Characteristics of the Near-Surface Boundary Layer within a Mountain Valley during Winter

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ABSTRACT

Within mountainous regions, estimating the exchange of sensible heat and water vapor between the surface and the atmosphere is an important but inexact endeavor. Measurements of the turbulence characteristics of the near-surface boundary layer in complex mountain terrain are relatively scarce, leading to considerable uncertainty in the application of flux-gradient techniques for estimating the surface turbulent heat and mass fluxes. An investigation of the near-surface boundary layer within a 7-ha snow-covered forest clearing was conducted in the Kananaskis River valley, located within the Canadian Rocky Mountains. The homogeneous measurement site was characterized as being relatively calm and sheltered; the wind exhibited considerable unsteadiness, however. Frequent wind gusts were observed to transport turbulent energy into the clearing, affecting the rate of energy transfer at the snow surface. The resulting boundary layer within the clearing exhibited perturbations introduced by the surrounding topography and land surface discontinuities. The measured momentum flux did not scale with the local aerodynamic roughness and mean wind speed profile, but rather was reflective of the larger-scale topographical disturbances. The intermittent nature of the flux-generating processes was evident in the turbulence spectra and cospectra where the peak energy was shifted to lower frequencies as compared with those observed in more homogeneous flat terrain. The contribution of intermittent events was studied using quadrant analysis, which revealed that 50% of the sensible and latent heat fluxes was contributed from motions that occupied less than 6% of the time. These results highlight the need for caution while estimating the turbulent heat and mass fluxes in mountain regions.

1. Introduction

From a hydrological perspective, there is substantial interest in estimating rates of snowmelt from mountainous regions, which are often the primary hydrological contributors to rivers that supply water to vast surrounding areas. Mountain environments have an inherently high degree of topographical variation and land cover heterogeneity, thus interactions between the land surface and the atmosphere are often complex and are difficult to predict. In support of this objective, there are numerous models available to simulate the energy and mass balances of snow that have been developed for various snow physics applications, including avalanche prediction (e.g., Brun et al. 1989; Bartelt and Lehning 2002), military research (e.g., Jordan 1991), and snowmelt (e.g., Marks and Dozier 1992; Tarboton and Luce 1996). While application of these models in complex terrain is met with numerous challenges, the focus of this study concerns the prediction of the turbulent fluxes of sensible and latent heat. In terms of the snow energy budget, the sensible heat flux can be an important source of melt energy, particularly during windy conditions (Marks et al. 2008; Jackson and Prowse 2009). The latent heat flux often represents a loss of snow water equivalent due to sublimation, which can be particularly significant under chinook winds (Golding 1978; Schmidt et al. 1998; Hayashi et al. 2005). Conversely, under humid conditions the latent heat flux can also be a significant source of melt energy due to condensation of water vapor on the snow surface (Marks et al. 1998).
All of the aforementioned energy balance models estimate the turbulent heat fluxes using the bulk transfer technique while applying Monin–Obukhov (M–O) similarity theory to account for the effect of atmospheric stability (as reviewed by Male and Gray 1981; Morris 1989; Andreas 2002). With this method the turbulent fluxes are considered to be proportional to the gradients measured between the snow surface and a reference height, which must be corrected to account for air density stratification. The significant advantage of this approach is that the turbulent fluxes can be estimated based solely upon local aspects of the flow and the interacting surface. These techniques have their theoretical origins in the ideal fully developed, steady-state atmospheric boundary layer that is hypothesized to exist over flat, homogeneous terrain. Notwithstanding, the bulk transfer technique is commonly applied in complex regions to estimate the turbulent energy fluxes (e.g., Munro 1989; Hood et al. 1999; Etchevers et al. 2004; Sicart et al. 2005). Application of this approach in nonhomogeneous terrain requires the wind speed, temperature, and water vapor gradients to be measured within an internal boundary layer in which the mean and turbulent properties of the flow are in equilibrium with the local surface, that is, have adjusted to the boundary conditions imposed by the local surface. The notion of an equilibrium boundary layer, as originally proposed by Townsend (1961), implies that close to the surface the balance of turbulent kinetic energy is dominated by the local rates of turbulence production and dissipation, such that the turbulent properties of the flow are essentially unaffected by processes elsewhere in the flow. Thus, for neutral boundary layers the production of turbulence is dominated by the generation of shear stress at the surface. The occurrence of an equilibrium layer is essential for first-order closure modeling methods to be used successfully. It is noted that, although higher-order turbulence closure models have been adapted for some atmospheric boundary layer applications (e.g., Wilson 1988; Denby 1999), they are infrequently used in operational land surface energy balance models and are not discussed further here.

In mountainous terrain, study sites are typically selected within open environments that exhibit some degree of local homogeneity, such as forest clearings, meadows, snow-covered frozen lakes, or broad terraces or ridges above the tree line. In ideal situations, the station is located such that the upwind fetch is maximized to ensure adequate distance for the development of an internal boundary layer. The obvious difficulty is that unobstructed fetch lengths are often very limited in mountainous terrain, and therefore siting of instrumentation is often a compromise between characterizing fluxes from important landscape units while attempting to minimize measurement errors (Pomeroy et al. 2003). In support of this approach, and for the use of first-order models in these environments, there have been a number of encouraging examples reported within the recent literature that have documented the presence of an equilibrium boundary layer within clearings located within complex terrain. Hammerle et al. (2007) using eddy covariance, measured surface fluxes within a meadow located on a 24° slope in the Stubai Valley, Austria, and compared them to fluxes observed in the open valley bottom. They observed that the degree of energy balance closure was similar at both sites; however, the quality of the data was sensitive to wind direction as a result of differences in upwind fetch characteristics. In a similar study, Hiller et al. (2008) reported good closure of the energy balance on a mountainside terrace that sloped 25° and did not reveal any disparaging issues with flux measurements associated with the complex terrain. In their study, the wind system was predominantly oriented along the valley axis. De Franceschi et al. (2009) examined near-surface turbulence statistics in a broad alpine valley in the Italian Alps that had a very homogeneous land cover and found that the turbulence statistics were well described by M–O similarity theory, regardless of wind direction and stability regime. Moraes et al. (2005) examined the turbulence characteristics within the Jacuí River valley in southern Brazil, which has surrounding ridges that were 350 m high, and found that the characteristics of the near-surface turbulence were strongly dependent on upstream flow conditions. In particular, when the wind directions were parallel to the valley and the flow had a long upwind fetch, an equilibrium layer was observed and M–O similarity theory was considered to be valid. However, for flow directions transverse to the valley, the flow could not be considered to be in equilibrium. These studies suggest that, under certain conditions within complex terrain, the wind flow near the surface can be interpreted in a similar manner as from more ideal terrain. However, the practitioner must also consider the cases where the terrain is configured in a manner in which the winds are subject to additional perturbations due to interaction between complex wind fields. For example, Weigel et al. (2007) studied the turbulence structure within the Riviera Valley in southern Switzerland and discovered that strong shear zones aloft were created as a result of the interactions between the valley flow, the katabatic flow from the valley slopes, and cross-valley circulations. The creation of turbulent kinetic energy was not strongest at the land surface, but in an elevated position. Similar studies regarding the turbulent flow structure near the surface in complex mountain basins are lacking.

Within this paper, the characteristics of the turbulent fluxes of momentum, heat, and water vapor as measured
in an open meadow in the bottom of a mountain valley are reported. The primary purpose of the study was to determine if the turbulent transfer processes within the valley were similar to the homogeneous environments from which much of the flux estimation theory is based upon. In particular, the degree to which the surrounding complex terrain influenced the flux generation processes was investigated.

2. Methods

a. Site description

The “Hay Meadow” mountain valley site is located at the Marmot Creek Research Basin (50°56’N, 115°08’W) in the Kananaskis River valley, in the eastern slopes of the Canadian Rocky Mountains. It is a relatively large (~7.5 ha) grass-covered clearing (Fig. 1a) that is bordered by the Kananaskis River to the east, mixed aspen and pine forest to the north and south (in which the trees are approximately 13 m tall), and sloping coniferous forest to the west. The vegetation in the clearing is mainly grass. The local terrain gently slopes to the southeast with an average grade of 2.5% but is flanked by steep mountains on both sides of the valley (Fig. 1b). The unobstructed fetch at this site was approximately 100–200 m, depending on wind direction.

b. Data collection

The instrumentation consisted of a Campbell Scientific, Inc., model CSAT3 sonic anemometer, which measures the sonic temperature $T_s$ and the longitudinal $u$, lateral $v$, and vertical $w$ velocity components, and a collocated Campbell Scientific KH20 krypton hygrometer, which measures the water vapor density $\rho_v$. These were both mounted at a height of 1.80 m above the ground surface. Radiation fluxes were measured with a Kipp and Zonen, Inc., CNR1 four-component radiometer that independently measures the incoming and outgoing shortwave and longwave radiation fluxes. Reference air temperature and humidity were measured using a Vaisala, Inc., HMP45 mounted in a non-aspirated 12-plate radiation shield. To measure the wind speed profile, four Met One, Inc., 50.5 two-dimensional sonic anemometers were mounted at heights of 0.35, 0.70, 1.70, and 3.40 m above the ground. The slow-response instrumentation was sampled every 5 s and was processed and recorded at 30-min intervals by a Campbell Scientific CR23x datalogger. A Campbell Scientific CR5000 datalogger was used to interrogate the eddy covariance instrumentation (sonic anemometer and krypton hygrometer) at a frequency of 20 Hz. Additional wind speed and direction data were also obtained from a station (50°56.6’N, 115°11.4W, elevation 2543 m), operated by Environment Canada, that is located on the southern flank of Mount Allan (directly west of the Hay Meadow).

This study presents meteorological and flux data collected between 9 February and 17 March 2006. The 2005/06 winter season was quite dry and the selected period contains a range of partial to complete snow cover. The entire record is used to characterize the winds, whereas the detailed turbulence characterization is based upon a particular period of data collected on 26 February, where the ground was completely covered with a uniform 0.07-m snow cover with the exception of a few exposed stems of grass.

c. Data processing

The raw data from the sonic anemometer and krypton hygrometer were despiked following a similar routine to that described in Vickers and Mahrt (1997). The second-order moments and eddy covariance products were then...
calculated using a block averaging method for discrete 30-min periods. The fluxes were rotated into the Cartesian coordinate system aligned with the mean wind using the planar-fit method (Wilczak et al. 2001). The sensible heat fluxes were corrected for the effect of humidity on the speed of sound measured by the sonic anemometer. The latent heat fluxes were corrected for the effect of O2 (van Dijk et al. 2003) and air density effects (Webb et al. 1980). Both sensible and latent heat fluxes were corrected for frequency response limitations due to pathlength averaging, block averaging, and, in the case of the krypton hygrometer, lateral separation, using the analytical method of Massman (2000, 2001). In all cases, the corrections were minor and had minimal effect upon the measured fluxes. Data were considered to be poor quality when the signal of the krypton hygrometer became degraded by frost forming on the lens. Instances of spikes within the data were usually associated with periods of precipitation or blowing snow.

d. Data analysis

In addition to providing a general description of the turbulent boundary at the point of measurement within the meadow, the analysis of the data addressed two specific purposes: 1) to test for the presence of an equilibrium boundary layer, and 2) to examine flux generation processes. With respect to the former, an equilibrium layer occurs when production of turbulence is approximately balanced by dissipation, which occurs when the advection and pressure transport terms in the budget of turbulent kinetic energy are negligible (Raupach et al. 1991). Direct measurements of the rate of turbulence dissipation were not made, so the following indirect indications of an equilibrium boundary layer were used: 1) evidence of logarithmic wind profile, 2) the value of the nondimensional wind speed gradient during neutral conditions, and 3) the behavior of the normalized standard deviation of the vertical wind velocity component.

After the wind encounters a change in surface roughness, in this case the transition from the forest to the clearing, an increase in the wind speed is expected near the surface in response to the smoother surface. The reestablishment of a logarithmic wind profile, observed during neutral conditions, can indicate that the flow has adjusted to the new surface (Garratt 1990). In this experiment, the wind speed profile from the four two-dimensional sonic anemometers was inspected for a log-linear profile. For neutral cases, a least squares approach was used to fit the measured wind speeds to the logarithmic height above the surface z using the log-linear equation

$$\ln(z) = \frac{\kappa}{u_*} w + \ln(z_{0m})$$

(1)

to obtain estimates of the friction velocity $u_*$ and the momentum roughness length $z_{0m}$. The value $\kappa$ is the von Kármán constant, which was taken as 0.4. For the case of steady, horizontally homogeneous wind flow, the friction velocity and the roughness lengths are expected to be very similar to the same properties derived from the momentum flux measured by the sonic anemometer:

$$z_{0m} = z \left( \exp \left( \frac{\kappa u}{u_*} \right) \right)^{-1}$$

(2)

$$u_* = \sqrt{\left( \frac{u'w'}{w} \right)^2 + \left( \frac{w'w'}{w} \right)^2}.$$  

(3)

More formally, if the momentum flux is related to the wind speed gradient as predicted by M–O similarity theory, the dimensionless wind shear $\phi_{u*}$ will equal approximately 1 under neutral conditions:

$$\phi_{u*} = \frac{\kappa z}{u_*} \left( \frac{\partial u}{\partial z} \right).$$

(4)

The metric of atmospheric stability used here is $\zeta$, which is formed by dividing the height above the surface by the Obukhov length $L$:

$$\zeta = \frac{z}{L} = -\frac{z \kappa g (w'\theta')}{\bar{w} u_*^3},$$

(5)

where $\theta$ is the potential temperature. For the conditions encountered in this study, the potential temperature is practically identical to the measured air temperature, so for simplicity the kinematic heat flux $\bar{w} \theta'$ is hereinafter referred to as $\bar{w} T'$.

A test commonly used to verify M–O similarity theory and the presence of an equilibrium boundary layer involves the standard deviations of the wind velocity components normalized by the friction velocity (i.e., $\sigma_u u_*^{-1}$, $\sigma_v u_*^{-1}$, and $\sigma_w u_*^{-1}$), which are sometimes referred to as integral turbulence characteristics (Foken and Wichura 1996) or as the turbulent energy-flux ratios (King 1990). In pure mechanical turbulence, the standard deviations of the velocity components are expected to be linear functions of $u_*$ (Panofsky and Dutton 1984). Thus, in neutral conditions, the normalized standard deviations often assume constant values. In stratified flow, M–O similarity theory states that the normalized standard deviations should be...
functions of $\zeta$ when measured in the surface layer. However, owing to nonlocal turbulence dynamics there are documented cases of larger normalized standard deviations that do not necessarily scale with $z$, and do not necessarily converge at a constant value when $z = 0$ (e.g., McNaughton and Brunet 2002). This has been observed primarily for the horizontal variances in complex terrain (e.g., Panofsky and Dutton 1984). Large eddies can have extensive horizontal length scales which can create the effect of a “turbulence memory,” where adjustment to the local surface is very slow (Beljaars 1987). On the other hand, the motions contributing to the vertical variance must have vertical length scales equal or less than the measurement height, so the vertical variance usually reaches a more rapid equilibrium with the local surface. Accordingly, the vertical standard deviation rarely deviates from the equilibrium conditions, even in more complex terrain. Examples from various studies are presented in Table 1. For the cases in which $s_{wu}^*$ values are significantly larger than 1.25, nonequilibrium boundary layers were observed because of complex flow considerations.

The data were further analyzed to better understand the flux generation mechanisms and to identify diverging processes from those observed in more homogeneous terrain. In support of this endeavor, the spectra and cospectra were calculated. The data from each 30-min period of interest were truncated to form $N = 2^{15} = 32,768$ point blocks. Spectra and cospectra were then calculated using the discrete Fourier transform. To prepare the spectral density estimates for graphical presentation, they were frequency weighted by multiplying the spectral density by the frequency $f$ and were then smoothed by averaging to form seven logarithmically spaced intervals per decade. Finally, the frequency scale was nondimensionalized by normalizing the product of the frequency and the measurement height by the mean wind speed; for example, $n = f z / \bar{u}$. Similar procedures were used to find the cospectral densities. The calculated spectra and cospectra for selected events were averaged to form composite curves, and were compared to the spectra and cospectra collected by Kaimal et al. (1972), using the parameterizations presented in Kaimal and Finnigan (1994).

To identify the contribution to the fluxes from intermittent events, the instantaneous kinematic fluxes of momentum $u'w'$, sensible heat $w'T'$, and water vapor $w'\rho'_w$ were subjected to quadrant analysis (Willmarth and Lu 1974). This is a conditional averaging technique that divides the instantaneous flux product into four quadrants depending on the sign of the flux components. The four quadrants are defined as

- **quadrant I**: $\alpha' > 0$, $w' > 0$,
- **quadrant II**: $\alpha' < 0$, $w' > 0$,
- **quadrant III**: $\alpha' < 0$, $w' < 0$, and
- **quadrant IV**: $\alpha' > 0$, $w' < 0$,

where $\alpha'$ represents the fluctuation of $u$, $T$, or $\rho_w$ from their mean values. Fluxes belonging to quadrants II and IV are referred to as ejections and sweeps, respectively, while quadrants I and III are respectively termed outward and inward reflections. The conditioning function is defined as

$$I_{k,H} = \begin{cases} 1, & \text{if } \alpha' w' \text{ lies within the } k \text{th quadrant and } |\alpha' w'| \geq H|\alpha' w'| \\ 0, & \text{otherwise,} \end{cases}$$

(6)
where the hyperbolic hole $H$ represents the fraction of the absolute instantaneous flux to the time-averaged flux for the 30-min period (as represented by the over-bar):

$$H = \frac{\left| \alpha' w' \right|}{\bar{\alpha} w'}.$$  (7)

The conditionally averaged flux (as represented by angled brackets) is made up of those instantaneous values that belong to quadrant $k$ and are larger than the hole size $H$:

$$\langle \alpha' w' \rangle_{k,H} = \lim_{\tau \to \infty} \frac{1}{\tau} \int_0^\tau \alpha' w'(t) I_{k,H} \, dt.$$  (8)

The total contribution from the samples belonging to a particular quadrant outside of a particular hole size is represented by the stress fraction:

$$F_{k,H} = \frac{\langle \alpha' w' \rangle_{k,H}}{\bar{\alpha} w'}.$$  (9)

In a similar manner, the fractional time occupied by the sum of all of the instantaneous fluxes belonging to a particular quadrant and larger than the specified hole size can be found as

$$T_{k,H} = \lim_{\tau \to \infty} \frac{1}{\tau} \int_0^\tau I_{k,H} \, dt.$$  (10)

Quadrant analysis has been most frequently used to examine the Reynolds stress within a plant canopy (e.g., Shaw et al. 1983; Baldocchi and Meyers 1988; Grant et al. 1986; Bergström and Högström 1989); however, Hayashi (1992) and Horiguchi et al. (2010) have used the technique to examine the effect of gusts.

3. Results

a. Wind conditions

The winds at the measurement site were typically calm and unsteady. During the period 9 February through 17 March 2006, 53% of the recorded 30-min mean wind speeds were less than 1 m s$^{-1}$, 30% were between 1 and 2.5 m s$^{-1}$, and the remaining 17% were between 2.5 and 5.0 m s$^{-1}$ (as measured by the sonic anemometer at 1.80 m). Despite the calm nature of the location, wind gusts were a characteristic feature. To support this, the turbulence intensity $T_{\alpha} = \sigma_{\alpha} \bar{\alpha}^{-1}$ that was bin averaged at evenly spaced wind speed intervals is shown in Fig. 2. At low wind speeds, the larger intensity values are an artifact of the exaggerated influence of single eddies upon the standard deviation; however, even at higher wind speeds the turbulence intensity remains around 35%. The nature of the unsteadiness of the wind, and how it affects the surface fluxes, forms the impetus for this paper.

b. Turbulence characteristics observed during 0900–1700 26 February 2006

The following results were collected during a single event in which there was complete snow cover and the meteorological conditions were comparatively stationary. This day was characterized by strong southwesterly flow conditions aloft in which the valley winds were effectively “decoupled” from the overlying flow, as suggested by Fig. 3, which compares the wind speed and direction measured in the valley bottom with those measured on an upwind ridgetop location (located approximately 1250 m above and 4.5 km west of the study site). During the early morning period, the valley surface winds were light and from the northwest and west, indicating katabatic flow from the adjacent mountain slopes. Between the hours of 0830 and 1930 local time the winds became stronger, slowly varying from south-southeast (150$^\circ$) to west-southwest (230$^\circ$). The winds received directly from the south were blowing down the main Kananaskis River valley, whereas the winds received from the southwest were likely due to the influence of the flow funneling through an adjacent tributary valley (which is visible in Fig. 1b). The air temperature in the valley bottom was approximately $-16^\circ$C during the early morning hours, and then began warming at 0330, rising to a mid afternoon maximum of 4$^\circ$C and then began to cool again at around 1930, later falling to $-13^\circ$C. Air temperatures observed at the adjacent ridgetop station rose steadily throughout the day from $-11^\circ$C to a value of $-4^\circ$C. During the selected portion of the
day (0900–1700) the sensible heat fluxes directed toward
the snowpack averaged 16 W m$^{-2}$, while the latent heat
fluxes averaged 24 W m$^{-2}$ (sublimating). The bulk
temperature difference between the reference height
(1.80 m) and the surface averaged around 4.8 C. However,
because of the stronger winds during the selected period,
near-neutral conditions were measured by the sonic ane-
mometer (0.003, z, 0.025). It was not possible to ob-
serve the strength of the inversion at higher levels within
the valley during that period.

The surrounding complex terrain was observed to
erect an influence upon many of the turbulence char-
acteristics measured by the sonic anemometer (Table 2).
As previously noted, the gustiness of the wind had a
predictable effect upon the horizontal velocity compo-
nents, resulting in a turbulence intensity ($\sigma_u / \bar{u}$) greater
than 30%. This resulted in a positively skewed distribution
of the longitudinal velocity (i.e., $u^3 \sigma_u^3 > 0$). Of greater
significance is that the standard deviation of the vertical
velocity was similarly affected by the wind gusts. The vertical
velocity distribution exhibited a large mean

\begin{table}
\centering
\begin{tabular}{lcccc}
\hline
 & $\sigma_u U^{-1}$ & $\sigma_u u^0$ & $\alpha u^3 \sigma_u^3$ & $\alpha u^4 \sigma_u^4$ \\
\hline
$\alpha = u$ & 0.32 (0.03) & 3.78 (0.24) & 0.44 (0.14) & 3.44 (0.32) \\
$\alpha = v$ & 0.39 (0.06) & 4.55 (0.54) & -0.06 (0.34) & 3.56 (0.47) \\
$\alpha = w$ & 0.13 (0.01) & 1.47 (0.15) & -0.20 (0.08) & 4.52 (0.25) \\
\hline
\end{tabular}
\caption{Turbulence characteristics of the selected event. Note:
values in parentheses are \pm 1 standard deviation.}
\end{table}

kurtosis value ($w^4 \sigma_w^{-4} > 3$) suggesting a strong influence
of outliers, and was negatively skewed ($w^3 \sigma_w^{-3} < 0$)
demonstrating a prevalence of downward motions. The negative
skewness also demonstrates the effect of blocking by the
surface, which physically constrains the size of the source
area for upward bursts of air.

c. Adjustment of the wind speed profile

For the wind directions observed during this event, the
unobstructed fetch was approximately 150–200 m. Be-
\begin{figure}
\centering
\includegraphics[width=\textwidth]{fig3.png}
\caption{Time series of (a) wind speed and (b) wind direction for
26 Feb 2006 as measured at the valley bottom (solid line) and on an
adjacent ridge (dashed line).}
\end{figure}

\newpage

beyond the open meadow, the terrain immediately upwind
was a mixture of shrubs (3–4 m tall), pine forest (10–12 m
tall), and open areas (right of way for electrical trans-
mission line). Throughout the selected period, the wind
speed profiles exhibited a well-defined logarithmic shape
(samples shown in Fig. 4). The reestablishment of the log-
linear wind profile downstream of the change in surface
roughness provides a positive indication that the internal
boundary layer had developed to a height of at least
3.3 m. However, upon comparison of the friction velocity
and momentum roughness length estimated from the
wind profiles and those same properties measured by the
sonic anemometer some remarkable differences were
observed. The mean friction velocity measured by the
sonic anemometer (0.34 m s$^{-1}$) was 31% larger than the
profile-derived value (0.26 m s$^{-1}$), while the momentum
roughness length $z_{0m}$ was an order of magnitude larger
(16.8 versus 2.2 mm, respectively). This suggests that the
actual momentum transfer to the surface was much larger
than was indicated by the mean wind speed profiles. Thus,
in terms of local scaling, the mean value of the dimen-
sionless shear $\phi_m$ was 0.76 rather than 1.0 as expected for
the neutral conditions encountered. Although the loga-
rithmic wind profile might suggest that the flow has ad-
justed to the local surface, the additional Reynolds stress
measured by the sonic anemometer is likely due to tur-
bulence advected from regions outside the internal
boundary layer, the presence of which negates the
possibility of this boundary layer being in equilibrium
(Kaimal and Finnigan 1994). This observation is consis-
tent with considerations of rough-to-smooth transitions in
the literature (e.g., Peterson 1969; Antonia and Luxton
1972; Rao et al. 1974; Chamorro and Porté-Agel 2009).
Downstream from a surface discontinuity, Antonia and
Luxton (1972) found that advection of turbulence was
comparable in magnitude to local turbulence production. For situations where the local production of turbulence is not closely matched with the dissipation, parameterization of eddy diffusivities based upon a mixing length will result in substantial errors (Peterson 1972).

d. Turbulence characteristics

The measured horizontal integral turbulence characteristics $\sigma_u u^*_{1}$ and $\sigma_w u^*_{1}$ (Table 2) were significantly larger than the typical flat terrain values summarized in Table 1. This was an anticipated result considering the gusty nature of the site and the complex surrounding topography. Of greater interest, is that the ratio for the vertical velocity component, $\sigma_w u^*_{1}$, was about 18% larger than the typical value of 1.25 for neutral conditions. The interpretation of the integral turbulence characteristic is aided by considering the linear correlation coefficient $r_{uw}$, which is related as

$$r_{uw} = \frac{\overline{u^* w^*}}{\sigma_u \sigma_w} = -\frac{u^*}{\sigma_u} \frac{w^*}{\sigma_w},$$

for which the mean value observed in the present study was $r_{uw} = -0.18$. The value of $r_{uw}$ can be as high as $-0.5$ for homogeneous shear flow in a wind tunnel boundary layer; however, for atmospheric flows $r_{uw}$ is commonly reported as $-0.3$ (Kaimal and Finnigan 1994). The reduction observed in the atmosphere is reportedly due to inactive motions that contribute to horizontal motions but not vertical motions (Högström 1990). The magnitude of $\sigma_w$ is limited because of the blocking effect at the surface, such that unsteadiness of the wind due to inactive motions will primarily influence the horizontal velocity components, and act to decrease the correlation coefficient.

With respect to the turbulence characteristics, the main departure between the present study and similar studies reported in the literature is that the fluctuations of the vertical velocity were larger than anticipated, which was presumably related to the dominating occurrence of wind gusts. The “microfront” of a wind gust can be associated with large longitudinal variations that have high correlation with the vertical velocity fluctuations, representing concentrated periods of strong momentum transfer (Mahrt and Gibson 1992).

e. Spectra and cospectra

The composite spectra of the velocity components are presented in Fig. 5, where they have been normalized by the kinematic momentum flux. The Kansas curves (Kaimal et al. 1972) are shown for reference purposes and are considered representative of the well-understood flat boundary layer. The spectra of all components contained significant energy at low frequencies, and when normalized by $u^*$, the effect of inactive motions can be seen quite clearly by the markedly higher magnitude (power) of the spectra relative to the Kansas examples. This is presumably due to the low degree of correlation observed between the horizontal and vertical velocity variances. As a consequence of the nonequilibrium conditions, the vertical velocity, which exhibits a spectral peak at around $n = 0.2$ (or $f = 0.44$ Hz), deviates from the expected curve much more than anticipated.

To gain an appreciation for how the momentum, heat, and vapor fluxes were affected by the perturbed airflow, the cospectra were similarly compared to the Kansas parameterizations (Fig. 6). Additional energy at lower frequencies is noted for all of the cospectra, but particularly for the $uw$ cospectrum. Thus, the wind gusts, which occupy the lower frequencies, are seen to have a marked effect upon the flux. The cospectral shapes of the sensible and latent heat fluxes are quite similar for the peak frequencies; however, the sensible heat displays a secondary increase in energy at lower frequencies. This could be related to warmer air being advected from aloft in association with the wind gusts. The presence of additional energy at lower frequencies is consistent with other observations in complex terrain. For example, the spectra and cospectra collected by Andreas (1987a) in a locally homogeneous field show a similar effect that was attributed to the effect of perturbations to the boundary layer that were introduced by surrounding hills. Above-canopy measurements made by Turnipseed et al. (2004) at Niwot Ridge, Colorado, also displayed enhanced low-frequency spectral energy due to mesoscale motions associated with mountain topography. The spectra and cospectra collected by Smeets et al. (1998, 2000) on the Pasterze Glacier in Austria also exhibited

![Fig. 4. Examples of logarithmic wind profiles measured on 26 Feb 2006.](image-url)
topographically generated low-frequency perturbations that affected the wind velocity variances and increased the fluxes.

To determine the effect of the low-frequency energy upon the calculated flux, the cumulative flux integral (Figs. 6d–f) was calculated as in Oncley et al. (1996):

\[
O_{g_{xy}}(f_0) = \int_{f_0}^{\infty} C_{xy}(f) \, df,
\]

where the cospectrum of two variables \( x \) and \( y \) was integrated between the highest measured frequency and 1 h. It is shown that the fluxes are reasonably well captured by the chosen 30-min averaging period, as evidenced by the leveling off of the individual curves. The latent heat flux reaches a near-steady value at slightly earlier times than the momentum and heat fluxes, which exhibit sporadic changes at longer averaging lengths. From examination of individual cospectra (not shown) it was apparent that the wind gusts could not be clearly identified as mesoscale motions, as they tended to overlap with the energy expected at turbulent time scales, thereby obscuring an obvious gap scale (Tyson 1968; Smeets et al. 1998, 2000; Cava et al. 2001; Turnipseed et al. 2004).

**f. Quadrant analysis**

To quantify the impact of the low-frequency enhancement of the fluxes, the instantaneous fluxes of \( u'w' \), \( T'w' \), and \( \rho'w' \) were subject to quadrant analysis (Fig. 7). By definition, for the case where the \( H = 0 \), the sum of the flux fraction from all quadrants must equal 1.0. Here it is noted that the measured flux is the relatively small sum of large gradient and countergradient motions, particularly for momentum and sensible heat. The contribution from countergradient quadrants (I and III), expressed as a percentage of the progradient (quadrants II and IV) contributions were \( u'w' = 60\% \), \( T'w' = 62\% \), and \( \rho'w' = 47\% \). With respect to the progradient motions, all of the flux quantities show near equal contribution from quadrants II and IV become more significant, thus indicating the importance of intermittent events. Quadrant analysis has most frequently been applied to study in-canopy turbulence, thus direct comparisons with literature values are not easily made. However, two particular observations that stand out from this study are 1) the large flux ratios and the significance of the countergradient fluxes (these are often observed to be less than 1; e.g., Hayashi 1992; Bergström and Högström 1989), and 2) the contribution toward the fluxes from samples beyond large values of \( H \)—for example, instantaneous fluxes in quadrant IV (sweeps) that are 30–40 times the average flux.

The cumulative frequency distribution of the flux fractions and the time fractions for increasing values of \( H \) are presented in Fig. 8. This graph shows the portion of the time-averaged fluxes that are contributed from

![Normalized spectra for (a) longitudinal, (b) latitudinal, and (c) vertical wind velocity components. Solid lines with circle points are measured data, composited from 30-min periods between 0900 and 1700 26 Feb 2006. Dashed lines are modeled curves representing flat terrain in Kansas (Kaimal et al. 1972).](image)
instantaneous fluxes larger than a particular value of \( H \), and also the percentage of time that those samples occupy. For example, 50% of the sensible heat flux is contributed from instantaneous fluxes that are greater than 15 times the mean value, yet only occupy 6% of the total time. The portion of the total flux that is contributed from large values of \( H \) is most significant for the momentum and sensible heat fluxes. However, the distributions of all of the flux quantities occurred at remarkably similar time distributions: 75% of the flux was generated by motions that occupied only 16%–17% of the time, 50% of the flux generated by motions that occupied 6%–6.5% of the time, and 25% of the flux was contributed by motions that occupied approximately 2% of the time. This demonstrates the importance of the contribution of the intermittent flow processes.

4. Discussion

a. Conceptual model of the near-surface boundary layer

The boundary layer structure observed at this site, despite the misleadingly normal appearance suggested by the logarithmic wind profile (Fig. 3), was strongly influenced by advection of turbulence from outside the clearing. The discord between the measured Reynolds stress and the wind speed gradient is consistent with observations reported in the literature of nonequilibrium boundary
layers occurring after rough–smooth surface discontinuities (e.g., Antonia and Luxton 1972). However, it is not believed that the forest edge is the only perturbation at this site: the boundary layer was additionally modified by wind gusts, which provided an important contribution to the fluxes. The boundary layer was only sampled near the surface, so the origin of the unsteadiness in the wind cannot be identified from measurements alone. However, based upon the near-surface turbulence characteristics and personal observations, it is hypothesized that perturbations arise from the complex interactions between the different flow mechanisms within the valley. Visual observations made by the authors from high-elevation vantage points throughout the surrounding basin noted that, during wind gusts, a visible pattern of intercepted snow was unloaded from the canopy of conifers. Within forest openings, the same process results in bursts of blowing snow. It is envisioned that this phenomenon is similar to the large-scale coherent motions or “honami” that have been observed in wheat fields (Finnigan 1979) or over water surfaces (e.g., the cats paws discussed by Hunt and Morrison 2000). The behavior is noted throughout the basin, suggesting that the effect is not local to the clearing at the valley bottom. One possible origin of the gusts is that they are vortices that are shed as the overlying flow separates over the upwind mountain ridge. Measurements and modeling by Revell et al. (1996) within a valley in the Southern Alps of New Zealand identify large eddies that were shed at the upwind ridge and were felt at the valley floor (located approximately 2000 m below the ridge). Measurements by Smeets et al. (1998, 2000) on the Pasterze Glacier in Austria reveal similar large-scale (low frequency) eddies that interact with motions on turbulent scales and increase the momentum and sensible heat transfer. These authors similarly hypothesized that the boundary layer perturbations were a result of large eddies shed by adjacent mountain ridges.

Another related casual observation was that the clouds above the valley were often seen to be rapidly moving in a direction perpendicular to the valley, while the conditions within the valley were very calm and interrupted only by the occasional gust. This suggests that a strong region of shear likely exists between the valley core and the overlying flow (as highlighted by Fig. 3), which could possibly be a source of unsteadiness in the wind. Another potential area of wind shear exists between the thermodynamically driven slope winds and the overlying flow within the valley core. Whiteman (1982) observed that, during temperature inversion breakups in snow-covered mountain valleys, the winds in the neutral core were often up-valley whereas the surface winds typically blew down the valley. Furthermore, he observed that inversion breakup was often caused by subsidence of warmer overlying air. This mechanism provides an important source of heat that can be further transported by advection within turbulent motions into the near-surface boundary layer. 

Regardless of the exact source, it is conceptualized that large-scale motions, which originate outside of the internal boundary layer within the clearing, are being superimposed upon the mean flow in the surface layer, transporting additional energy that must be locally dissipated as well as transferring scalar properties from aloft. The intermittent nature of these events plays a critical role in the generation of heat and mass fluxes at this site. It is recommended that future research efforts in this region examine the wind and thermodynamic structure of the valley to identify any shear zones or temperature irregularities. This will be critically important for refinement of operational snow physics modeling in this area.
b. Practical implications

The advection of additional turbulent energy into the surface layer from aloft negates the possibility of this site being considered an equilibrium boundary layer. The observed discrepancy between the measured momentum flux and that derived from the wind profiles is evidence of the breakdown of the mixing length parameterization for the eddy diffusivity. Although the measurement site was locally homogeneous, the resulting turbulence structure was quite dissimilar to the homogeneous terrain from which flux gradient and M–O similarity theory were intended. Thus, with respect to bulk transfer models, the typical practice of scaling the momentum transfer model using \( z_{\text{ref}} \) derived from wind profiles or textbook values representing a smooth snow surface will substantially underpredict the momentum flux. Furthermore, invoking Reynolds analogy or similar approaches (e.g., Brutsaert 1975; Andreas 1987b) to scale the sensible and latent heat flux upon the momentum flux will be similarly misguided. Application of the bulk transfer approach for estimating the fluxes of sensible and latent heat in this environment has been investigated by Helgason and Pomeroy (2005) and Helgason (2009). It was found that, during neutral conditions, the fluxes were still proportional to the local gradients of the mean constituent; however, the bulk transfer coefficient for momentum was 6–7 times those of heat and water vapor. In effect, the advection of turbulence due to upstream topographical complexities results in bluff-roughness effects (Hignett 1994), where the scaling parameters depend on factors external to the local surface. The practical problem then becomes the determination of appropriate “effective” roughness lengths for use in models.

The complex surrounding topography also presents a challenge for measurement of surface fluxes using the eddy covariance technique. Since the wind gusts are the dominant flux-enhancing feature, it is important to capture enough individual gusts in the averaging period so their effect can be adequately represented in the calculated fluxes. However, the presence of these large-scale motions also increases the number of nonstationary records (largely because of increased random error), particularly for the momentum flux. The contribution of low-frequency motions to the calculated fluxes is captured by the flux ogives (Fig. 6b). The trade-off between stationarity and complete accounting of the flux must be carefully considered, particularly for lower wind speeds.

The advection of energy and the associated enhancement of the momentum flux tend to ensure that the surface layer remains well mixed and neutral. This leaves some uncertainty for modeling applications where a measure of stability is required. The M–O parameter \( \zeta \) likely reflects the situation most accurately since it tends to neutrality because of the appearance of \( u_{\text{*}}^{2} \) in the denominator. However, the use of the bulk Richardson number or gradient Richardson number will overestimate the true stability since these approaches assume that the only origin of mechanical turbulence is due to wind shear, as represented by the near-surface wind speed gradient, and thus does not account for mixing due to advected turbulence. The practitioner should be particularly careful when applying empirical relationships derived using estimates of stability in similar environments.

5. Conclusions

Valley bottoms are the normal location for meteorological measurements in mountain regions in Canada and many other countries; however, the winter micrometeorology of these sites shows some very complex behavior, which makes interpretation of near-surface atmospheric observations both uncertain and problematic. Observations in a 7-ha homogeneous meadow in the Kananaskis River valley within the Canadian Rocky Mountains reveal that the winter wind regime is characterized by extensive calm periods interspersed with intermittent wind gusts. Investigation of the near-surface boundary layer within the clearing showed that wind gusts transport additional turbulent energy into the clearing. As a result, the boundary layer could not be considered to be in equilibrium, as reinforced by striking differences between the momentum flux measured by the sonic anemometer and that estimated from the logarithmic wind speed profile during neutral conditions. Thus the momentum flux was not representative of the local aerodynamic roughness but rather reflected perturbations introduced by surrounding topography and land surface discontinuities. The intermittent nature of the gusts were evident in spectra and cospectra where the peak energy was shifted to lower frequencies as compared to those observed in homogeneous flat terrain. The contribution of intermittent events was studied using quadrant analysis, which revealed that 50% of the flux was contributed from motions that occupied less than 6% of the time.

There is a critical requirement for accurate prediction of the convective heat fluxes available for snowmelt in mountain regions. Most available models that are being applied in these environments utilize a bulk transfer approach for estimating the turbulent fluxes of sensible and latent heat (e.g., Etchevers et al. 2004; Rutter et al. 2009). Relative to typical terrain found in the region, the measurement site was considered to be an ideal environment for testing the application of flux-gradient techniques. The meadow was locally homogeneous, had a relatively
extensive fetch, and displayed an exemplary logarithmic wind profile. However, examination of the boundary layer characteristics revealed that the turbulence could not easily be related to the local gradients. Thus application of flux-gradient estimation techniques should be approached with considerable caution, in particular, the heat transfer coefficients (roughness lengths), which must be used take on an “effective” role as the turbulence is reflective of nonlocal scales. Therefore the selection of appropriate transfer coefficients cannot be based upon existing approaches, and will need to be empirically selected upon consideration of the topographical features and wind flow patterns of the area. At present, there remains a considerable gap between the first-order turbulence parameterizations available in the modeling tools and the more complex higher-order numerical approaches that would be required to capture the nonlocal scaling of the turbulence in these environments. Thus, continued study of the processes responsible for generating surface energy fluxes in complex terrain will help to inform the adjustment of the modeling tools for more accurate use in mountain basins.

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