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Numerical Simulation of Nocturnal Drainage Flow Properties in a Rugged Canyon

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ABSTRACT

A two-dimensional, time-dependent flow model coupled with a radiative transfer module has been applied to examine the characteristics of nocturnal flow in a steep canyon in the Rocky Mountains in Colorado. The effect of nighttime surface cooling on drainage flow is examined and compared with observations. In a complementary study, tracer data have been analyzed to estimate the mass flux from a tributary canyon and to examine processes of transport and diffusion. Simulations indicate that the strength and structure of the drainage wind are controlled mainly by terrain features, ambient wind conditions, and effective radiative cooling rates. The transport of tracer from a lower secondary vortex to an upper primary vortex is largely controlled by diffusional processes; removal of tracer from the canyon is controlled by the primary vortex and its interaction with the ambient wind. Differences between mass fluxes from model simulations and those calculated from experiments involve uncertainties in both the structure of the model and the analysis of data.

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1. INTRODUCTION

Experimental analysis of the patterns of nocturnal circulation in canyons [e.g., by *Coulter et al.*, 1989] have improved our understanding of the mechanisms by which nocturnal winds evolve. The analysis by *Coulter et al.* [1989] is based on observations from the field study of the U. S. Department of Energy's Atmospheric Studies in Complex Terrain (ASCOT) program over the Brush Creek Valley (BCV) area in western Colorado [*Clements et al.*, 1989]. Figure 1 presents the topographic features of the BCV drainage area. The contour values have been converted to relative altitude in meters (actual altitude minus 1,717 m, the lowest altitude in the target area). Pack Canyon (PC), a major tributary that enters BCV perpendicularly from the northeast, slopes 12° upward from the confluence for about 700 m before terminating in a steep (40°) bluff 500 m high. The canyon floor is quite narrow (~ 35 m). Atop the bluffs at the head of PC, the terrain continues upward with a 4° slope for about 1.5 km. The experimental analysis by *Coulter et al.* [1989] show that drainage flow can be clearly identified in PC in the absence of upslope intrusion of drainage flow from the main canyon and entrainment of mesoscale flow.

The objective of this study is to model detailed wind structure of the nocturnal boundary layer in the canyon and to evaluate its dispersion characteristics in terms of tracer source strength, ambient meteorological conditions, and canyon geometry. To meet the purposes, a three-dimensional, time-dependent model developed at Argonne National Laboratory, Argonne, Illinois, for simulations ranging from dynamics and microphysical processes of mesoscale [*Lee and Park*, 1994a] to microscale pollutant transport and dispersion in urban street canyons [*Lee and Park*, 1994b] has been modified to a two-dimensional, time-dependent flow model and has been coupled with a simplified radiative transfer module for simulations of the nocturnal flow properties over the complex terrain. While the model spatial dimensions have been reduced, additional fields of tracer concentration and surface temperature must be calculated to examine the effects of topography and radiative transfer processes on transport and dispersion properties and on radiative cooling of the surface.

Figure 2 presents the two-dimensional model domain representing a southwest-northeast cross section of the oriented along the PC centerline as shown in Figure 1. The external winds are assumed to be southwesterly (i.e., opposite to gravity-driven drainage flow in PC). The sidewalls of PC rise 640 m to the bluff. For simulations of momentum, tracer concentration, and air temperature, the model top is 900 m above the PC-BCV confluence; but for calculations of radiative cooling, the model is extended to approximately 8 km above the canyon floor. At steady state, the total air mass passing through the outflow region above PC (640–900 m) should be

equivalent to that coming through the inflow region. Two-step computations are employed. The steady-state momentum fields are computed first, without consideration of radiative transfer processes; and then the transient fields for momentum, temperature, and tracer concentrations due to the effects of radiative transfer processes are calculated. The simulated fields are compared with ASCOT observations. The mass flux profile above the centerline of PC (extended) is calculated to diagnose the concentration fields in terms of dispersion within the boundary layer that results from mechanical and convective instability during the nighttime; results are compared with the analysis of Coulter *et al.* [1989].

2. EQUATIONS AND COMPUTATIONAL PROCEDURES

The equations for momentum and the continuity are written as

$$\frac{\partial}{\partial t}(\rho \vec{V}) + \nabla \cdot \rho \vec{V} \vec{V} = -\nabla p + \nabla \cdot (\mu_m \nabla \vec{V}) + \vec{S}_m + \rho g \vec{k} \cdot B \quad (1)$$

and

$$\frac{\partial \rho}{\partial t} + \nabla \cdot \rho \vec{V} = 0, \quad (2)$$

where

$$\vec{V} \equiv (u, w),$$

$$\nabla \cdot \equiv \left(\frac{\partial}{\partial x}, \frac{\partial}{\partial z} \right),$$

$$\vec{S}_m \equiv \left(\nabla \cdot \mu_m \frac{\partial}{\partial x} \vec{V}, \nabla \cdot \mu_m \frac{\partial}{\partial z} \vec{V} \right),$$

ρ is the density of air, μ_m is the viscosity of air, and p is the dynamic air pressure. In the last term of (1), B represents the net buoyancy due to the virtual temperature differences minus the drag of liquid water.

The conservation equations for air temperature, water vapor mixing ratio, and species concentrations are written as

$$\frac{\partial}{\partial t}(\rho T) + \nabla \cdot (\rho \vec{V} T) = \nabla \cdot (K \nabla T) + \frac{L_v}{c_p} \left(\frac{\delta(\rho q)}{\delta t} \right)_{NA} + \frac{\rho T}{c_p} Q, \quad (3)$$

$$\frac{\partial}{\partial t}(\rho q) + \nabla \cdot (\rho \vec{V} q) = \nabla \cdot (K \nabla q) + \left(\frac{\delta(\rho q)}{\delta t} \right)_{NA}, \text{ and} \quad (4)$$

$$\frac{\partial}{\partial t}(\rho C) + \nabla \cdot (\rho \vec{V} C) = \nabla \cdot (K \nabla C), \quad (5)$$

where T , q , and C denote the air temperature, water vapor mixing ratio, and species concentration, respectively, L_v is the latent heat of vaporization, c_p is the specific heat of air, and K is the diffusivity of air. In the last term of (3), Q denotes the heating rate per unit mass of air. In (3) and (4), the term $[\delta(\rho q)/\delta t]_{NA}$ represents the changes in water vapor mixing ratio due to nonadiabatic processes. In a cloud-free environment, $[\delta(\rho q)/\delta t]_{NA} = 0$. At present, the values for eddy viscosity are determined diagnostically by following the approaches of *Deardorff* [1973] and *Sommeria* [1976].

The radiative forcing is calculated following the approach of *Oh* [1989]. The spectrum of solar radiation is divided into three contributions: Rayleigh scattering, ozone absorption, and water vapor absorption. The spectrum of long-wave radiation is divided into four bands depending on absorbers: water vapor band centers, water vapor band wings, CO_2 band, and O_3 band. For this study, skies are assumed to be clear. A two-stream, delta-Eddington method, economical and suitable for simulations of atmospheric processes under variable atmospheric conditions [*King and Harshvardhan*, 1986], is employed for calculations of solar radiation fluxes. The flux equations are written as

$$\frac{dF^\uparrow(\vartheta)}{d\vartheta} = a_1 F^\uparrow(\vartheta) - a_2 F^\downarrow(\vartheta) - a_3 \frac{S(\vartheta)}{\mu_o}, \quad (6)$$

$$\frac{dF^\downarrow(\vartheta)}{d\vartheta} = a_2 F^\uparrow(\vartheta) - a_1 F^\downarrow(\vartheta) + a_4 \frac{S(\vartheta)}{\mu_o}, \text{ and} \quad (7)$$

$$\frac{dS(\vartheta)}{d\vartheta} = -\frac{S(\vartheta)}{\mu_o}, \quad (8)$$

where $F^\uparrow$ and $F^\downarrow$ denote the upward and downward solar flux, respectively; ϑ is the optical depth; μ_0 is the value of μ ($= \cos \zeta$, where ζ is the zenith angle) at the top of the atmosphere; S is the direct downward solar radiation; and a_i ($i=1, 2, 3, 4$) is the coefficients. The a_i can be determined if the values for zenith angle, scattering optical thickness, and single scattering albedo are known. The optical properties are determined for Rayleigh scattering ($0-0.44 \mu\text{m}$ and $0.44-0.69 \mu\text{m}$), O_3 absorption ($0-0.44 \mu\text{m}$ and $0.44-0.69 \mu\text{m}$), CO_2 absorption ($0.69-3.95 \mu\text{m}$), and H_2O absorption ($0.69-3.85 \mu\text{m}$) by using formulas from the literature [Lacis and Hansen, 1974; Wiscombe and Evans, 1977; Fouquart, 1988; Oh, 1989].

The downward and upward fluxes of long-wave radiation at level z may be expressed in terms of blackbody radiation πB and fractions of spectral bands f_j [Oh, 1989] as

$$R_i^\downarrow = \pi B_i + \sum_{j=1}^J \int_{\pi B_i}^0 f_j(\pi B) \tau_j(\Delta\chi) d(\pi B), \text{ and} \quad (9)$$

$$R_i^\uparrow = \pi B_i + \sum_{j=1}^J \int_{\pi B_i}^{\pi B} f_j(\pi B) \tau_j(\Delta\chi) d(\pi B), \quad (10)$$

where πB_z is the blackbody flux at the surface; τ_j denotes the transmission function corresponding to f_j ; and $\Delta\chi = |\chi - \chi_z|$, with χ and χ_z denoting the absorber amounts to the surface and to the level z , respectively. The fractions for spectral subdivisions ($\sum f_j = 1$, where $j = 1, 2, 3, 4$) are parameterized as a function of T . Spectral band 1 accounts for the H_2O band centers ($0-340 \text{ cm}^{-1}$ and $1,380-1,900 \text{ cm}^{-1}$); band 2 accounts for the H_2O band wings and H_2O continuum ($340-540 \text{ cm}^{-1}$, $800-980 \text{ cm}^{-1}$, $1,100-1,380 \text{ cm}^{-1}$, and $1,900-3,000 \text{ cm}^{-1}$); band 3 accounts for the H_2O $15\text{-}\mu\text{m}$ overlap, H_2O continuum, and CO_2 ($540-800 \text{ cm}^{-1}$); and band 4 accounts for the H_2O continuum and O_3 ($980-1,100 \text{ cm}^{-1}$). The transmission functions (τ_j) corresponding to spectral bands (f_j) are calculated using parameterizations by Oh [1989]. Oh's formulas are derived on the basis of the analyses of Chou and Peng [1983], Kneizys et al. [1983], and Chou [1984].

At the surface, the changes in surface temperature T_s by radiative cooling is determined from a prediction equation [Arakawa, 1972]:

$$C_s \frac{\partial T_s}{\partial t} = F_s - R_s - H_s - LE_s, \quad (11)$$

where, at the surface, F_s is the net downward flux of solar radiation, R_s is the net upward flux of long-wave radiation, H_s is the upward flux of sensible heat, LE_s is the upward flux of latent heat, and C_s denotes the heat capacity of surface material. The value of F_s is zero during the nighttime; but R_s , H_s , and E_s vary strongly with T_s . The heat capacity is expressed as

$$C_s = \sqrt{\frac{\lambda c_s}{\omega}},$$

where λ denotes the thermal conductivity of soil, c_s is the heat capacity per unit volume, and ω is the angular frequency of the diurnal change. Sensible and latent heat fluxes at the surface are calculated using bulk aerodynamic methods.

A staggered grid system is employed for vector (u and w) and scalar (T , q , C , p , and ρ) quantities, as shown in Figure 2. The scalar quantities are defined at the centers of the grid cells (solid circle), while the wind vectors are defined at the midpoints of cell boundaries. Both the vector and scalar quantities are defined at the boundaries as well. As illustrated in Figure 2, we assume that the upwind ambient horizontal velocity u is constant, that the forced-flow boundary condition is applied at the outflow region, that shear-free flow is assumed at the top of the model, and that nonslip boundary conditions are applied to all solid surfaces.

The set of governing equations is solved on the staggered grid by a finite-volume method (finite area in two dimensions). Finite-difference equations are derived by integration of the governing equations over an elementary control area or cell surrounding a grid node. The solution algorithm used for the momentum equation is the semi-implicit method for pressure-linked equations outlined in Patankar [1980]. The momentum and pressure fields are determined through iterative adjustment. The discretized momentum equations are solved initially with a pressure field obtained from the previous step; then the pressure field is recalculated from a pressure Poisson equation to yield the pressure change needed to adjust the velocity field to ensure mass continuity. Iteration continues until required adjustments in the momentum and pressure fields are sufficiently small. After the velocity field is obtained, solution fields for air temperature, water vapor mixing ratio, and tracer concentration are calculated from conservation equations for T , q , and C discretized by a finite-volume method and solved by the successive overrelaxation method.

3. MODEL SIMULATIONS AND DISCUSSIONS

Baseline simulations have been carried out to examine the flow properties in PC with inflow strength $u_0 = 5$ m/s and with surface properties as presented in Table 1. The horizontal grid size is set to 100 m. The vertical grid size is set to 30 m up to the shear-free level (the model top for drainage flow analysis) (900 m) and increases progressively thereafter to about 500 m. Initially, the temperature and pressure at the bottom of the model are assumed to be 280 K and 820 mbar, respectively; and the vertical profiles of air temperature T , air pressure p , and air density ρ below 900 m are calculated from

$$T_k = T_{k-1} - (z_k - z_{k-1}) \frac{g}{c_p}, \quad (12)$$

$$p_k = p_{k-1} \exp\left(-\frac{g(z_k - z_{k-1})}{RT}\right), \text{ and} \quad (13)$$

$$\rho_k = \frac{p_k}{RT_k}, \quad (14)$$

where R is the gas constant for dry air, and subscript k denotes the vertical grid index. Above the 900 m level, the air temperature is assumed to decrease with height at a rate of 6.5 °C/km.

A steady-state solution field for momentum is obtained without considering radiative transfer processes. A small time step (0.001 s) is applied initially, and then the time step is increased progressively up to 10 s as the momentum field approaches a steady state. Transient fields for momentum, for thermal and moisture structure, and for tracer concentration are then calculated by considering radiative transfer effects. The time step is set to 10 s during the transient-flow calculations.

Figure 3 presents the time evolution of surface temperature, sensible heat flux, and latent heat flux at $x = 2.75$ km. The surface temperature decreases during nighttime by 0.3–0.6 °C/h, and increases after sunrise by 0.5–0.9 °C/h to a maximum temperature of about 19 °C at 1400 local standard time (LST). During nighttime, sensible and latent heat fluxes maintain negative values of about -6 W m⁻² and -13 W m⁻², respectively. As the sun rises, both sensible and latent heat fluxes increase rapidly. The sensible heat flux reaches a maximum of 208 W/m² at noon, while the latent heat flux reaches a maximum of 170 W/m² at 1300 LST. Daily amplitude of surface temperature is about 10 °C.

Momentum fields calculated without and with radiative transfer effects are presented in Figures 4a and 4b, respectively. The radiation-transfer calculations cover the period 0000 to 0400 LST. The aspect ratio, defined as the ratio of the height to the width of the BCV Canyon, is about 0.4. The maximum wind in Figure 4a represents 6.1 m/s. We see clearly that two vortices are developed in the canyon: a primary clockwise (viewed from down canyon) vortex in the upper right part of the canyon (D in Figure 4a) and a secondary counterclockwise vortex in the lower left part of the domain opposite PC (C in Figure 4a). The primary vortex is formed because of the deflection of flow by the PC wall, while the secondary vortex is created because of the separation of the reverse flow formed in the lower portion of the primary vortex as the flow is deflected by the western slope. The strength of the secondary vortex, which is controlled mainly by the flow characteristics of the primary vortex, is weaker than that of the primary vortex by about a factor of 5. We find that the strength of the primary vortices is much smaller than the ambient wind speed and that the strength of the secondary vortices is yet smaller than that of the primary vortices by about the same factor. We see downslope winds along the PC wall, as well as in the lower portion of the western wall, resulting from the two-vortex steady-state canyon flow. We also notice that weak vortices or wakes are generated near the western rim region (A and B in Figure 4a), while orographic updrafts are developed beyond the eastern rim region (E and F in Figure 4a). The flow field becomes more complicated as changes in pressure gradient caused by differential temperatures that result from radiative cooling or heating near the surface affect the mechanical entrainment (Figure 4b). The drainage winds are clearly developed on both walls, and several vortices of nearly equal strength are formed (E, F, and H in Figure 4b) in the canyon; updraft regions (lower middle portion of the canyon, D and G in Figure 4b) are developed; and the wakes along the western rim region (A and B in Figure 4a) are replaced by downslope winds (A and B in Figure 4b). The primary vortex formed at steady state (D in Figure 4a) is separated into two vortices (E and H in Figure 4b) by the forced convection (G in Figure 4b), while the secondary vortex (C in Figure 4a) is pushed toward the middle of the canyon because of the new circulation developed by drainage winds in the upper portion of the western slope (C in Figure 4b) and the forced convective flow away from the western slope (D in Figure 4b). The orographic updrafts remain intact as in the steady-state case in the eastern rim region (I and J in Figure 4b).

Figures 5a and 5b show vertical velocity fields at steady state (Figure 5a) and after 4 hours of radiative transfer effects (0000 to 0400 LST) (Figure 5b). By plotting the vertical wind component, we see clearly the outstanding cellular features associated with updrafts and downdrafts. We notice that the broad downdraft region along PC at steady state narrows in Figure 5b because of radiative cooling. The development of downslope winds on both sides generates updraft cells right next to the drainage regions. A second downdraft region on the left side of the

canyon is not reflected on the PC side, however, because of the stronger drainage flow from PC that results in more convergence at the BCV-PC interface. The weak updraft region in the middle of the canyon extends about 400 m from the canyon floor. The vertical velocity field has a wavelength and amplitude of about 700 m and 400 m, respectively.

In a complementary study, tracer dispersion has been simulated and analyzed to examine exchange processes between external and canyon air masses and to investigate the time constant of inert contaminants within the canyon. For the former case, concentrations of four grid points at the inflow boundary are held constant (simulating a continuous source), and transient fields of momentum, temperature, and species concentrations are calculated as a boundary value problem. For the latter case, however, a uniform initial concentration field in the canyon is assumed, and the momentum and thermodynamic fields in the model domain are assumed to be invariable, and then the concentration fields are computed as an initial and boundary value problem. Computational procedures examine the relationship between concentration patterns and drainage wind for the former case and analyze the removal efficiency of pollutants in the canyon for the latter case. Figure 6 shows the concentration field obtained after 4 hours of real-time time simulation for the former case, while Figures 7a and 7b show the concentration fields at 0 and 4 hours for the latter case. All contour lines are scaled by the maximum concentration in the field at that time.

In Figure 6, we see that the location of the maximum concentration is controlled by the source location, but relatively high concentrations extend downward along the drainage regions, while relatively low concentrations extend above the PC-BCV confluence to about 350 m. Our analysis of the concentration evolution (not shown) indicates that the concentrations along both drainage slopes increase with time, while the vertical extension of the low concentration tongue in the middle of the canyon increases with time, reaching almost to the canyon height (~ 600 m) after 8 hours. Figure 7a for the case of uniform initial concentrations within the canyon without drainage flow effects shows that the locations of maximum concentration coincide closely with the vortex centers (C and D Figure. 4a), indicating efficient removal of pollutants via convection. (The concentration field reaches a structural steady state because concentrations are normalized by the maximum value. Actual values would continue to decrease.) Transport of pollutants from inner streamlines to outer streamlines is affected by the magnitude of the diffusivity of the pollutants. Once the pollutants are moved to outer streamlines, the pollutants are entrained into fresh air flow; however, in this case (Figure 7a), the maximum concentration remains in the lower left portion of the canyon (about 25% height), indicating that the pollutants are largely trapped in the lower vortex. The location of the secondary maximum concentration coincides with the center of the vortex in the upper right portion of the canyon (D in Figure 4a). The transport of pollutants from

the lower vortex to the upper vortex depends largely on diffusion, a process that is much less efficient than the processes of convection and entrainment at the top of the canyon. As drainage winds start to develop, the maximum concentration shifts toward the center of BCV, and multiple cells develop because the forced convections at various locations form vortices and complicated streamlines. Nonetheless, Fig. 7b shows that the maximum concentrations are located fairly close to the centers of the vortices (see E and H in Figure 4b), again indicating efficient removal of pollutants from the valley via convection. The high concentration cell at $(x, z) = (2.9 \text{ km}, 100 \text{ m})$, which cannot be identified clearly from the momentum field (Figure 4b), is due to the entrainment between the vortex at E (Figure 4b) and the easterly stream of the vortex F (Figure 4b). The high concentration remains in the lower southwestern portion of the canyon (about 50% height).

By assuming a width of PC as a function of height [Coulter *et al.*, 1989], mass fluxes can be computed along the PC drainage region by integrating the flow field. Figure 8 presents the time evolution of mass fluxes in the PC drainage region. Our analyses show that the streamlines around the primary vortex (D in Figure 4a) transport fresh air into the canyon initially with a maximum value of about 110 t/s at 400-m height of PC (open circles). As the surface temperature decreases by radiative cooling, the flow field in the canyon becomes complicated because of drainage winds, forced updrafts, and reflected winds off the canyon walls (see Figure 4b). At this stage, the drainage winds, which depend on inflow strength and cooling rate, are formed on both sides of the canyon slopes. We notice that the maximum values of the drainage mass flux decrease rapidly during the first 2 hours to about 50 t/s (solid circles in Figure 8) and then remain unchanged during the next 4 hours (open and solid squares in Figure 8). As the sun rises at about 0700, the flow field becomes chaotic initially (open triangles in Figure 8), and then undergoes rapid adjustment to convective instability (solid triangles in Figure 8).

Two additional calculations have been made to examine the effects of inflow strength on flow fields. The inflow values are set to 2.5 and 10 m/s for two additional runs. We find that the flow fields obtained with different inflow strength show similar structures of the baseline run, with wind velocities and vortex strengths almost linearly proportional to momentum influx. The vertical velocity field has a wavelength and amplitude of about 700 m and 400 m, respectively, in all cases studied, indicating that the wave characteristics are determined mainly by topographic structure, at least in this two-dimensional simulation. Figure 9 presents the vertical profiles of mass fluxes and drainage area at 0400 LST for different inflow strengths. The vertical profile of drainage area represents the case when $u_o = 5 \text{ m/s}$. Note that the changes of drainage area show similar structures in all cases. The maximum mass flux increases with the increase of ambient wind strength; however, our analysis shows that the effect of the inflow strength on canyon flow

development can be limited and overcompensated by the drainage wind generated by radiative cooling. For example, our analysis show that the relationship between inflow strength (u_o in meter per second) and maximum mass flux (M in metric tons per second) can be expressed as $M = 5.15 u_o + 25.7$ to give $M = 25.7$ t/s when $u_o = 0$, indicating that a significant portion of the downslope mass flux can be attributed to drainage wind generated by radiative cooling.

Observational analysis [Coulter *et al.*, 1989] for cases of southwesterly wind (the conditions are similar to the flow conditions applied in our baseline simulations) shows that the mass fluxes are large ranging from 44 to 66 t/s at about 100 m of height; and thereafter the values decrease to 10–17 t/s¹ at about 200 m of height and to 2–3 t/s at the upper PC. On the other hand, the simulated fields show that the mass flux increases with height reaching a maximum of about 50 t/s at about 300 m and then decreases thereafter; however, the computed mass fluxes cannot compare directly with observations by Coulter *et al.* [1989] because of uncertainties on both model simulations and experimental analysis. For example, the two-dimensional simulations cannot account for the y-directional effects on the flux calculations (i.e., effects of BCV on PC flow). The experimental analysis assumed a constant drainage area to integrate mass fluxes, while the model results show that the drainage area changes significantly with height, as shown Figure 9.

4. CONCLUSIONS AND FUTURE STUDIES

Results produced from a two-dimensional, time-dependent flow model show that the properties of the drainage wind can be simulated and interpreted in terms of flow conditions and topographical configurations. The strength of the drainage wind and its distribution are controlled mainly by the ambient wind conditions and effective radiative cooling rates. Simulated contaminants in the canyon are controlled by the primary vortex formed in the upper portion of the canyon and augmented dispersion and strong ambient winds, while the transport of pollutants from the lower vortex to the upper vortex is largely controlled by diffusional processes.

More numerical studies are planned with an improved model. Extension of the model from two dimension to three dimensions will allow the inclusion of y-directional effects in the flux calculations, while the addition of a cloud module [Lee and Park, 1994a] will allow examination of flow properties under cloudy conditions.

Acknowledgments. Thanks are due to Drs. J. D. Shannon and R. A. Reck at Argonne National Laboratory and Mr. Ronald Meyers of White Sands Missile Range for useful discussions

and support. This work was supported by the Office of Health and Environmental Research of the U.S. Department of Energy Office of Energy Research, under contract W-31-109-ENG-38.

REFERENCES

- Arakawa, A., *Design of the UCLA General Circulation Model*, Technical Report No. 7, Department of Meteorology, University of California, Los Angeles, 1972.
- Chou, M.-D., Broadband water vapor transmission functions for atmospheric IR flux computation, *J. Atmos. Sci.*, **41**, 1755-1778, 1984.
- Chou, M.-D., and L. Peng, A parameterization of the absorption in 15 mm CO₂ spectral region with application to climate sensitivity studies, *J. Atmos. Sci.*, **40**, 2183-2192, 1983.
- Clements, W. E., J. A. Archuleta, and P. H. Gudikson, Experimental design of the 1984 ASCOT field study, *J. Appl. Meteor.*, **28**, 405-413, 1989.
- Coulter, R. L., M. Orgill, and W. Porch, Tributary fluxes into Brush Creek Valley, *J. Appl. Meteor.*, **28**, 555-568, 1989.
- Deardorff, J. W., The use of subgrid transport equations in a three-dimensional model of atmospheric turbulence, *J. Fluid Eng.*, **95**, 429-438, 1973.
- Fouquart, Y., Radiative transfer in climate modeling, in *Physically-Based Modelling and Simulations of Climate and Climate Change*, edited by M. E. Schlesinger, NATO Advanced Study Institute Series, pp. 223-283, Kluwer, Dordrecht, the Netherlands, 1988.
- King, M. D., and Harshvardhan, Comparative accuracy of selected multiple scattering approximations, *J. Atmos. Sci.*, **43**, 784-801, 1986.
- Kneizys, F. X., E. P. Shettle, W. O. Gallery, J. H. Chetwynd, Jr., L. W. Abreu, J. E. A. Selby, S. A. Clough, and R. W. Fenn, *Atmospheric Transmittance/Radiance: Computer code LOWTRAN 6*, Optical Physics Division, Hanscom AFB, Bedford, MA, 1983.
- Lacis, A. A., and J. E. Hansen, A parameterization for the absorption of solar radiation in the earth's atmosphere, *J. Atmos. Sci.*, **31**, 118-133, 1974.
- Lee, I. Y., and H. M. Park, Comparison of microphysics parameterizations in a three-dimensional dynamic cloud model, *Atmos. Environ.*, in press, 1994a.
- Lee, I. Y., and H. M. Park, Parameterization of pollutant transport and dispersion in urban street canyons, *Atmos. Environ.*, in press, 1994b.
- Oh, J.-H., Physically-based general circulation model parameterization of clouds and their radiative interaction, Ph.D. Dissertation, Oregon State University, Corvallis, Oregon, 1989.
- Patankar, S. V., *Numerical Heat Transfer and Fluid Flow*, McGraw-Hill, New York, 1980.
- Sommeria, G., Three-dimensional simulation of turbulent processes in an undisturbed trade wind boundary layer, *J. Atmos. Sci.*, **33**, 216-241, 1976.

Wiscombe, W. J., and J. W. Evans, Exponential-sum fitting of radiative transmission functions,
J. Comp. Phys., 24, 416-444, 1977.

- Fig. 1.** Topographic map of the BCV drainage area. Contours are in meters above sea level, and the straight line shows the projection of the cross section for two-dimensional analysis. Universal Transverse Mercator (UTMX and UTMY) coordinates are shown along the abscissa and ordinate, respectively.
- Fig. 2.** Schematic diagram for the nocturnal drainage flow analysis and staggered grid system to locate vector and scalar variables.
- Fig. 3.** Time evolution of surface temperature, sensible heat flux, and latent heat flux over a forested surface.
- Fig. 4.** Steady-state momentum field obtained without radiative cooling (4a) and transient momentum field obtained after 4 hours of real-time simulation (0000 to 0400 LST) with radiative cooling (4b). The maximum value represents 6.1 m/s.
- Fig. 5.** Vertical velocity fields obtained at steady state (5a) and after 4 hours of real time simulation (0400 LST) with radiative cooling (5b).
- Fig. 6.** Concentration field obtained after 4 hours of real-time time simulation (0400 LST).
- Fig. 7.** Concentration field obtained under steady-state conditions (7a) and that obtained under transient conditions after 4 hours of real-time simulation (0400 LST) with radiative cooling (7b).
- Fig. 8.** Time evolution of mass fluxes in the PC drainage region.
- Fig. 9.** Vertical profiles of mass fluxes after 4 hours of real-time simulation (0400 LST).

TABLE 1. Initial Soil Properties Applied to the Model Calculations

Julian day	162
Surface albedo, %	14
Soil moisture content, %	2
Surface roughness length, m	1.0
Soil density, g cm ³	1.65
Specific heat of soil, cal g ⁻¹ °C ⁻¹	0.5
Thermal conductivity of soil, cal s ⁻¹ cm ⁻¹ °C ⁻¹	0.001

Topography (m)

Fig. 1.

Fig. 3.

Case Study B: Canyon Flow at 00:00 LST

Fig. 4a.

Case Study B: Canyon Flow at 04:00 LST

Fig. 4b.

Case Study B: Vertical Velocity at 00:00 LST (cm/s)

Case Study B: Vertical Velocity at 04:00 LST (cm/s)

Case Study B: Relative Concentration at 04:00 LST

Case Study B: Relative Concentration at 00:00 LST

Case Study B: Relative Concentration at 04:00 LST

Fig. 8.

Fig. 9.

**DATE
FILMED**

7/18/94

END