A Numerical Model Study of Turbulent Airflow in and Above a Forest Canopy

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Abstract

A simplified second-moment turbulence closure model, which has been reasonably well tested in various geophysical problems, is used to simulate effects of a tall tree canopy on air circulations in the atmospheric boundary layer. Qualitative simulation of a canopy flow, with nearly constant and low wind speeds in a canopy, but large wind shears near a treetop, and unstable (stable) temperature layers within a canopy during the night (day) are all satisfactory. Strong couplings between the mean and turbulence variables are obvious when simulations performed with and without a canopy in the model are compared with one another.

1. Introduction

Mean wind speeds within a forest canopy are known to be considerably smaller than those over trees due mainly to the surface drag induced by leaves, stems and branches (e.g., Oliver, 1981). Tall trees also modify temperatures within and above a canopy. For example, during the day initial warming of the upper canopy results in an unstable layer above, and a stable layer within the canopy (e.g., Hosker, et al., 1974). The stability structure is reversed during the night because the rate of cooling due to longwave radiation is maximum in the upper canopy.

Observations and analyses of wind speeds and temperatures within and above tall trees are difficult to make comparing to those over a bare soil (or short vegetation). The latter surfaces have been conventionally characterized by "roughness lengths", and empirical expressions of profiles based on Monin-Obukhov similarity theory are widely accepted. However, observations appeared in the past are not sufficient to obtain similarity profiles of wind speeds and temperatures within and over a canopy. A simple exponential profile may fit wind speeds reasonably well in the upper part of a canopy, but fails to represent the profiles near the ground where wind maxima are often observed. If trees were considered to be large roughness elements, empirical expressions for the profiles similar to those over a bare soil could be used. The similarity profiles are, however, valid only in the limited region. The lower limit might be slightly above the top of trees, and the upper limit might be the height where turbulence is significant. Thus, the upper limit varies diurnally, and is significantly higher during daytime than nighttime. These limits are defined loosely, and so may have very little significance in practice.

Even less is known about effects of trees on air circulations over sloped surfaces (Bergen, 1969). The Atmospheric Studies in Complex Terrain (ASCOT) program of the U.S. Department of Energy has sponsored field experiments during the last three summers over an area designated as the California Known Geothermal Resources area. The objectives of the ASCOT program (Dickerson and Gudiksen, 1980) are:

a) to improve fundamental knowledge of transport and dispersion processes in complex terrain, and

b) building on this improvement in the understanding of the physics, provide a methodology for performing air quality assessments.

Extensive observations were made during summer 1980 over an area of 7 km×10 km (Dickerson, 1981); instrumentation included five tether-sondes, one pibal, four acoustic sounders, five laser anemometers, ten 10 m towers, one 30 m tower and twenty-nine surface stations. Measurements of spatial and temporal distribution of
pollutant concentration also were obtained using as tracers heavy methane, perfluorocarbon, SF6, and balloons. Most wind and temperature data have been processed, and the tracer data will become available shortly. High concentration of pollutants are expected to occur during nocturnal periods since turbulent mixing is suppressed, and the drainage flows that frequently develop are shallow. Therefore, the majority of experiments were conducted during the nighttimes.

The ASCOT program also supports modeling of transport and diffusion of energy-related pollutants over terrain typical of areas of energy development. For example, Yamada (1981) used a three-dimensional mesoscale hydrodynamic model based on simplified second-moment turbulence closure equations to simulate development of nocturnal drainage flow over the California Geysers area. The effects of trees were not considered in the model for reasons of simplicity. Qualitative simulation of drainage flow, horizontal convergence of winds in the valley, and resulting vertical motions were satisfactory when compared with initial data collected during 1979 summer. On the other hand, observed wind speeds were much smaller and more uniform with height. It was almost certain that observed drainage flows had been decreased considerably by drag induced by trees surrounding most observational sites. Therefore, it was concluded that in order to improve model prediction, effects of tall trees must be considered in the model.

Earlier numerical models for canopy flow simulations (e.g., Inoue, 1963; Cionco, 1965; Barr 1971) have used mixing-length and eddy viscosity concepts. These models could obtain mean wind speeds which agree well with observations if appropriate forms of eddy viscosity and mixing length profiles are known. A closure problem, i.e., to express Reynolds stresses (in equations of mean motion) in terms of mean wind has been a central issue and an as yet unresolved problem in the turbulence theory. Alleviating difficulty in obtaining appropriate eddy viscosity profiles for a canopy flow, Wilson and Shaw (1977) have proposed to use second-moment turbulence closure equations (e.g., Donaldson, 1973; Meller, 1973) which are reasonably well established for various geophysical fluid problems. Wilson and Shaw still have to employ mixing-length concept to calculate turbulent length scales.

All of the above canopy models have assumed neutral stratification and sought steady-state solutions. Neutral stratification in the lower atmospheric boundary layer, if it occurs, lasts only for a short period of time during the transitions that take place around sunset and sunrise. Steady state may be approximately satisfied for several hours around midafternoon and midnight. However, steady state and neutral stratification rarely occur simultaneously. Furthermore, previous models have been applied only in the very low part (up to twice the height of a canopy) of the atmospheric boundary layer whose height could become as high as 2 km or more during daytime.

In the present study, effects of a tall tree canopy on wind speeds, temperatures and turbulence throughout the entire depth of the atmospheric boundary layer are examined with a simplified turbulence closure model (Yamada, 1981). Diurnal variations of the above variables are computed by taking solar radiation balance in a canopy into consideration. Model equations, initial and boundary conditions, and results are discussed in the following sections.

2. Model equations

Model equations are similar to those used in Yamada (1981) except that form drag due to trees and a heat energy balance in a canopy are added as discussed in detail in the following sections.

a) Form drag due to trees

The equations for the mean motion and Reynolds stresses with the Boussinesq approximation are written as (e.g., Busch, 1973; Meller, 1973),

\[ \frac{\partial U_i}{\partial t} + \frac{\partial}{\partial x_k} (U_k U_i + u_k u_i) + \varepsilon_{k} f_k U_i \\
= - \frac{1}{\rho} \frac{\partial P}{\partial x_i} + \left[ 1 - \beta (\Theta_e - \langle \Theta_e \rangle) \right] g_i \\
+ \nu \frac{\partial^2 U_i}{\partial x_i^2} - \frac{\partial P}{\partial x_i} + \nu \frac{\partial^2 u_i}{\partial x_i^2} \\
+ f_k (\varepsilon_{k} u_k u_j + \varepsilon_{k} u_i u_j) \\
- \beta (g U_i \theta_e + g u_i \theta_e) + \frac{P}{\rho} \left( \frac{\partial U_i}{\partial x_i} + \frac{\partial u_i}{\partial x_i} \right), \quad (1) \]

and

\[ \frac{\partial u_{ij}}{\partial t} + \frac{\partial}{\partial x_k} \left( U_k u_{ij} + u_k u_{ij} - \nu \frac{\partial}{\partial x_k} u_{ij} \right) \\
= \frac{1}{\rho} \frac{\partial}{\partial x_k} (p u_{ij}) + \frac{1}{\rho} \frac{\partial}{\partial x_i} (p u_{ij}) \\
+ f_k (\varepsilon_{k} u_k u_{ij} + \varepsilon_{k} u_i u_{ij}) \\
= - u_{ki} \frac{\partial U_j}{\partial x_k} - u_{kj} \frac{\partial U_i}{\partial x_k} \\
- \beta (g u_i \theta_e + g u_i \theta_e) + \frac{P}{\rho} \left( \frac{\partial u_i}{\partial x_i} + \frac{\partial u_j}{\partial x_j} \right). \]
where the variables given by the lower case letters are fluctuations from the mean values represented by the upper case letters, $U_i$ is the wind component in $x_i$ direction, $\theta_v$ virtual potential temperature, $p$ pressure, $\langle \rho \rangle$ air density averaged over a horizontal plane, $g_l=(0, 0, -g)$ the acceleration due to gravity, $f_k=(0, f_y, f)$ the Coriolis parameter, $\beta$ the coefficient of thermal expansion, $\varepsilon_{ijk}$ alternation tensor, and the over-bars indicate ensemble averages.

The last two terms on the right hand sides of Eqs. (1) and (2) would vanish if $p$ and $u_i$ were continuous functions, which is the case without a canopy. Wilson and Shaw (1977) argue that these variables are not necessarily continuous across a solid boundary and impose a form drag on the fluid motion. Wilson and Shaw interpreted the overbars in Eqs. (1) and (2) as either spatial or temporal averages. However, their argument appears to be equally valid if the overbars signify the ensemble averages. Realizations (time record of a variable over an interval of time large enough to include all time scales of significant fluctuations) of $p$ and $u_i$ obtained in a canopy flow should show discontinuities since solid obstacles (leaves, stems and branches of trees) flutter around a sensor.

For reasons of simplicity, Wilson and Shaw assumed that pressure forces contributed the major portion of the total drag of a canopy, and the form drags appearing in (1) and (2) were modeled as

$$\frac{\partial p}{\partial x_i} = C_{d}(z) U_i |U|,$$  \hspace{1cm} (3)

where $C_d$ is a drag coefficient for a canopy, $a(z)$ the plant area density and the absolute sign insures that the direction of the drag force is always opposite to the wind direction.

The second-moment equation (2) contain third-order moments such as pressure wind-shear correlations, triple moments, and dissipation. Following the discussion in Mellor (1973), we model these higher moments in terms of second moments. Rotta's hypothesis (Rotta, 1951) for correlation of pressure and wind-shear is used, along with Kolmogoroff's isotropy hypothesis (Kolmogoroff, 1941) for dissipation, and a down-gradient diffusion assumption for triple moments. The resulting equations are similar to Eqs. (19)-(21) in Mellor except that the form drag terms are added here. Subsequently Mellor and Yamada (1974) proposed a hierarchy of turbulence-closure models. More recent studies by Yamada (1977, 1981) indicate that a simple model, which solves prognostically only a turbulence energy equation and a turbulence length scale equation, simulates well the Wangara data (Clark et al., 1971) and nocturnal drainage flows in the California Geysers area. The model is tentatively referred to as "the Level 2.5 model" since the complexity of the Level 2.5 model lies between the Level 2 and Level 3 models of Mellor and Yamada. The Level 2.5 model equations with boundary layer approximation have been given elsewhere (Yamada, 1981), but they are repeated here since form drag terms are added.

For reasons of simplicity, a horizontally homogeneous canopy is considered. Then, the governing equations are reduced to one-dimensional in vertical direction. Equations of mean motion are

$$\frac{\partial U}{\partial t} - fV = - \frac{1}{\langle \rho \rangle} \frac{\partial p}{\partial x} + \frac{\partial}{\partial z} (-uw) - \gamma C_{d}(z) U |U|,$$  \hspace{1cm} (4)

and

$$\frac{\partial V}{\partial t} + fU = - \frac{1}{\langle \rho \rangle} \frac{\partial p}{\partial y} + \frac{\partial}{\partial z} (-vw) - \gamma C_{d}(z) V |V|,$$  \hspace{1cm} (5)

where $U$ and $V$ are mean wind components in the $x$ and $y$ directions, respectively, $uw$ and $vw$ are the Reynolds stresses; $\gamma$ is a fraction of an area covered with trees (Fig. 1).

A turbulence energy equation is given by

$$\frac{\partial (q^2/2)}{\partial t} = \frac{\partial}{\partial z} \left[ q_l S_q \frac{\partial (q^2/2)}{\partial z} \right] - uw \frac{\partial U}{\partial z} - vw \frac{\partial V}{\partial z} + g \frac{q_b}{B_1} - 2 \gamma C_{d}(z) |U|^3 + |V|^3,$$  \hspace{1cm} (6)

and a turbulence length scale $l$ is obtained from

$$\frac{\partial (q^2 l)}{\partial t} = \frac{\partial}{\partial t} \left[ q_l S_q \frac{\partial (q^2 l)}{\partial z} \right] + l E_1 \left[ -uw \frac{\partial U}{\partial z} - vw \frac{\partial V}{\partial z} + g \frac{q_b}{B_1} \right] - q^2 \frac{1}{B_1} \left[ 1 + E_1 \left( \frac{l}{k_z} \right)^2 \right] + 2 \gamma C_{d}(z) |U|^3 + |V|^3,$$  \hspace{1cm} (7)

where $q^2 = u^2 + v^2 + w^2$ is twice the turbulence kinetic energy, $w \theta_v$ kinematic turbulence buoy-
ancy flux, \( \theta_v \), the fluctuation part of virtual potential temperature, and \( E_1, E_2, S_q \) and \( B_1 = (1.8, 1.33, 0.2, \text{ and } 16.6) \) empirical constants determined from laboratory experiments (Mellor and Yamada, 1981).

b) Solar radiation in a canopy

The net solar radiation \( R_{Nh} \) at treetop is given by

\[
R_{Nh} = (1 - \alpha_t)S + R_{Lh} \downarrow - R_{Lh} \uparrow,
\]

where \( \alpha_t \) is the tree albedo, \( S \) the direct solar radiation, \( R_{Lh} \downarrow \) the incoming longwave radiation, and \( R_{Lh} \uparrow \) the outgoing longwave radiation. \( R_{Lh} \uparrow \) is computed from

\[
R_{Lh} \uparrow = \varepsilon T^4 + (1 - \varepsilon_t)R_{Lh} \downarrow,
\]

where \( \varepsilon_t \) is the emissivity of trees, and \( T \) is the Stephan-Boltzman constant. The net radiation in a canopy is assumed to be that given by Uchijima (1961), quoted in Ross (1975, p. 52).

Now Eq. (11) may be integrated using (12a, b, c, d). A schematic profile of \( L(z) \) is also shown in Fig. 2.

The heat energy balance equation within a canopy may be given as

\[
H = \text{Sensible heat flux}, \quad LE = \text{Latent heat flux}, \quad S_t = \text{Energy stored in main stems and branches}, \quad P_s = \text{Energy used in photosynthesis processes in plants}.
\]

Bradley et al. (1980) indicate that the heat storage in a plantation of 37-year-old pine trees could be as large as 10% of the net radiation. The heat energy stored is expected to warm the air during the nighttime in much the same way as heat energy stored in soil heats the air. As a first approximation, however, the storage term is neglected by assuming that all the energy stored in the stems and branches becomes available immediately to heat adjacent air. Similarly the energy used in photosynthesis could become a significant portion of the net radiation, but is neglected here for simplicity.

Under the conditions assumed, the internal heat energy equation is written as
\[
\frac{\partial \Theta}{\partial t} = \frac{\partial}{\partial z} \left(-w \Theta\right) + \frac{(1-\gamma)}{\rho C_p} \frac{\partial R_N}{\partial z} + \frac{\gamma}{\rho C_p} \left(1 + \frac{1}{B}\right)^{-1} \frac{\partial R_{NP}}{\partial z}, \quad (14)
\]
where the Bowen ratio \(B = -H/LE\) in a canopy is assumed to be constant with height which appears to be supported by the observations of Bradley et al. (1980). The longwave radiation flux \(R_N/\rho C_p\) without tree is computed according to Sasamori (1968). The last term on the right hand side of Eq. (14) is obtained first by differentiating (10) with respect to \(z\) then substituting \(L(z)\) from (11). The results are obtained after some algebraic manipulations as,

\[
\frac{\partial R_{NP}}{\partial z} = \begin{cases} 
\gamma R_{NH}[k \exp(-kL(z))][a_{L,\max}(1-z_i)^{-1} + a_{S,\max}(1-z/h)] + \exp(-kL(0)) \gamma/h & \text{for } z_i < z/h \leq 1.0, \\
\gamma R_{NH}[k \exp(-kL(z))][a_{L,\max}(h-z_i)(1-z_i)^{-1} + a_{S,\max}(1-z/h)] + \exp(-kL(0)) \gamma/h & \text{for } z_i < z/h \leq z_2, \\
\gamma R_{NH}[k \exp(-kL(z))][a_{S,\max}(1-z/h)] + \exp(-kL(0)) \gamma/h & \text{for } 0 \leq z/h \leq z_2,
\end{cases}
\]

where

\[
L(z/h) = \begin{cases} 
(h/2)[a_{L,\max}(1-z_i)^{-1} + a_{S,\max}(1-z/h)^2] & \text{for } z_i < z/h \leq 1, \\
(h_z a_{L,\max}/2)[(1-z_i) + (z_i^2 + z_i z_2 - z/h)^2 + z_2(z/h)(1-z_i)] + (h a_{S,\max}/2)(1-z/h)^2 & \text{for } z_i \leq z/h \leq z_2, \\
(h_z a_{S,\max}/2)(1-z_2) + (h a_{S,\max}/2)(1-z_2)^2 & \text{for } 0 \leq z/h \leq z_2.
\end{cases}
\]

A conservation equation for the mixing ratio of water vapor is given by

\[
\frac{\partial Q_w}{\partial t} = \frac{\partial}{\partial z} (-w Q_v) + S, \quad (17)
\]
where \(S\) represents a source of water vapor due to evapotranspiration of plants. Distribution of \(S\) varies with various factors such as kind, age and structure of trees, amount of soil moisture, stomata openings, photosynthetic activities and turbulent air motion adjacent to leaves. We feel that there are still too many unknowns and uncertainties to include \(S\) in the present study, thus it is neglected here.

Relationships between the turbulent fluxes in Eqs. (4), (5), (6), (7), (14) and (17) with the variables \(U, V, \theta, Q_v, q\) and \(l\) are discussed in detail elsewhere (e.g., Yamada 1978), and are not repeated here.

3. Boundary conditions

Surface boundary conditions for Eqs. (4), (5), (6), (7), (14), and (17) are constructed from the empirical formulas by Dyer and Hicks (1970) for the nondimensional wind and temperature profiles (see the Appendix in Yamada, 1981). These formulas are valid only for horizontally homogeneous surfaces. It is assumed, however, that the same relations are fair approximations in a canopy, provided that the formulas are applied sufficiently close to the surface (\(z < 3\) m). It should be noted that, when \(0 < \gamma < 1\), the values computed are the average values over the surface where trees and open areas are uniformly distributed.

The soil temperature \(T_s\) is obtained by solving a heat conduction equation,

\[
\frac{\partial T_s}{\partial t} = \frac{\partial}{\partial z} \left( k_s \frac{\partial T_s}{\partial z} \right), \quad (18)
\]
where \(z_s\) is positive downward, and \(k_s\) is the conductivity of soil. Measurements during the O'Neil Experiment (Lettau and Davidson, 1957) indicate that soil conductivities vary almost linearly with depth. An average profile determined from all seven observational periods is adopted here;

\[
k_s = (0.15 + 4z_s) \times 10^{-6}, \quad (19)
\]
where \(k_s\) is in \(m^2 s^{-1}\) and \(z_s\) in \(m\). Boundary conditions for Eq. (18) are the heat energy balance at the soil surface and specification of the soil temperature at 50 cm. The heat-energy balance at the surface is given by

\[
R_{NG} = \gamma R_{NH} \exp(-kL(0))(1-\gamma) + (1-\gamma)(1-a_G)S + R_{LG\downarrow} - R_{LG\uparrow} = H_G + L_{E_G} + G_S, \quad (20)
\]
where \(R_{NG}\) is the net radiation, \(a_G\) the surface albedo, \(R_{LG\downarrow}\) the incoming longwave radiation, \(R_{LG\uparrow}\) the outgoing longwave radiation, \(H_G\) the heat flux, \(L_{E_G}\) the latent heat flux and \(G_S\) the soil heat flux. The subscript "G" indicates the
values at the ground.

The direct solar radiation $S$ in Eqs. (8) and (20) is computed from

$$ S = S_m \cos Z, \quad (21) $$

where $S_m$ is the near surface, direct solar radiation, and $Z$ is the zenith angle of the sun. The zenith angle is determined from the following formula;

$$ \cos Z = \sin \phi \sin \delta + \cos \phi \cos \delta \cos H, \quad (22) $$

where $\phi$ is the latitude, $\delta$ is the declination of the sun, and $H$ is the solar hour angle, positive clockwise from apparent noon. Since the maximum change in the solar declination $\delta$ in 24 hr. is less than 0.5 degrees, $\delta$ is assumed to be constant during a given day. Spencer (1971, quoted in Paltridge and Platt, 1976, p. 63) provides a formula to compute $\delta$ in radians,

$$ \delta = 0.006918 - 0.399912 \cos \theta_0 + 0.070257 \sin \theta_0 $$

$$ - 0.006758 \cos 2\theta_0 + 0.000907 \sin 2\theta_0 $$

$$ - 0.002697 \cos 3\theta_0 + 0.001480 \sin 3\theta_0, \quad (23) $$

where the angle $\theta_0$ in radians is related to the Julian date $J_d$ by

$$ \theta_0 = 2\pi (J_d - 1)/365, \quad (24) $$

Equation (23) estimates $\delta$ with a maximum error of 0.0006 radians. Solar hour $H$ can be obtained if the longitude, clocktime and the equation of time are known. The equation of time is the difference between the local apparent time and a fixed mean solar time which is derived from the motion of a celestial equation at a rate equal to the average movement of the sun. The solar hour angle $H$ is given in radians by

$$ H = \pi(t_s - 12)/12, \quad (25) $$

where $t_s$ is the true solar time (local apparent time) in hours. The true solar time is obtained from

$$ t_s = t_{c.t.} + \Delta t_{long} + t_{eq}, \quad (26) $$

where $t_{c.t.}$, $\Delta t_{long}$ and $t_{eq}$ are the clocktime, the longitude correction and the equation of time, respectively. The longitude correction accounts for the difference between the local meridian and a standard meridian, and is positive if the local meridian is east of the standard. The equation of time is provided by Spencer (quoted in Paltridge and Platt, 1976, p. 63) as follows;

$$ t_{eq} = \frac{12}{\pi}(0.000075 + 0.001868 \cos \theta_0 $$

$$ - 0.032077 \sin \theta_0 - 0.014615 \cos 2\theta_0 $$

$$ - 0.040849 \sin 2\theta_0), \quad (27) $$

where $t_{eq}$ is in hours and $\theta_0$ is defined by Eq. (24). Equation (27) has a maximum error, compared with values tabulated in the National Almanac, of 35 s in time.

The amount of solar radiation reaching the surface is much less than that at the top of the atmosphere due to many factors, including molecular scattering and absorption by permanent gases such as oxygen, ozone, and carbon dioxide. This effect is parameterized by Atwater and Brown (1974), who modified the original form by Kondrat'yev (1969) to include the effect of the forward Relyeigh scattering. The expression is

$$ G = 0.485 $$

$$ + 0.515 \left[ 1.041 - 0.16 \left( \frac{0.000949 P + 0.051}{\cos Z} \right)^{1/2} \right], \quad (28) $$

where $P$ is pressure in mb. Other important factors that also modify the amount of incoming solar radiation include water vapor, clouds and air-borne particles. Parameterizations for these factors are not included in the present model.

Currently $S_m$ in (21) is calculated from

$$ S_m = S_0 G, \quad (29) $$

where $S_0$ is the incoming radiation flux at the top of the atmosphere and $G$ is given by Eq. (28).

The outgoing longwave radiation $R_{LG\uparrow}$ in (20) is computed from

$$ R_{LG\uparrow} = \varepsilon \sigma T^4 + (1 - \varepsilon)R_{LG\downarrow}, \quad (30) $$

where $\varepsilon$ is the emissivity of the ground.

The surface heat flux $H_G$, latent heat flux $L_E G$ and soil heat flux $G_S$ are given by

$$ H_G = - \rho_0 C_p u_\ast T_\ast, $$

$$ L_E G = - \rho_0 L_u Q_\ast, $$

and

$$ G_S = - k_s \partial T_s/\partial z_s|_G, \quad (31a, b, c) $$

where $\rho_0$ is the air density, $u_\ast$ the friction velocity, $T_\ast$ the temperature scale and $Q_\ast$ the water vapor scale. Substituting (31a, b, c) into (20) we obtain

$$ \gamma R_{NH} \exp \left[ - kL(0) \right] (1 - \gamma) $$

$$ + (1 - \gamma) \left[ (1 - \alpha_G)S + \varepsilon \sigma R_{LG\downarrow} - \varepsilon \sigma T^4 \right] $$

$$ = - \rho_0 C_p u_\ast T_\ast (1 + B) - k_s \partial T_s/\partial z_s|_G, \quad (32) $$

where Eq. (30) is substituted for $R_{LG\uparrow}$. Equation (32) is equivalent to Eq. (21) in Yamada (1981) if $\gamma = 0$ (no trees). Equation (32) provides the surface boundary condition to solve the heat conduction equation (18). Procedures for nu-
Numerical solutions have been discussed in detail between Eqs. (21) and (29) in Yamada (1981), and are not repeated here except to provide the final expressions to replace Eqs. (30) and (31) in Yamada, i.e.,

$$E_i = (k_i / \Delta z_i) / [4 \varepsilon G(1 - \eta) \alpha (T_0)^{\eta}] + m (P_0 / P_C)^{R/C_p} + k_i / \Delta z_i, \quad (33)$$

and

$$F_i = [\eta (1 - \eta) R_{NG} \exp (-k L(0)) + (1 - \eta) [1 - \alpha \sigma + \varepsilon G R_{LG} \sigma] \varepsilon G \sigma (T_0)^{\eta} + m \theta (z_i)] \sigma (T_0)^{\eta} + m (P_0 / P_C)^{R/C_p} + k_i / \Delta z_i], \quad (34)$$

It should be noted that the incoming longwave radiation $R_{Lh}$ at the treetop and $R_{LG}$ at the surface are computed according to Sasamori (1968).

Boundary conditions for $U$, $V$, $T$, $q$, and $l$ along the upper computational boundary ($z = 2,000m$) are

$$\langle U, V \rangle = \langle U_g, V_g \rangle, \quad (35a, b)$$

where $U_g$ and $V_g$ are geostrophic wind components defined as

$$\langle U_g, V_g \rangle = (1/f) (-\partial P / \partial y, \partial P / \partial x). \quad (36a, b)$$

Potential temperature and the mixing ratio are specified, and turbulence is assumed to vanish along the upper boundary. Soil temperature at 50 cm below the surface is also specified. The constant values used in the simulations are discussed in the following section.

4. Initial values

Horizontal wind component $U$ is logarithmic ($u_s = 0.4 \text{ m s}^{-1}$, and $z_0 = 0.01 \text{ m}$ are assumed initially) until it reaches a westerly geostrophic wind of 10 ms$^{-1}$, and is constant above that level. The north-south wind component $V$ is assumed initially zero. The air temperature is adiabatic in the first 1,600 m above the ground (303 K, initially) and increases above at a rate of approximately 10 K km$^{-1}$. The initial water vapor content is computed assuming the relative humidity of 30% using the initial temperature profile and the hydrostatic approximation. In the subsequent simulations, the surface water vapor is computed by assuming a relative humidity of 50%. The geostrophic wind is assumed to be westerly 10 ms$^{-1}$ at 2 km and 12 ms$^{-1}$ at the ground, i.e., a thermal wind of $-1 \text{ m s}^{-1} \text{ km}^{-1}$ is assumed. The geostrophic wind is constant with time. The initial turbulence length scale $l$ increases with height ($= k z$ where $k$ is Von Karman's constant, 0.4) until it reaches 50 m, and then decreases linearly with height, and vanishes at the upper computational boundary. All the turbulence variables are initialized by using the Level 2 model of Mellor and Yamada (1974). The Level 2 model is an algebraic model obtained by neglecting time dependent, advection and diffusion terms in the second-moment equations.

Numerical values of the parameters $\alpha_1$, $\alpha_2$, $\varepsilon_0$, $\varepsilon_1$, $\varepsilon_2$, etc. used in the actual simulations are given following a definition of each parameter in the Nomenclature.

5. Numerical procedures

Mean wind, temperature and turbulence vary greatly with height near the surface and treetops, but much less so away from the canopy. In order to resolve these variations, vertical grid levels are determined according to the following relationship between the grid level ($\xi$) and the vertical coordinate ($z$):

$$z = a (\xi - 1) \quad \text{for} \quad 1 \leq \xi \leq \xi_{\text{lin}},$$  

and

$$z = a \xi^2 + b \xi + c \quad \text{for} \quad \xi_{\text{lin}} < \xi \leq \xi_{\text{max}},$$

where the constants $(a, b, c) = (0.3869, -6.512, 44.814)$ are determined to satisfy a continuity at $\xi = \xi_{\text{lin}}$ and a boundary condition $z = z_{\text{max}}$ at $\xi = \xi_{\text{max}}$. Numerical values of $\xi_{\text{lin}} = 11$, $\xi_{\text{max}} = 80$ and $a = 2$ are used, resulting in a constant grid interval (2 m) between the surface and 20 m above the ground. The grid interval gradually increases above that level and becomes 45 m near the upper boundary (Table 1). In order to increase the accuracy of finite-difference approximations and to suppress computational noise, mean and turbulence variables are defined on a staggered grid. Turbulence is defined at the levels given by $z$ in Table 1 while the mean variables are at $z_m$ which is also determined from Eqs. (37a, b) by substituting $\xi = 1.5, 2.5, \ldots$ etc. The locations of the vertical grid levels for the soil layers are determined according to a log-linear relationship between the grid level and the vertical coordinate (Table 1).

The prognostic equations (4, 5, 6, 7, 14, and 17) are approximated by a finite-difference expression by Laasonen (Eq. 31 in Yamada and Mellor, 1975). The finite difference equation is solved by a standard Gaussian elimination method (Richtmyer and Morton, 1967, p. 198).
6. Results and discussions

a. Mean variables

The horizontally homogeneous canopy assumed here is an exception rather than a rule in the real atmosphere. Temperature differences of the order of 1°C in the average air temperature at crown height over a separation of the order of 30m are often observed over an apparently uniform forest (e.g., Bergen, 1974). Such temperature differences could generate local wind speeds of the order of 10 cm s\(^{-1}\), and the assumption of horizontal homogeneity becomes questionable. Temperature distributions would be much more inhomogeneous over a forest in complex terrain area, a situation where most forests are located. Therefore, the present results should be interpreted with caution as the values averaged over a relatively large area so that small scale, local inhomogeneities are minimized. The results should be compared only qualitatively with observations. Nevertheless the model results appear to be quite reasonable and useful, particularly in understanding the complex effects of tall trees on air circulation in the lower atmosphere. The role of tall trees is clearly demonstrated when model results are compared with those obtained without trees.

All the simulations presented here assume solar radiation in late July (Julian day=204). Integration started at 0000 LST and lasted for four diurnal cycles. After the second day, diurnal variations become similar but not exact; for example, the upper boundary layer is warmed constantly (\(\sim 2 \text{ K day}^{-1}\)) since warming due to solar radiation exceeded the cooling due to longwave radiation. Constant warming would not have occurred if advection terms had been included. On the other hand, wind speeds and particularly turbulence kinetic energy appear to attain similarity in diurnal variations much more quickly than potential temperature. Simulations shown here are for the last two days of a four-day simulation. All the results discussed here are, unless noted otherwise, obtained by assuming \(\gamma=0.9\) for the canopy and \(\gamma=0\) for bare soil. The Bowen ratio in a canopy was estimated to be 1.5 from the vertical profiles of sensible and latent heat fluxes reported by Bradley et al. (1981). Initial and boundary conditions are exactly the same for all cases and given in the previous sections. The height of a tree is 20 m.

Diurnal variations of computed potential temperature over bare soil (Fig. 3) display deep unstable layers during the day and shallow stable layers near the ground during the night, which agree with observations (e.g., Clarke et al., 1971). On the other hand, potential temperatures for a canopy case (Fig. 4) show unstable layers within a canopy during the period between sunset (1900 LST) and sunrise (0500 LST). This results from rapid cooling of treetops by longwave radiation. For a denser forest (\(\gamma=1\)), potential temperatures (Fig. 5) show stable layers within a canopy during the day, presumably caused by reduced solar radiation at the surface. Maximum temperatures occurred at approximately 6 m below the treetop where the heating function \(\frac{\partial R_{\text{net}}}{\partial z}\) (Eq. 15) in the present model attains a maximum. The location of the maximum heating is a function of an extinction coefficient in the exponential profile assumed for the net radiation in a canopy (Eq. 15) and the profile of a leaf area density function \(a_L(z)\) in Eq. 12. The height of maximum temperature remains almost constant with time during the most of the daytime, and a strong stable layer develops below the maximum. The temperature difference between the maximum and that at 1 m above the ground is on the order of 10°C in this case. Such strong temperature gradients (\(\sim 8^\circ\text{C}\)) have been observed in a very dense loblolly pine plantation (Hosker et al., 1974). Whether the temperature maximum occurs or not and how long it lasts appear to strongly depend on the amount of tree coverage. For

<table>
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Fig. 3 Diurnal variations of simulated potential temperatures with $\gamma = 0$ for the last 48 hours of a 4-day simulation. The values on the contour lines ($\Theta - 273$) are in units of degree K.

Fig. 4 Same as in Fig. 3 except $\gamma = 0.9$.

Fig. 5 Same as in Fig. 3 except $\gamma = 1.0$.

example, a simulation with tree coverage of 90% did not produce maxima (Fig. 4). Significantly high turbulence generated in the upper part of a canopy transported heat from above, resulting in uniform temperature profiles within a canopy. The role of turbulent mixing will be discussed in detail in the latter part of this section. Since potential temperatures are expected to be sensitive to the Bowen ratio used in the model, an additional simulation is performed (Fig. 6) where the same conditions as in Fig. 5 are used except that the Bowen ratio is reduced to 0.5 from 1.5. The lower Bowen ratio means less heat energy is available to heat the air since more heat energy is used for evaporation. As expected, the computed potential temperatures for $B = 0.5$ (Fig. 6) are significantly lower (4-5°C) than those for $B = 1.5$ (Fig. 5).

Concurrent diurnal variations of wind speed for a bare soil case (Fig. 7) show nocturnal low-level wind maxima of 12-14 ms$^{-1}$ due to inertial oscillation of wind vectors around the geostrophic wind (Blackadar, 1957). The inertial oscillation is retarded during the daytime by strong vertical
turbulent mixing resulting in relatively uniform wind profiles. Results for a canopy case (Fig. 8) are considerably different from the bare soil case. Wind speeds within a canopy show almost no diurnal variations. Computed wind speeds in the mixed-layer during the day are much lower, and nocturnal low-level wind maxima are much stronger than the counterparts for a bare soil case. Computed strong wind shears near the treetop are consistent with observations (e.g., Bergen, 1971; Oliver, 1971). Wind shears are enhanced for a canopy case since wind speeds within a canopy are reduced considerably due to tree drag, and computed nocturnal low-level wind maxima are stronger. The magnitude of the wind maxima are determined, if the other conditions remain the same, by the difference between the actual wind speed at sunset and geostrophic wind. This difference determines a diameter of inertial oscillation of wind vectors (Blackadar, 1957). Computed wind speeds (Fig. 8) during the day for a canopy case are much smaller than those for a bare soil case (Fig. 7) since considerably more mean kinetic energy than...
in the case of bare soil is transferred to turbulence kinetic energy and dissipated in heat to the atmosphere. Diurnal variations of computed wind speeds for \( \eta = 1 \) differ only slightly from those in Fig. 8 (\( \eta = 0.9 \)), and are not shown here.

### b. Turbulence variables

Since most observations for a canopy flow were designed mainly for the measurements of the mean variables, very little data are available to compare with the computed turbulence. Nevertheless the computed turbulence information is useful in enhancing understanding the role of tall trees on air circulation in the lower atmosphere. We also believe that no significant improvements in prediction of mean variables can be made without correct turbulence parameterization since mean and turbulence variables are strongly coupled with each other, as demonstrated in the present study.

Diurnal variations of computed (twice) the turbulence kinetic energy are shown in Fig. 9 (\( \eta = 0 \)) and Fig. 10 (\( \eta = 0.9 \)). Maximum turbulence kinetic energy for a canopy case (Fig. 10) is approximately twice as large as that for a bare soil case (Fig. 10). Direct comparison of computed turbulence kinetic energy with observations is not possible, but the profiles which increased almost linearly with height within a canopy are consistent with a \( \sigma_u \) profile obtained in a pine forest (Bradley et al., 1980). On the other hand, computed turbulence kinetic energy for a bare soil case is much more uniform with height (Fig. 9). Significant level of turbulence in a canopy is obtained even during the night. Turbulence is generated and maintained by the strong wind shears near the treetop (Fig. 8) and by unstable temperature layers within a canopy (Fig. 4). In other words, a tall tree is an efficient generator of mechanical as well as thermal turbulence.

Total Reynolds stress \( \tau = [-u'w' + \bar{u}' \bar{w}']^{1/2} \) is computed, and the results are shown in Fig. 11 (\( \eta = 0 \)) and Fig. 12 (\( \eta = 0.9 \)). The Reynolds stress for a canopy case is, as for the turbulence kinetic energy, much larger than that for a bare-soil case. The large Reynolds stress and strong wind shear at treetop are responsible for generation of mechanical turbulence, particularly during the night. Similarly computed eddy viscosity coefficients \( K_M = -u'w'/U/ \partial \bar{U} / \partial z \) are shown in Fig. 13 (\( \eta = 0 \)) and Fig. 14 (\( \eta = 0.9 \)). Computed eddy viscosity coefficients in Figs. 13 and 14 shows considerable similarities in varia-
tions and magnitudes despite the fact that both turbulence kinetic energy and Reynolds stresses are markedly different for the two cases. However, there is a slight difference in the values during the day in the lower part of a canopy where a canopy case is smaller than a bare-soil case due to differences in stability. Finally, computed turbulence length scales are shown in Fig. 15 ($\gamma=0$) and Fig. 16 ($\gamma=0.9$). Both figures show that the length scale increases with height, and reaches $\sim 20\text{ m}$ at approximately $200\text{ m}$ above the ground. The length scale decreases above that level and vanishes at the upper computational boundary, in accordance with the boundary condition imposed there. A vertical profile of a length scale has yet to be determined observationally. Limited data appear to suggest a maximum at between $150$ and $200\text{ m}$ above the ground followed by either a more or less constant value (Blackadar, 1962) or a decrease with height (Clarke, 1970). Blackadar's interpolation formula for a neutral atmosphere (Blackadar, 1962) and a modified expression for a stratified atmosphere (Yamada and Mellor,
have been successfully used to simulate some observational data. However, direct application of the interpolation formulas for a canopy case appears to be unphysical since the interpolation formulas mentioned above consider no effects of mechanical and thermal turbulence produced by trees. On the other hand, the present equation for a turbulence length scale (Eq. 7) has more physics than the interpolation formulas, and apparently worked well in previous simulations (Briggs et al., 1977; Yamada and Mellor, 1979). In general the prognostic length scale equation has produced the results that are less sensitive to diurnal variations than the interpolation formula previously used (e.g., see Fig. 23 in Yamada and Mellor, 1975).

7. Summary and conclusions

The motivation of the present study is the need to improve model simulations of transport and diffusion of pollutants over complex terrain area covered with tall trees, particularly the California Geyser area where the ASCOT field experiments have been conducted. As a first step toward the
goal, a three-dimensional hydrodynamic model was used to simulate terrain effects on the air circulations in the lower atmosphere (Yamada, 1981). In the present paper, effects of trees are considered separately from the terrain effects.

The present model differs considerably in various aspects from earlier canopy models. First the model is based on simplified second-moment turbulence closure equations which have been reasonably well tested in various geophysical fluid problems (Mellor and Yamada, 1981), thus alleviating the difficulty in formulating an eddy viscosity expression appropriate for a canopy flow. Second the model simulates complete diurnal variations of wind, temperature and turbulence. Most earlier models sought steady state solutions in a neutral atmosphere. Third simulations are conducted for the entire depth of the atmospheric boundary layer, differing earlier studies that were applied only in the surface layer (<50 m).

Surface vegetations, particularly tall trees, present considerable difficulties in formulating proper boundary conditions needed to solve governing equations. Difficulties are associated mainly with uncertainties in a momentum balance (additional surface drag induced by leaves, stems and branches), a heat energy balance (interruption of incoming solar radiation from reaching the ground, energy stored in a canopy, energy used in photosynthesis), and a water vapor balance (evapotranspiration, soil moisture). Only the momentum and heat energy balances are considered in this study.

The model results are applied only for horizontally homogeneous canopies which are difficult to find in the real atmosphere. Therefore, the present results should be interpreted with caution as the values averaged over a relatively large area so that small scale, local inhomogeneities are minimized, and are compared only qualitatively with observations. Nonetheless model results appear to be quite reasonable: nearly constant and low wind speeds in a canopy, large wind shears near treetops and unstable (stable) temperature layers during the night (day) within a canopy are all reproduced.

Interesting results are obtained when the model simulations with and without a canopy are compared with each other. Computed wind speeds within a canopy are, as expected, always much smaller due to surface drag than those without a canopy. However, nocturnal low-level wind maxima over a canopy are found to be considerably larger than those without a canopy. The magnitudes of the wind maxima are determined by the difference between the actual wind speed at sunset and geostrophic wind (Blackadar, 1957). Computed wind speeds during the day with a canopy are much smaller than the counterparts without a canopy, since considerably more mean kinetic energy than in the case of a bare-soil is transferred to turbulence kinetic energy and dissipated in heat to the atmosphere. This clearly demonstrates a strong coupling between the mean winds and turbulence, reiterating our belief that no significant improvement in prediction of mean winds can be made without a proper parameterization of turbulence. It is not necessary to emphasize a strong coupling existing between potential temperature profiles and turbulence.

Computed turbulence kinetic energy, Reynolds stress, eddy viscosity coefficients and a turbulence length scale are also presented, but no serious efforts were made to compare them with data.

The present model is by no means complete due to simplifications employed in the momentum and heat energy balance, arbitrariness in parameters used (leaf surface area density, drag coefficient, extinction coefficient, soil conductivity) and negligence of a water vapor balance in a canopy. However, the objective of the present study is fulfilled if this paper could generate some discussions on the subject and encourage experimentalists to make additional measurements of turbulence in a canopy.

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**List of Symbols and Numerical Values Used**

\[ a(z) : \text{a plant area density} = a_L(z) + a_s(z) \]
\[ a_L(z) : \text{a leaf surface area density} \]
\[ a_{L\text{max}} : \text{a maximum value of} \ a_L(z), = 0.8 \text{ m}^2/\text{m}^3 \]
\[ a_s(z) : \text{a non-leaf surface area density} \]
\[ a_{s\text{max}} : \text{a maximum value of} \ a_s(z), = 0.05 \text{ m}^2/\text{m}^3 \]
\[ B : \text{Bowen ratio} (=H/LE) \text{ in a canopy}, = 1.5 \text{ is used except in Fig. 5.} \]
\[ C_d : \text{a drag coefficient}, = 0.2 \text{ is used} \]
\(C_p\) : specific heat capacity of dry air at constant pressure, 1.003 joules g\(^{-1}\) K\(^{-1}\)

\(f\) : Coriolis parameter, \(=9.374 \times 10^{-5}\) s\(^{-1}\) for 40 deg. N is used

\(g\) : acceleration of gravity

\(G_s\) : soil heat flux

\(h\) : tree height, =20 m is used

\(H\) : sensible heat flux

\(k\) : an extinction coefficient in the exponential profile of the net radiation in a canopy, =0.6 is used

\(k_s\) : soil conductivity

\(L(z)\) : a leaf area index defined by Eq. (11)

\(LE\) : latent heat flux

\(p\) : pressure fluctuation

\(P\) : pressure

\(P_0\) : a reference pressure, 1,000 mb

\(q^2\) : twice the turbulence kinetic energy, \(=u'^2+v'^2+w'^2\)

\(Q_e\) : mixing ratio of water vapor

\(Q_{\text{water vapor scale}}\)

\(R\) : gas constant for dry air, 0.28704 joules g\(^{-1}\) K\(^{-1}\)

\(S_0\) : solar constant, =1368.07 W m\(^{-2}\)

\(T\) : absolute temperature

\(T_s\) : soil temperature

\(T_*\) : temperature scale

\(u_i\) : velocity fluctuation \((u, v, w)\)

\(U_i\) : mean velocity \((U, V, W)\)

\(u*\) : friction velocity

\(x_i\) : coordinates \((x, y, z)\)

\(z_s\) : vertical coordinate for soil layers, positive downward

\(z_0\) : roughness length of the ground, =0.01 m is used

\(z_1, z_2\) : height parameters to define \(a_L(z)\), \((z_1, z_2) = (0.5, 0.4)\) are used

\(a_G\) : a ground albedo, =0.3 is used

\(a_t\) : a tree albedo, =0.1 is used

\(\beta\) : thermal expansion coefficient, =1/\(\Theta_0\)

\(\varepsilon_G\) : emissivity of the ground, =0.98 is used

\(\varepsilon_{alt}\) : alternating tensor

\(\varepsilon_t\) : emissivity of a tree, =0.98 is used

\(\eta\) : a fraction of the area covered with trees, \(0 \leq \eta \leq 1\)

\(\Theta_0\) : virtual potential temperature, \(=(P_0/P)^{R/c_p} (1 + 0.61 Q_e) T\)

\(\langle \rho \rangle\) : horizontally averaged air density

\(\phi\) : latitude, =40°N is used

References


数値モデルによる森林内外の大気流に関する研究

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大気境界層内の空気の流れに与える森林の影響をシミュレートするために、簡単化した乱流 closure モデルを使った。このモデルは既に様々な流れ場をシミュレートするのに応用され比較的成果を挙げている。森林内での風速分布、森林上層部に起こる大きな wind shear、夜間（昼間）森林内に起こる温位不安定層（安定層）等、定性的には観測と合う結果が得られた。森林がある場合とない場合の計算結果を比べることにより、乱流と平均流との間に強い相関がある事が明白に示された。