NUMERICAL SIMULATION OF FLOW IN BRUSH CREEK VALLEY, COLORADO

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1. INTRODUCTION

Brush Creek Valley is located 55 km northeast of Grand Junction, Colorado. It is a 25 km long valley between Brush Mountain and Skinner Ridge oriented northwest of southeast. It slopes into Roan Creek Valley which then merges with the Colorado River Valley. The valley floor of Brush Creek slopes approximately 14 m/km and is 650 m deep near the mouth. The sidewalls are steep, with slopes of 30-40 degrees. Except for some short box canyons, the valley has no tributaries. In the fall of 1985, this was the site of a major observational program sponsored by the Department of Energy (DOE) as part of its Atmospheric Studies in Complex Terrain (ASCOT) program.

In conjunction with the experimental program the DOE is also supporting the development of models for simulating flows in complex terrain. In this paper, we present some results from our three-dimensional, non-hydrostatic, finite element model applied to simulations of flow in Brush Creek Valley. These simulations are not intended to reproduce any particular experiment, but rather are to evaluate the qualitative performance of the model, to explore the major difficulties involved, and to begin sensitivity studies of the flows of interest.

2. NUMERICAL MODEL

The numerical model used in these simulations solves the following set of three-dimensional, non-hydrostatic, Boussinesq equations.

\[
\begin{align*}
\frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} + w \frac{\partial u}{\partial z} &= -\Theta \frac{\partial E}{\partial x} + f(v - u) \\
&+ K_H \left( \frac{\partial^2 u}{\partial x^2} + \frac{\partial^2 u}{\partial y^2} + \frac{\partial^2 u}{\partial z^2} \right) + K_V \frac{\partial^2 u}{\partial z^2} \\
\frac{\partial v}{\partial t} + u \frac{\partial v}{\partial x} + v \frac{\partial v}{\partial y} + w \frac{\partial v}{\partial z} &= -\Theta \frac{\partial E}{\partial y} - f(u - v) \\
&+ K_H \left( \frac{\partial^2 v}{\partial x^2} + \frac{\partial^2 v}{\partial y^2} + \frac{\partial^2 v}{\partial z^2} \right) + K_V \frac{\partial^2 v}{\partial z^2} \\
\frac{\partial w}{\partial t} + u \frac{\partial w}{\partial x} + v \frac{\partial w}{\partial y} + w \frac{\partial w}{\partial z} &= -\Theta \frac{\partial E}{\partial z} - \frac{(\Theta - \overline{\Theta})}{\overline{\Theta}} - f(u - w) \\
&+ K_H \left( \frac{\partial^2 w}{\partial x^2} + \frac{\partial^2 w}{\partial y^2} + \frac{\partial^2 w}{\partial z^2} \right) + K_V \frac{\partial^2 w}{\partial z^2}
\end{align*}
\]

(1)

\[
\begin{align*}
\frac{\partial \Theta}{\partial t} + u \frac{\partial \Theta}{\partial x} + v \frac{\partial \Theta}{\partial y} + w \frac{\partial \Theta}{\partial z} &= \\
&+ K_H \left( \frac{\partial^2 \Theta}{\partial x^2} + \frac{\partial^2 \Theta}{\partial y^2} \right) + K_V \frac{\partial^2 \Theta}{\partial z^2} \\
\frac{\partial \overline{\Theta}}{\partial t} + u \frac{\partial \overline{\Theta}}{\partial x} + v \frac{\partial \overline{\Theta}}{\partial y} + w \frac{\partial \overline{\Theta}}{\partial z} &= \\
&+ K_H \left( \frac{\partial^2 \overline{\Theta}}{\partial x^2} + \frac{\partial^2 \overline{\Theta}}{\partial y^2} \right) + K_V \frac{\partial^2 \overline{\Theta}}{\partial z^2}
\end{align*}
\]

(2)

where \( \overline{\Theta} \) represents differentiation with respect to time, \( M \) is the "lumped" mass matrix, \( C \) is the gradient matrix, \( C^T \) is the divergence matrix, \( N \) is the advection matrix, \( K \) is the diffusion matrix, and \( f \) is a vector which includes the geostrophic and buoyancy forcing terms. The unknowns are the vectors of nodal velocities, \( U \), the element Exner functions \( P \), and the nodal potential temperatures, \( T \). (The definitions of these matrices and the details of their construction can be found in Gresho et al. 1984.)

Rather than solving the coupled set of momentum and continuity equations, we operate on equation 6 with \( C^T M^{-1} \) and use the equivalent form of equation 7 given by \( C^T U = 0 \).
to derive the consistent discrete Poisson equation for the Exner function:

\[(C^T M^{-1}) P = C^T M^{-1} A \]  \hspace{1cm} (9)

where \( A \) is the right hand side of equation 6. This equation together with equations 6 and 8 form the set of equations we solve.

The basic time discretization scheme employed in the model is a variant of the forward Euler scheme (see Gresho et al., 1984). In order to allow for the very stringent timestep limitation of the standard forward Euler scheme for our advection-dominated problems of interest, we developed a modified Euler method based on the introduction of a balancing tensor-diffusivity term (BTD) term. We have also eliminated an instability associated with the forward Euler treatment of gravity wave (buoyancy) terms by employing a forward-backward integration procedure. The time integration algorithm used in the model can be summarized as follows: For a given \( U \) and \( T \) at time \( n \), solve equation 8 via forward Euler with a BTD term for \( T \) at time \( n+1 \); build \( A \) (the right-hand-side of 6) using \( U \) at \( n \) and \( T \) at \( n+1 \); solve equation 9 by employing an incomplete Cholesky conjugate gradient method for \( P \) at \( n \); then form \( A \cdot C^T P^n \), and solve equation 6 via forward Euler with BTD to obtain \( u \) at time \( n+1 \).

3. GENERAL PROBLEM DEFINITION:

The problem domain was chosen to be 9375 m by 10875 m in the horizontal and is shown in Fig. 1. The terrain data for each surface grid point was taken from the Defense Mapping Agency Planar Digital Terrain Data. This data was then smoothed to minimize small scale grid noise. The top of the domain was set at a constant value of 4000 m above sea level.

The grid contains 10875 (25x29x15) elements and 12480 node points. The horizontal element size is constant with \( \Delta x = \Delta y = 375 \text{ m} \) while the vertical element size expands from \( \Delta z = 20 \text{ m} \) in the lowest element.

![Figure 1. The horizontal computational domain showing geography. The contours represent nodal surface elevations. The contour interval is 100 m.](image)

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For the simulations reported in this paper the turbulence model is a simple constant \( K \) model with \( K_U = 100.0 \text{ and } K_V = 10.0 \text{ m}^2/\text{sec} \) and there is no the geostrophic wind, i.e. \( u_x=v=w=0 \).

The velocity boundary conditions on the top and bottom of the grid are no slip, i.e. \( u=v=w=0 \). On the lateral boundaries the "natural" boundary conditions are set equal to zero, i.e. \( \sigma K_{ij} \frac{\partial_n u}{\partial n} = 0 \) where \( n_j \) is the outward pointing boundary normal vector.

The boundary conditions on the potential temperature were zero normal gradient on the lateral boundaries, \( \Theta = \text{constant} \) on the top and \( \Theta = \lambda \times 1.0 \text{ ft} \) (in hours) on the bottom.

The initial conditions are described by a linearly stratified atmosphere at rest, i.e. \( u=v=w=0 \) and \( \Theta = \lambda z \).

4. SIMULATION 1

For the first simulation we chose \( \lambda \) to be 2 K/km. Figure 2 shows the horizontal velocity fields at approximately 45 and 150 above ground level one hour into the simulation. This clearly shows the development of the nocturnal flows with horizontal divergence off the ridges and convergence into the valley regions. Figure 3 is the corresponding vertical cross section near the mouth of Brush Creek, \( y=6.75 \text{ km} \). (In this figure, the vectors are the two-dimensional velocities in the plane of the cross section while the contours are isobars of the velocity component perpendicular to the cross section.) The sidewall drainage is shown to be rather shallow with a local return flow above each sidewall. This suggests that at early time at least, the down-valley flow is not sufficient to remove all the mass flowing into the valley off the side walls. Figure 4 shows the horizontal velocities two hours into the simulation. The maximum velocity at 45 m has increased from 2.0 m/s at one hour to 4.4 m/s at two hours while at 150 m the velocity increased from 1.8 m/s to 4.0 m/s. Figure 5, the vertical cross section at 2 hours, shows that the sidewall flow has deepened slightly and doubled its speed. Also, the sidewall recirculation zones are no longer in evidence. The down-valley flow has filled the entire valley and increased from 0.8 m/s to 2.2 m/s. Since, in this simulation, the vertical velocity over the center of the valley is very small and upward, the air in the down-valley wind is pushed off the ridge tops and flows down the side walls rather than entering through subsidence over the valley center.

During the ASCOT experiment, a Doppler lidar was placed in the Brush Creek Valley and made measurements of vertical sections of the radial component, "down-valley," wind at various places in the valley. Observations which were made during low synoptic wind conditions showed that the down-valley wind maximum was located at approximately 100 m above the valley floor and the down-valley wind did not fill the valley to ridge top. In contrast, our simulation suggests that the down-valley flow fills the maximum wind occurring between 165 and 200 m above the valley floor. We believe that this is a direct result of our constant K turbulence model which gave a vertical diffusion coefficient that was too large in the levels above the drainage jet.

Another important cause of difficulty with the numerical simulation is the limited area covered by the grid. In an earlier simulation, the computational domain was truncated in the center of Roan Creek. This led to flow coming off the steep walls on the north side of Roan Creek and flowing unhindered out of the southern end of the domain rather than turning and flowing down the creek. This also enhanced the flow in Brush Creek significantly. In the current simulations, up-valley flow can be seen in the first 2 kilometers of Brush Creek in Fig. 2b.
5. SIMULATION 2

Since the slope of Brush Creek is only 1 degree, the down-valley flows observed during the night can not be simple katabatic slope flows but, rather, are the result of mass convergence from the slope flows off the sidewalls. In a recent paper, Nappo and Rao (1987) showed that two-dimensional katabatic flows on an ideal finite-length slope are weakened as the background stratification of the atmosphere increases. If this relationship holds for the sidewall flows in Brush Creek, then the down-valley flow should also be weaker due to the decrease in mass convergence. To test this hypothesis, we repeated the simulation using a neutral background atmosphere, $\lambda = 0$. Figures 6 and 7 show the results after one hour and Figs. 8 and 9 show the two hour results. As predicted, both the sidewall flow and the down-valley flow are stronger than in the previous case and the location of the down-valley maximum wind has moved upward.

6. SIMULATION 3

Our final simulation was with a lapse rate of 4 K/km. The results after two hours are depicted in Figs. 10 and 11. It is again apparent that the increased stability of the atmosphere has reduced the magnitude of the nocturnal flow.

7. CONCLUSIONS

We have demonstrated that, in spite of the crude physics which has been employed in the model, many of the qualitative features of the flows observed at Brush Creek have been captured in our simulations. Our sensitivity study of the dependence of drainage flows on the background stratification for these three-dimensional simulations appears to support and extend the results of Nappo and Rao which were based on idealized, two-dimensional slope flows. For flows over reasonably complex terrain areas such as Brush Creek, we conjecture that our abrupt truncation of the computational domain is perhaps the major limiting factor in the success of our simulations.

REFERENCES


AUSPICES

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Figure 4. As in Fig. 2 except after 2 hours.

Figure 5. As in Fig. 3 except after 2 hours.

Figure 6. The horizontal velocity vectors after 1 hour with neutral stratification. a) at approximately 45 m AGL. b) at an approximately 150 m AGL.

Figure 7. The vertical cross-section at $y = 6.75$ km after 1 hour with neutral stratification. The vectors represent the 2D velocity in the plane of the cross-section and the contours are isotachs of the normal velocity component.
Figure 6. As in Fig. 6 except after 2 hours.

Figure 10. The horizontal velocity vectors after 2 hours with 4 K/km stratification. a) at approximately 45 m AGL. b) at approximately 150 m AGL.

Figure 8. As in Fig. 8 except after 2 hours.

Figure 11. A vertical cross-section at y = 6.75 km after 2 hours with 4 K/km stratification. The vectors represent the 2D velocity in the plane of the cross-section and the contours are isolachs of the normal velocity component.

Figure 9. As in Fig. 7 except after 2 hours.