosphere. Consequently, steep slopes cause less advective CO₂ fluxes.

The thermal distribution and stability occurring within the canopy are substantially different from the ambient atmosphere. The thermal stability and its influence on the canopy flow in hilly terrain are explored in calm and stably stratified conditions. The stability around the canopy is characterized by stratification with a primary super stable layer above the top of the
canopy, a secondary super stable layer in the lower canopy, and an unstable layer within the canopy. The thermal stratification around the canopy primarily drives the airflow in the canopy layer on the slope, which can be expressed as the upper-canopy drainage flow (UDF) layer in the upper canopy inversion layer and lower-canopy drainage flow (LDF) layer in the inversion layer in the lower canopy. On gentle slopes, air in UDF sweeps horizontally to join the shallow LDF on the slope surface, while on steep slopes, the stagnated flow in LDF jumps upwards, perpendicular to the slope, from the lower canopy layer to join the shallow UDF in the upper canopy layer. The generation and direction of the shifting-wind structure within the canopy are induced by the hydrostatic buoyancy force and baroclinic instability on the slope, which are functions of the inclination of the slope.

The turbulent properties of the stratified canopy flow are closely associated with thermal and dynamic conditions on the slope. The downward transport of momentum in the canopy is reduced due to strong stability. Upward transport of momentum occurs in the deep canopy. The heat flux is predominantly transported downward with the minimum heat flux in the thermal transition zone in the middle of the canopy. TKE is available near the top of canopy on the midslope and downslope. The region experiencing strong wind shifts is on the lower slope where the wind shear is strong. The largest TKE values are found in the region of wake vortices across the canopy edge. Buoyancy production of TKE is a principal sink of TKE under stable conditions, which suppresses turbulence significantly near the top of the canopy and in the deep canopy. TKE is generated by shear production near the top of the canopy and by wake production in the canopy. The transport of TKE by pressure perturbation, which is of the same order as the production terms, supplies TKE where the buoyancy suppression is very strong and
extracts TKE where wake production is dominant. Our results suggest that the relative importance of pressure perturbation is enlarged as the slope increases.

The three-dimensional canopy processes resulted from the interplay between local thermo-topographic flow and large-scale synoptic flow is investigated in forested complex terrain at Renon, Italy. In the absence of synoptic-scale forcing, thermal-driving topographic flow dominates, which is characterized by weak down-slope northerly winds. The down slope winds blow throughout vegetation canopy and contribute to rich CO$_2$ accumulation within canopy in the downwind direction. The modification of local slope winds by synoptic northerlies is limited in deep canopy. However, the northerlies significantly intensify local slope winds. Recirculation is developed in N-S direction when prevailing northerlies blow down-slope into regrowth forest which has concentrated leaves in the lower canopy exerting large drag force on canopy flow. The wind direction of local slope winds can be modified by the weaker southerlies all through the canopy and up to triple canopy height above that the local slope winds disappear. The interactions of local slope wind and southerlies result in reduced up-slope southerlies on the upper slope and leading down-slope wind on the lower steep slope where the local down-slope winds exceed the southerlies.

Temperature distribution indicated that dense mature and re-growth forest canopies are subject to stronger cooling in the upper canopy. The upper canopy cooling has minor influence on the thermal conditions in deep canopy because the layer with dense foliage density reduces vertical heat transfer, leading to very weak inversions in deep canopy. In contrast, cooling in the canopy overwhelms the weak heat flux on ground surface in the sparse grassland, resulting in strong inversions near the ground surface. Radiative cooling is the primary driving force of the within-canopy down-slope winds. The correlation between thermal condition and wind
microstructures in the canopy is obvious during northerly wind period, which is depicted as the cooler recirculation in N-S direction and updraft occurring when winds blow from cooler region to warmer region in the E-W direction.

CO$_2$ transport is primarily determined by the wind. The local down-slope winds throughout vegetation canopy contribute to rich CO$_2$ accumulation within canopy in the downwind direction. During northerly wind period, recirculations in N-S direction are characterized with high concentration of CO$_2$ up to the top of reversed flow, while CO$_2$ is built up in a very shallow layer in E-W direction where flow subsides from east to west. The updraft motion above the subsided flow does not affect CO$_2$ in the shallow surface layer, as the main CO$_2$ source is the ground surface. Compared with northerly wind and local slope wind periods, there’s no extensive CO$_2$ accumulation during southerly wind period, which is attributed to the cross-slope winds causing better mixing. However, a shallow layer with relatively high CO$_2$ concentration is found homogeneous near the ground surface, which is attributed to the nocturnal air sinking.

Our CFD application in three-dimensional canopy However, this CFD model shows a weaker representative of large-scale synoptic influences. Downscaling technique, such as coupled meteorology and CFD model, is necessary to improve the prediction of large scale forcing. In addition, high-resolution topography and vegetation data will be advantages to solve the thermo-topographic airflows and its related ecosystem-atmosphere exchanges.

6.2 Suggestions for Future Work

Our numerical experiments represent a breakthrough in the modeling of canopy flow and its related mass and energy transport processes under interactions between local thermo-topographic flow and large-scale synoptic flow. Conclusions drawn from the numerical studies
provide insights into the issues caused by complex terrain and canopy structure in eddy-flux measurements. However, these numerical results also need to be justified by intensive field experiments in the future.

The CFD model applied in local scale shows a weaker representative of large-scale synoptic influences. Downscaling technique, such as coupled meteorology and CFD model with parameterization of ecological processes, is necessary to improve the prediction of large-scale forcing. In addition, high-resolution topography and vegetation data will be advantages to solve the thermo-topographic airflows and its related ecosystem-atmosphere exchanges which could be achieved by data assimilation of FLUXNET and satellite observations into the model to improve its prediction ability and extend its applicability.
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