A Numerical Study on the Diurnal Variation of Low-Level Wind in the Lee of a Two-Dimensional Mountain

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Abstract

A series of numerical experiments was made to search for the basic features of diurnal wind variation in the lee of a mountain range using a two-dimensional dry version of the MRI non-hydrostatic model. The features of diurnal wind variation were classified into four regimes according to prevailing wind speed with respect to the characteristic behavior of the lee convergence zone (LCZ). Under moderate prevailing wind, the LCZ propagates downwind far into the leeside plain during the daytime accompanied by a surge of strong wind. This is in qualitative agreement with the daytime advance of downslope wind observed in the Canterbury Plains in New Zealand, and the Kanto plain in Japan.

1. Introduction

Local wind systems induced by orography are divided into two categories: thermally driven mountain/valley breezes, and dynamically controlled airflow over mountains (*e.g.*, Atkinson, 1981). The former is typically observed under no or weak prevailing wind, while the latter dominates under strong prevailing wind. As a result they have been studied separately with little attention to their interaction. Most of the studies on the airflow over mountains have neglected diurnal thermal forcing.

However, there is some observational evidence that strong downslope winds in the lee of mountains are affected by the diurnal effect. Climatological studies have revealed that downslope windstorms tend to be more frequent in the nighttime than in the daytime (Brinkmann, 1974; Arakawa *et al.*, 1982; Akiyama, 1956; Atkinson, 1981), while sur-

face winds over leeside plains tend to intensify in the daytime as the nocturnal inversion layer disappears (Ninomiya *et al.*, 1985; Yamagishi and Katoh, 1996). Banta (1984, 1986) described the evolution of the boundary layer in the lee of a low mountain range, where a shallow upslope wind was driven as the nocturnal inversion layer was destroyed after sunrise, and then replaced by the downslope wind in the late morning. In recent years, some numerical studies incorporating diurnal surface heating have shown its importance on the behavior of low Froude number flow passing mountains (Reisner and Smolarkiewicz, 1994; Georgelin *et al.*, 1996; Sun and Chern, 1993).

The Southern Alps Experiment (SALPEX), which commenced in 1993, is designed to improve understanding of the role of the New Zealand Southern Alps, which extends for 400–500 km in the northeast-southwest direction (Fig. 1), on mesoscale atmospheric systems (Wratt *et al.*, 1996; see also <http://www.niwa.cri.nz/salpex>). SALPEX is a research project coordinated by New Zealand's National Institute of Water and Atmospheric Research

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Fig. 1. Topography of the Canterbury Plains, New Zealand. The area enclosed by a square in (a) is shown in (b). Hatching indicates the area of 300 m, 600 m and 1200 m above m.s.l. with increasing tones. Station names are Christchurch Airport (CHA; 37 m above m.s.l.) and Craigieburn Forest (CRF; 914 m above m.s.l.), and Sites 1–7 where anemometers were installed in the 1996 SALPEX field campaign. The north-west-south-east line shows the baseline (x -axis) of the cross section in Fig. 6.

(NIWA) with collaboration from many other organizations. One of the targets is understanding the leeside effects of the Southern Alps on the wind fields over the Canterbury Plains, which is characterized by complicated wind fields including northwesterly foehn and diurnally driven circulations (Ryan, 1987; McKendry *et al.*, 1986; McGowan and Sturman, 1996; Cherry, 1972). Ryan (1987) notes that strong diurnal variations in wind strength are observed over the Canterbury Plains when the large-scale flow onto New Zealand is from the northwest. This is especially the case in winter, when a nocturnal inversion containing a katabatic flow often forms below the overlying northwesterly. Ryan (1987) also notes a three-dimensional feature, in which some low-level air deviates around the top of the South Island, flows through Cook Strait (which separates the North and South Islands), and is drawn into the lee trough on the southeastern side of the Southern

Fig. 2. Diurnal variation in the frequency of northwesterly wind (280° – 340°) in each speed range for four summer months (Jan., Feb., Nov., Dec.) obtained from hourly surface data at Christchurch Airport (1961–1988) and Craigieburn Forest (1974–1978). Time is in NZST (= UTC + 12).

Alps as a northeasterly wind. So far three field campaigns have been made in SALPEX, including the biggest one in 1996 (14 Oct.–3 Nov.). Observation was mainly focused on the upwind side of the Southern Alps, so that numerical studies are important in understanding the features of wind fields over the Canterbury Plains.

The present paper focuses on the basic features of diurnal variation of wind in the lee of a mountain range, by using the two-dimensional version of the non-hydrostatic mesoscale model developed in the Forecast Research Department of the Meteorological Research Institute (MRI). Ridley (1991) has compared two- and three-dimensional dry hydrostatic model simulations for 10 ms^{-1} and 20 ms^{-1} northwesterly flows onto New Zealand's South Island. He concludes that "quantitative aspects of the results from the two-dimensional simulations remain valid in three dimensions," but there are quantitative differences due to three-dimensional aspects of New Zealand's orography. While a two-dimensional simulation may be less realistic than a three-dimensional one, we expect it to give us good qualitative guidance on some of the aspects influencing diurnal flow variation. A two dimensional simulation allows us to make a large number of experiments by changing physical parameters. We take this advantage by covering a wide range of prevailing wind speed in comparison to previous studies (Banta, 1986; Bossert and Cotton, 1994; Wolyn and McKee, 1994; Qi and Fu, 1993), so that we can see how the diurnal variation depends on the prevailing wind speed and other physical parameters.

2. Diurnal variation of leeside winds in the Canterbury Plains

A climatological analysis of the northwesterly wind was made by using the climatic database (CLIDB) at NIWA. Figure 2 shows the diurnal variation in the frequency of northwesterly wind (280° –

Fig. 3. Half-hourly time series of wind on 20–21 October 1996 obtained from the anemometers installed in the SALPEX field campaign.

340°) at the Christchurch Airport (CHA) on the plains, and at Craigieburn Forest (CRF) in the mountains for the summer months (Jan., Feb., Nov. and Dec.) in 1960–1988 and 1974–1978, respectively. At CHA, strong northwest winds are more frequent in the daytime than at night. At CRF, there is less diurnal variation.

The diurnal behavior of the Canterbury northwest wind was investigated in detail using a network of seven anemometers installed in the 1996 field campaign. Figure 3 shows half-hourly time series of wind on 20–21 October, on which moderate northwesterly wind developed over the South Island in a pre-cold-frontal situation. At Site 1, which was on the eastern edge of the mountain area, the northwesterly wind blew almost continuously for two days. In the Canterbury Plains (Sites 2–6), the northwesterly wind ceased or weakened in the nighttime, and intensified again in the morning of 21 October. The onset of the northwest wind tended to be later at a site farther downwind: around 11 NZST at Site 2 and 13 NZST at Site 6. This agrees with the description of Lamb (1974), who gave a brief report on the “advance” of northwest wind over the Canterbury Plains.

A similar diurnal feature has been found in some other regions of the world. In the case studied by Banta (1984, 1986), the dominance of downslope wind in a shallow valley was preceded by a downwind advance of a “lee convergence line” which formed in a few hours after sunrise, although the horizontal scale of his analysis was much smaller (~ 10 km) than the Canterbury case. A daytime surge of northwesterly wind is observed also in the Kanto plain, which is to the east of the Central Mountains of Japan, as shown in Fig. 4. In this case, a surge line can be traced from hourly record of AMeDAS (Automated Meteorological Data Acquisition System) as a line of rapid increase in wind speed and temperature. Figure 5 shows the time-height section of wind and potential temperature at the Meteorological Tower of the MRI (213 m high).

The passage of the surge line can be recognized by an increase of wind speed and a temperature rise between 1230 and 1300 JST ($\longleftrightarrow$ in Fig. 5). The statistical study of Fujibe (1992) shows that the northwesterly surge in the Kanto plain is most frequently observed in the late morning subsequently to the disappearance of the nocturnal inversion layer, suggesting the importance of diurnal surface heating on the surge.

In view of these observational background, our main interest in the numerical experiment is whether, and how, the model can reproduce daytime wind intensification and downstream advance over the leeside plain.

3. Model and experimental design

3.1 Outline of the control experiment

The numerical model used in this study is a two-dimensional, fully compressible dry version of the MRI mesoscale non-hydrostatic model. The primitive version of the model was firstly developed at the Forecast Research Department of MRI, and a comprehensive description has been given in the Technical Report of MRI (Ikawa and Saito, 1991). A series of studies on “Yamaji-kaze,” a downslope windstorm in a region of western Japan, has been made using an anelastic version of the model (Saito and Ikawa, 1991; Saito, 1993, 1994). A fully compressible version was developed by Saito (1997), where the buoyancy term is directly evaluated from the density perturbation. Semi-implicit time integration scheme, which treats sound waves implicitly not only in the vertical direction, but also in the horizontal direction, is used.

In the lower boundary, a surface boundary layer is assumed, where the resistance law gives surface heat and momentum fluxes. This is based on Monin and Obukhov’s similarity law, with Sommeria’s (1976) formula. The ground temperature is predicted by solving the equations of surface heat balance and heat conduction at four levels in the ground (Kato,

Fig. 4. Surface wind and temperature fields at 12 JST 20 March 1987 (JST = UTC + 9). The area enclosed by a square in (a) is shown in (b). Dotted lines show hourly positions of the northwesterly mesofront defined from hourly time series at each station.

1996). To determine the diffusion coefficients, the turbulent closure model is employed. The formulation is based on Deardorff (1980) and Saito (1993), while it is slightly modified as described in 3.2. As the computational diffusion, a time filter and a fourth-order linear damper which suppresses mainly two-grid-scale noise are employed.

Table 1 shows the parameters used in the control experiment of the study. The horizontal and vertical coordinates in the model domain are denoted by x and z , respectively, and the coordinate perpendicular to the domain by y . The velocity components in the x , y and z directions are denoted by u , v and w , respectively. The horizontal grids are equally spaced with an interval of $\Delta x = 4$ km in the whole domain. The vertical grids are placed on the z^* system defined by

Fig. 5. Time-height section of wind and potential temperature (here defined by $T + \Gamma_d z$, where T is temperature and Γ_d is the dry adiabatic lapse rate) at the MRI tower.

$$z^* = \frac{z - z_S}{z_T - z_S} z_T, \quad (1)$$

where z_S is surface altitude and z_T is the model top height at 16.8 km. The vertical grid intervals are variable, with $\Delta z^* = 20$ m at the lowest level and $\Delta z^* = 800$ m at the top. As a staggered grid system is used, the lowest level for u and v is $z^* = 10$ m. The calculation of horizontal advection was made with fourth-order (Saito, 1998).

The prevailing wind was given by assigning a uniform and time-independent pressure gradient in the y -direction. Hereafter we denote the corresponding geostrophic wind speed by G . The value of G was changed between 0 ms^{-1} and 16 ms^{-1} according to runs.

The mountain shape in the control experiment was given by a function

$$z_S = H \cos^2 \left(\frac{\pi x}{2x_0} \right), \quad \text{if } |x| \leq x_0$$

$$z_S = 0, \quad \text{if } |x| \geq x_0 \quad (2)$$

where H is the mountain height and x_0 is a parameter of its width (40 km for the control run). Figure 6 shows this function together with a northwest-southeast section of the New Zealand South Island on the line shown in Fig. 1. The model topography has a shape close to that of the southeastern side of the Southern Alps. The coastline of the experiment was placed at $x = \pm 240$ km, which is much farther from the mountain in comparison to the real topography of the South Island. This was in order to avoid complexities arising from interaction of a sea breeze and orographically generated wind systems.

The initial wind field was given by uniform distribution of $u = G$ and $v = 0$. The initial stratification was given by a constant lapse rate of θ in order to avoid complicating the interpretation of the results, although a downslope windstorm often takes place in the presence of an inversion layer in the low

Table 1 Parameters in the control experiment.

Calculation domain	
Horizontal grids	182 grids with 4 km intervals
Vertical grids	42 grids with variable intervals (see text)
Ground layer	4 grids at 0.04 m, 0.15 m, 0.4 m and 0.6 m below the surface
Boundary condition	
Side	Open boundaries with radiation condition (Orlanski, 1976; Ikawa and Saito, 1991)
Top	Rigid boundaries with Rayleigh damping for the upper eight levels ($z^* > 10$ km)
Initial condition	
Stratification	$\theta = \Theta + \Gamma z$, where $\Theta = 15^\circ\text{C}$ and $\Gamma = 3.5^\circ\text{C/km}$
Surface temperature	Same as air temperature at the surface altitude
Local time	00 LT
Ground parameters	
Thermal diffusivity	$7 \times 10^{-7} \text{ m}^2\text{s}^{-1}$
Specific heat capacity	$1 \text{ MJ m}^{-3}\text{K}^{-1}$
Surface roughness	$z_0 = 0.04 \text{ m}$
Bowen ratio	$B = 0.5$ (for positive sensible heat), ∞ (for negative sensible heat)
Geographical parameters	
Latitude	40°N , f -plane
Topography	Uniform in the y -direction Cosine-shaped mountain ($H = 1200 \text{ m}$ and $x_0 = 40 \text{ km}$ in Eq. (2)) with sea beyond $x = \pm 240 \text{ km}$ from the peak
Solar angle	At the equinox
Atmospheric physics	
Moisture	Considered only for the calculation of surface longwave and sensible heat fluxes (see text)
Turbulent process	Turbulent closure based on Klemp and Wilhelmson (1978) and Saito (1993) with some modifications (see text)
Prevailing wind	Constant and uniform pressure gradient in the y -direction (dp/dy)

level. The value $d\theta/dz = 3.5^\circ\text{C/km}$ corresponds to a Brunt-Vaisala frequency of $N = 0.011 \text{ s}^{-1}$, so that the critical wind speed $u_c = NH$ is 13 ms^{-1} for $H = 1200 \text{ m}$. The roughness length $z_0 = 0.04 \text{ m}$ is the same as the value used by McKendry *et al.* (1988) in a numerical study of thermal circulation in the Canterbury Plains. The latitude was set to be 40°N in view of application to leese side plains in the middle latitude of both hemispheres. The season was assumed to be at the equinox, so that sunrise and sunset were at 06 LT and 18 LT, respectively.

The calculation was made for three days (= 72 hours) with 00 LT as the initial time. The flow pattern was found to be sufficiently periodic after the first 24 hours except for limited cases mentioned later (Section 5.4). In the following chapters, the flow patterns on the third day (48–72 hours from the initial time) are presented.

3.2 Modifications to the original model

The original model makes no consideration of atmospheric water vapor in evaluating the downward longwave flux ($L^\downarrow$) at surface. This results in un-

realistically strong radiational cooling at the land surface when it is used as a dry version. The model does not allow for surface evaporation, so that the sensible heat flux is too large. In order to avoid these problems, $L^\downarrow$ was evaluated by assuming a time independent profile of water vapor concentration, based on a linear distribution of relative humidity (h)

$$h = 0.6 \frac{z_h - z}{z_h}, \quad (\text{for } z < z_h)$$

$$h = 0, \quad (\text{for } z \geq z_h) \quad (3)$$

where z_h was set to be 10 km. The mixing ratio was calculated from Eq. (3) with the initial temperature profile, and was kept constant throughout the simulation. The surface heat flux was evaluated by assuming a Bowen ratio (B) of 0.5 in the unstable case and ∞ in the stable case, in which latent heat flux is generally small and can be either positive or negative.

Another change to the model was the upper limit of the turbulent mixing length l_∞ . In the two-

Fig. 6. Northwest-southeast cross section of the South Island obtained from a weighted average

$$\overline{z_S(x)} = \frac{\iint \exp[-\{(x'-x)/X\}^2 - \{(y'/Y)\}^2] z_S(x', y') dx' dy'}{\iint \exp[-\{(x'-x)/X\}^2 - \{(y'/Y)\}^2] dx' dy'}$$

where x and y are the coordinates parallel and perpendicular, respectively, to the line shown in Fig. 1, z_S is the altitude above m.s.l., $X = 1$ km and $Y = 5$ km. The thick solid line shows the two-dimensional topography assigned in the control experiment in the numerical study. The dashed and dotted lines show the topography assigned in the " $x_0 = 80$ km" and the "plateau" experiments, respectively.

dimensional version of the original model, l_∞ is given by

$$l_\infty = \sqrt{\Delta x \Delta z}, \quad (4)$$

which is frequently used in large-eddy simulations (Deardorff, 1974; Klemp and Wilhelmson, 1978). This is not suitable for the present study because Δx is as large as 4 km. Therefore Eq. (4) was replaced by

$$l_\infty = 100 \text{ m}, \quad (5)$$

in view of the suggestion of Gambo (1978).

In the simulation, the daily total net sensible heat flux is positive for the diurnal variation of the

ground temperature, while the model does not contain the atmospheric longwave radiative cooling process. This inconsistency produces secular warming of the boundary layer by a few degrees per day, as in many other numerical models for local circulations. In order to remove this warming, a weak Newtonian cooling was assigned in the form

$$\frac{d\theta}{dt} = [\text{physical terms}] - \frac{\theta - \theta_i}{t_N}, \quad (6)$$

where θ is potential temperature, θ_i is its initial value, and t_N (= 12 hours in the whole domain) is the e^{-1} -folding time of relaxation.

3.3 Sensitivity experiments

In addition to the control experiment described above, some test experiments were made on the conditions listed in Table 2.

The "no Q_H " and the "fixed T_S " experiments were made in a search for steady states in the absence of diurnal variation. The "no Q_H " experiment was made by removing surface heat flux, and the "fixed T_S " experiment was made by fixing the ground temperature at the initial value. The results of these experiments will be shown in Chapter 4, before that of the control experiment (Chapter 5), in order to see the features of the wind fields in the absence of diurnal temperature change.

Other experiments were made in order to examine the effects of physical parameters such as the mountain height and shape, surface heating intensity and surface roughness. The "const K_M " experiment was made by eliminating the stability dependence of eddy viscosity in order to find the effect of the diurnal variation of frictional force. The results of these experiments are shown later in Chapter 6.

4. Steady state solutions

4.1 The "no Q_H " experiment

Figure 7 shows the x - z cross sections of u , w and θ obtained from the "no Q_H " experiment for $G = 2$,

Table 2. List of sensitivity experiments.

Experiment	Difference from the control experiment
No Q_H	No surface heat flux
Fixed T_S	Ground surface temperature fixed at the initial value
$H = 600$ m	The mountain height $H = 600$ m
$x_0 = 80$ km	The mountain width $x_0 = 80$ km (dashed line in Fig. 6)
Plateau	A plateau 40 km long inserted on the upwind side of the peak (dotted line in Fig. 6)
$B = 0.1$	The Bowen ratio $B = 0.1$ instead of 0.5 in the unstable case
$z_0 = 0.2$ m	The roughness length $z_0 = 0.2$ m instead of 0.04 m
Const K_M	A stability-independent formulation of turbulent momentum transfer, where the diffusion coefficient was fixed at $10 \text{ m}^2 \text{ s}^{-1}$ and the surface momentum flux was calculated by using the formulation for neutral stability

Fig. 7. Distribution of u and w (arrows) and θ (isentropes with units in $^{\circ}\text{C}$) obtained from 48 hour runs of the “No Q_H ” experiment.

Fig. 8. An x - G diagram of surface wind (u and v at $z^* = 10$ m) and vertical velocity (w at $z^* = 370$ m) for 48 hour runs of the “No Q_H ” experiment. Contours are drawn for $w = 0, \pm 20 \text{ cms}^{-1}, \pm 40 \text{ cms}^{-1}, \pm 80 \text{ cms}^{-1}$ and $\pm 160 \text{ cms}^{-1}$ with dashed lines for negative values.

5, 10 and 16 ms^{-1} . Figure 8 shows the x - G diagram of surface wind (i.e., u and v at $z^* = 10$ m) and w at $z^* = 370$ m.

The flow pattern for $G = 2 \text{ ms}^{-1}$ is characterized by a shallow layer of downslope wind on the slope with a speed of $\sim 10 \text{ ms}^{-1}$ and a depth of a few hundred meters. The layer of the downslope wind separates from the ground near the mountain foot with a vertical velocity of $\sim 10 \text{ cms}^{-1}$, and an area of stagnation spreads for a few tens of kilometers over the plain on the downwind of the separation.

Fig. 9. Surface pressure deviation (δp) for 48 hour runs of the “No Q_H ” experiment.

As G increases the downslope wind is deeper and an upward motion at its edge is stronger. For $G = 5 \text{ ms}^{-1}$ and 10 ms^{-1} , the upward speed is 30 cms^{-1} and 100 cms^{-1} , respectively. However, the extent of the downslope wind is maximum at $G \cong 4 - 5 \text{ ms}^{-1}$, and it retreats toward the mountain foot as G increases above this range.

This upward motion may be worth calling a “jump,” which is often used in literature on the dynamics of airflow over mountains. In the presence of diurnal heating, however, a thermally driven motion can also generate a convergence zone on the leeside of the mountain, as shown in Chapter 5. In some range of G , a thermally generated convergence and a dynamically induced jump are so closely coupled that it is impractical to define them separately. Hereafter we use a generic expression “lee convergence zone (LCZ)”.

For $G = 16 \text{ ms}^{-1}$, the LCZ disappears and the flow pattern becomes a supercritical type. The x - G diagram in Fig. 8 indicates a sharp transition of the flow regime between $G = 14 \text{ ms}^{-1}$ and 16 ms^{-1} , which is in rough agreement with the critical wind speed (13 ms^{-1}).

Figure 9 shows the distribution of surface pressure

Fig. 10. Same as Fig. 7 but for the “fixed T_s ” experiment.

Fig. 11. Same as Fig. 8 but for the “fixed T_s ” experiment.

deviation from the initial state (δp). The LCZ at the mountain foot corresponds to a pressure depression, which will be called a “lee trough” hereafter. The lee trough deepens as G increases, but it suddenly weakens at $G = 16 \text{ ms}^{-1}$ with the disappearance of LCZ.

4.2 The “fixed T_s ” experiment

Figures 10 and 11 show the results of the “fixed T_s ” experiment. In comparison to the “no Q_H ” experiment, the LCZ is nearer the peak by the order of 10 km. The difference from the “no Q_H ” experiment is largest for moderate prevailing wind ($G \approx 5 \text{ ms}^{-1}$).

The difference of the two experiments corresponds to that of the stratification of the boundary layer

over the plain. In the “no Q_H ” case, the lowest few hundred meters is nearly neutral reflecting the strong turbulent mixing associated with the strong downslope wind. On the other hand, a thin stable layer exists in the lowest few hundred meters in the “fixed T_s ” case, because air temperature is higher than in the initial state whereas the ground temperature is fixed. This air-surface temperature difference is accompanied by negative heat flux (negative Q_H) of $5\text{--}10 \text{ Wm}^{-2}$, $15\text{--}20 \text{ Wm}^{-2}$ and $20\text{--}30 \text{ Wm}^{-2}$ for $G = 5, 10$ and 16 ms^{-1} , respectively. These are a small fraction (typically one third or so) of the nocturnal cooling in the diurnal variation cases.

5. Diurnal wind variation in the control experiment

A series of experiments for different G values reveals that the features of diurnal wind variation can be classified into four regimes with respect to the behavior of LCZ in the daytime. A detailed description of each regime is given in the following.

5.1 Upslope wind regime: $G \leq 2 \text{ ms}^{-1}$

Figure 12 shows the cross section of u , w and θ from 06 LT to 21 LT for the case $G = 2 \text{ ms}^{-1}$. Figure 13 is a x - t diagram of surface wind and w at $z^* = 370 \text{ m}$.

At sunrise (06 LT), the wind field is characterized by a downslope flow on the leeside slope and a cold air layer with a weak reverse flow over the plain. This flow pattern is similar to that of the “fixed T_s ” run. Not only in this case but also for other G values, the nocturnal flow patterns are similar to those obtained from the “fixed T_s ” condition. In the remaining part of this chapter, we shall focus our attention on the daytime flow fields.

Fig. 12. Distribution of u , w and θ for the case $G = 2 \text{ ms}^{-1}$.

Fig. 13. An x - t diagram of surface wind at $z^* = 10 \text{ m}$ and w at $z^* = 370 \text{ m}$ for the case $G = 2 \text{ ms}^{-1}$. Contours are drawn for $w = 0, \pm 10 \text{ cm s}^{-1}$ and $\pm 20 \text{ cm s}^{-1}$.

By 09 LT, the cold air layer has almost disappeared over the plain, while upslope wind of $\sim 4 \text{ ms}^{-1}$ develops on the lower part of the slope. It converges with the downslope wind blowing down the upper part of the slope, and forms a well-defined

LCZ. As the upslope wind intensifies, the LCZ shifts from $x = 25\text{--}30 \text{ km}$ at 09 LT, to $x \cong 20 \text{ km}$ at 12 LT with an enhanced upward motion of $\sim 40 \text{ cm s}^{-1}$. After 12 LT, the upslope wind gradually weakens and the LCZ moves back down the slope. The daytime wind field is characterized by an upward and downward movement of a LCZ on the slope, corresponding to the development and weakening of the upslope wind.

In the late afternoon, the LCZ moves onto the plain with an increasing propagation speed. At the passage of the LCZ, the area near the mountain foot experiences a surge of downslope wind. This may look like a nocturnal mountain breeze, but is much stronger and deeper than an ordinary drainage wind. The LCZ weakens within a few hours of sunset as an inversion layer forms over the plain.

5.2 Propagating LCZ: $3 \text{ ms}^{-1} \leq G \leq 6 \text{ ms}^{-1}$

For the cases $G \geq 3 \text{ ms}^{-1}$, the daytime upslope wind is not generated but the leeside slope is dominated by the downslope wind throughout a day. If G is not too large ($3 \text{ ms}^{-1} \leq G \leq 6 \text{ ms}^{-1}$), the layer of strong downslope wind spreads far into the plain during the daytime.

Figures 14 and 15 show the result for $G = 5 \text{ ms}^{-1}$. With the disappearance of the nocturnal inversion layer over the plain, a well-defined LCZ develops at

Fig. 14. Distribution of u , w and θ for the case $G = 5 \text{ ms}^{-1}$.

Fig. 15. Same as Fig. 13 but for $G = 5 \text{ ms}^{-1}$. Contours are drawn for $w = 0, \pm 20 \text{ cms}^{-1}, \pm 40 \text{ cms}^{-1}$ and $\pm 80 \text{ cms}^{-1}$.

the edge of the strong wind layer and then moves downstream during the daytime. This feature is similar to the situation around sunset in the case $G = 2 \text{ ms}^{-1}$. The LCZ continues to move downwind into the plain, so that a sudden increase in wind

Fig. 16. Surface pressure deviation (δp) for the case $G = 5 \text{ ms}^{-1}$.

speed occurs over a distance of more than 100 km from the mountain foot by 18 LT. The propagation speed of the LCZ is about 4 ms^{-1} , which is slightly lower than the prevailing wind speed. After sunset, the wind over the plain rapidly weakens as an inversion layer forms.

The thermal structure of LCZ changes with time. Within a few hours of the sunrise (06 and 09 LT), the LCZ is a boundary between the cold air pool over the plain and warmer air of the downslope wind (Figs. 14a and 14b). By 12 LT, however, the LCZ changes into a warm-core structure as seen from isolated 22°C isentropes in Figs. 14c–14d. Figure 16 shows the diurnal variation of δp distribution. The well-defined lee trough in the morning disappears

Fig. 17. Track of the LCZ (the position of $w_{\max}$) for each case of $G \leq 6 \text{ ms}^{-1}$.

in the midday, while the LCZ agrees with a weak trough at $x \cong 80 \text{ km}$ at 12 LT and $x \cong 130 \text{ km}$ at 15 LT. This structure implies a feedback mechanism in an unstable boundary layer, where the low-level convergence at the LCZ pumps up the warm air, and the resulting pressure depression maintains the convergence.

In order to compare the upslope wind regime and the propagating LCZ regime, Figure 17 shows the track of the LCZ for each value of G from 0 ms^{-1} to 6 ms^{-1} . The position of the LCZ was defined by the maximum of w (hereafter $w_{\max}$) in the lowest 1 km in the domain $0 \leq x \leq 180 \text{ km}$; a LCZ was regarded as absent if $w_{\max} < 10 \text{ cm s}^{-1}$. The $G = 0 \text{ ms}^{-1}$ case corresponds to a pure mountain/valley

breeze circulation which is symmetric with respect to the peak. The position of $w_{\max}$ stays above the peak from 10 LT to 18 LT due to the convergence of upslope winds blowing from both sides of the mountain (the LCZ at $x = 20 \text{ km}$ at 09 LT is superficial because it is in fact an upslope wind along the slope). In the cases of $G = 1 \text{ ms}^{-1}$ and 2 ms^{-1} , the LCZ moves up and down on the slope during the daytime as described in 5.1, while it moves unidirectionally away from the mountain foot in the cases $G \geq 3 \text{ ms}^{-1}$. The LCZ tends to move faster under a stronger prevailing wind.

5.3 Oscillating LCZ: $7 \text{ ms}^{-1} \leq G \leq 14 \text{ ms}^{-1}$

The results for $G \geq 7 \text{ ms}^{-1}$ are largely different from those presented above. The LCZ does not propagate far into the plain, but only exhibits an oscillatory movement near the mountain foot.

Figures 18 and 19 show the result for $G = 10 \text{ ms}^{-1}$. The LCZ at the edge of strong wind layer advances from $x \cong 35 \text{ km}$ at 06 LT to $x \cong 55 \text{ km}$ at 15 LT, and then gradually retreats toward the mountain foot during the evening. The wind outside this region remains weak throughout a day, although the surface wind speed in the far downstream ($x \geq 80 \text{ km}$) increases in the daytime as the nocturnal inversion layer disappears.

Figure 20 shows the diurnal variation of δp distribution for the $G = 10 \text{ ms}^{-1}$ case. The lee trough

Fig. 18. Distribution of u , w and θ for the case $G = 10 \text{ ms}^{-1}$.

Fig. 19. Same as Fig. 13 but for $G = 10 \text{ ms}^{-1}$. Contours are drawn for $w = 0, \pm 40 \text{ cm s}^{-1}$ and $\pm 80 \text{ cm s}^{-1}$.

Fig. 20. Surface pressure deviation (δp) for the case $G = 10 \text{ ms}^{-1}$.

which is much deeper than in the $G = 5 \text{ ms}^{-1}$ case (Fig. 16) stays near the mountain foot ($x \cong 40 \text{ km}$) with only a slight downwind bulge in the daytime. This indicates that the dynamical effect is strong enough to keep the LCZ from propagating far downwind.

Fig. 21. Track of the LCZ for each case of $G \geq 5 \text{ ms}^{-1}$.

Figure 21 shows the LCZ track defined in the same way as in Fig. 17. As G increases the amplitude of the oscillation decreases, so that the flow pattern becomes closer to a steady state.

5.4 Supercritical flow regime: $G \geq 16 \text{ ms}^{-1}$

As in the “no Q_H ” and “fixed T_S ” runs, the LCZ disappears at $G = 16 \text{ ms}^{-1}$. Figures 22 and 23 shows the result for $G = 16 \text{ ms}^{-1}$. The flow pattern is almost independent of time, although the surface wind on the slope is stronger in the nighttime than in the daytime.

In order to summarize the features of the four regimes, the x - G diagrams of surface wind and w ($z^* = 370 \text{ m}$) at 06 LT and 15 LT are shown in Fig. 24. The feature for 06 LT is similar to that of the “fixed T_S ” experiment as mentioned in 5.1. The daytime intensification of wind over the plain is most conspicuous in the propagating LCZ regime ($3 \text{ ms}^{-1} \leq G \leq 6 \text{ ms}^{-1}$). The time change of wind fields in this regime is in qualitative accordance with the daytime advance of the downslope wind described in Chapter 2. If we confine our attention to the area within a few tens of kilometers of the mountain foot, the features of the oscillating

Fig. 22. Distribution of u , w and θ for the case $G = 16 \text{ ms}^{-1}$.

Fig. 23. Same as Fig. 13 but for $G = 16 \text{ ms}^{-1}$. Contours are drawn for $w = 0, \pm 40 \text{ cms}^{-1}$ and $\pm 80 \text{ cms}^{-1}$.

LCZ regime ($7 \text{ ms}^{-1} \leq G \leq 14 \text{ ms}^{-1}$) is also acceptable as a model of daytime advance of strong downslope wind.

As seen from Fig. 24, the transition of regimes occurs in a narrow range of G . This means that the flow pattern for this range is very sensitive to the change in G and, quite possibly, to the change in low-level atmospheric conditions (such as stratification) as well. In fact, it was observed during the experiments that the flow patterns for transitional G values (e.g., $G = 7 \text{ ms}^{-1}$ and 14 ms^{-1}) were somewhat "unstable" in the sense of incomplete periodicity of the diurnal cycle, probably because of slight day-to-day differences in the low-level stratification. For other ranges of G , however, the diurnal cycle is satisfactorily regular as stated before.

6. Sensitivity experiments

In the six sensitivity experiments listed in Table 2, a propagating LCZ was commonly produced

Fig. 24. Same as Fig. 8 but for (a) 06 LT and (b) 15 LT of the control experiment.

Fig. 25. Track of the LCZ obtained from each experiment for $G = 5 \text{ ms}^{-1}$.

Fig. 26. Track of the LCZ obtained from each experiment for $G = 10 \text{ ms}^{-1}$.

Fig. 27. Same as Fig. 8 but for (a) 06 LT and (b) 15 LT of the " $H = 600$ m" experiment.

for $G = 5 \text{ ms}^{-1}$ and an oscillatory LCZ appeared for $G = 10 \text{ ms}^{-1}$ except for the $H = 600$ m case. Figures 25 and 26 show the track of the LCZ obtained in each experiment. Apart from some differences in the propagation speed and/or location, there is an essential similarity in the behavior of the LCZs for each G values. This indicates that a daytime advance of strong wind is one of the general features in the lee of mesoscale mountain area. In the following, some detailed description is given on the comparison of each experiment and the control case.

6.1 The " $H = 600$ m" experiment

Figure 27 shows the x - G diagram of surface wind, and w at $z^* = 370$ m for the " $H = 600$ m" experiment. The features of diurnal variation for weak and moderate prevailing wind ($G \leq 5 \text{ ms}^{-1}$) are similar to those obtained from the control experiment (Fig. 24), although the LCZ is located nearer the mountain by ~ 20 km (Fig. 25). However, the LCZ disappears for stronger prevailing winds ($G \geq 7 \text{ ms}^{-1}$) instead of the oscillating LCZ in the control experiment. This is evidently because of the halved critical wind speed ($u_c = NH \cong 7 \text{ ms}^{-1}$). In other words, the flow pattern changes from the propagating LCZ regime to the supercritical regime as G increases above 5 ms^{-1} , without an oscillatory LCZ regime.

6.2 The " $x_0 = 80$ km" and the "plateau" experiments

The flow feature of the " $x_0 = 80$ km" experiment is quite similar to that of the control case. The propagation speed of the LCZ differs from the control experiment only slightly in the $G = 5 \text{ ms}^{-1}$ run. In the "plateau" experiment, the upslope wind is more intense than in the control experiment under a weak prevailing wind ($G \leq 2 \text{ ms}^{-1}$; not shown), but the flow patterns for moderate and strong prevailing winds ($G \geq 5 \text{ ms}^{-1}$) do not show essential difference from the control case. These results indicate that the flow pattern is insensitive to the mountain shape within the two-dimensional restriction.

6.3 The " $B = 0.1$ " experiment

As expected, the daytime advance of LCZs is slower in this experiment than in the control one ($B = 0.5$). A propagating LCZ appears for $3 \text{ ms}^{-1} \leq G \leq 5 \text{ ms}^{-1}$, but the upward motion is weaker and penetration distance is much shorter than in the control case. The LCZ shows an oscillatory behavior for $G \geq 6 \text{ ms}^{-1}$, namely at a smaller value of prevailing wind speed than in the control experiment. For $G = 10 \text{ ms}^{-1}$, the LCZ is almost as conspicuous as in the control experiment, but the amplitude of the oscillation is nearly half of that of the control case. These results confirm the importance of daytime heating in the advance of the downslope wind.

6.4 The " $z_0 = 0.2$ m" experiment

In comparison to the control experiment ($z_0 = 0.04$ m), the LCZ is located nearer the peak at any time of day. This corresponds to previous findings that surface friction prevents the downwind spread of the downslope wind (Richard *et al.*, 1989; Saito, 1993). The amplitude of the LCZ oscillation in the $G = 10 \text{ ms}^{-1}$ case is substantially smaller than in the control experiment. This indicates that surface friction weakens the diurnal oscillation of LCZ, although the effect on the propagating LCZ appears to be small.

6.5 The "const K_M " experiment

The downwind propagation of LCZ is produced for $G = 5 \text{ ms}^{-1}$ with only a slight decrease in the propagation speed. On the other hand, the downwind shift of the LCZ in the $G = 10 \text{ ms}^{-1}$ run is nearly half of that in the control case. These results indicate that the diurnal variation of the frictional force is not essential to the behavior of downslope winds, although it intensifies the amplitude of the LCZ movement in the oscillatory LCZ regime.

7. Summary and remarks

A series of two-dimensional numerical experiments was made on diurnal wind variation in the lee of a mountain range. The daytime wind fields obtained from the control experiments, in which the mountain height was set to be 1200 m, were found to change systematically according to the prevailing wind speed (G). They were classified into following four regimes with respect to the behavior of the lee convergence zone (LCZ).

- (1) The upslope wind regime ($G \leq 2 \text{ ms}^{-1}$): The upslope wind develops and weakens on the lower part of the leeside slope, while the downslope wind dominates on the upper part of the slope. The LCZ between the two wind systems moves up and down on the slope during the daytime, then onto the plain in the late afternoon.
- (2) The propagating LCZ regime ($3 \text{ ms}^{-1} \leq G \leq 6 \text{ ms}^{-1}$): The downslope wind dominates on the leeside slope throughout a day. In the daytime it extends far downstream over the plain following an advancing LCZ.
- (3) The oscillating LCZ regime ($7 \text{ ms}^{-1} \leq G \leq 14 \text{ ms}^{-1}$): The downslope wind dominating over the leeside slope spreads onto the plain in the daytime, but not far from the mountain foot. The LCZ exhibits an oscillatory movement near the mountain foot, with a smaller amplitude under a larger G value.
- (4) The supercritical regime ($G \geq 16 \text{ ms}^{-1}$): The LCZ disappears and the flow pattern shows little diurnal variation, except for some day-night changes in wind speed.

The features were produced with only a minor changes in sensitivity experiments on mountain height and shape, surface heating intensity, surface roughness and the stability dependence of eddy viscosity. In the case of a low mountain (600 m high), the oscillating LCZ regime did not occur, but the flow feature changed from the propagating LCZ regime to the supercritical regime at a value of $G \cong 5\text{--}7 \text{ ms}^{-1}$.

This study has shown the importance of diurnal surface heating on the behavior of leeside winds. In particular, the propagating and oscillating LCZ are in qualitative agreement with the daytime advance of downslope wind observed in leeside plains such as the Canterbury Plains in New Zealand and the Kanto plain in Japan. It is encouraging that the model was able to reproduce some observed features, although the real topography is much more complex than was assigned in the model. We hope that our findings will provide useful knowledge in future studies, including a realistic simulation based on a three-dimensional model.

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2次元の山の風下側における下層風日変化の数値的研究

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山脈の風下側における下層風の日変化について、気象研究所非静力学モデルの2次元・非湿潤過程版による一連の数値実験を行った。風の日変化の特徴を一般風速に応じて次の4つのレジームに分けることができた。(1) 谷風レジーム：昼間に山麓部で谷風が吹き、これと山頂を吹き越えるおろし風とが収束する状態、(2) LCZ 侵入レジーム：昼間におろし風が、前面に収束帯 (lee convergence zone=LCZ) を伴いつつ風下側の平野へ侵入していく状態、(3) LCZ 振動レジーム：LCZ が山の風下側で振動する状態、(4) 超臨界レジーム：多少の風速の日変化を除いてほぼ定常な状態。このうち(2)と(3)は昼間に平野で地上風速の急増をもたらす。これはニュージーランドのカンタベリー平野や関東平野における観測事実と定性的に合う。