Dear Dr. Liu,

We appreciate all insightful comments from reviewers and thank you very much for handling this process! We have made a major revision. We rewrote abstract and final section and addressed all comments raised by the reviewers. We believe that we have clarified all of reviewers´ concerns and the manuscript is ready to accept for publication.

Thank you very much for your consideration!

Sincerely,

Chuixiang Yi et al

Response to Interactive comment on "Stably stratified canopy flow in complex terrain" by X. Xu et al.

-------------------Reviewer #1--------------------

Reviewer #1

General Comments: The manuscript reported numerical studies of stably stratified canopy flow with complex terrain using Reynolds averaged Navier-Stokes (RANS) equations and RNG k-ε turbulent model. Thermal stratification in canopy flow is a long-standing problem for numerical modeling using available computational fluid dynamics (CFD) techniques. The major challenge is the intermittent turbulence associated with canopy flow under thermal stratification conditions. Advanced tools such as large eddy simulations (LES) cannot accurately capture the intermittent turbulence feature. On the other hand, direct numerical simulation (DNS) needs prohibitive computational resource makes it un-realistic to handle this problem. So the studies conducted in this work represent the latest effort in this area, in particular, the quantification and budget analysis of turbulent kinetic energy (TKE) under stably stratified canopy flow in complex terrain. Overall, the paper is a very good study and this reviewer recommends for publication with ACP after minor revision.

Authors:

We are grateful for reviewer's positive comments!

Reviewer #1

Special Comments: Page 28488 L22-25: the authors mentioned the computational domain size and grid used. On Page 28489 L4 provided the surface roughness height to be 0.01m. Did the authors conduct mesh independence studies to make sure the mesh sizes chosen in the simulation were fine enough so that the numerical results obtained were not sensitive to the mesh size? Please clarify this issue. Also please clarify what mesh size used close to the ground of the terrain and mesh size used in the area far away from the ground.

Authors:

We have tested the sensitivity of mesh grids in this numerical modeling. The vertical resolution of canopy structure (Leaf area density profile) in this numerical simulation is at a 1m resolution. We tested the mesh sensitivity of mesh grids at a finer resolution (0.2-1m) than canopy structure resolution. The results have low dependency on grid spacing lower than 0.6m at ground surface. The mesh size from the ground to the top of canopy is 0.5m and stretched from 0.5m at the top of canopy to about 1.5m at the top boundary of the domain. The grid spacing is described in L25-26, page 28488 and L1-3, page 28489.

Reviewer #1

Page 28486 L1-5: This manuscript mentioned super-stable layers near leaves. The authors provided reference of Yi et al. (2005). Are there any research reports to show the turbulent flow field in this region? It is a tough task and might be out of the scope of this paper, just curiosity. A friendly reminder here is that the paper should emphasize that the flow is fully turbulent, even in the region of the super-stable layer. Even though RANS can still handle super-stable layer, the reason is that no matter RANS or large-eddy simulation (LES), turbulent intermittence is a significant challenge for accurate numerical simulations which might finally rely on experimental measurement.

Authors:

The super stable layer has been demonstrated by a few more eddy-flux measurements (van Gorsel et al., 2011; Alekseychik et al., 2013) since Yi et al. (2005) reported it. However, to our knowledge, there is no research on the details of full turbulence development in the super-stable layer. As the reviewer pointed out, studying turbulent intermittence around the super stable layer would be interesting. We will keep this in mind and address this point in our future study.

References:

van Gorsel, E., Harman, I. N., Finnigan, J. J., Leuning, R.: Decoupling of air flow above and in plant canopies and gravity waves affect micrometeorological estimates of net scalar exchange. Agric. For. Meteor. 151, 927-933, 2011.

Alekseychik, I. Mammarella, P., Launiainen, S., Rannik, Ü., Vesala, T.: Evolution of the nocturnal decoupled layer in a pine forest canopy. Agric. For. Meteor. 174-175, 15-27, 2013.

Reviewer #1

Page 28496 L10-15 about discrete Richardson number for stability of canopy flow, is there any way to output your RANS results of turbulent intensity, which might be directly used to check the consistence with the predicted Richardson numbers shown in Fig. 4? Answer to this question is optional (the paper extensively discussed the Ri and temperatures (Fig 3 and Fig 4)).

Authors:

We appreciate reviewer's suggestion. Unfortunately, we did not have the calculated turbulent intensity.

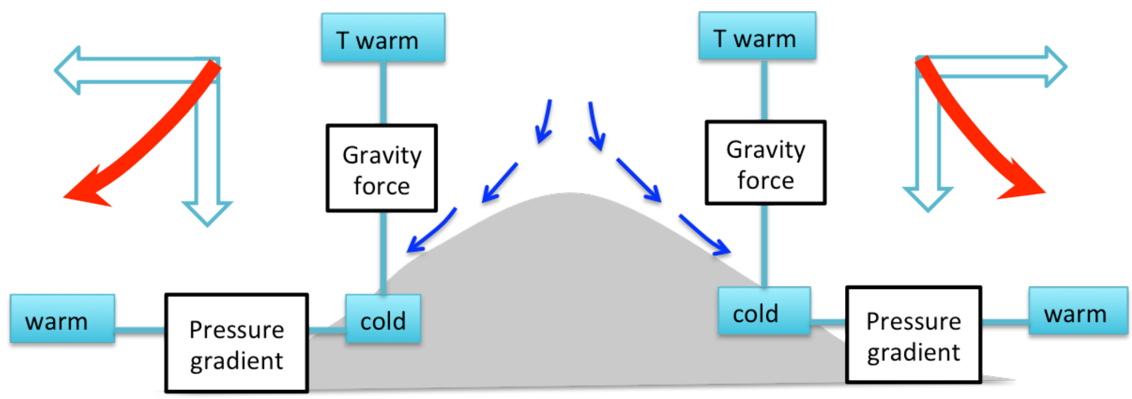
Reviewer #1

Page 28497 L25-30 about wind flow structures, there are two major factors to form the scenario, "converges towards the hill," the terrain and the cooling rate. Which factor is the dominant term? If cooling rate removed, will still generate this phenomenon? Please explain.

Authors:

The sinking flow above hill is initialized by the buoyancy force. However, if there is no slope, the descending wind would be very weak and there will be no convergence above the summits of hills. The temperature gradient in vertical and cooling on the slope surface and canopy layer leads to gravity force in vertical. In addition, the cooling on slope surface leads to positive pressure gradient in horizontal from the sloped surface. The symmetric terrain makes the convergence symmetric. A schematic figure below shows the net force along the slope resulting from gravity force in vertical and pressure gradient in horizontal.

The forces that control the flows can be described as: hydrodynamic pressure gradient $F_{hd} = d\Delta p/dx \approx U^2H/L^2$ and buoyancy force (or hydrostatic pressure gradient) $F_{hs} = g(\Delta\theta/\theta_0)\ \sin\alpha \approx g(\Delta\theta/\theta_0\ )H/L$, where α is the slope angle, $\Delta\theta$ is the potential temperature difference between the ambient air and the colder slope flow, θ_0 is the ambient potential temperature (Belcher et al., 2008). Froude number Fr=U/NL=$\sqrt{F_{hd}/F_{hs}}$, where $N = (g/\theta_{\ 0})d\theta/dz$, L is the hill length scale. When Fr <<1, the hydrostatic pressure gradient is large enough to initiate drainage flow.



Reviewer #1

Page 28501 L12-24 about turbulent kinetic energy budget analysis in Figures 9-10, is there any way to validate the each term, e.g. any available experimental data about these? Which of these terms can be determined through experimental measurement? In such a way, it might be helpful to validate CFD results in the future (based RANS methodology).

Authors:

Meyers and Baldocchi (1991) have analyzed TKE budget in a deciduous forest under near neutral to slightly unstable condition. They simplified the calculation of vertical profile and ignored the buoyancy production. However, Leclerc et al. (1990) showed the importance of buoyancy production in both stable and unstable condition by observation in a deciduous forest. Direct measurements of turbulent pressure fluctuations in plant canopies are extremely difficult. Surface pressure fluctuations are used to approximate the anemometers' level pressure fluctuations to infer the role of pressure fluctuations in canopy turbulence (Maitani and Seo, 1985; Shaw et al. 1990; Shaw and Zhang, 1992). However, the calculation with surface pressure fluctuation fails to separate pressure diffusion and return-to-isotropy components of pressure-velocity interactions (Dwyer et al., 1997). The canopy flows are coupling results of topography and vegetation. The TKE fields are sensitive to the terrain's aspect ratio (Katurji et al., 2011), which are also shown in our numerical results.

References:
Meyers, T.P., Baldocchi, D. D.: The budgets of turbulent kinetic energy and Reynolds stress within and above deciduous forest. Agric. For. Meteor. 53, 207-222, 1991.
Meyers, T. P. and Paw U, K. T.: 1986, Testing of a Higher-Order Closure Model for Flow Within and Above Plant Canopies, Boundary-Layer Meteorol. 37, 297–311, 1986.
Leclerc, M. Y., Beissner, K. C., Shaw, R. H., den Hartog, G., Neumann, H. H.: The influence of atmospheric stability on the budgets of the reynolds stress and turbulent kinetic energy within and above a deciduous forest. J. Appl. Meteor. 29, 916–933, 1990.
Katurji, M., Zhong, S., Zawar-Reza, P.: Long-range transport terrain-induced turbulence from high-resolution numerical simulations. Atmos. Chem. Phys., 11, 11793–11805, 2011.
Wilson, N. R. and Shaw, R. H.: A Higher Order Closure Model for Canopy Flow, J. Appl. Meteorol. 16, 1197–1205, 1977.
Dwyer, M. J., Patton, E. G., and Shaw, R. H.: Turbulent kinetic energy budgets from a large-eddy simulation of airflow above and within a forest canopy, Bound.-Lay. Meteorol., 84, 23–43, 1997.

Reviewer #1

Other comments: Overall this study brings lots of transition phenomena due to thermal stratification and the complex terrain interaction each other. For instance, the wind shear changes from the case H/L=0.6 to H/L=1.0 are very interesting (Fig. 6 and Fig. 7 of the paper), look forward to seeing experimental measurement for this variation. It is a very good work.

Authors:

To validate the numerical results, the necessary intensive measurements of wind and temperature fields below and above the top of canopy under different atmospheric stability are required. We look forward to this as well!

-------------------------- **Reviewer #2**----------------------------

Reviewer #2

This manuscript presents numerical simulations of stably stratified flow within and above a canopy over an isolated, idealized two-dimensional hill. The numerical model uses the Renormalized Group (RNG) k-epsilon turbulence model. The topic is an interesting and important one. Flow decoupling and drainage flows under stable conditions are important in controlling nighttime fluxes from forest canopies. While the simulations are potentially interesting, I found the discussion of them rather confusing and not particularly enlightening. I also have a number of questions about the model itself. This is compounded by the English language. A number of sentences just didn't make sense and I was unable to work out what you were trying to say. In addition to my scientific suggestions below, the language in the manuscript needs careful proof reading. Given my long list of questions I recommend major revisions for this manuscript.

Authors:
We thank referee #2's for positive comments and constructive criticisms.

Reviewer #2
Major comments
1) I'm not quite sure what the aim of the paper is. The abstract suggests that it is just showing that the model can successfully simulate stable canopy flows. It does simulate canopy flows which look reasonable and qualitatively reproduces some features seen in field observations, but since the simulations are very idealised there is no data to quantitative compare with and so it is impossible to be sure that the model really is "accurate".

Authors:
The canopy layer is an interface between land and atmosphere, which directly influences mass and energy exchange between vegetation and atmosphere. The complexity of nocturnal canopy flows in hilly terrain has been challenging the accurate measurement of mass and energy flux in FLUXNET community. The huge difficulty and problem (e.g. advection error) in eddy-flux measurements has been caused by stably-stratified canopy flow over complex terrain so that the measured data must be trashed under conditions with weak turbulence and very stably stratified flow (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., 2010; Aubinet and Feigenwinter, 2010; Queck and Bernhofer, 2010; Siebicke et al., 2012). This is a well-known and long-standing problem in FLUXNET that consists of more than 500 towers around the world. We want to understand this difficult condition from a modeling aspect. We agree with the reviewer that this is not accurate prediction since measurements that can be used to compare with our modeling results are not available yet. However, we want to use CFD modeling to qualitatively understand the main features of stably-stratified canopy flow that have been captured piecemeal by separate field observations (Alekseychik et al., 2013; van Gorsel et al., 2011; Jacob et al., 1992; Leclerc et al., 1990;Yi et al., 2005; Zhang et al., 2010). We rewrote the abstract to clarify reviewer´s concerns.

References:
Alekseychik, I. Mammarella, P., Launiainen, S., Rannik, Ü., Vesala, T.: Evolution of the nocturnal decoupled layer in a pine forest canopy. Agric. For. Meteor. 174-175, 15-27, 2013.
Aubinet, M., Feigenwinter, C.: Direct CO2 advection measurements and the night flux problem. Agric. For. Meteor. 150, 651-654, 2010.
Aubinet, M., Heinesch, B., Yernaux, M.: Horizontal and vertical CO2 advection in a sloping forest. Bound.-Layer Meteor. 108, 397–417, 2003.
Feigenwinter, C., Montagnani, L., Aubinet, M.: Plot-scale vertical and horizontal transport of CO2 modified by a persistent slope wind system in and above an alpine forest. Agric. For. Meteor. 150, 665–673, 2010.
van Gorsel, E., Harman, I. N., Finnigan, J. J., Leuning, R.: Decoupling of air flow above and in plant canopies and gravity waves affect micrometeorological estimates of net scalar exchange. Agric. For. Meteor. 151, 927-933, 2011.
Goulden, M. L., Daube, B. C., Fan, S., Sutton, D. J., Bazzaz, A., Munger, J. W., Wofsy, S. C.: Physiological response of a black spruce forest to weather. J. Geophys. Res. 102, 28987–28996, 1997.
Jacobs, A.F.G., van Boxel, J.H., Shaw, R.H.: The dependence of canopy layer turbulence on within-canopy thermal stratification. Agric. For. Meteorol. 58: 247-256,1992.
Leclerc, M. Y., Beissner, K. C., Shaw, R. H., den Hartog, G., Neumann, H. H.: The influence of atmospheric stability on the budgets of the reynolds stress and turbulent kinetic energy within and above a deciduous forest. J. Appl. Meteor. 29, 916–933, 1990.
Montagnani, L., Manca G., Canepa, E., Georgieva, E., Acosta, M., Feigenwinter. C., Janous, D., Kerschbaumer, G., Lindroth, A., Minach, L., Minerbi, S., Mölder, m., Pavelka, M., Seufert, G., Zeri, M., Ziegler, W.: A new mass conservation approach to the study of CO2 advection in an alpine forest. J. Geophys. Res. 114, D07306. doi:10.1029/2008JD010650, 2009.

Queck, R., Bernhofer, C.: Constructing wind profiles in forests from limited measurements of wind and vegetation structure. Agric. For. Meteor. 150, 724-735, 2010.
Sun, J., Burns, S. P., Delany, A. C., Oncley, S. P., Turnipseed, A. A., Stephens, B. B., Lenschow, D. H., LeMone, M. A., Monson, R. K., Anderson, D. E.: CO2 transport over complex terrain. Agric. For. Meteor. 145, 1–21, 2007.
Staebler, R. M., Fitzjarrald D. R.: Observing subcanopy CO2 advection. Agric. For. Meteor. 12, 139-156, 2004.
Siebicke, L., Hunner, M., Foken, T.: Aspects of CO2 advection measurements, Theor. Appl. Climatol. 109, 109–131, 2012.
Yi, C., Monsoon, R. K., Zhai, Z., Anderson, D. E., Lamb, B., Allwine, G., Turnipseed, A. A., Burns, S. P.: Modeling and measuring the nocturnal drainage flow in a high-elevation, subalpine forest with complex terrain. J. Geophys. Res. 110, D22303. doi:10.1029/2005JD006282, 2005.
Zhang, G., Leclerc, M. Y., Karipot, A.: Local flux-profile relationships of wind speed and temperature in a canopy layer in atmospheric stable conditions. Biogeosciences 7, 3625-3636, 2010.

Reviewer #2

2) Justification of the RNG closure. Two references are given to support the use of the RNG model over complex terrain. These are both for neutral flow. Only the previous paper by the authors has a canopy included (Xu and Yi, 2013), but this contains no validation of the model or comparison with observations. Has the RNG model been validated for canopy flows? Has it been validated for stable flows? The reader needs some evidence the model is correct before believing the results from this study.

Authors:

We gave a detailed discussion of current models and their limitations applied in canopy flows in the section of 'Introduction'. RNG turbulence models have been applied to vegetation canopy flows on hilly terrain (Xu and Yi, 2013; Pattanapol et al., 2006) and urban canopy flows (Kim and Baik, 2004; Kim et al., 2012; Koutsourakis et al., 2012; Cheng et al., 2009). Pattanapol et al. (2006) compared the numerical results with measurements on hill slopes. The representation of source/sink terms of canopy (drag force, wake production in momentum and TKE budget) shows a more accurate prediction than those using roughness parameters. The RNG model is validated for urban canopy flows under neutral and unstable condition. Due to the difficulties in measuring night-time small scale turbulence and strong variability in topography and canopy structure, we can only have some qualitative comparison between our model prediction and field measurement. As we do this pioneer study of stably stratified canopy flow in hilly terrain, we expect more results from other model simulations, wind tunnel experiments and accurate quantification of the high resolution topography and vegetation morphology and intensive measurements of canopy flows to validate our modeling results.

Cheng, W. C., Liu, C. H., Leung, D. Y. C.: On the correlation of air and pollutant exchange for street canyons in combined wind-buoyancy-driven flow. 43, 3682-3690, 2009.
Kim, J. J., Baik, J. J.: A numerical study of the effects of ambient wind direction on flow and dispersion in urban street canyons using the RNG k-e turbulence. 38, 3039-3048, 2004.
Kim, M.J., Park, R. J., Kim, J. J.: Urban air quality modeling with full O3-NOx-VOC chemistry: Implications for O3 and PM air quality in a street canyon. Atmos. Environ. 47, 330-340, 2012.
Koutsourakis, N., Bartzis, J. G., Markatos, N. C.: Evaluation of Reynolds stress, k-e and RNG k-e turbulence models in street canyon flows using various experimental datasets. Environ. Fluid Mech. 12, 379-403, 2012.
Pattanapol, W., Wakes, S. J., Hilton, M. J., Dickinson, K. J. M.: Modeling of surface roughness for flow over a complex vegetated surface. World Acad. Sci. Eng. Technol. 26, 271-291, 2006.
Xu, X., Yi, C.: The influence of geometry on recirculation and CO_2 transport over forested hills. Meteorol. Atmos. Phys. 119, 187-196, 2013.

Reviewer #2

3) In the description of the RNG model you state that "Tp is calculated as the residual of all other terms". How can you do that as you don't know what dk/dt is? Or do you assume dk/dt

= 0 (implied later on in section 3.5)? In that case this is only a steady state turbulence closure model, but is applied to a time-varying model? Seems like a major limitation to me. Can you comment on this?

Authors:
In our simulation, steady state ($\partial k/\partial t = 0$ and $\partial \varepsilon/\partial t = 0$) is assumed. We corrected the typos in the eq. (13) and (14), thank you! The condition of steady state flow under stable condition was proposed by Mahrt (1992). We evaluated our model setup in section 2.2 that meets the steady state assumption.

Reviewer #2
4) It is stated that the flow is sufficiently forced to ensure the flow remains turbulent. I find this a bit hard to believe with such stable layers. Accurately simulating stable flow is hard - and this comes back to my comments above about whether the model is really validated.

Authors:
We think that the simulation of stable flow on flat terrain is hard. But the buoyancy flow on slope can be simulated by CFD model. Luo and Li (2011) have simulated buoyancy driven slope flow and wall flow with even stronger cooling on slope surface (-100 Wm^{-2}) and wall surface (-100 Wm^{-2}) under calm conditions (no synoptic winds). The stronger cooling intensity leads to stronger down-slope wind, thus more down-slope air mixing.

Luo, Z., Li, Y.: Passive urban ventilation by combined buoyancy-driven slope flow and wall flow: Parametric CFD studies on idealized city models. Atmos. Environ. 45, 5946-5956, 2011.

Reviewer #2
5) Key to interpreting these idealised simulations seems to be the drawing down of air into the canopy near the summit due to continuity. This is in part due to the idealised topography, and also the complete absence of any background flow. It would be interesting to know how more complex terrain and / or a weak (but non-zero) wind would modify the results. Is this something you have considered? It would at least be worth commenting on.

Authors:
In this study, our focus is on canopy flow driven purely by thermal and orographic forcing. The idealized simulations are better to interpret different mechanisms that regulate the small-scale turbulence in canopy layer. We also have the curiosity to see how synoptic weather condition (background wind) affects the local canopy flow. We have just finished one more manuscript in which we did three-dimensional simulation of canopy flows with real vegetation and terrain under three different synoptic weather conditions. The results are much different because of interactions between background wind and local slope wind, redistributed heat flux by heterogeneous vegetation and topographic effects.

Reviewer #2
6) These simulations should be entirely symmetric (in fact you state they are at the start of section 3), but the streamlines in figure 1 are not symmetric. Why? What breaks the symmetry?

Authors:
The asymmetry is very tiny. We believe that this is okay because any small numerical

perturbation will cause the solution to instead go towards a stable asymmetric solution, the so-called Coanda effect. The most important is that we have tested the grid-independence.

Reviewer #2

7) Section 3.2. The pooling of cold air at the bottom of the slopes seems to be important in decelerating the flow. How is this influenced by the model geometry? Would the results differ with a wider domain? Did you test sensitivity to this? How would this translate to the real world with 3-d valleys? (As an aside, in order to reach a steady state, the cold air must go somewhere. I assume that there is outflow from the lateral boundaries?)

Authors:

We agree with the reviewer that the pooling of cold air at the bottom of slopes decelerates the down-slope flow. We found that on the same elevation, the temperature inversion on steeper slope is slightly stronger than that on gentler slope. We didn't make the experiments of wider domain. But we believe that the wider domain would not affect the cold air pool in our single hill case because of the long enough lateral domain and open lateral boundary. The 3-D valleys will lead to a different cold air pool because a closed valley would block the ventilation of cold air, which cannot be interpreted by our single hill modeling.

Reviewer #2

8) End of section 3.2. The effects of slope here controlling whether flow penetrates to the bottom of the canopy or not are interesting. You imply that this is due to the buoyancy force, which is in part true. I think there is more to it than that though. Even on a shallow slope there is a downslope drainage flow, and so by continuity some air must be drawn down deep into the canopy to compensate. I think this needs a more careful analysis to explain what is happening. It may also be amenable to some scaling analysis to show how the slope effect scales? Similarly I do not fully understand what causes the differences in the regions of baroclinicity at the bottom of the slope, and hence the differences in circulation. In particular the upslope flow in the mid canopy over the gentle slope seems odd. Is there any observational evidence of this? How much is this controlled by the cooling of cold air? These are the kind of details which may be sensitive to the turbulence parametrisation - which again comes back to the question of how well validated the model is.

Authors:

We agree with reviewer that the downslope drainage flows even occur on shallow slope. The drainage flow is controlled by competition of hydrodynamic pressure gradient $F_{hd} = d\Delta p/dx \approx U^2H/L^2$ and buoyancy force (or hydrostatic pressure gradient) $\mathrm{F_{hs}} = \mathrm{g}(\Delta\theta/\theta_0)\ \sin\alpha \approx \mathrm{g}(\Delta\theta/\theta_0)\mathrm{H/L}$, where α is the slope angle, $\Delta\theta$ is the potential temperature difference between the ambient air and the colder slope flow, θ_0 is the ambient potential temperature (Belcher et al., 2008). Froude number Fr=U/NL=$\sqrt{F_{hd}/F_{hs}}$, where $N = (g/\theta_0)d\theta/dz$, L is the hill length scale. When Fr <<1, the hydrostatic pressure gradient is large enough to initiate drainage flow. We have two schematic figures below showing temperature and pressure field (Figure 1), which indicate the region of baroclinicity. The canopy winds in the baroclinicity region shift direction from deep drainage flow to shallow drainage flow (Figure 2). For the concern of the upslope flow in the mid canopy, we could not find a quantitative analysis of it. However, we did a smoke experiment in calm conditions on a vegetated slope in 2011 and found this updraft motion (Figure 3). The link of our experiment video is https://www.youtube.com/watch?v=ljZ88QZWPw0.

Reference:

Belcher, S. E., Finnigan, J. J., Harman, I. N.: Flows through forest canopies in complex terrain. Ecol. Appl. 18, 1436-1453, 2008.

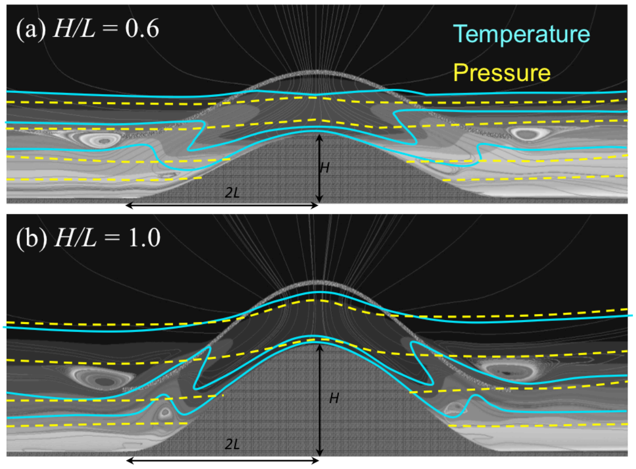


Figure 1 The temperaure and pressure fields indicating the baroclinicity regions.

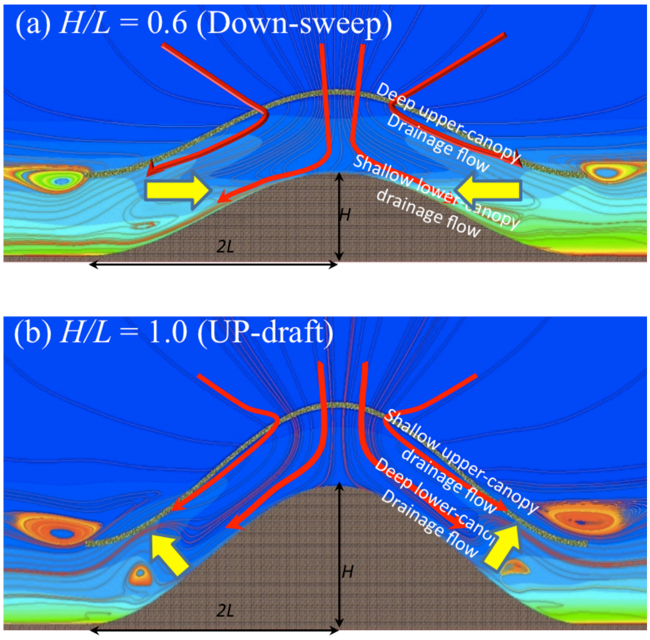


Figure 2 The drainage flow pattern and the direction of wind shifting in canopy

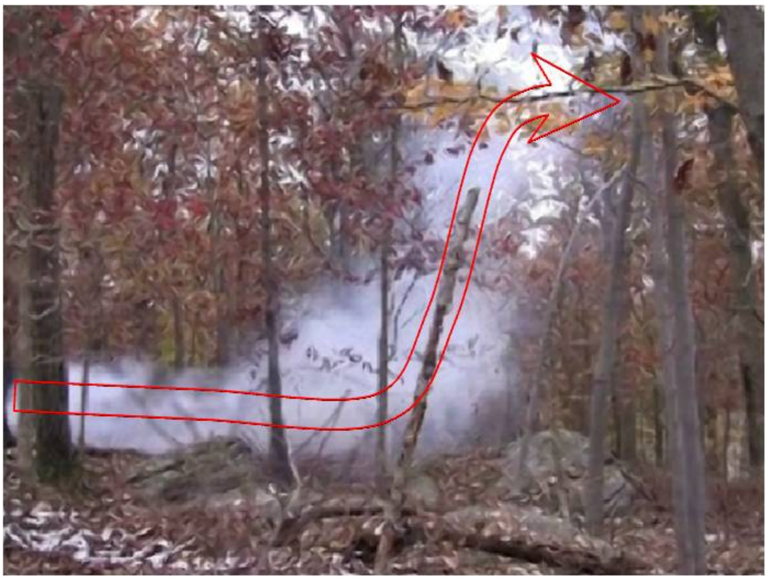


Figure 3 Updraft motion detected in forest canopy in the early morning by smoke experiment.

Reviewer #2

9) I wasn't entirely clear from the text whether the canopy is only on the slope and not on the flat ground. This seems to be implied by Figure 2. This may have a significant effect in controlling what happens at the bottom of the slope and is an added complication. In particular, I wonder if this controls the vortices seen near the bottom the slopes. Did you try experiments with a fully forested domain?

Authors:

In this simulation, canopy is covering the slope. We have the description in section 2.1. The vortices near the bottom of slopes are caused by the edge of canopy. Strong shear and large TKE are shown in vortex regions. We did not make experiments with fully forested domain.

Reviewer #2

10) Section 3.3 I found to be rather unsurprising. The results seem entirely consistent with the observed mean flow and much of the section is just repeating other studies.

Authors:

The major purpose of this study does not focus on turbulence in very stable stratification but complexity of local flows resulting from interactions between canopy, terrain, and cooling effect. We believe that these simulations for difficult conditions in eddy-flux measurements will be useful for FLUXNET community to understand their advection problems and issues. So we are happy that we can reproduce the shear stress and turbulent heat flux profiles in agreement with measurements and qualitative expectations.

Reviewer #2

11) Section 3.4 is potentially interesting, but given the questions raised above about the RNG scheme and how well it has been validated in stable / canopy flows it is hard to have too much faith in the conclusions, particularly about the importance of the pressure term. Other observational studies do seem to suggest this is important though and it would be interesting to pin this down.

Authors:

We agree with the reviewer that pressure perturbation is uncertain to TKE budget. We have added the discussion of the uncertainties in calculating and measuring pressure perturbation in 3.4.

Reviewer #2

12) In the concluding remarks you say that no comparison with field observations is possible. There may not be detailed measurements of all the relevant terms in the TKE budget, but there are (limited) measurements of mean flow and turbulent fluxes from multi-tower, multi-level experiments as stated. The model could, and should, be compared with these to validate it. The other source of data is wind-tunnel experiments which are generally better observed and more controlled. There have been recent experiments at the CSIRO Pye Lab wind tunnel with stable canopy flows. These are not yet published, but if and when the data is published this would be another valuable source of validation data.

Authors:

We agree with the reviewer that the numerical model should be validated. However, we still cannot get enough information from the limited measurement data. Current analysis of the stable canopy flow lacks detailed description of either canopy structure, topography, or atmospheric stability. Wind tunnel experiments would be a good way to validate our idealized model due to controlled setting. We know that there are research groups that made wind tunnel experiments of canopy flow. But these experiments are under neutral or unstable condition (Endalew et al., 2009; Segalini et al., 2013; Poggi et al., 2004). We are happy to hear that CSIRO Pye Lab has done wind tunnel experiments of canopy flow under stable condition. We look forward to seeing these data published and compared with our model results.

Endalew, A. M., Hertog, M., Delele, M. A., Baetens, K., Persoons, T., Baelmans, M., Ramon, H., Nicolai, B. M., Berboven, P.: CDF modeling and wind tunnel validation of airflow through plant canopies using 3D canopy architecture. INT. J. *HEAT. FLUID* FL. 30, 356-368, 2009.
Segalini, A., Fransson, J. H. M., Alfredsson, P. H.: Scaling laws in canopy flows: A wind-tunnel analysis. Bound.-Layer Meteorol. 148, 269-283, 2013.
Poggi, D., Porporato, A., Ridolfi, L., Albertson, J.D., Katul, G. G.: The effect of vegetation density on canopy sub-layer turbulence. 111, 565-587, 2004.

Reviewer #2

Minor comments

1) p28488, line 14. You state that the benefit of the RNG model is the lack of any tuneable empirical parameters. This is not true as the model contains 7 empirical constants (see p28493, lines 14-15).

Authors:

The empirical parameters are derived from Renormalized group method. The parameters are numerically derived and not subject to experimental adjustment (Yakhot and Orszag, 1986 ;Yakhot, 1988). We rephrased this sentence to make the statement more accurate.

Yakhot, V. and 5 Orszag, S. A.: Renormalization group analysis of turbulence, Phys. Rev. Lett., 57, 1722–1724, 1986.
Yakhot, V.: Propagation Velocity of Premixed Turbulent Flames, Combustion Science and Technology, 60:1-3, 191-214, 1988.

Reviewer #2

2) p28490, line 9. What is $\theta\infty$? How does this differ from θ_{00} defined on the previous page?

Authors:

$\theta_0(z) = \theta_{00} + \gamma z$ is defined as the ambient temperature, where θ_{00} is surface temperature and γ *is* positive temperature gradient . θ_∞ is buoyancy reference temperature. It is specified as the mean temperature of the domain. The buoyancy force in eq. 3 is a linear function of fluid thermal expansion and the local temperature difference compared with buoyancy reference temperature.

Reviewer #2

3) Eqs 7-9. Why is this form of the drag force taken rather than the more usual F = CDau|U|?How does this compare?

Authors:

In numerical modeling of canopy flow, the drag coefficient Cd is usually set as a constant value for uniform canopy (Patton et al., 2003; Dupont et al., 2008; Katul et al., 2006). We applied the leaf area density profile from a real forest without a direct measurement of drag coefficient. Thus, we use the empirical relationship between leaf area density and resistance coefficient (eq. 8 and 9).

Reference:

Katul, G. G., Finnigan, J. J., Poggi, D., Leuning,R., Belcher, S. E.: The influence of hilly terrain on canopy–atmosphere carbon dioxide exchange. Boundary-Layer Meteorology 118, 189–216.

Dupont, S., Brunet, Y., Finnigan, J.: Large-eddy simulation of turbulent flowover a forested hill: validation and coherent structure identification. Q .J. Roy. Meteorol. Soc. 134, 1911–1929, 2008.

Patton, E.G., Sullivan, P.P., Davis, K.J.: The influence of a forest canopy on top=down and bottom-up diffusion in the planetary boundary layer. 129, 1415-1434, 2003.

Reviewer #2

4) Section 2.3. What is the Prandtl number taken as? No value is given in the text.

Authors:

The value of Prandtl number is given in section 2.2 (line 10-11 page 28495).

Reviewer #2

5) p28495, line 19. Should be "The Richardson number..."

Authors:

We corrected.

Reviewer #2

6) p28496, line 1. Should be "with" not "With" at the start of the line.

Authors:

We corrected. Thanks.

Reviewer #2

7) p28496. There are several references to subfigures 4a - 4f. Figure only contains 4 subfigures though, and these are not actually labelled. Do you mean the profiles a-f in the figures? If so, this is a very confusing notation. Please change.

Authors:
We made two corrections (Fig. 4 locations a-f) and (Fig. 4 locations e and f). Thanks.

Reviewer #2
8) p28496, line 23. Why is the depth of the secondary super stable layer "due to strong temperature inversion"? Is the strong inversion not just part of the super stable layer? I found this sentence confusing.

Authors:
We wanted to emphasize that the super stable layer is deep on the lower slope consistent with deep and strong inversion layer. We rephrased this sentence. Thanks.

Reviewer #2
9) p28497, lines 1-2. Do you really mean stronger entrainment at the summit? Why just there? I interpret entrainment to be mixing due to turbulence. Is it not the mean flow, i.e. air being drawn down into the canopy over the summit to balance the downslope flow which is suppressing the secondary super-stable layer?

Authors:
We corrected the word ‘entrainment’ and add a reference.

Reviewer #2
10) p28497, lines 8-9. I don’t see the point of this sentence. Previous studies have already observed the stable canopy layer and linked it to decoupling. How does this clarify that?

Authors:
We replace ‘clarify‘ with ‘explain’.

Reviewer #2
11) p28497, line 10. "van Gorsel" not "Gorsel".

Authors:
We corrected. Thanks.

Reviewer #2
12) p28497, line 21. "from the terrestrial ecosystem."

Authors:
We corrected. Thanks.

Reviewer #2
13) p28497, lines 25-29. I found these sentences rather unclear. The phrases "under- goes direction shift within canopy." is odd. The English needs improving here to make the meaning clearer.

Authors:
We rephrased this sentence.

Reviewer #2

14) p28498, line 10. What do you mean by "lateral sides"? This sentence is unclear.

Authors:
We rephrased this sentence.

Reviewer #2
15) p28498, lines 10-11. This sentence is also very unclear. Why is the sinking motion diverted? What do you mean by top canopy layer?

Authors:
Here we mean that the sinking motion changed its direction from descending into the canopy to following the shape of canopy. We rephrased this sentence.

Reviewer #2
16) p28498 and figure 5. Again confusing whether Fig 5.d is referring to a subfigure or the location of a particular profile. Figure 5 seems to contain two subfigures labelled a), b) etc. I would suggest using a different naming convention for the profile locations to avoid confusion.

Authors:
We have two panels of subfigures in Figure 5, clarified as streamwise wind velocity and vertical wind velocity.
We corrected the captions of the figures to indicate the labeled (a)-(f) and their locations on the slope.

Reviewer #2
17) p28498, line 19 and figure 5. How can the velocity maximum be below the lower stable layer? The model description implies a no-slip lower boundary (the roughness length is given), but Figure 5 seems to show a non-zero velocity at the surface, in fact a velocity maximum occurs there. How can this be? Is the lower boundary actually free slip? Please explain, and if free slip then justify this choice.

Authors:
Our model is set with no-slip boundary. The wind velocity is not zero at the surface because the centers of the bottom grid cells are not exactly at the surface. We added this note to the figure caption.

Reviewer #2
18) p28498, lines 24-25. This sentence doesn't make sense. What do you mean by "...determines the shift direction within canopy."

Authors:
We rephrased this sentence. Thanks.

Reviewer #2
19) p28505, lines 3-4. ".. with additional strong non-linear terms". What additional terms? Do you mean the RNG turbulence closure? I don't see the point of this sen- tence anyway.

Authors:
Corrected by rewriting the section of concluding remarks.

Reviewer #2
20) p28505, lines 13-14. "... at the ultra-short wave scale in the whole spectrum of atmospheric turbulence study." This sentence doesn't make sense. Do you mean you are looking at very small-scale flows?

Authors:
Corrected by rewriting the section of concluding remarks.

Reviewer #2
21) Figure 1. Can you mark the canopy on this figure? Figure is not very good quality, and is difficult to read when printed.

Authors:
We updated figure 1. Thanks.

Reviewer #2
22) Figure 2. Caption mentions green dashed lines, but lines appear to be white to me?

Authors:
The white dashed line is the top of canopy. The cyan blue lines highlight the 'fish-head' temperature distribution. We corrected the caption. Thanks.

Reviewer #2
23) Figure 3. The second sentence in the caption is very poorly phrased. When you say "... which is normalized by the half length scale L" you presumably mean the locations. I would split this sentence and say "The locations of the size sections are labelled as (a-f). Horizontal distances are normalized by the half length, L, of the hill." or something similar. The caption mentions a green curve. It looks more like light blue to me?

Authors:
We corrected the caption. Cyan blue is more appropriate to name the color in figure 3. Thanks.

Reviewer #2
24) Figure 4. Resolution is not sufficiently good when printed. Are these bitmaps rather than vector graphics?

Authors:
We updated Figure 4. Thanks.

Reviewer #2
25) Figure 9 and 10. Plot the y-axis on the edge of the figures, not on the x = 0 line for clarity.

Authors:
We updated Figures 9 and 10. Thanks.

-----------------------------Reviewer #3 ----------------------------

Reviewer #3
This study investigated 2-D stably stratified mean and turbulent flows above and within plan canopies in complex terrain by using the Renormalized Group (RNG) k-e turbu- lence model, an important topics in studying canopy micrometeorological processes. The stable stratification was generated by imposing persistent constant heat flux at the ground surface and linearly increasing cooling rate in the upper canopy layer. The terrain-induced influence in dynamic part was carried out by using a gentle hill and a steep hill. Mean and turbulence flows were then characterized by analyzing profiles of mean and turbulence quantities at different locations over the hill slopes. The model approach was sound enough for such a study; the analysis was comprehensive; the information was updated; and the results were unique. I would recommend its publica-tion with minor revisions.

Authors:
We are grateful for reviewer's positive comments!

Reviewer #3
(1). Introduction: Might be good to mention limitations of linear analysis models in studying stratified canopy flows over complex terrain.

Authors:
We added the limitations of linear analysis models to Introduction.

Reviewer #3
(2). Lines 5-10 on page 28487. The last half sentence seems not belong to this speculation.

Authors:
We rephrased this sentence.

Reviewer #3
(3). Lines 13-17 page 28488. Rewrite this long sentence

Authors:
We rephrased this long sentence.

Reviewer #3
(4). Lines 15-16 page 28489. Justify if these rates are commonly observed rates.

Authors:

The ambient temperature gradient of 6 K km^{-1} is appropriate to represent the typical nocturnal inversion strength, although the measured gradient of ambient temperature can be larger or smaller depending on surface condition and background wind. The data from Vertical Transport and Mixing (VTMX) experiment showed the gradient of ambient potential temperature of 10 K km^{-1} in the Utah's Salt Lake Valley in October 2000. Texas Air Quality Study II in August and September 2006 revealed inversion gradient at peak inversion strength of 8-16 K km^{-1}. The Slope Experiment near La Fouly (SELF-2010) showed a potential temperature gradient of 5.8. km^{-1} in the Alpine Slope in Val Ferret, Switzerland. On the onset of downslope flow, the gradient of potential temperature can be as low as 4 K km^{-1}, which is measured in September, Australia.

Reviewer #3
(5). Lines 16-17 page 28490. Was the surface cooling included in the boundary conditions instead of a source term?

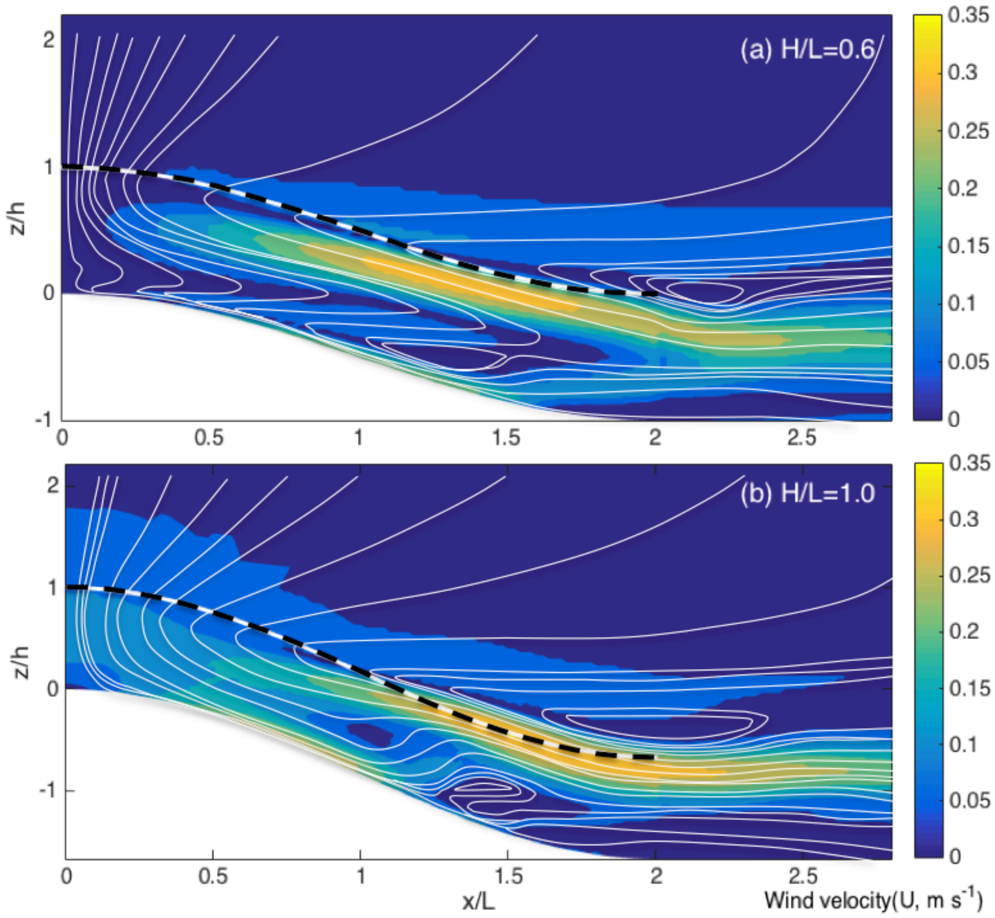
Authors:
Yes, the surface cooling is boundary condition.

Reviewer #3
(6). Page 28495. Explain in more detail why a strong inversion layer was found across the lower jar.

Authors:
Three contributions can be contributed to the strong inversion across the lower jaw: (1) the air in the lower jaw is initially colder than the air in upper jaw, (2) the slope surface has a stronger cooling rate (-15Wm^2), (3) the cool dense air near slope surface is flowing down the slope and accumulates on the cooler slope flow below.

Reviewer #3
(7). Figure 1. Label the streamlines with wind speeds so as to give a clue about the distributions of wind fields.

Authors:
We have tried to add the wind speed as a background in the figure 1 as suggested by the reviewer #3 but the new figure looks a little bit overcrowded. We agree that the addition of wind speed makes the dynamics more clear, so we added the following figure as a new figure (Figure 6) in the revision.



Reviewer #3
(8). Section 3.1. How the dynamics influences the thermal structures is analyzed in some degree. A clear summary might be necessary to help readers construct a clear structure about these linkages.

Authors:
We summarized the interaction of the flow dynamics and thermal structure in the Concluding remarks.

Reviewer #3
(9) How much did the flow structures impact the formation of the primary/secondary stable layers?

Authors:
The radiative cooling in the canopy and slope surface is the primary driving force of the flow structure and temperature inversion. The flow structure with upper-canopy drainage flow (UDF) and lower-canopy drainage flow (LDF) determines the location of super stable layers at levels with $\partial\bar{U}/\partial z = 0$. The drainage flows intensify the temperature inversion $\partial\bar{\theta}/\partial z$ down the slope, thus intensify the stability of super stable layers.

Reviewer #3
(10). Explain more why UDF was deeper than the LDF on the gentle slope.

Authors:
The drainage flow is controlled by competition between hydrodynamic pressure gradient $F_{hd} = d\Delta p/dx \approx U^2 H/L^2$ and buoyancy force (or hydrostatic pressure gradient) $F_{hs} = g(\Delta\theta/\theta_0)\ \sin\alpha \approx g(\Delta\theta/\theta_0\ )H/L$, where α is the slope angle, Δθ is the potential temperature difference between the ambient air and the colder slope flow, θ_0 is the ambient potential temperature (Belcher et al., 2008). Froude number Fr=U/NL=$\sqrt{F_{hd}/F_{hs}}$, where $N = (g/\theta_0)d\theta/dz$, L is the hill length scale. When Fr <<1, the hydrostatic pressure gradient is large enough to initiate drainage flow. On the slope, the flow is flowing along the shape of canopy layer forming the UDF. The LDF is formed by the air drainage from the hillcrest. Since the intensity of drainage flow is proportional to slope angle, the sinking motion is weaker on the gentle slope. Less air sinking to the deep canopy results in LDF shallower than UDF.

Reference:

Belcher, S. E., Finnigan, J. J., Harman, I. N.: Flows through forest canopies in complex terrain. Ecol. Appl. 18, 1436-1453, 2008.

Reviewer #3
(11). It looked that the TKE was larger on the gentle slope than on the steep slope. Was that due to the small wake production?

Authors:
The larger TKE on gentle slope is mainly due to the supply of TKE transport of pressure perturbation to compensate for TKE loss by buoyancy consumption. The uncertainty remains because the pressure perturbation is calculated as a residue term. There is evidence that pressure perturbation is important to provide TKE in lower canopy under unstable condition (Dwyer et al., 1997). We do not have data to validate if buoyancy can induce larger pressure perturbations under stable conditions.

Reference:

Dwyer, M. J., Patton, E. G., Shaw, R. H.: Turbulent kinetic energy budgets from a large-eddy simulation of airflow above and within a forest canopy. Bound.-Layer Meteor. 84, 23-43, 1997.

Reviewer #3
(12). Given the closely packed iso-streamlines in figure 1, it seems that horizontal advection played a role in TKE. Please explain.

Authors:
The advection transport of TKE is relatively large near the top of canopy than other heights, especially at the mid- and lower-slope. However, advection of TKE is not important when compared with buoyancy and wake production in our stably stratified canopy flow.

Reviewer #3
(13). Section 4 could be largely shorted and just emphasize the major conclusions

Authors:
We shortened the section 4 and summarized the interactions of flow dynamics and thermal structure for comment (8).

Reviewer #3

(14). It could be better to enlarge Figure 4.

Authors:

Figure 4 is updated.

Stably Stratified Canopy Flow in Complex Terrain

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Abstract

Stably stratified canopy flow in complex terrain has been considered a difficult condition for measuring net ecosystem–atmosphere exchanges of carbon, water vapor, and energy. A long-standing advection error in eddy-flux measurements is caused by stably-stratified canopy flow. Such a condition with strong thermal gradient and less turbulent air is also difficult for modeling. To understand the challenging atmospheric condition for eddy-flux measurements, we use the Renormalized Group (RNG) k-ε turbulence model to investigate the main characteristics of stably stratified canopy flows in complex terrain. In this two-dimensional simulation, we imposed persistent constant heat flux at ground surface and linearly increasing cooling rate in the upper canopy layer, vertically varying dissipative force from canopy drag elements, buoyancy forcing induced from thermal stratification and the hill terrain. These strong boundary effects keep nonlinearity in the two-dimensional Navier-Stokes equations high enough to generate turbulent behavior. The fundamental characteristics of nighttime canopy flow over complex terrain measured by the small number of available multi-tower advection experiments can be reproduced by this numerical simulation, such as: (1) unstable layer in the canopy and super-stable layers associated with flow decoupling in deep canopy and near the top of canopy, (2) subcanopy drainage flow and drainage flow near the top of canopy in calm night , (3) upward momentum transfer in canopy, downward heat transfer in upper canopy and upward heat transfer in deep canopy and (4) large buoyancy suppression and weak shear production in strong stability.

Keywords: Canopy flow, complex terrain, stable stratification, Richardson number, turbulent kinetic energy, CFD, RNG k-ε model

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 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 (Edburg et al., 2010), and others.

The typical patterns of forest 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:

$$l_m = \kappa \left| \frac{dU/dz}{d^2U/dz^2} \right|$$

where κ 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, 2007a).

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., 2010; 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 characterize considering only a super-stable layer. Canopy structure (quantified by 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. We speculate that interaction between thermal stratification and terrain slopes and vegetation canopy may result in multiple super-stable layers. The complicated thermal and flow patterns cause difficulties in understanding the mechanisms and rates of exchange of mass and energy between the terrestrial biosphere and the atmosphere (Alekseychik et al., 2013; Burn et al., 2011; Yi et al., 2005).

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, the analytical model (Finnigan and Belcher, 2004) is limited to neutral condition and hills of gentle slope. The analytical model is developed based on the linearized perturbation theory for the flow over a rough hill (Jackson and Hunt, 1975), which assumes that the mean flow perturbations caused by the hill are small in comparison to the upwind flow. Poggi and Katul (2007b) and Ross and Vosper (2005) have shown that the analytical model fails to model the flow pattern on dense canopies on narrow hills. Third, 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 and its related scalar transport in neutral and unstable cases (Shen and Leclerc, 1997; Wang, 2010; Mao et al., 2008). 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 resolution is high enough, 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-ε turbulence model to investigate stably stratified canopy flows in complex terrain. The RNG k-ε turbulence model was developed by Yakhot and Orszag (1986a) using the renormalized group methods and prescribes the turbulent length scale related to transport of turbulent kinetic energy and dissipation rate (Yakhot and Orszag, 1986b; Smith and Reynolds, 1992). Compared to standard k-ε turbulence model, the numerically derived parameters are not subject to experimental adjustment in RNG k-ε turbulence model. The rate of strain term in the dissipate transport equation is important for treatment of flows in rapid distortion limit, e.g. separated flows and stagnated flows (Biswas, and Eswaran, 2002) which commonly occur in vegetated hilly terrain. The initial successes in applying the RNG k-ε turbulence model to generate airflows in hilly terrain have been demonstrated by Kim and Patel (2000) and Xu and Yi (2013).

2 Method

2.1. 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 is 100m long covered with a 15m tall homogeneous forest canopy, which extends from 650m of the 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 0.01m.

In this study, the topography is specified with a ridge-like sinusoidal hill, infinite in the unsimulated third dimension. The shape function of the hill in 2D is defined as

$$H(x) = \frac{H}{2} cos\left(\frac{\pi x}{2L}\right) + \frac{H}{2} \tag{1}$$

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 (canopy height $h = 15$m) is designed horizontally homogeneous along the slope. Leaf area density profile $a(z)$ is specified as values from observation of an actual forest (Yi et al., 2005) with the maximum leaf area density at about 8m. Leaf area index (LAI) is 3.3. The ambient temperature is $\theta_0(z) = \theta_{00} + \gamma z$, where $\theta_{00} = 288K$, is the potential temperature at $z = 0$, γ is ambient lapse rate, set to -6°C km^{-1}. The cooling rate at ground surface is set to -15 Wm^{-2}. Since we are most interested in calm night-time conditions, no wind in the domain is initially specified. The fixed pressure boundary condition (open boundary) is applied to lateral boundaries and top-boundary, where the pressure is close to 0.0 Pa, relative to the external pressure.

2.2. Conservation of mass and momentum

The flow is assumed to be steady and the Boussinesq approximation is applied. The mass, momentum, and energy balance equations in the canopy sub-layer can be written as:

$$\frac{\partial \bar{u}_j}{\partial x_i} = 0 \tag{2}$$

$$\bar{u}_j \frac{\partial \bar{u}_i}{\partial x_j} = -\frac{1}{\rho}\frac{\partial P_*}{\partial x_i} + \nu \frac{\partial^2 \bar{u}_i}{\partial x_i x_j} - \frac{\partial}{\partial x_j}\left(\overline{u_i' u_j'}\right) - g_i \beta(\bar{\theta} - \theta_\infty) - F_{Di} \tag{3}$$

$$\bar{u}_j \frac{\partial \bar{\theta}}{\partial x_j} = \Gamma \frac{\partial^2 \bar{\theta}}{\partial x_i x_j} - \frac{\partial}{\partial x_j}\left(\overline{\theta' u_j'}\right) + \frac{1}{\rho c_p} Q_{source} \tag{4}$$

where $\bar{u}_i$ and $\bar{u}_j$ are the mean velocity components along x_i and x_j direction, respectively, $\bar{\theta}$ is the mean potential temperature, u_i', u_j' and θ' are the fluctuations from their mean value $\bar{u}_i$, $\bar{u}_j$ and $\bar{\theta}$, ρ is the air density, ν is kinematic viscosity of air, P_* is the deviation of pressure from its reference value, β is the thermal expansion coefficient of air, and θ_∞ is the reference temperature, g_i is the gravity acceleration in i direction, $\Gamma = \nu/P_r$ is thermal diffusion coefficient, turbulent Prandtl number P_r is 0.5 in canopy layer and 1 above the canopy. P_r = 0.5 is close to the values used in large-eddy simulations of stably stratified atmospheric boundary layer turbulence (Basu and Porté-Agel, 2006; Stoll and Porté-Agel, 2008). In most of the region above the canopy (except very near the top of canopy), turbulence is very weak. In this region, molecular effects are dominant, especially in conditions without synoptic wind. Q_{source} is the energy source. When the atmosphere is stably stratified, $Q_{source} < 0$ indicating radiative cooling of the canopy elements and ground surface. The constant cooling rate at the surface can drive a steady state stable boundary layer on flat and sloped terrain (Brost and Wyngaard, 1978), so we set $Q_{source} = 0$ in the lower canopy layer (0-8m) and then linearly decreased to -8 Wm^{-3} at the top canopy layer. The thermal conditions are efficient to drive fully developed turbulent flows, according to dimensional analysis of the bulk Reynolds number :

$$Re_b = \frac{h_i U}{\nu} = \frac{O(10^1 m)\times O(10^{-1} ms^{-1})}{O(10^{-5} m^2 s^{-1})} = O(10^5)$$

where h_i is the depth of boundary layer, U is bulk velocity and ν is kinematic viscosity.

The steady state assumption is satisfied with condition proposed by Mahrt (1982),

$$\mathrm{F}\widehat{\mathrm{H}}/\widehat{\mathrm{T}} \ll 1 \tag{5}$$

where F is the Froude number, $\widehat{\mathrm{H}}$ is the ratio of the average flow depth H to the surface elevation drop ΔZ_s, and $\hat{T}$ is the ratio of the time scale T to the Lagrangian time L/U. The Froude number is defined as

$$F = U^2 / \left(g \frac{\Delta\theta}{\theta_0} H \right) \tag{6}$$

where U is downslope velocity scale (= $O(10^{-1})$ m s^{-1}), g is gravity acceleration (= 9.81m s^{-2}), $\Delta\theta$ is scale value for potential temperature deficit of the canopy layer (= $O(10^0)$ K), θ_o is the basic state potential temperature (= $O(10^2)$ K), H is the flow depth scale, chosen to be the depth of significant temperature deficit which coincides with the layer of enhanced thermal stratification (= $O(10^1)$ m). In this simulation setting, $F = O(10^{-2})$. $\widehat{H} = H/\Delta Z_s$, where $\Delta Z_s = L \sin\alpha$, L is downslope length scale (= $O(10^1)$ m), $\sin\alpha(\%) = O(10^1)$. Thus, $\widehat{H} = O(10^0)$. $\hat{T} = TU/L$, where $T = O(10^4)$ s is suggested by Mahrt (1982) to represent the order of magnitude of temporal accelerations associated with the diurnal evolution of drainage circulations. In our simulation, $\hat{T} = O(10^2)$. Thus, $F\widehat{H}/\hat{T} = O(10^{-4}) \ll 1$.

F_{Di} is the drag force exerted by the canopy elements in i direction,

$$F_{Di} = \frac{1}{2} K_r u_i |U| \tag{7}$$

where K_r is the resistance coefficient, which is derived from an empirical relationship given by Hoener (1965),

$$K_r = \frac{1}{2}\left[\frac{3}{2\phi} - 1\right]^2 \tag{8}$$

where ϕ is porosity of the canopy layer, which can be obtained from leaf area density profile $a(z)$ (Gross, 1993),

$$\phi(z) = \frac{\sqrt{1 + 4a(z)} + 1}{2a(z)} \tag{9}$$

F_{Di} is zero above the canopy.

2.3. RNG *k*–*ε* model

The RNG model was developed by Yakhot and Orszag (1986a; b; Yakhot et al., 1992) using Re-Normalization Group (RNG) methods. The RNG *k*-*ε* turbulent model has been successfully applied in reproducing topographic and canopy related flows (Kim and Patel, 2000; Xu and Yi, 2013; Pattanapol et al., 2007).

In RNG *k*-*ε* model, the Reynolds stress in Eq. (3) and turbulent heat flux in Eq. (4), respectively, are solved by turbulent viscosity, as:

$$-\overline{u_i' u_j'} = \mu_t \left(\frac{\partial \bar{u}_i}{\partial x_j} + \frac{\partial \bar{u}_j}{\partial x_i}\right) - \frac{2}{3}\delta_{ij} k \tag{10}$$

$$-\overline{\theta' u_j'} = \mu_\theta \frac{\partial \bar{\theta}}{\partial x_j} \tag{11}$$

Where μ_t and $\mu_\theta = \mu_t / P_r$ are the turbulent viscosities of momentum and heat, respectively, δ_{ij} is Kronecker delta, and k is the turbulent kinetic energy.

RNG *k*-*ε* model assumes that turbulence viscosity in Eq. (10) is related to turbulence kinetic energy k (TKE) and dissipation ε:

$$\mu_t = \rho C_\mu \frac{k^2}{\varepsilon} \tag{12}$$

where k and ε are determined from the transport equations for k and ε; C_μ is a dimensionless constant.

The steady state transport equations for k and its dissipation ε are written as:

$$\bar{u}_i \frac{\partial k}{\partial x_i} = \frac{\partial}{\partial x_i}\left(\frac{\mu_t}{\sigma_k}\frac{\partial k}{\partial x_i}\right) + P_s + P_b + P_w + T_p - \varepsilon \tag{13}$$

$$\bar{u}_i \frac{\partial \varepsilon}{\partial x_i} = \frac{\partial}{\partial x_i}\left(\frac{\mu_t}{\sigma_\varepsilon}\frac{\partial \varepsilon}{\partial x_i}\right) + C_{\varepsilon 1}\frac{\varepsilon}{k}P_s - \rho C_{\varepsilon 2}\frac{\varepsilon^2}{k} - S \tag{14}$$

where P_s is shear production, given by:

$$P_s = \mu_t \frac{\partial \bar{u}_i}{\partial x_j}\left(\frac{\partial \bar{u}_i}{\partial x_j} + \frac{\partial \bar{u}_j}{\partial x_i}\right) \tag{15}$$

P_b is buoyancy production, given by:

$$P_b = -\mu_\theta g_i \beta \frac{\partial \bar{\theta}}{\partial x_i} \tag{16}$$

P_w is wake production caused by canopy elements as (Meyers and Baldocchi, 1991):

$$P_w = \bar{u}_i F_{Di} = \frac{1}{2} K_r |U| \bar{u}_i^2 \tag{17}$$

T_p is pressure collection term, which is calculated as residual of other TKE components, S is a volumetric source term which includes the rate-of-strain, given by:

$$S = \frac{C_\eta \eta^3 \left(1 - \frac{\eta}{\eta_0}\right) \varepsilon^2}{(1 + \beta_0 \eta^3) k} \tag{18}$$

$$\eta = \frac{k}{\varepsilon}\left[\frac{P_s}{\mu_t}\right]^{1/2} \tag{19}$$

where the empirical constants C_μ, σ_k, σ_ε, $C_{\varepsilon 1}$, $C_{\varepsilon 2}$, β_0, and η_0 are 0.0845, 0.7194, 0.7194, 1.42, 1.68, 0.012, and 4.38, respectively (Yakhot and Orszag, 1986a; b).

3 Results and discussion

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 (Fig. 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 Fig. 1) to explore the thermal and mechanical processes that govern the airflow structures.

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 (Fig. 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 sect. 3.2). In accordance with the fish-head temperature distribution, temperature profiles are shown in three layers (Fig. 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 (Fig. 3e, f). The temperature difference from the hill surface to the top of the canopy at the hill crest 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.

The 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,

$$Ri_g = \frac{(g/\bar{\theta})(\partial\bar{\theta}/\partial z)}{(\partial\bar{U}/\partial z)^2} \tag{20}$$

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),

$$Ri_b = \frac{g}{\bar{\theta}}\frac{\theta(z_2) - \theta(z_1)}{[U(z_2) - U(z_1)]^2}(z_2 - z_1) \tag{21}$$

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,

$$Ri_l = \frac{g}{\theta_{m,n}} \frac{(\theta_{m,n} - \theta_{m,n-1})(z_{m,n} - z_{m,n-1})}{(u_{m,n} - u_{m,n-1})^2 + (w_{m,n} - w_{m,n-1})^2} \quad (22)$$

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 (Fig. 1). Ri_l is found to be extremely large (~ 10^5) just above the canopy on the upper to middle slope (Fig. 4, locations a-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 (Fig. 4 locations e and 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. The deep secondary super stable layer is consistent with deep and strong temperature inversion layer where wind is stagnated. The absence of a secondary super stable layer on the summit could be explained by stronger mixing of warmer air from above-canopy, because stronger drainage flow promotes the penetration of warm air from aloft when cold air moves down the slope (Zängl, 2003). 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 dominate the resultant atmospheric profile.

The nocturnal stable canopy layer could be used to explain the occurrence of within- and above- canopy flows decoupling observed in prior studies. van 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 the 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.

3.2 Wind flow structures

Figure 1 shows that air above the canopy sinks and converges towards the hill and then shifts direction 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 2.6*L* in horizontal and 1.3*h* 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 lateral sides towards slopes of the hill. However, instead of further descending into the canopy, the sinking motion is diverted to follow the shape of the top-canopy layer as it reaches the top of the canopy (Fig. 1 and 6). 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 0.3 meters per second (m s^{-1}) at location (Fig. 5d and 6a) on the gentle slope and 0.35 m s^{-1} at location (Fig. 5e and 6b) 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 (Fig. 5d) for a gentle slope (0.18 m s^{-1}) and at location (Fig. 5c) for a steep slope (0.29 m s^{-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 regulates the direction of wind shifting within canopy (Fig. 1 and 6). On the gentle slope (H/L = 0.6), UDF is much thicker compared with LDF (Fig. 6a). 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, Fig. 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 Fig. 5 (down-sweep). When the slope is steep (H/L = 1.0), UDF is much shallower than LDF on the upper slope. Air in LDF accelerates on the upper slope (Fig. 5a-c), followed by deceleration and stagnation. The stagnated flow jumps perpendicularly from the 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 Fig. 1 and 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: $\mathrm{F_{hs}} = \mathrm{g}(\Delta\theta/\theta_0)\sin\alpha$, where α is the slope angle, $\Delta\theta$ is the potential temperature difference between the ambient air and the colder slope flow, θ_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 α 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.

3.3 Turbulent fluxes of momentum and heat

Fig. 7 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 (Fig. 7e, f), where the wake vortices are formed. Shear stress is positive above the canopy indicating a downward transfer of momentum that 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 on steep slope. 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 very common, occurring in stable atmospheric conditions (Zhang et al., 2010). The opposite sign in momentum transfer near the slope surface on steep and gentle slope can be explained by the strength of LDF on the slope.

The dominant positive turbulent heat flux, $-\overline{w'\theta'}$ indicates downward heat transfer above and within the canopy (Fig. 8). Heat transfer on the upper slope (Fig. 8a, 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 ($-\overline{w'\theta'} > 0$) 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 upward turbulent heat flux (Fig. 8) 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.

3.4 Turbulent Kinetic Energy (TKE) budget

In steady state, the TKE budget Eq. (13) can be written as:

$$0 = T_a + T_t + T_p + P_s + P_b + P_w - \varepsilon \qquad (23)$$

where 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 ε 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 (Fig. 9). 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 (Fig. 10 and 11). 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 and Leclerc (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.

The positive shear production P_s indicates the net transfer of kinetic energy from the mean flow to the turbulent component of the flow (Fig. 10 and 11). 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 (Fig. 10 and 11). TKE is weakly transported by turbulence upward near the canopy top ($T_t < 0$) and downward ($T_t > 0$) 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 (Fig. 10). 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 (Fig. 11d-f), but the relative significance of wake production becomes more prominent. On the upper slope (Fig. 11a-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.

4 Concluding remarks

Stably stratified canopy flows in complex terrain are investigated by a RNG turbulent model, with emphasis on strong boundary effects, including persistent thermal forcing from ground and canopy elements, damping force from canopy drag elements, and buoyancy effects from temperature stratification and topographic character.

The fundamental characteristics of nighttime canopy flow over complex terrain are addressed by this numerical simulation as follows:

(1) **Multiple layering of thermal stratification** The stability around the canopy is characterized by stratification with super stable layers above the top of the canopy and in the lower canopy, and an unstable layer within the canopy (Fig. 2, 3, 4).

(2) **Bifurcation of thermal-driven drainage flows** The drainage flow above the canopy is mainly driven by thermal stratification, being separated into two streams in the canopy: the upper-canopy drainage flow (UDF) layer and the lower-canopy drainage flow (LDF) layer (Fig. 1, 5, 6).

(3) **Buoyancy suppression of turbulence** The downward transport of momentum and heat flux in the canopy is reduced due to strong stability and reversed to be upward in the deep canopy (Fig. 7, 8). Buoyancy production suppresses turbulence significantly near the top of the canopy and in the deep canopy (Fig. 10, 11).

The thermal stratification and nocturnal drainage flows are interactive. The drainage flows, initiated by thermal stratification, result in the formation of super stable layers. In addition, the drainage flows intensify the temperature inversion down the slope, thus intensifying the stability of super stable layers. The properties of momentum and heat transfer may be related to the ‘shear-driven’ and ‘buoyancy-driven’ coherent structures that can lead to decoupling between the lower and upper canopy (Dupont et al., 2012). *Although* unstable layer is more likely to occur during the foliated period (Dupont et al., 2012) and may only have influence on the small-scale motions within the canopy (Jacob et al., 1992), the super-stable layers associated with flow decoupling have direct influence on a larger scale soil, within- and above-canopy exchange processes (Alekseychik, et al., 2013).

The canopy flow behavior presented in Fig.1 is expected to be measurable directly by multiple eddy-flux towers that are equipped with multi-level micrometeorological instruments (Feigenwinter et al., 2010; Baldocchi, 2008). Some turbulent exchange processes remain uncertain and require further study, including (i) how the varied vegetation structure, strength of background wind and ambient stability influence the within-canopy stratification and turbulence, and (ii) how the complicated flows regulate scalar transfer within the canopy and scalar exchange between the vegetation and atmosphere aloft.

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Figure 2. Contours of potential temperature (K) along the right slope: (a) $H/L = 0.6$; (b) $H/L = 1.0$. The 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 cyan blue dashed lines highlight the 'fish-head' temperature distribution.

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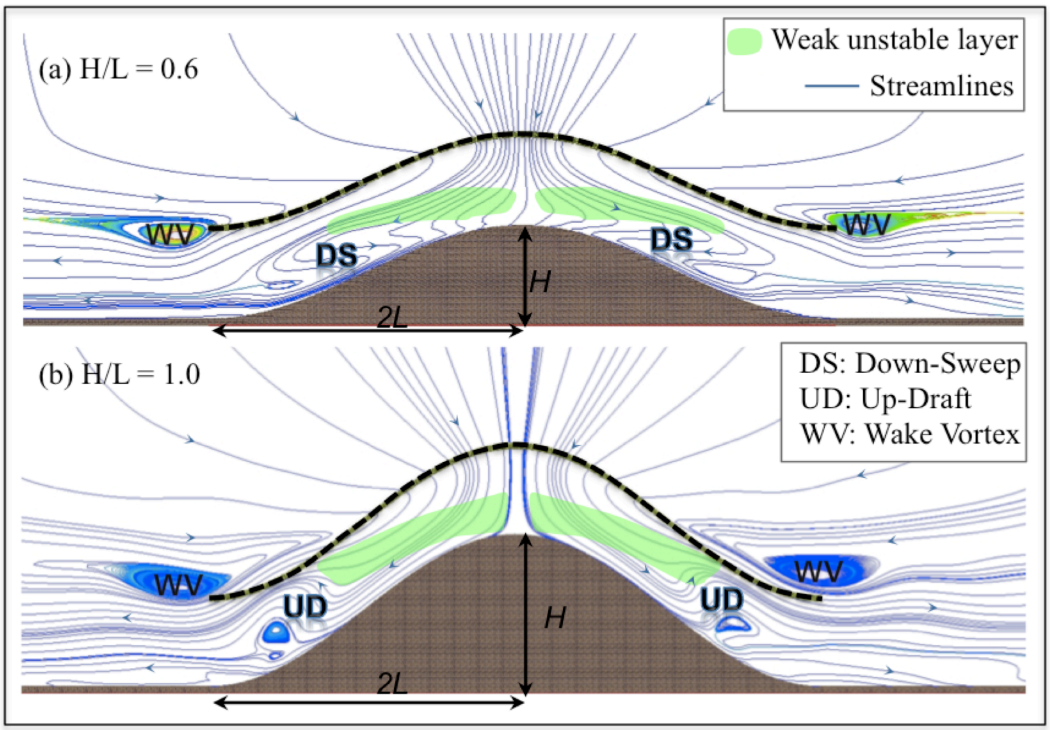

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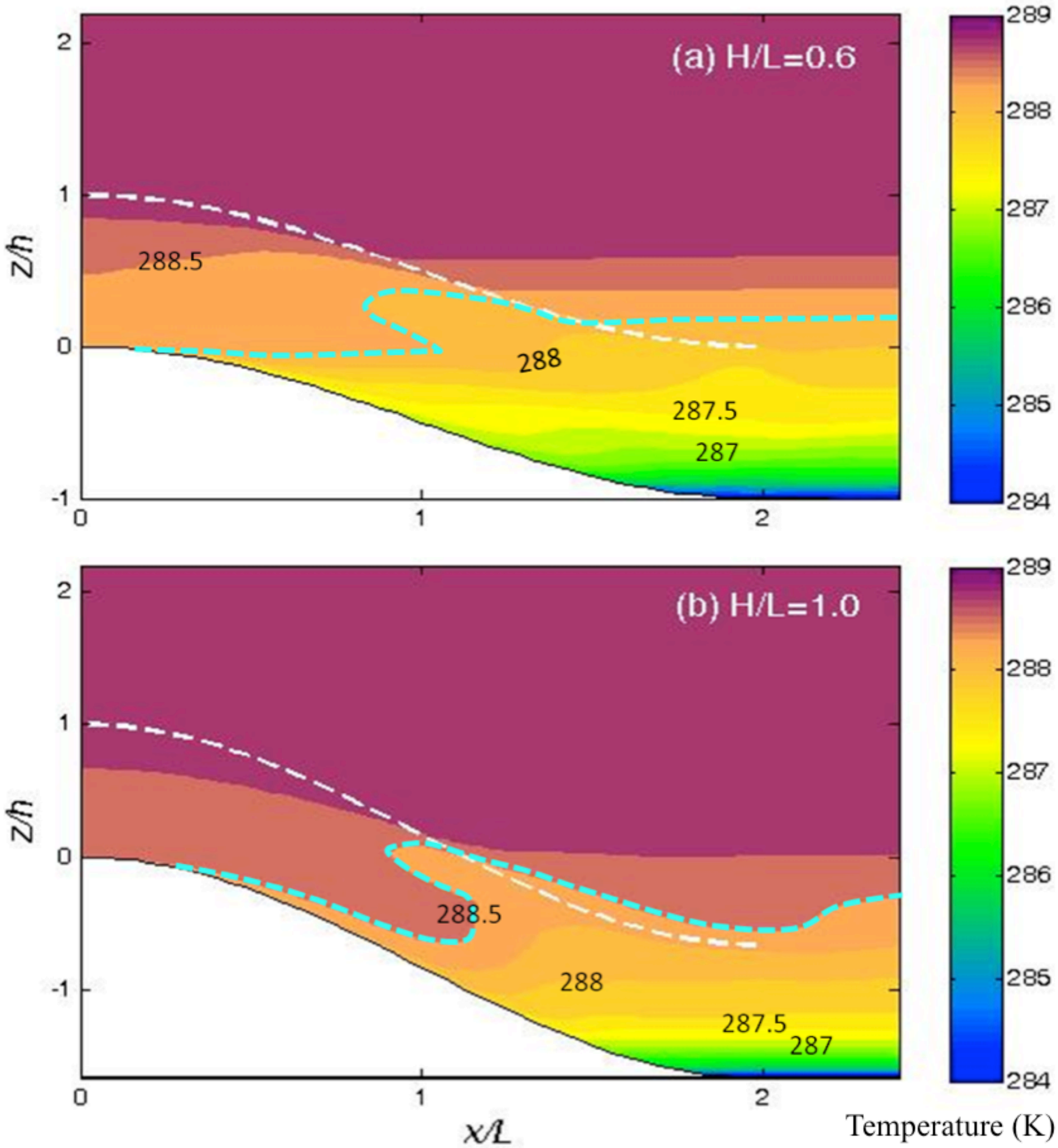

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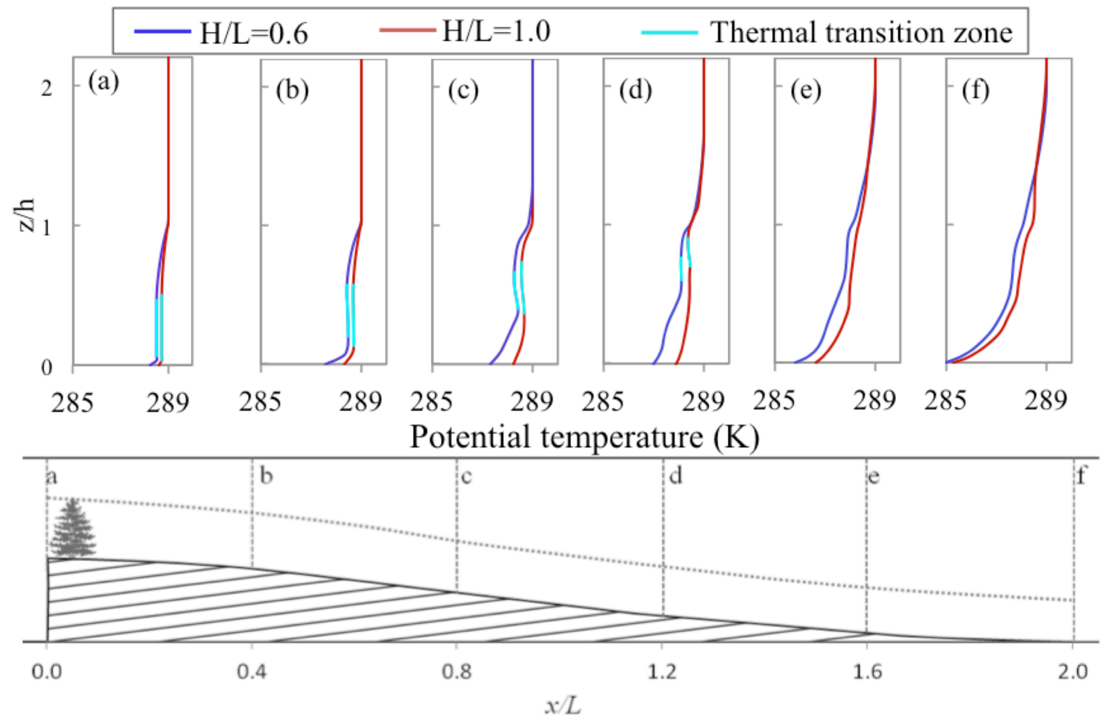

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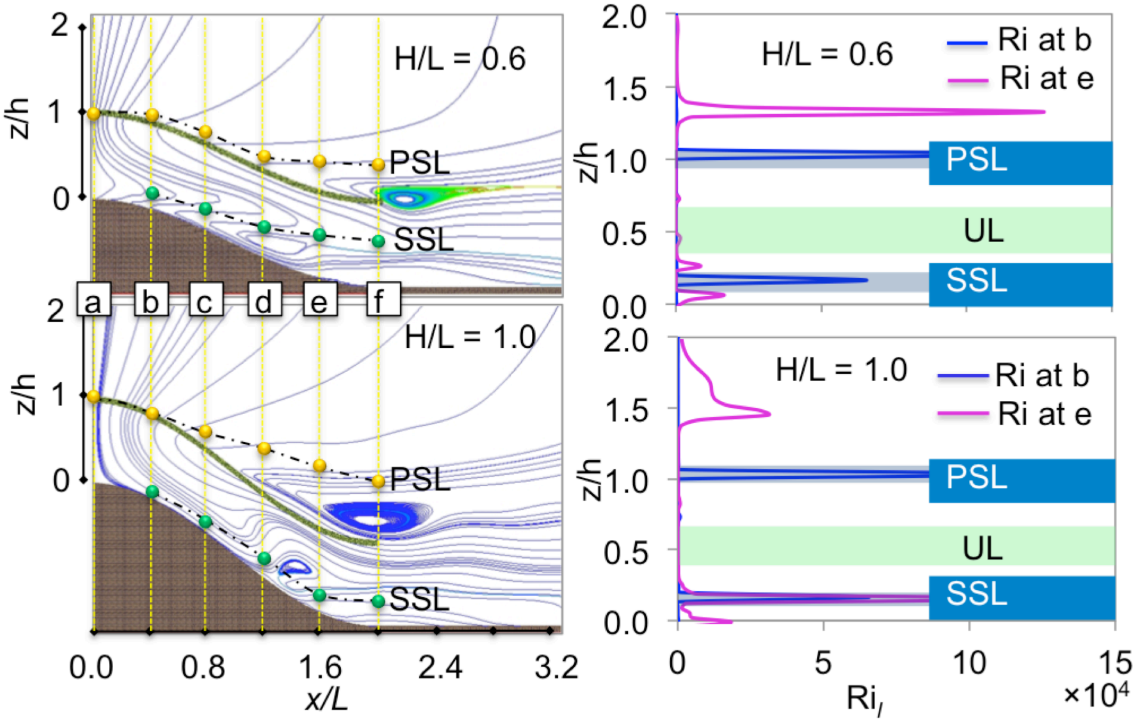

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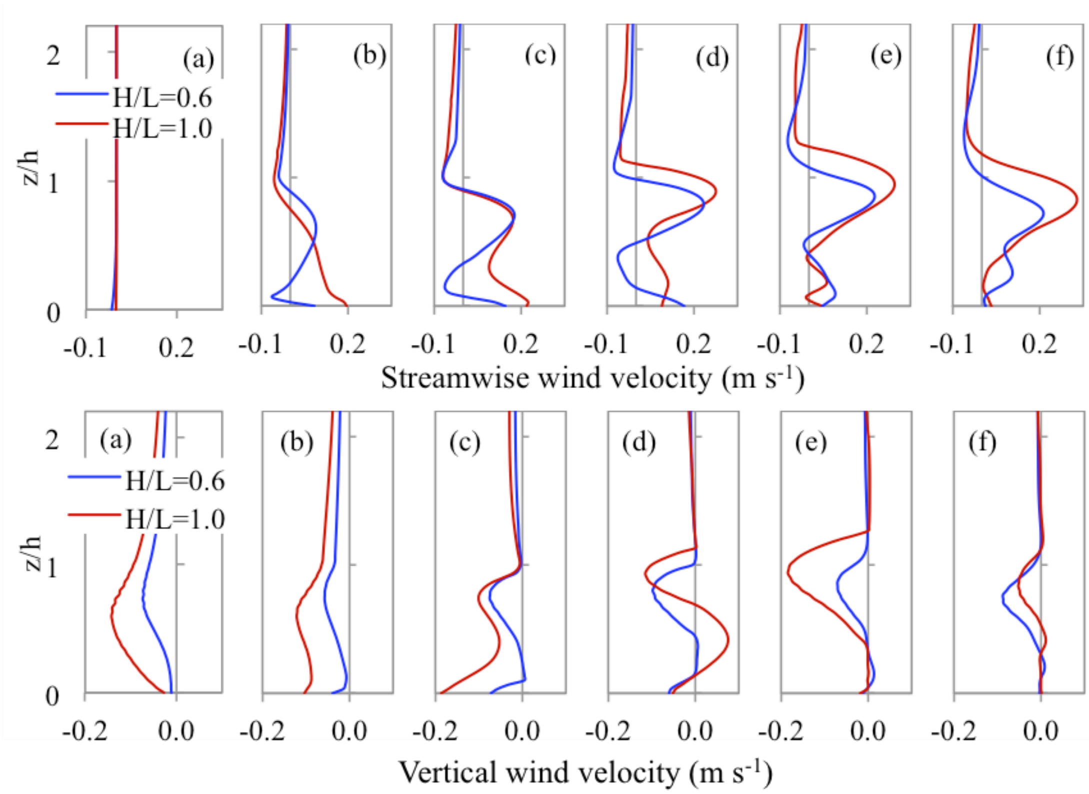

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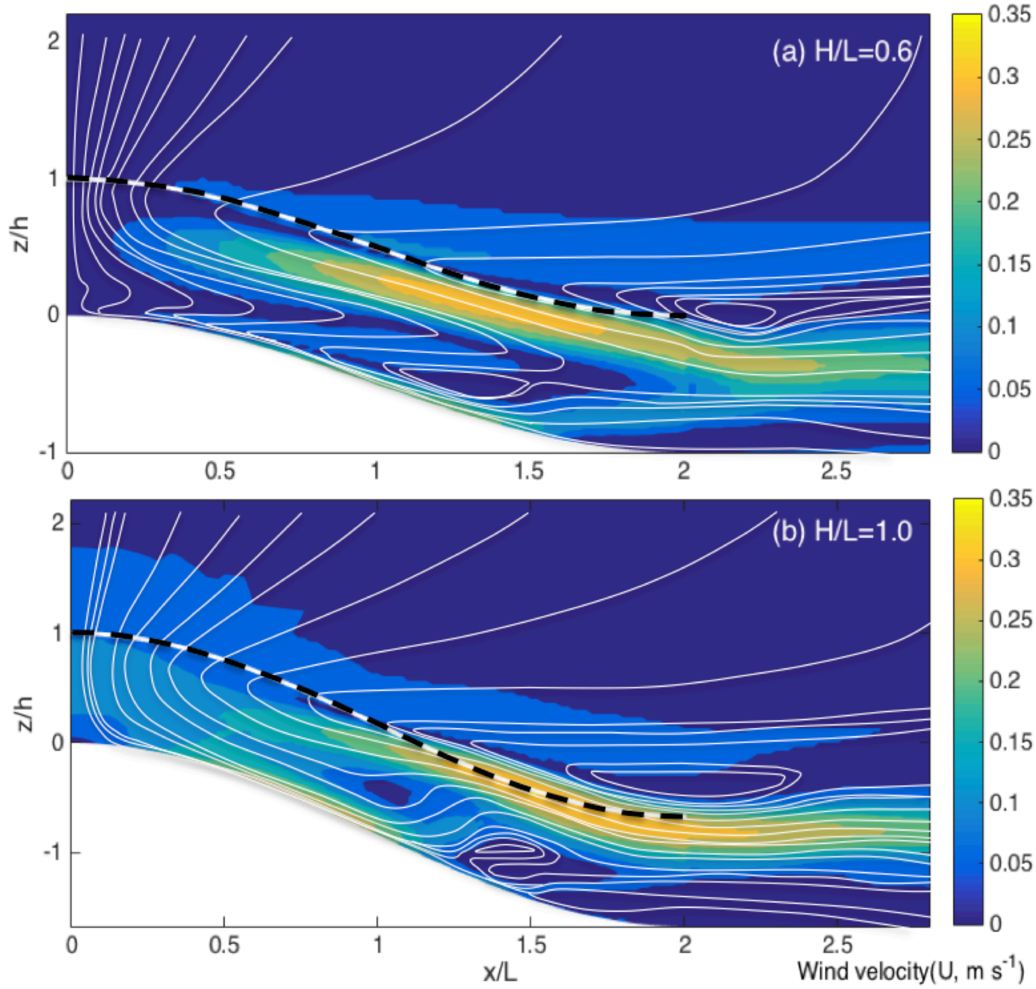

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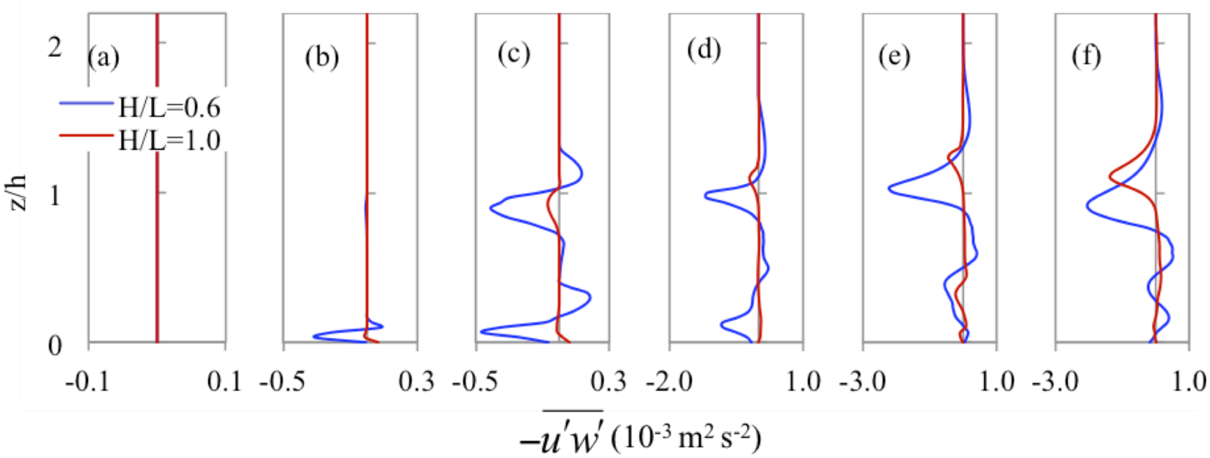

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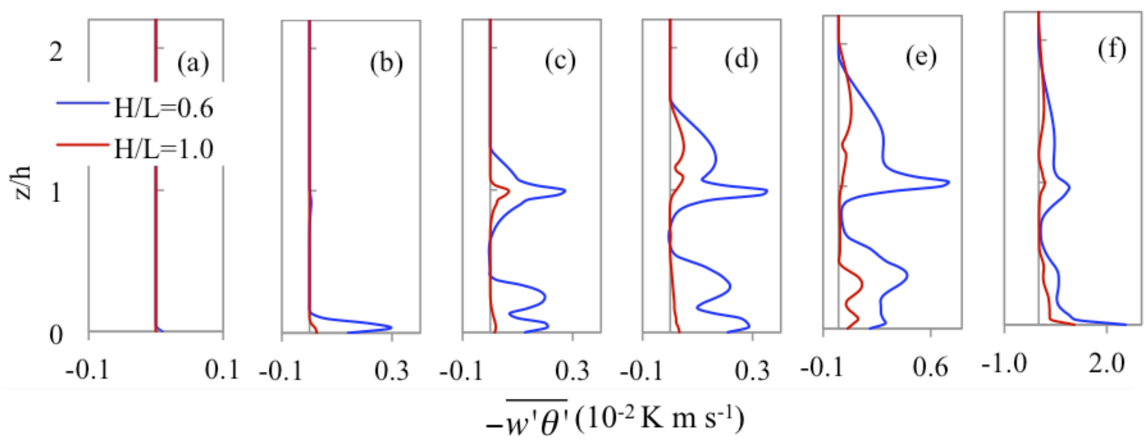

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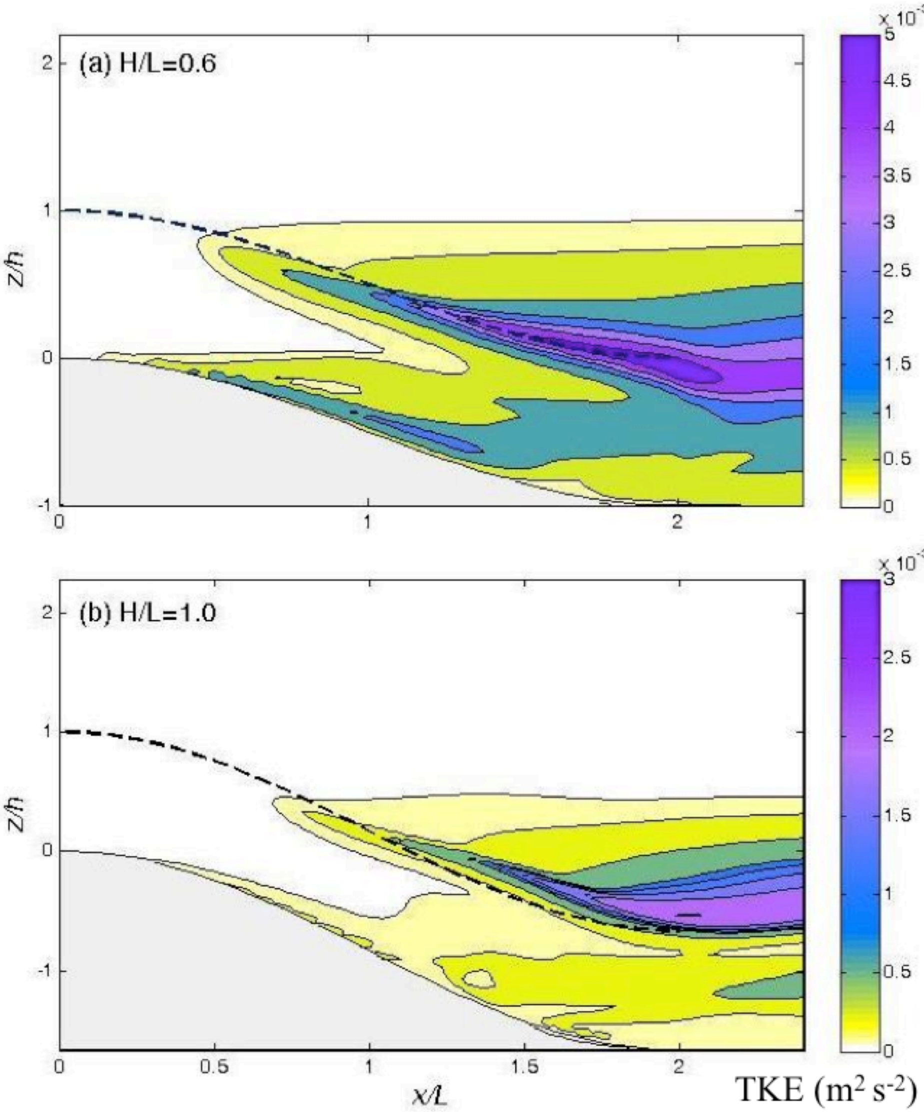

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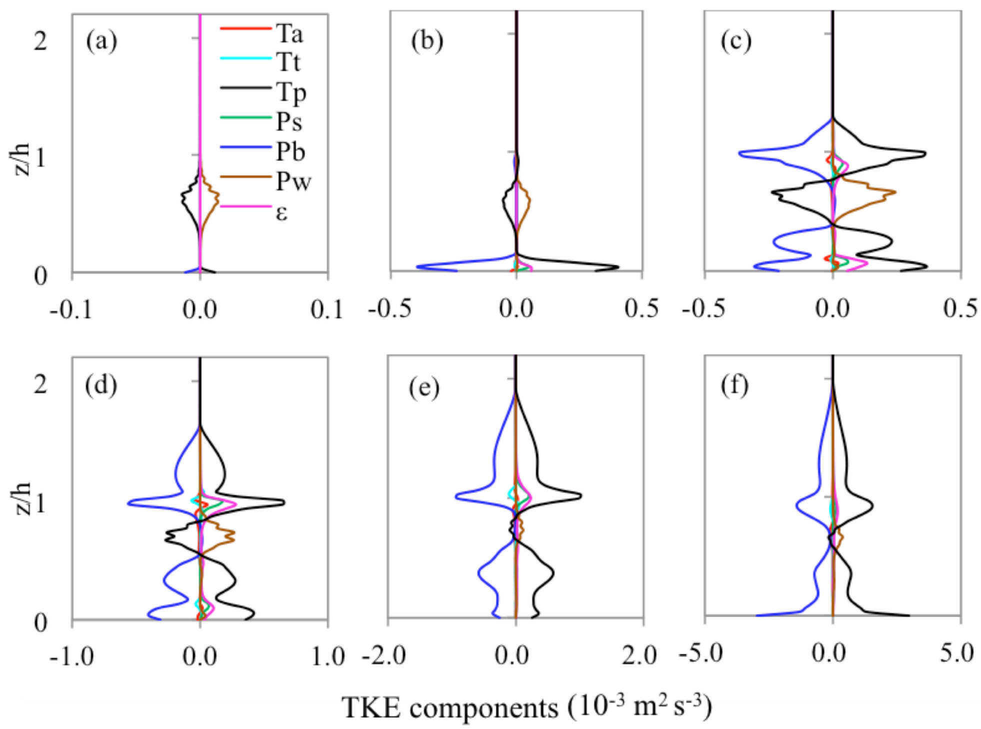

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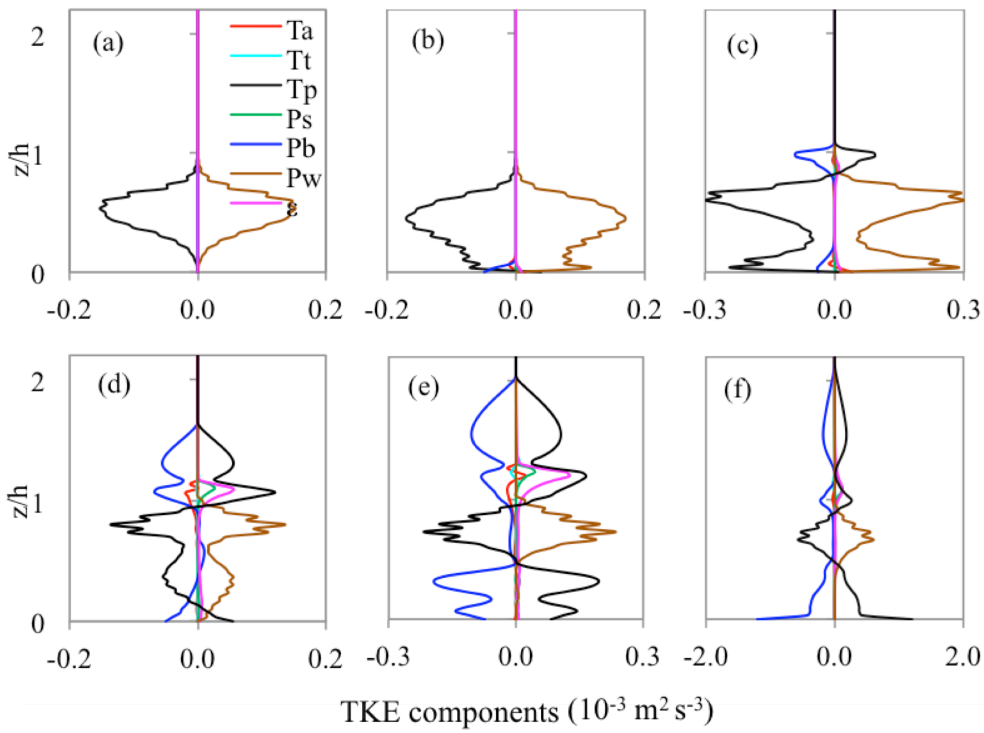

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