Prediction of Flow and Diffusion Over Vegetated Regions on Hills and Ridges

Robert N. Meroney
Fluid Mechanics and Wind Engineering Program
Civil Engineering Department
Colorado State University
Fort Collins, CO 80523

David E. Neff
Fluid Mechanics and Wind Engineering Program
Civil Engineering Department
Colorado State University
Fort Collins, CO 80523

George Wu
ENECOTECH, Inc.
1580 Lincoln Street, Suite 1000
Denver, CO 80203

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INTRODUCTION

Mountainous ridges fully or partially covered by high-roughness vegetation can influence wind speeds and dispersion in the atmospheric boundary layer. Removal of the forest over ridgetops can substantially increase crest wind speeds, reduce turbulence, and inhibit flow separation. Consequently plume dispersion can also be modified substantially by forest clearcut operations. This paper reviews the behavior of flow over vegetated complex terrain and consolidates recent information concerning the amplification (or reduction) of groundlevel concentrations due to the presence of hills and vegetation covered terrain.

Wind Flow Over Hills and Ridges

Hills or ridges are known to cause wind "speed-up" associated with streamline convergence. Convergence will occur in neutrally stratified flows at hill crest; whereas maximum streamline convergence will often occur on the lee of a hill or ridge when the wind flows between the surface and an elevated inversion. Tall vegetation such as woods, wind-breaks, or forests may degrade the wind environment over a hill. Eddies created by a tree canopy can enhance surface mixing reducing near-surface wind speeds and increase the likelihood of flow separation. The eddies themselves may also cause gustiness which increases pollutant dispersal.

General Wind Speed and Turbulence Characteristics. Single hills and isolated ridges are known to produce higher wind speeds at a given height over the crest than far upwind during high speed neutrally stratified air flow. The increase in wind speed should be of the order of h/L, where h is hill height above the surrounding terrain and L is some characteristic horizontal hill width (say the distance from the crest to half-hill height). Crest-top wind speed does not seem to be sensitive to hill shape for the same average slope as long as flow separation does not occur. These values may be reduced by the presence of forests, local undulations, and regions of flow separation. In particular, when hill slopes exceed 0.3 it is likely that flow separation may occur at hill crest decreasing wind speeds and inducing large gustiness. Hill slope is more significant than hill shape in defining wind profiles (as long as the elevation is not so abrupt as to generate separation regions downwind). Three dimensional hills are found to produce lower wind speed increases than similar cross-section two-dimensional ridges.

Thermal stratification will also change the air flow over single hills and mountains substantially. Stable stratification may cause low level winds to move laterally around a hill barrier decreasing crest top winds but enhancing hill side winds, or alternatively an elevated inversion just above hill crest may induce very strong winds on the downslope side of a hill or ridge. Blumen (1990) has edited a series of articles about such flows in hills, mountains and valleys into a monograph on atmospheric processes over complex terrain.[1] In particular chapter 4 on mountain waves and downslope winds by D.R. Duran, chapter 5 on perturbation solutions to flow over hills by D.J. Carruthers and J.C.R. Hunt, and chapter 7 on physical modeling of flow over hills and mountains by R.N. Meroney are relevant reading for the air-pollution specialist interested in air-pollution meteorology over hills and mountains.

Turbulence behavior over hills depends upon upwind fetch, strength of stream line convergence over the crest, and regions of increased shear. For small shallow hills such that surface flows have limited time to come into local equilibrium with the hill presence the upwind turbulence is primarily advected along streamlines with minor changes; hence, changes in wind shear play a small role. On the other hand for somewhat steeper hills convergence and divergence of streamtubes may lead to "rapid-distortion" or stretching, twisting, or shortening of turbulent vortex elements. Stretching distortions can lead to enhanced local vorticity and increase in turbulence. When the local turbulent velocity is scaled by local mean speed at the hill crest, the net effect may be a decrease in turbulent intensity although in absolute terms the turbulent fluctuations are greater. Of course if separation over the crest occurs then elevated regions of increased turbulence downwind will be present.

Hills or mountains with small slope or long upwind fetch conditions permit the near ground flow to move toward local equilibrium with local shear. An inner boundary layer, l_z grows upward in which these effects are significant. To a first order the depth at crest should be related to $l_z \ln[l_z / z_o] = 2\kappa^2 L$, where z_o is surface roughness and L is characteristic hill half-width. For most conditions $l_z = 0.05$ L when h/L is of order one. In such conditions over moderate slope hills analytic and numerical methods based on linear-perturbation concepts work quite well.

Field and Laboratory Measurements. As a result of environmental and energy concerns there now exists a large number of field data sets related to wind flow fields over hilly and mountainous terrains. Some were performed to answer questions about nuclear power safety, others were concerned with atmospheric transport of power plant plumes, and some were posed to evaluate wind-energy potential over hill crests. Articles in Blumen (1990) refer to the U.S. Department of Energy ASCOT test series (Geysers, CA; Brush Creek, CO), the Environmental Protection Agency CTMDP test series (Cinder Cone Butte,ID; Hog Back Ridge, NM; Tracy Power Plant, NV), and various international measurement efforts like the Askervein hill project (US, NZ, UK, CN).[1] Meroney (1978, 1980) reviews pre-1980 field studies which had laboratory counterpart experiments (eg. Rakaia River Gorge, NZ; Gebbies Pass, NZ; Kahuku Point, Oahu, HA).[2,3] During the 1980s a number of field studies over relatively smooth isolated hills and ridges were performed specifically to validate wind flow models proposed to predict wind-energy potential (Askervein Hill, Blashaval, Brent Knoll, Great Dun Fell, Nyland Hill, UK; Kettle Hill, CAN). There now exists at least 69 laboratory studies for 31 cases of which comparable field measurements are available.

Almost all of the isolated hill and ridge studies examined terrain with very small surface roughness (1-3 cm); hence, there are few published data for wind flow over simple terrain shapes incorporating forests, woods, shelterbelts, clearings or clearcut areas. (This is not to say individual meteorological measurements do not exist in clearings and clearcut areas in hills, but no extensive sets of measurements were made in such situations to document the overall influence of variations in surface vegetation.) One exception was a two-dimensional ridge study performed in Australia.[4]

The ASCOT Brush Creek study involved a 650 m deep valley in western Colorado where the southwest-facing walls were dry, and barren, while the northeast-facing walls were moist and brush covered. The vegetation was found to make a significant difference in diurnal absorption of thermal radiation, and the wind flows which developed in the valley. The ASCOT Geyser area studies were performed over tree covered hills and meadow covered valleys. Again inclusion of the vegetation in analytic and numerical models was necessary to reproduce the wind flows observed. Unfortunately, no neutrally stratified wind conditions were observed in either ASCOT study, for the principal program goal was to evaluate the mechanisms of night time and daytime drainage flow in open-ended valleys.

An extensive set of physical model experiments were also performed during the 1980s. The Askervein Hill Project was a collaborative study of boundary-layer flow over low hills.[5] Two field experiments were conducted during fall 1982 and 1983 near and around Askervein, a 116 m high hill on the west coast of the island of South Uist in the Outer Hebrides of Scotland. Over 50 towers were deployed and instrumented for wind measurements. Most were cup anemometers mounted on simple 10 m posts, but two 50 m, one 30 m, one 16 m, and thirteen 10 m towers were instrumented for three-component turbulence measurements. Subsequently, wind tunnel simulations of the hill were carried out at three different length scales (1 to 800, 1 to 1200, and 1 to 2500) in two wind tunnel facilities.[6] The wind-tunnel results compared well with each other and with full scale data Changes in mean flow speedup over the physical models were reproduced very well, including those due to small local terrain features that may be physically small at model scale.

Relaxation of the aerodynamic roughness criterion ($Re_* = u_* z_o/\nu > 2.5$) affected the flow only on the lee side of the model hills. Turbulence changes induced by the hill did not seem to depend on the nature of the surface roughness (suggesting the inner boundary length was quite small). An excessively smooth surface, however, resulted in some overestimation of hill crest wind speeds. Simulations in two facilities using three models at three different scales showed a gratifying degree of consistency. The only effect of model scale was a predictable increase in difficulty in making measurements very close to the surface as the size of the model decreased. The depth of the turbulent inner layer was similar to the value predicted by Jensen et al. (1984).[7]

The general behavior of wind flow over simple and complex terrain is qualitatively well understood. Measurements in field or laboratory situations have been made under an amazingly broad range of conditions.

Nonetheless, the possible combinations of hill shape, slope, surface roughness, stratification conditions, upwind approach conditions, surrounding terrain undulations, and unsteadiness associated with the diurnal cycle and weather result in few quantitatively reliable estimators for hill wind speeds. Actual measurements of wind flow over vegetation covered terrain which can be used to specify the effect of forest edges, clearings or clearcuts on wind speed and dispersion are minimal. The few studies identified will be discussed in the following section.

Wind Flow Over Vegetative Canopies

As early as 1893, a German scientist, Metzger, investigated the effects of wind action on trees. Subsequently, a variety of studies have been made of the behavior of winds well inside and directly above a forest canopy.[8-19] Some measurements are available for the variation of the wind at the edge of a forest.[20-22] Much of this data is accumulated in periodically issued reviews and books.[23-26]

Flow Perturbed by Tree Barriers. Laboratory simulation of canopy flow in the wind tunnel has been used by the forest meteorologist in his efforts to understand the complex nature of flow generated by the tree--a permeable, random shaped, elastic object. Tiren, 1927, attempted to estimate crown drag from conifer branch-drag measurements made in a wind tunnel as part of his study of stem forms.[18] Wind-breaks have been studied by models to determine soil erosion and blow-down characteristics.[27-30] Others performed studies to characterize the transport of scalar products into, within and above vegetation.[31-34] These studies were conducted to deduce the qualitative behavior of tree barriers for specific problems. The investigators apparently made no attempt to scale dynamically the character of a live tree except to compensate intuitively for shape and porosity.

To model completely the complex geometry and structural characteristics of a live tree is obviously not practical; however, measurements made on coniferous and deciduous trees in the wind tunnel and the field suggest that equivalence of drag and wake characteristics between model and prototype trees should be sufficient to study the general flow phenomenon.[17, 35-38] Subsequently, a number of studies have been completed with greater attention to the flow characteristics of individual canopy elements.[34, 39-42]

Above-canopy Forest Flow Field. The atmospheric boundary layer (ABL) is that portion of the atmosphere where surface drag due to the motion of the air relative to the ground modifies synoptic-scale motions caused by horizontal pressure gradients, Coriolis forces, and buoyancy. The depth of the ABL is highly variable (50 to 2000 m), but it generally increases with proximity to the equator, with wind speed, and as the earth surface roughens, but it decreases at night, and is strongly modified by thermal winds, inversions, and stratification. Counihan (1975) reviewed all adiabatic ABL data taken between 1880 to 1972.[43] For high wind speeds (U_{10} > 5-7 ms^{-1}) Counihan recommended 600 m as a reasonable average boundary layer depth for both rural and urban cases independent of wind speed and roughness.

The lowest 10% of the atmospheric boundary layer is called the surface layer. It is characterized by the sharpest variations of wind speed, temperature, humidity, and turbulence characteristics with height. Counihan concluded that the surface (or constant flux) layer would be about 100 m deep during adiabatic conditions. In stratified situations the surface layer depth is about equal to the absolute value of the Monin Obukhov length, $L_{mo} = -Tu_*^3 / (\kappa g\ w't')$. For a summary of surface layer behavior for both neutral and stratified flows combined with both smooth and rough surfaces see Panofsky and Dutton (1984).[44]

Logarithmic Velocity Profile Models. Within the surface layer the mean wind-speed profile is commonly described by logarithmic expressions. For situations when stratification has only a minor influence, a modified logarithmic law has been proposed:

$$u(z) = (u_* / \kappa) \ln_e [(z - d + z_o)/ z_o],$$

where $u_* = (\tau/\rho)^{1/2}$ is the surface friction velocity, d is the zero-plane displacement, κ is Von Karman's shear layer constant, and z_o is the surface roughness. The displacement thickness, d, is important for tall roughness elements such as agricultural crops, forests, and cities. When the roughness elements are short, such that z_o < 0.2 m, one can set d = 0. The parameters can be determined from representative field measurements or models. Fitting an expression which permits three free parameters to field measurements of wind speed in agricultural canopies is not trivial. It is not uncommon for some least-square fitting routines to produce negative displacement heights--which is, of course, inappropriate.

No exact definition of high roughness has been offered, but roughness of a height exceeding 10% of the surface layer is generally viewed as high roughness. (Alternatively, whenever the logarithmic expression with d set equal to zero fails to fit measured wind distributions, the full expression may be justified.) Generally, the von Karman universal constant κ is assumed equal to 0.4 based on extensive experimental study of fully developed turbulent flow through pipes and its relationship to the Kolmogorov dissipation constant. Some experimentalist treat the constant as another free parameter to improve curve fit to data; hence, values ranging from 0.15 to 0.5 have been recorded. Nonetheless, it is customary to accept the initial value of 0.4 unless there are very persuasive arguments to do otherwise.

Some derivations of the logarithmic expression depend upon the assumption that shear stress is nearly constant with height above the surface. Matching of inner and outer similarity solutions to a boundary layer demonstrates, however, that such an assumption is not really necessary to the existence of a region which depends logarithmically on displacement above a ground plane. Nonetheless, the shear stress may be expected to vary substantially above the canopy roof; hence, it would be best to associate the friction velocity with the average drag produced by the wind on the forest. In order to avoid negative displacement height values it is customary to assume the von Karman constant $\kappa = 0.4$, to prespecify displacement height as some fraction of the forest canopy depth (say $d = 0.67$ h) and to solve for friction velocity and surface roughness height by fitting the modified logarithmic expression to measured data.

The effective values of the parameters may vary locally when the surface roughness is non-homogeneous. A non-homogeneous surface occurs when the ground surface changes from water to land, urban to rural, or cleared to forested. Such changes may make it appear that effective surface roughness, surface friction, and displacement height vary with height within the velocity profile. This aspect of the flow field will be discussed further in another section.

Surface roughness estimates have been estimated by many scientists for flow data obtained over different agricultural crops and forests. There is a wide variance in results even for flow over the same surface. Frequently experimentalist fail to obtain data above the wake region of individual roughness elements ($z > 1.5h$); sometimes the data are taken during non-neutral conditions; and often upwind nonhomogenuities distort the measured profiles. Several sets of tabulated data are available.[43, 45-49]

Jaeger (1965) recorded wind speed measurements over a ten year period over stands of Scotch pine located in southern Germany as they grew from 3 to 8 m height.[50] He made estimates of the variation in u_*, z_o, d, β (Deacon parameter), and Richardson number, Ri, from wind and temperature data collected from meteorological towers placed within the forest stand. He found that the following correlations described the measurements:

$d = 0.63$ h, regression coefficient, $r = 0.73\text{-}0.93$;

$z_o = 0.174\text{ h} + 0.227$, regression coefficient, $r = 0.44$; and

$u_* = (0.027\text{ h} + 0.062)\, U_{9.6m} + b$, regression coefficient, $r = 0.84$.

The expressions for d and z_o are seen to be similar to those derived from examination of under-canopy flows. However, the correlation for z_o is rather poor, and in a personal communication Massman suggested universal expressions for friction velocity are not reliable.[51] The variation of above-canopy wind speeds for typical forests suggest values of displacement height and roughness, $d = 0.67$ h, $z_o = 0.125$ h, and $u_* = u(h) \int (C_f/2) = 0.316u(h)$, respectively. Given these dimensions one finds that the modified logarithmic law becomes:

$$u(z) = 0.316(u(h)/\kappa)\ln_e[8z/h - 4.34].$$

Of course, this expression is only approximate, since it assumes independence from drag coefficient, leaf area index, and foliage distribution variations. The expression should not be applied below $z = 1.5$ h.

Power-law velocity profile models. In an alternative empirical approach to describe the wind variation with height the velocity variation is described by a simple power law of elevation. It is widely used in describing the wind shear in the atmospheric surface and internal boundary layers in view of its simple format and engineering expediency. The general form of the expression used is:

$$u(z)/u_{ref} = (z/z_{ref})^{\alpha},$$

where u_{ref} is the reference wind at a reference height z_{ref}, and α is the power law index (exponent). The effect of turbulence induced by the surface roughness upon the wind shear is accounted for by the magnitude of the power law index, whose magnitude is normally smaller than unity but larger than zero. Often the power law index is determined empirically by fitting the expression above to measured data; however, it is also possible to match the magnitude of predicted velocity and shear at a specified height and relate the power law index, α, to logarithmic profile parameters such as z_o, d, and $L_{Monin-Obukhov}$). For neutral flow the expression is simply:

$$\alpha = z_m/[(z_m - d + z_o)(\ln_e [(z_m - d + z_o)/z_o)/z_o],$$

where z_m is the matching or mid-height over which both profiles are presumed valid.

Empirical expressions which relate power law index and surface roughness length have been proposed by Counihan (1975) and Baron (1982).[43,52] Counihan's expression was developed by fitting logarithmic and modified logarithmic profiles to 70 different sites over data to a height of 100 m:

$$\alpha = 0.096 \log_{10} [z_o] + 0.016 (\log_{10} [z_o])^2 + 0.24,$$

for $0.001 \leq z_o \leq 5$. Baron fit a similar relationship to the nomogram proposed by Davenport[53] such that:

$$\alpha = 0.125 \log_{10} [z_o] + 0.0004/z_o + 0.336,$$

for a roughness range $0.01 \leq z_o(m) \leq 5.5$. However, these last two functions produce significantly different estimates. For example Baron's expression produces power index values 17 to 38% greater than Counihan's expression over the range from smooth to rough roughness . This variation may simply be the result of using different data sets, the influence of stratification, or it may be that displacement height was not considered in a similar manner for the two data sets.

Baron examined a wide cross-section of field and laboratory data and created figures which predict power law index in terms of roughness element height, h, and effective roughness height, z_o respectively.[52] Given canopy heights, h, ranging from 10 to 30 m in depth, associated roughness length, z_o, varying from 1.25 to 3.75 m in size, one expects power law index, α, to vary from 0.45 to 0.52.

<u>Wind Flow Near Clearings, Clearcuts, and Forest Edges</u>. When airflow passes from a cleared area into a forest, winds initially penetrate into the canopy space, but then the streamlines are lifted upward to the canopy roof . The penetration distance among the trunk space in the canopy understory may persist for 5 to 10 tree heights. Subsequently the wind rises above a recirculation region and re-enters the forest about 20h from the windward forest edge. Cionco (1981) sketched how such entrance flows might look from the perspective of smoke plumes on the battle field. [54] When the airflow passes from a forest canopy to a cleared area the under canopy flow begins to accelerate as much as 5 tree heights upwind as streamlines move toward the ground, but downwind of the forest edge low-level winds may require substantial distance to readjust to the new smaller surface roughness. Meroney (1968) observed the effect of initial wind penetration at the windward forest edge, the low speed recirculating zone, and the flow acceleration before the downstream forest edge on canopy drag.[36]

Eimern (1964) considered the aerodynamics of shelterbelts and summarized the influence of density, shape, surface roughness, thermal stratification, wind angle and tree arrangement on downstream wind speed, turbulence, soil moisture, etc..[55] Behind porous objects the velocity defect generally persists twice as far downwind; however, the turbulence intensity excess is diminished. Maximum length of shelter will occur for long shelterbelts of near 50% permeability, and 30% velocity defect may still exist at 0.5h for x/h > 50. The wind will return to its undisturbed condition in about half the distance if the shelter belt is only twice as long as it is high. Wind approaching a shelter belt at an angle increases its effective porosity, decreases the shelter, and increases turbulence excess.

The micrometeorology of shelter belts and forest edges are reviewed by McNaughton (1989).[56] He notes that although extensive studies have been performed downstream of thin shelterbelts the effects of forest edges have received far less attention. Indeed with respect to wind flow downstream of forest edges he notes that "this discussion is more a summary of our ignorance than of our knowledge." There are similarities as well as differences between flow downstream of thin shelterbelts and forest edges. The foliage density of the forest canopy replaces the porosity used for narrow shelterbelts. Upwind profiles must be characterized by the upwind forest roughness, displacement height, forest friction velocity, and foliage density.

McNaughton sought a comparison to the flow over a forest canopy edge and the flow that occurs when a boundary layer passes over a solid backward facing step. For solid steps a recirculating eddy occurs of downwind extent of about 6 h. But permeability often allows the wind to penetrate the forest upwind of the forest edge. For example, during wind-tunnel experiments performed over plastic model trees Meroney noticed winds increased above and within the canopy over the last 10 h.[36] In coniferous forests researchers detect upwind penetration over several heights upwind, but in a denser foliage other researchers see little penetration at all. Nonetheless, little evidence exists to support the presence of a recirculating eddy downwind of the forest edge. The flow velocities and surface shear appear to adjust to the immediate absence of the forest edge by 20 h; however, the wind continues to accelerate over a longer distance as a deeper layer of the atmosphere adjusts to the change of surface roughness.

There appear to be very few measurements of actual winds made above and below forest canopies near clearings or forest edges. Leahey and Hansen (1987) report measurements made on meteorological tower located 60 m from a 0.5 km^2 in a forest in Alberta, Canada.[57] Trees were primarily pine and aspen with heights ranging from 18 to 24 m growing on flat terrain. Measurements were taken at 10 and 20 m levels on a 24 m tower using a Gill U-V-W anemometer. They identified strong horizontal jets of air and large vertical velocities during unstable conditions, but rather normal conditions under stable stratification. Winds in excess of 6 ms^{-1} occurred about 15% of the time. Their measurements suggest that clearing a ridge may produce strong convergence toward ridge lines, which could modify ridge top conditions.

Fowler et. al. (1987) examined the effects of shelterwood cutting (30-percent canopy removal) and clearcutting clearings from 0.8 to 8.5 ha on climatic variables of the High Ridge Evaluation Area within the Umatilla National Forest in northeastern Oregon.[57] Areas were harvested in 1976 after nine years of prelogging calibration. The authors concluded that hydrological effects of the cuttings were surprisingly small, but wind passage and velocities increased dramatically with removal of the forest cover. The authors compared data from roughly equivalent 11-month periods during pretreatment and posttreatment. Little change was noted for the Watershed No. 4, height 20 m case, but at all other locations the wind speeds increased substantially in all classes! One should note that the weather station in Watershed No. 4 was within the uncut area; whereas the station in Watershed No.1 was in the middle of a clearcut region. Indeed winds at 6 m height in Watershed No. 1 exceeded winds at 20 m height in Watershed No. 4!

Elliott and Barnard ,1990, discuss a field experiment to examine the effect of scattered groves of trees and grass on the variability of wind speed and turbulence at the Goodnoe Hills, WA, wind-power site.[58, 59] Two permanent towers and seven portable towers were used. The two permanent towers measured wind at heights from 15 to 107 m and 15 to 59 m above the ground. Wind speed measurements were taken from nine bivane anemometers sampled every second and averaged every minute. The site contains a broad ridge on which the MOD-2 turbines were installed. Terrain is gently sloping to the west and north, but drops abruptly off to the Columbia River Gorge to the south. Vegetation is mostly low sage brush and grass and scattered groves of scrub oak, western juniper and ponderosa pine.

Two towers were about 200-300 m downwind of a grove of 10-18 m high trees at which 20-30 percent reductions of wind speed and a 2-3 times increase in turbulence were measured at a height of 32 m. Wind gusts also increased at 30 m, but by heights of 60 m or distances of 500 m downwind tree effects were considerably reduced (25 to 50 h).

Change of Surface Roughness

It has long been observed that when the wind flows from one surface texture to another a transition takes place in wind speed and turbulence within an inner-boundary-layer, l_z, that grows in depth with downstream distance from the surface change. When the surface change is associated with roughness height, and downstream wind profiles are plotted semilogarithmically with height, then a distinct "kink" in the slope of the plot is observed which can be associated with this inner-boundary-layer depth, l_z. The wind profile near the ground will adjust to surface roughness changes as it moves downwind from the ground cover transition. Above l_z the profiles will correspond to the wind profile for the roughness before the change in cover.

Change of Roughness Models. A number of different analytic and numerical models exist to predict the resultant variation in wind profiles which exist at different fetch distances downstream of a transition of roughness. The subject is extensive enough that a literature review has been prepared on the topic by Hunt and Simpson (1982).[60] They also provide a table summarizing field and laboratory change of roughness experiments used to verify various models. Unfortunately, little data exists for roughness variations as large as the abrupt change that occurs from a forest edge to a meadow or a clearcut region. Most models grew from the perturbation analysis originally proposed by Townsend in 1976. Subsequent researchers have modified assumptions, boundary conditions, definition of perturbation variables and scaling lengths, but the basic concepts have remained the same. This same perturbation approach has subsequently been applied to predicting the effects of surface elevation, surface temperature, surface heating, surface humidity, and stratification on atmospheric boundary layer wind profiles and turbulence.

A presentation by Jensen (1978) is widely accepted.[61] Given an upwind roughness, z_{o1}, a downwind roughness, z_{o2}, a corresponding up- and downwind surface friction velocity, u_{*1} and u_{*2} , and a distance downwind from the roughness change, x, then:

$$u(x) \approx (u_{*1}/\kappa)\ln[z/z_{o1}] + (u_{*1}/\kappa)\ln[z_{o2}/z_{o1}](\ln[z/z_{o2}]/\ln[l_z/z_{o2}]) - 1\}$$
$$u_{*}/u_{*1} \approx 1 + \ln[z_{o2}/z_{o1}]/\ln[l_z/z_{o2}]$$
$$l_z \ln[l_z/z_{o1}] = 2\kappa^2 x$$

where κ is the von Karman constant commonly set to 0.4. These expressions suggest that the perturbation shear stress and velocity decrease slowly and inversely with $\ln[l_z/z_{o2}]$ and are proportional to $\ln[z_{o2}/z_{o1}]$. By normalizing l_z and x on the larger of the up- or downwind roughness, z_o, an almost universal plot of inner boundary layer growth was prepared from field and model data. An empirical fit to the nonhomogeneos equation for l_z might be $l_z/z_o = 0.3\,(x/z_o)^{0.8}$

Multiple Change of Surface Roughness Models. By superposition of the linear-perturbation solution for a one-dimensional change in surface roughness, one can create a method to predict the effect of arbitrarily distributed surface roughness on wind profiles. Belcher, Xu and Hunt (1990) proposed such a model to predict the effect of non-homogeneous two-dimensional roughness on wind profiles and surface stress.[62] Due to the lack of field or model data they compare their results to higher-order turbulence closure solutions of similar boundary conditions. The perturbation approach produces quite good correlation for roughness changes as large as $|\ln[z_{o2}/z_{o1}]|$ of order one.

Derickson and Peterka (1992) have also developed a method to correct anemometers for multiple changes of upwind roughness.[63] Their ad hoc approach follows earlier work on roughness changes proposed by Deaves (1981) and Cook (1985).[64,65] Whereas the Jensen approach is limited to correcting wind profiles in the lower regions of the boundary layer, this method corrects for the eventual adjustment of the gradient wind profile at all elevations to the change in surface conditions. This method is probably an over-kill for estimating the effects of forest clearings and clearcut regions of finite extent.

Dispersion Over Vegetative Canopies

Dispersion over model vegetative canopies has been extensively examined in wind tunnels. Meroney (1968) considered wind flow, turbulence and diffusive behavior over a finite-fetch model deciduous forest.[36]

Sources were located upwind and along the forest border. Plumes were observed to initially penetrate the forest boundary, loft above the canopy, and subsequently descend along the forest lee boundary. Mixing was significantly enhanced over the rough canopy. Pendergrass and Arya (1984) also considered plume behavior over a step change in surface roughness.[66,67]

Recently, Wu (1993) examined the behavior of plumes released upwind and at forest boundaries.[68] Several transition situations were considered including smooth-to-rough and rough-to-smooth roughness changes. All results were compared with the dispersion over homogeneous surface roughness of similar scale. When the plume fell below the inner boundary layer produced by the roughness transition ground level concentration variations were detected, but when the plume height exceeded the inner boundary layer height little ground effect was noticed. Wu found that for elevated stacks the ground-level concentrations continue to follow curves similar to those found when the entire surface is covered with upstream height roughness. For plumes released from stacks shorter than local inner-boundary layer height concentrations fall between ground-level concentration (glc) curves associated with releases over homogeneous surfaces of upwind and downwind roughness.

Table 1 summarizes available information concerning the resultant amplification (or reduction) of observed maximum concentrations in terms of an amplification ratio, A. The ratio A is defined as the ratio of the observed maximum glc in the presence of roughness variation to the maximum glc observed for a similar release over the homogeneous upwind surface roughness. Subscript 1 refers to conditions upwind of the roughness change, whereas subscript 2 identifies conditions downstream from the roughness change. Sources were placed upwind of the roughness change or directly above the transition location. A negative roughness ratio, M, reflects a smaller upstream roughness condition.

The results suggest that in the absence of terrain height variations an increase in vegetation height may sometimes lead to a 10 to 60% increase in groundlevel concentrations. But a decrease in surface roughness generally leads to a decrease in surface concentrations.

VEGETATIVE EFFECTS ON WIND FLOW OVER HILLS

Complex hilly terrain may exist with a variety of vegetative surface cover. For example, the approach terrain and the hill itself may both be either bare or vegetation covered. Alternatively, the upwind surface may be smooth (farmed plains or meadows) and the hills may be rough (tree covered), or the upwind surface may be rough (tree covered) and the hill itself bare. In some cases only portions of the hill may be bare due to selective shelterwood cutting or clearcutting. The presence or absence of high roughness may lead to lower /higher wind speeds, higher/lower turbulence, or attached/separated streamline flow. These kinematics determine polution levels from plumes released upwind, over, and downwind of the terrain.

Homogeneous Surface Roughness Over Hills

The upwind surface roughness induces different approach velocity profiles which can lead to variations in hill crest wind profiles. The approximate effects of such profile changes on the fractional speedup have been examined using physical modeling, inviscid rotational numerical models, 2nd-order turbulent closure models and linear-perturbation analysis.

Field and Laboratory Data. Bouwmeester et al. (1978) performed wind tunnel measurements over triangular hill shapes, with equal hill heights and slope, but surface roughness varying between cases by a factor of ten ($z_o/h_{hill} = 0.0013$ versus 0.0178).[69] The rough hill produced larger speedup values at all heights; however, since the reference wind speed at a given height is usually less for rough surface flows, the actual crest height wind speeds are less.

Bradley (1978) measured wind flows over a tree-covered ridge in Australia.[4] The ridge height, h_{hill}, was 170m, the upwind ridge length was L_u = 550 m and the downwind length somewhat longer, L_d = 600 m.; hence, the average h_{hill} /L ratio equals about 0.29. The hill was covered with 10 m tall trees, and the associated surface roughness, z_o, and displacement height, d, were estimated to be 1.0 m and 7.0 m respectively; hence,

$z_o/h_{hill} = 0.006$. The atmospheric boundary layer was estimated to be between 600 and 800 m. Bradley obtained wind data at various heights up to 100 m above the crest during neutral conditions. A separation region was believed to exist downwind of the crest, where vertical heights have been corrected for forest displacement above the ridge ground level. The roughened wind-tunnel model was relatively slightly rougher than the Bradley forested hill.

<u>Inviscid -Rotational Numerical Model Results</u>. Bouwmeester et al. (1978) used the inviscid-rotational potential flow model proposed by Derickson and Meroney (1977) to predict fractional wind speed up over hills for different approach flow profiles.[69,70] Nine velocity distributions were compared for different combinations of z_o/h_{hill} and h_{hill}/δ. The resultant fractional speedup ratios were essentially independent of roughness for $h_{hill}/\delta = 4$, and there was only a slight dependency on roughness for smaller h_{hill}/δ values. The upper flow perturbations proposed by Jackson and Hunt (1975) are also essentially inviscid potential flow solutions, and they also imply primary dependence on hill slope and no variation with surface roughness, that is $\Delta S = (h_{hill}/L)\sigma(x,y)$.[71] These results seem inconsistent with the measurements reported earlier. One reason might be that despite the range of roughnesses specified for the upwind profiles, the absolute roughness magnitudes were less than $z_o/h_{hill} = 10^{-4}$. The differences might also result because the inviscid model does not correct for the inner-boundary layer which increases as $l_{z1}/l_{z2} = (z_{o1}/z_{o2})^{0.2}$. Thus, if the roughness length increases by a factor of ten the inner-boundary-layer increases by almost two..

<u>Turbulence Model Insights</u>. Frost, Maus and Fichtl (1974) solved the turbulent boundary layer equations over a plane surface using mixing-length closure for a horizontal pressure distribution equal to that along the $\psi = 0.6$ streamline of inviscid potential flow around an elliptical cylinder.[72] Numerical solutions were carried out for aspect ratios of 2:1 and 4:1 and for various surface roughness lengths. Fractional speed-up at inner-boundary-layer height appears to remain nearly constant as roughness increases.

Taylor and Gent (1974) solved for flow over a hypothetical hill using second-order turbulence closure methods.[73] Bouwmeester et al. (1978) determined that surface shear stress predicted by inviscid models underestimated the Taylor-Gent nonlinear model values by up to 300%.[69]

<u>Linear-perturbation Model Insights</u>. Linear-perturbation models provide for the influence of surface roughness in an inner layer, l_z, such that the fractional speed-up becomes:[61,71,74]

$$\Delta S(z) = (h_{hill}/L)\sigma(x)\,(\ln[L/z_o]/\ln[l_z/z_o])^2,$$

or

$$\Delta S(z) = (h_{hill}/L)\sigma(x)\,(\ln[L/z_o]/(0.8\ln[L/z_o]))^2 \nrightarrow f(z_o).$$

Thus, most analysis predicts that variation in homogeneous surface roughness upwind and over the hill will produce only small changes in the fractional speed-up, but consequently large variations in the actual mean profiles.

Change in Roughness Effects on Flow Over Hills

An early observation made during wind-tunnel measurements was that when surface roughness was reduced over the steeper models the mean velocity on the lee side actually increased, and the flow did not separate, even intermittently, though the flow remained turbulent. Conventional wisdom for flow around bluff bodies usually proposes the addition of surface roughness to inhibit separation not its removal. Britter et. al. (1981) explained this paradox by arguing that the delay of separation is induced because the surface velocity near the separation point is increased as the flow accelerates over the smoother hill surface.[75] Thus the boundary layer can penetrate further into the adverse pressure gradient on the lee side of the hill. (The boundary layer is energized by the descent of streamlines toward the wall after a decrease in surface roughness.)

<u>Linear-perturbation Model Insights</u>. Jensen and Petersen (1978) discussed the possibility of adding the linear-perturbation solutions for boundary layer response to changes in surface roughness and elevation.[74] Since the solutions are separately linear their perturbations should be additive; thus

$$u(z)_{hill\ \&\ roughness} = u_o(z) + \Delta u(z)_{roughness} + \Delta u(z)_{hill}.$$

Hunt (1978) applauded this step, and he concluded that the maximum perturbation induced by change of roughness would occur at height z_{o2} whereas the maximum perturbation induced by the hill would occur at height l_z. Consequently, it is not really possible to cancel out the changes induced by one effect by the other. Hunt proposed the ratio:

$$\frac{|\Delta u(l_z)_{hill}|}{|\Delta u(z_{o2})_{rough}|} = \frac{(h_{hill}/L)(\ln[L/z_{o1}])^2}{\ln[l_z/z_{o2}]\ln[z_{o2}/z_{o1}]}$$

Thus, where there is a large change in roughness of say from 1 m to 10 cm at the hill half-width and it is accompanied by a change in slope say 1 in 4 over a hill width of 1000 m, the maximum hill slope effect is comparable to the change in roughness effect! On the other hand, if the change of roughness is from 2 m to 2 cm under otherwise similar conditions, then the maximum hill slope effect is five times less than the maximum roughness effect!

Field and Fluid Model Data. Britter et al. (1981) performed wind-tunnel experiments over a polynomial shaped hill for which $h = 0.1$ m, $L = 0.25$ m, $z_o = 0.002$ m, and $u_* / U_\delta = 0.0685$.[75] The authors suggested that at a model scale of 1: 500, the surface roughness corresponded to a value of about 1 m (forest canopy). A second experiment was performed where the roughness ended 1 m upwind of the hill crest. This resulted in significant acceleration in the lower part of the boundary layer as a direct result of the change in roughness. Finally, velocity profiles were also measured at the equivalent location of the hill crest downwind of the roughness change after the hill was removed. Examination of the figure reveals that the increase in wind speed of the smooth hill over the rough hill is almost exactly equal to the perturbation induced by the change in roughness alone.

Laboratory Measurements of Vegetation Covered Terrain

Extensive tree shelter belts were planted over the Rakaia Gorge, NZ, terminal moraine and river plain studied by Meroney et al. (1978).[76] These 10 m high dense tree belts were planted by farmers to protect sheep paddocks. Comparison of field and physical model measurements revealed that the shelter belts played a dominant role in determining near surface winds. Measurements made over vegetation free models of the Rakaia Gorge over-estimated wind speeds and under-estimated turbulence levels at a 10 m measurement height . When vegetation was modeled field and laboratory wind speeds agreed at sample correlation coefficient levels from 0.68 to 0.78 and rank correlation coefficient levels from 0.78 to 0.95. (Sample correlation coefficient levels for field measurements at the same sites taken on independent days was only 0.68. This suggests that there is an inherent limitation to the paired replication of any single realization of a wind flow pattern by a model whether physical or numerical.

Recently Gong and Ibbetson (1989) reported physical model measurements made over cosine shaped hills and ridges of slope 15°.[77] They added a uniform roughness to the hills made of a rubber sheet having flat-topped circular cylinders 3 mm high and 2 mm diameter at a uniform spacing of 3.6 mm between centers. The surface Reynolds number, Re_* , was about 5 ,which implies a rough model surface during simulations. Hill height was 31 mm and half hill width was 100 mm. The effective surface roughness, z_o, was determined to be 0.17 mm. If we assume a scale ratio of 1:10,000 then field scale hill height would be 310 m, roughness height would be 30 m, and surface roughness, z_o, equals 1.7 m. This would be typical of many forest covered hills/mountains found in nature. Gong and Ibbetson recorded extensive mean velocity, shear, and turbulence information. Comparisons of their data against linear-perturbation models was excellent on upwind hillsides and higher levels. Measurements over the two-dimensional ridge and the circular hill suggests that the mean flow and turbulence over a circular hill resembles those over two-dimensional ridges of similar cross-section, but with reduced perturbation amplitudes.

The removal of vegetation upwind of the crests of hills has been shown to substantially increase hill-top winds and reduce the probability of separation and consequent gustiness.[78,79] Models based on linear-perturbation principles appear to predict correctly the order of magnitude of combined elevation and roughness change effects on wind speed profiles. Inviscid flow models are not expected to account for the effects of roughness change over hilly terrain.

VEGETATION EFFECTS ON DISPERSION OVER HILLS

Many dispersion experiments have been performed in complex terrain situations which include both terrain undulation and vegetation; however, few have focused on the specific dispersive influence of vegetation. Qualitatively, the effects of plume downwash, trapping in separation regions, and enhanced mixing downwind of ridges, bluffs and hills are well recognized.[1,80] Some field diffusion experiments were performed in the presence of forested hills (AVOCET Geysers-1981 and Brush Creek-1984 studies), but most scrutiny was placed upon diurnal drainage flow mechanisms. A few numerical programs incorporated boundary conditions which acknowledge the presence of such vegetation and its effect upon scalar transport, but there has been no systematic evaluation of the influence of clear cuts and forest extent on diffusion.[81]

Arya et al., 1981, considered the influence of a steep triangular ridge on steady plumes released upwind, over and downwind of a model ridge crest in a wind tunnel.[82] The large separated flow region and the associated shear layer in the lee of the ridge dominated plume trajectories and concentration fields. The authors examined dispersion over four smooth-profile model hills for plumes released at various locations. The objective of this research was an evaluation of the effect of hill slope on dispersion over low hills. The steeper hills produced separation cavities and the same general streamline behavior as steep triangular hills. The mildest slope hill avoided separation, but downwind groundlevel concentrations were enhanced by streamline descent downwind of the hills.

Three-dimensional model hill shapes were examined by Arya and Gadiyaram, 1986, and Snyder and Britter, 1987.[82,83] Plumes released upwind of a three-dimensional hill tend to follow streamlines closer to the hill crest than for similar two-dimensional hills. Ground level concentrations (glc) typically decrease for hill crest releases, but the glc increase for release locations downwind of the crest. Although maximum glc are greater downwind in such situations; further downwind surface concentrations are actually less due to enhanced vertical mixing in the hill wake.

Table 2 summarizes the influence of hill configuration in terms of a relative terrain amplification factor, A. The amplification factor, A, is the ratio of the maximum glc in the presence of the hill to the maximum glc without the hill present. In this table although surface roughness varied from case to case, it was uniform over the hill and upstream. The influence of surface vegetation roughness is noted primarily through its tendency to encourage flow separation over the hill. Ridges of limited lateral extent or three-dimensional hills seem to produce larger amplifications in concentration than similar cross-section two-dimensional ridges, especially on the upwind side of the hill..

CONCLUSIONS

The development of wind profiles over different homogeneous surface roughness conditions can be predicted with fair accuracy. Measured profiles of wind speed both below and above vegetative canopies follow analytic models well. The actual values for surface shear and roughness length will depend upon the total area averaged, especially in areas where surface elevation varies. Surface roughness, z_o, and displacement height, d, can be related to canopy foliage density and average tree height. Surface shear can be predicted with somewhat less confidence.

The presence of openings, cleared areas, shelterwood clearings, and clearcut regions within forests produce motions which are qualitatively anticipated. Unfortunately, there are very little field data from forests available to verify any analytic or numerical models for such situations. The downwind effects of roughness change can be predicted by linear-perturbation models if roughness displacement height variations are not too large and flow separation is absent.

Gaseous plumes released upwind, over, or downwind of a change in surface roughness produce amplification or reduction of maximum surface level concentrations ranging from 0.4 to 1.6. Plumes released upwind, over or downwind of two or three dimensional hills produce amplification factors ranging from 1.1 to 4.0, 0.3 to 2.5, and 1.7 to 15.0, respectively. The largest variations occur when surface layer flow separation occurs downwind of the hill.

When surface roughness variation (forest clear cut operations) and hills are present together, the resulting glc amplification factor ratio will depend upon the combined effects of hill slope and upwind fetch to roughness changes. Over hill slopes greater than 0.3, the hill slope should dominate, but when large roughness changes occur upwind of modest slope hills, the inner boundary layer effects of roughness modifications may also be significant.

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Table I. Concentration amplification factors for flow over a change in roughness.

Author (Date)	Roughness Ratio M = $\ln(z_{o1}/z_{o2})$	Displ. Ratio (d_1/z_{o1})	Displ. Ratio (d_2/z_{o2})	Source Height (H_s/z_{o1})	Amplification Factor (A) Upwind	Factor (A) Edge Release
Meroney et.al (1968)	-8.76	0.00	2.86	9.0 x 10^3	—	0.71
				1.8 x 10^4		0.43
Yeh and Nickerson (1970)	-4.35	0.00	0.00	0.00	1.15	—
Pendergrass and Arya (1984)	-2.35	14.7	25.35	0.00	—	1.14
				5.3 x 10^2	—	1.25
				1.1 x 10^3	—	1.33
				1.4 x 10^3	—	1.57
				2.0 x 10^3	—	0.94
				2.9 x 10^3	—	1.00
				3.7 x 10^3	—	1.23
Wu (1993)	-2.88	0.00	0.00	2.0 x 10^3	1.040	1.054
				1.0 x 10^4	1.074	1.154
	+2.88	0.00	0.00	1.1 x 10^2	0.916	0.863
				5.6 x 10^2	1.022	0.928
	-5.03	0.00	0.00	2.0 x 10^3	1.012	1.063
				1.0 x 10^4	1.160	1.077
	+5.03	0.00	0.00	1.3 x 10^1	0.714	0.784
				6.5 x 10^1	0.978	0.838

Table 2. Concentration amplification factors for flow over hills and ridges.

Hill Slope (H/L)	Hill Roughness (H/z_o)	Hill Aspect Ratio (L/H)	Source Height (H_s/H)	Factor (A) Upwind	Amplification Factor (A) Hill Top	Factor (A) Downwind
2.0 (63°)	384	∞	1.50	—	2.50	10.0
Triangular			2.50	—	1.17	2.67
Arya et al., 1981						
0.13 (13°)	440	∞	0.25	2.8	0.3	7.5
Russian			0.50	2.5	0.7	6.4
			1.00	1.8	0.9	10.8
			1.50	1.9	0.9	7.8
			0.25	2.0	0.5	15.0
0.20 (16°)			0.50	2.0	0.6	8.0
Russian			1.00	1.7	0.9	5.6
			1.50	1.2	1.0	2.9
			0.25	1.5	0.9	3.4
			0.50	1.1	0.6	3.0
0.33 (26°)			1.00	1.5	0.8	2.4
Russian			1.50	1.2	0.8	1.7
Arya et al., 1987						
0.5 (26.5°)	3357	2.0	0.50	3.3	—	—
Snyder and			1.00	1.8	—	—
Britter, 1979						
0.5 (26.5°)	3357	2.0	0.50	—	—	6.0
Castro and			1.00	—	—	3.0
Snyder, 1982			1.50	—	—	2.1
0.3 (17.5°) Cone	3357	3.2	0.125	—	0.59	4.0
Arya and			0.25	—	0.35	5.9
Gadiyaram,			0.50	—	0.45	5.6
1986			1.00	—	0.51	2.6
0.5 (26.5°)	227 or	2.0	0.50	3.30	—	—
Triangular	smooth	3.9	0.50	2.00	—	—
Snyder and		5.9	0.50	1.90	—	—
Britter, 1987		∞	0.50	1.50*	—	—
		2.0	1.00	1.75	—	—
		3.9	1.00	1.25*	—	—
		5.9	1.00	1.00*	—	—
		∞	1.00	0.60*	—	—
1.0 (45°) Bell	smooth	2.0	0.50	4.00	—	—
Snyder and			1.00	1.25	—	—
Britter, 1987						
0.31 (14°)	182	3.23	0.42	1.0	—	—
Cos squared			0.67	1.0	—	—
Gong, 1991						

* These values may be too low since concentrations were not measured downwind at the anticipated reattachment streamline locations.