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Air motion within and above forest vegetation in non-ideal conditions

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Abstract

Gaining a good knowledge of air motion in forest vegetation is a necessary step towards a better understanding of a number major abiotic impacts on the trees such as wind risk, pollutant and nutrient deposition, frost, material dispersion and transport, and energy, water and carbon exchanges. In a recent survey study, Raupach et al. [Raupach, M.R., Finnigan, J.J., Brunet, Y., 1996. Bound.-Layer Meteorol. 78, 351–382] reviewed the current state of knowledge about air flow under ideal conditions (neutral to slightly unstable conditions, homogeneous and extensive canopy, flat terrain). This paper extends the knowledge by employing advances in our understanding of the flow in 'non-ideal' situations. The paper is divided into four topic areas: canopy flow under stable stratification, disturbed flow (forest edge, forest clearing, sparse canopy), canopy flow over complex terrain, and extreme wind events. Discussion of the latter two topics is of limited scope because of the scanty literature. A detailed account is given to the nighttime canopy wave phenomenon, broad patterns of the transitional flow across forest edges, and models of various complexities of the disturbed flows. Both observational and modeling aspects are discussed wherever possible. This synthesis study has identified a number of important questions in need of further research.
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1. Introduction

Trees suffer many abiotic risks during their long lifespan. From a meteorological perspective these include frost, drought, fire, climate induced insect and disease outburst, snow and ice damage, exposure to air pollution and the adverse impact of climate change on various forest biomes. The literature is huge on these subjects (A search of Science Citation Index database with keywords 'drought and forest' returned 500 entries for the year 1997 alone!). To keep the paper within a reasonable length, the author will limit the discussion to a focused subject —air motion in forest vegetation. The choice also reflects the belief that a useful review paper should emphasize the subject most familiar to the author.

Gaining a good knowledge of air motion in forests is a necessary first step towards a better understanding of a number of major abiotic agents. The assessments of static wind load (Blackburn et al., 1988; Mayer, 1989), the dynamic coupling between air flow and the trees (Gardiner, 1995; Wood, 1995; Peltola, 1996) and damage arising from the combined effect of snow and wind loads (Mayer, 1989; Peltola et al., 1997; Päätalo et al., 1999) all require detailed information about the vertical wind profile and its spectral characteristics. The takeoff and spread of seeds, pollens, and disease-bearing spores, and the dispersion of pesticide and

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fungicide spray plumes are influenced to a great extent by the nature of air motion within and immediately above the forest (Greene and Johnson, 1989; Di-Giovanni and Kevan, 1991; Aylor et al., 1993; Miller et al., 1993). It is also recognized that the level of success of experimental studies on deposition of air pollutants (Fuentes et al., 1992; Guenther et al., 1996; Buzorius et al., 1998), trace metals (Lindberg et al., 1997) and the exchange of greenhouse gases (Wofsy et al., 1993; Grace et al., 1995; Jarvis et al., 1997) is dependent upon how well the canopy flow is understood. In fact, lack of good understanding of the nocturnal phase of the flow is a major obstacle in these studies.

The literature is rich on the subject of air motion in vegetation. In a review paper presented at the 1993 IUFRO Wind and Trees Conference (Finnigan and Brunet, 1995) and in a more recent review of studies of canopy flow (Raupach et al., 1996), Finnigan and Brunet (1995) and Raupach et al. (1996) demonstrated that canopy air motion is far from random, with major contributions to turbulent motions arising from coherent eddies. Most of the studies they reviewed are directed towards flow in ideal situations (neutral to slightly unstable conditions, horizontally extensive and uniform canopy, flat terrain). These studies have adopted the traditional strategy, long used by forest meteorology and micrometeorology communities, of examining the simplest possible cases in order to isolate important governing mechanisms. This paper will extend the knowledge by employing advances in our understanding of the flow in 'non-ideal' and hence more realistic conditions. To put the review in perspective, Section 2 presents the full governing equation of motion and the assumptions needed for the ideal situations. Section 3 examines the flow in forests at night when the air is stably stratified, giving a detailed account of the canopy wave phenomenon. Section 4 examines three cases of disturbed flow, namely flow across forest edges, in forest clearings and in sparse canopies. The next two sections (Sections 5 and 6) discuss two important topics: canopy flow on undulating terrain and extreme wind events. The discussion is however of very limited scope because of the scanty literature on these two topics. Indeed they are major knowledge gaps in need of substantial research. Both observational features and modeling studies are reviewed wherever possible.

2. Governing equation

To put the review in perspective, it is useful to consider the full governing equation of the longitudinal velocity component of the air motion

$$\frac{\partial u}{\partial t}+u\frac{\partial u}{\partial x}+v\frac{\partial u}{\partial y}+w\frac{\partial u}{\partial z} = -\frac{1}{\rho_o}\frac{\partial p}{\partial x}+g\frac{\theta}{\theta_o}\sin\alpha+fv-\frac{\partial\overline{u'w'}}{\partial z}-C_dAu|\vec{U}| \tag{1}$$

where x, y, z define a Cartesian coordinate with x being in the longitudinal, y lateral and z vertical direction, u, v, and w are the velocity components in the x, y, and z directions, respectively, ρ_o air density, p atmospheric pressure, g gravitational acceleration, θ perturbed potential temperature, θ_o background potential temperature, α terrain slope angle, f Coriolis parameter, $\overline{u'w'}$ Reynolds stress in the z direction, C_d a drag coefficient, A leaf area density and $\vec{U}$ wind vector (Mahrt, 1982; Raupach and Shaw, 1982; Wilson, 1985). Terms on LHS represent acceleration of the air parcel and terms on RHS are forces exerted on the moving air. Assuming steady state (term 1 of Eq. (1)=0), horizontal homogeneity (term 2=0), one dimensional flow (terms 3 and 4=0), no horizontal pressure gradient or baroclinic forcing (term 5=0), flat terrain (term 6=0), and ignoring the Coriolis force (term 7=0), Eq. (1) reduces to a balance between the vertical divergence of the Reynolds stress (term 8) and the canopy drag force (term 9)

$$\frac{\partial\overline{u'w'}}{\partial z}+C_dAu|\vec{U}|=0 \tag{2}$$

Eq. (2) is a key equation in all modeling studies of flow in homogeneous canopies (Wilson and Shaw, 1977; Li et al., 1985; Meyers and Paw, 1986; Wilson, 1988). Solutions for the mean velocity, regardless of the closure schemes (see Section 4), all show logarithmic and exponential patterns over and within the canopy, respectively, and an inflection point near the canopy top. The dynamic instability linked to the inflected velocity profile produces energetic turbulence in the canopy air layer. For flow regimes examined below, one or more of the assumptions that lead to Eq. (2) are violated.

3. Canopy flow under stable stratification

3.1. Observed features

Stable stratification usually occurs shortly after sunset and persists till sunrise. Nocturnal air motion in forests can be classified into three regimes: weakly stratified turbulence which appears stationary, sporadic or intermittent turbulence, and turbulence modulated strongly by wavelike motions. Their intrinsic dynamics in transporting materials and in interacting with the trees are quite different. Weakly stratified turbulence, which tends to occur at high winds ($>3\,\mathrm{m\,s^{-1}}$ over the forest) under cloudy conditions (net radiation flux over the forest $R_{\mathrm{n}} > -20\,\mathrm{W\,m^{-2}}$, where the negative sign denotes loss of radiation energy from the forest), is relatively well understood. Coherent eddy motions are evident in the form of, for example, inverse temperature ramps, at multiple levels within and above forests (Bergström and Högström, 1989; Gao et al., 1989; Paw U et al., 1992). The associated eddies are thought to result from a continuous hydrodynamic instability linked to the inflected velocity profile in the upper canopy (Finnigan and Brunet, 1995; Raupach et al., 1996). Modeling frameworks of the turbulence (Wilson and Shaw, 1977; Raupach and Shaw, 1982) and our understanding about the interaction with the trees (Gardiner, 1995; Peltola et al., 1997; Flesch and Wilson, 1999b) established under neutral and unstable conditions is directly applicable here.

Under the weakly stratified regimes, the level of turbulence is reduced somewhat by thermal stratification as compared to the daytime unstable regimes (Baldocchi and Meyers, 1988; Shaw et al., 1988). The mean wind profiles follow a well-defined logarithmic pattern above the canopy and an exponential function within the canopy (Raupach and Thom, 1981). The two aerodynamic parameters that appear in the logarithmic expression, roughness length and displacement height, are functions of canopy morphology (averaged tree height, leaf area index, tree geometry, etc., Shaw and Pereira (1982), Rosenberg et al. (1983), Raupach (1992) and Warland (1996)). The empirical coefficient in the exponential function is primarily a function of leaf area index: working formulae were established from simulations of canopy flow using a higher order closure model (Lee, 1997) and from simple mixing length arguments (Raupach and Thom, 1981). The exponential model is adequate in the canopy layer but is not a good model near the ground within a forest with an open trunk space that permits a sub-canopy jet (Shaw, 1977).

Strong radiative cooling ($R_{\mathrm{n}} < -60\,\mathrm{W\,m^{-2}}$) under clear skies coupled with weak ambient winds ($<1.5\,\mathrm{m\,s^{-1}}$ over the forest) will lead to sporadic or intermittent turbulence. Brief outbursts of turbulence interrupt otherwise smooth air motion. The intermittent occurrence of turbulence is interpreted by Nappo (1981) as evidence of overturning of the air in the planetary boundary layer, a phenomenon he observed over simple and complex terrains. Mixing is very weak except during the brief turbulent periods, leading to a huge accumulation of CO_2, water vapor and air pollutants within the forest. CO_2 mixing ratio up to 800 ppm is a common occurrence near the forest floor. The elevated concentration is not completely diluted until 2–3 h after sunrise, even in a sparse forest (Yang, 1998). The high CO_2 concentration may enhance seedling growth through a CO_2 fertilizing effect (Bazzaz and Williams, 1991), but this diurnal pattern, to the author's knowledge, has not been considered by previous CO_2 enrichment experiments which all maintain constant CO_2 concentrations under controlled conditions. The cooler air layer in the canopy is essentially decoupled from the warmer layer above. Radiative frosts are most likely to occur under these conditions (Avissar and Mahrer, 1988; Blennow, 1988; Blennow and Persson, 1998).

Under very stable conditions, the vertical gradient of the Reynolds stress (term 6, Eq. (1)) is small and therefore the horizontal pressure gradient (term 5, Eq. (1)), either associated with synoptic weather systems (Lee et al., 1994) or baroclinic forcing due to even a slight terrain slope (Wyngaard and Kosovic, 1994), is relatively large and cannot be ignored. The usual assumption that the Reynolds stress gradient is balanced by the canopy drag force is not valid. The logarithmic/exponential wind model is no longer a good one in describing the wind speed profile within and above the canopy (Fig. 1).

The third air motion type is characterized by periodic patterns in the time series of temperature or other scalars and the velocity components (Fig. 2). The periodic pattern is indicative of canopy waves (buoyancy waves in the canopy). A recent study of canopy

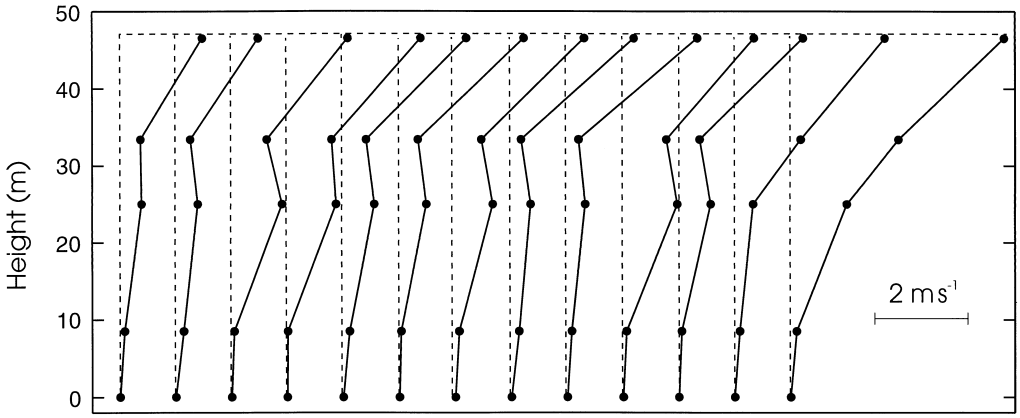

Fig. 1. Successive 30 min averages of the longitudinal wind speed measured with sonic anemometers in the mixed deciduous forest at Camp Borden during period 0:30–7:30 local daylight time, October 5, 1998. The mean tree height was 22 m and LAI was 2.0. Details of the experiment can be found in (Fuentes and Wang, 1999).

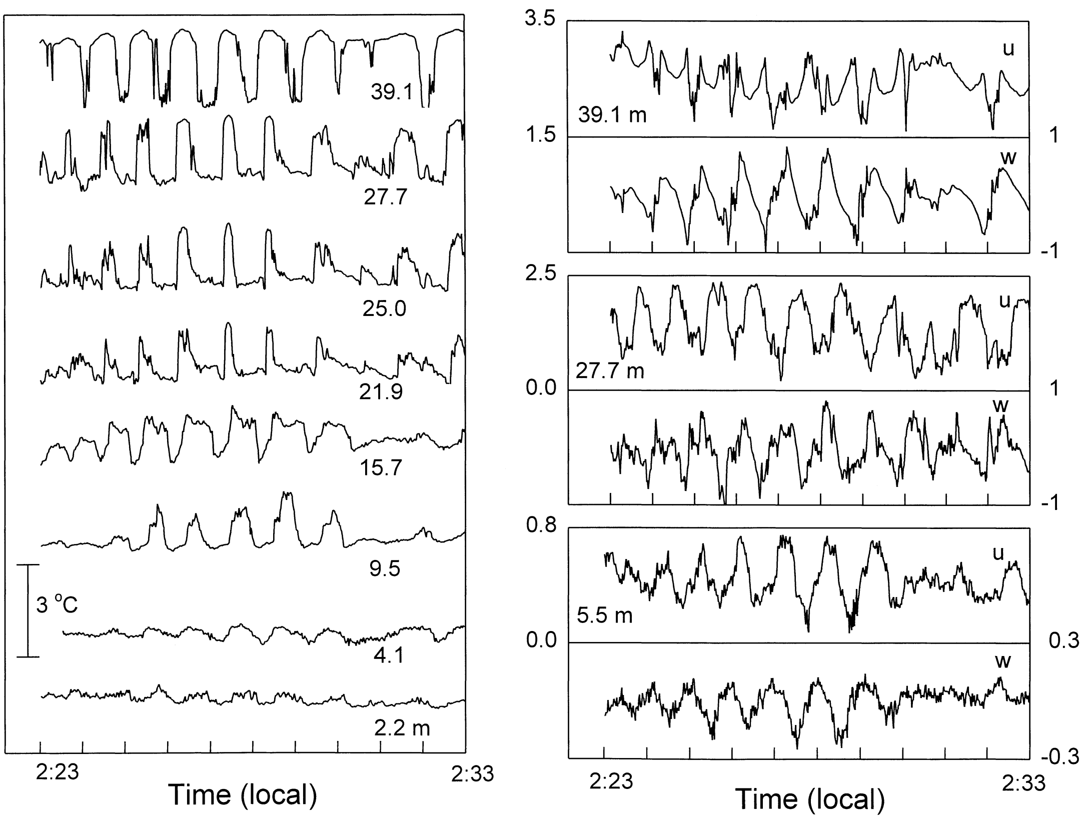

Fig. 2. Ten-minute time series of temperature (°C), longitudinal (u, ms^{-1}) and vertical velocity (w, ms^{-1}) during a canopy wave event (2:22–3:22 local standard time, July 13, 1994) in a boreal aspen forest. Heights of observation are indicated. The average tree height was 21 m and LAI was 1.5.

wave climatology shows that the wave motion is a common form of air motion in forest vegetation at night (Lee and Barr, 1998). These waves are shear-generated. They propagate in the direction of the mean wind at a speed slightly faster than the wind speed at the treetops. The median values of wave speed, wavelength, wave vertical displacement, and wave frequency were found to be 1.6 $\mathrm{m\,s^{-1}}$, 75 m, 10 m and 0.021 Hz, respectively, based on a 5-month observation in a boreal forest.

It is shown from a linear analysis of the shear instability that momentum flux in the air layer undergoing the instability process is not constant with height (Davis and Peltier, 1976). It can also be shown that scalar fluxes will reverse sign across the critical level, the height where the wave speed matches the mean wind speed. This may be the reason why fluxes observed over forests behave erratically in the presence of canopy waves (Lee et al., 1996). The lack of a constant flux layer also introduces some difficulty in relating the average wind load on trees to the measured momentum flux above the canopy (Gardiner, 1995).

The mechanisms by which the forest vegetation interacts with the atmosphere in the presence of canopy waves are yet to be fully explored. Canopy waves accounted for significant release of CO_2 from a tropical forest at night (Fitzjarrald and Moore, 1990). The interpretation of fluxes of momentum, sensible heat and CO_2 over the forest vegetation as derived from the conventional time or Reynolds averaging (Stull, 1988) is however not straightforward (Lee et al., 1996). Unlike the daytime gusty wind that penetrates deep into the forest (Denmead, and Bradley, 1985) and favors rapid settling of pesticide spray droplets to the canopy thus enhancing the catch efficiency (Miller et al., 1993), oscillatory motions will reduce the settling velocity of those particulates whose response time is much shorter than the oscillation period (Stout et al., 1995). A consequence of this could be that certain seeds can spread farther out if released into the wavy air at night. Certain fungus spores are released from the forest floor at night (Schmidt and Wood, 1972). Pressure fluctuations at the soil surface induced by the wave motion aloft may help the takeoff through a pressure pumping effect. One general character of the wave motion is that it is persistent in time. The resonant response of stems and the whole tree to the motion may compound the wind stress.

3.2. Modeling

Modeling the canopy flow under very stable conditions (intermittent turbulence regime) is a challenging task. One problem arises from the difficulty in finding an interval suitable for performing the Reynolds averaging when the turbulence is globally intermittent (Mahrt, 1999). Because static stability plays an important part in the flow dynamics, a successful model must include a proper parameterization of the heat exchange between the air and the plants. As noted above, the horizontal pressure gradient must also be included in the governing equations. The pressure gradient is, however, associated with atmospheric circulations at scales much larger than the scale of micrometeorology and therefore the information is not readily available. For these reasons, very few modeling studies of nocturnal canopy flow can be found in the literature (Yamada, 1982; Schilling, 1991).

In comparison with the intermittent turbulence regimes, we are in a better position to model canopy waves. Theories of buoyancy wave dynamics, well established by studies of laboratory flows (Drazin and Reid, 1981) and flows in the atmosphere (Gossard and Hooke, 1975) and the ocean (Turner, 1973; Thorpe, 1987), can be extended to investigate the canopy wave phenomenon. For example, Lee (1997) developed a two-dimensional inviscid linear model whose essential feature is a Taylor–Golstein equation modified to account for the canopy drag effect. As with any other linear models (Gossard and Hooke, 1975), Lee's governing equations keep the first-order linear perturbation terms and ignore all non-linear terms. The study shows that the nocturnal canopy flow indeed permits an unstable mode that shares common features of a Kelvin–Helmholtz instability (Davis and Peltier, 1976; Lalas and Einaudi, 1976): a phase speed equal to the background wind near the center of the shear layer, a horizontal wavelength proportional to the depth of the shear layer, and an amplitude that decays rapidly away from the region of shear.

The linear model is only a small step towards a full understanding of the canopy wave dynamics because in reality shear-generated waves can quickly grow out of the linear phase. One feature evident of almost all canopy wave events is the existence of a monochromatic wave frequency (Fig. 2). This makes it possible to separate the wind and scalar fields into mean, wave and turbulence components using a triple decomposition procedure involving a phase averaging (Finnigan et al., 1984). The interactions among the three components can then be investigated with ensemble models that use closure schemes such as the E–ϵ scheme proposed by Fua et al. (1982) for waves in the upper boundary layer.

It is also possible to investigate the wave generation, its non-linear growth, the secondary instability and possibly wave breaking using LES (large eddy simulation). Several LES studies of canopy turbulence under neutral stability have been published (Shaw and Schumann, 1992; Dwyer et al., 1997; Shen and Leclerc, 1997). LES of stably stratified flow encounters the difficulty of separating the resolvable scale motion from the sub-grid scale motion (Mason and Derbyshire, 1990). In the case of wave motions, the problem is less severe at the stage before and possibly beyond the initial wave breaking, as shown by the successful two-dimensional LES simulation of wave generation in the stratosphere (Schilling and Janssen, 1992). It is hypothesized, based on the fact that the wave motion is the normal state of affairs in forests, that wave breaking and secondary instability are the principal mechanisms of turbulence generation in forests at night. Studies of this motion type should greatly improve our understanding of nocturnal canopy turbulence and transport processes.

4. Disturbed canopy flow

4.1. Overview

This section reviews the literature on air motion in inhomogeneous landscapes characterized by forest edge transitions, forest clearings, and patchy canopies. The literature on windbreak flow, another important case of the disturbed flow, has been reviewed elsewhere (Plate, 1971; Wilson, 1985; McNaughton, 1989; Judd et al., 1996). A feature common to all these disturbed flow situations is that the one-dimensionality adopted by experimental and modeling studies of flow in horizontally uniform forests is now lost (terms 2–7 of Eq. (1) not equal to zero).

Most of the observational and modeling studies cited below are restricted to the simple case of flow perpendicular to 2D edges. There is evidence to show that windbreak flow is not sensitive to the wind angle in the range $\pm 25^\circ$ of the perpendicular angle (Wilson, 1997). In contrast, flow across a forest edge is essentially three dimensional, because changes in the pressure gradient (term 5, Eq. (1)) and the Coriolis (term 7, Eq. (1)) forces across the edge will lead to a large swing in wind direction (Wilson and Flesch, 1999). We lack data for more complex (and more realistic) geometries (e.g., irregular clearcut patterns) to guide proper logging operation to reduce wind stress. Obviously, wind tunnel and computer simulations are the most economical ways of addressing this knowledge gap (Flesch and Wilson, 1998).

4.2. Forest edge

4.2.1. Exit flow

The first comprehensive field study of air flow across a forest edge was reported by Raynor (1971). His observation, even though limited only to the mean wind field, proved to be quite useful in validation of models of various complexities (Shinn, 1971; Li et al., 1990; Wilson and Flesch, 1999). In later observational studies, both in the field (Gash, 1986; Miller et al., 1991; Irvine et al., 1997; Flesch and Wilson, 1999a) and in wind tunnels (Chen et al., 1995), fast response anemometers were deployed to quantify higher-order Reynolds statistics and turbulence spectra across the edge.

Examination of exit flow (flow from forest to open field) across a two-dimensional edge can be posed in the conceptual framework modified from the one proposed for windbreak flow (e.g., Judd et al. (1996)), as shown in Fig. 3. The wind profile approaching the edge (profile A) is identical to that in the interior of the forest. At the forest edge (profile A1) the profile retains the broad pattern of profile A (e.g., a strong inflection point near the treetops), with perhaps a slight acceleration near the ground. The flow field in the lee of the forest edge can be divided into three zones:

1. Quiet zone (D): The portion of the profile within this zone does not differ much from the exit profile A1. At the far end of zone D, the inflection point disappears from the profile. Limited data suggests that the zone extends to a distance of 4–7h, where

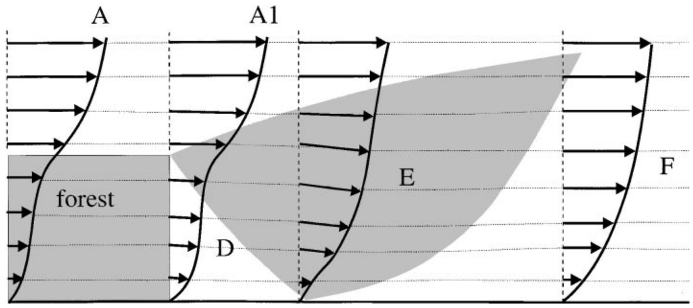

Fig. 3. Schematic diagram of exit flow across a two-dimensional forest edge. The flow field in the open is divided into quiet zone (D), mixing zone (E) and re-equilibration zone (F).

h is the mean tree height (Raynor, 1971; Chen et al., 1995).

2. Mixing zone (E): Streamlines in zone E are tilted slightly downward, a feature supported by most studies (Raynor, 1971; Li et al., 1990; Green, 1992; Stacey et al., 1994; Chen et al., 1995) except Fritschen (1985). Mixing is enhanced near the height $z = h$ in this zone, in comparison to that further downstream, possibly by the Kelvin–Helmholtz instability similar to that found in a plane mixing layer. The plane mixing layer analogy has been used to explain turbulent structures downstream of windbreaks (Plate, 1971; Zhuang and Wilson, 1994; Judd et al., 1996). Direct signatures of the instability such as coherent structures and Kelvin–Helmholtz billows are yet to be reported for the edge flow, but one could consider the enhanced turbulent kinetic energy up to $x/h = 5$ near the treetop level (Liu et al., 1996) and intermittent recirculating eddies (Bergen, 1975) as indirect evidence for the analogy, where x is distance from the forest edge. Unlike the flow in the interior of the forest where persistent strong wind shear due to the canopy drag maintains the instability and the subsequent turbulence, here the instability is self-destructing because it acts to smooth out the mean shear that supports the instability. Hence the inflected profile is quickly replaced by a logarithmic one.
3. Re-equilibration zone (F): The profile is fully equilibrated with the new surface. The distance at which this occurs decreases with increasing surface roughness of the open field. In a wind tunnel study that simulates air flow from a forest to a very smooth open field, Chen et al. (1995) found that even at a distance of $x = 22h$ from the forest edge, the mean wind speed is still lower than the potential value measured with no upstream obstacles. Observations by Raynor (1971) in an open field downwind of a coniferous forest shows that the logarithmic profile established at $x = 5h$.

4.2.2. Flow into forest

Turning attention now to flow from the opening to the forest, we note two key features as summarized by McNaughton (1989): (1) a rapid adjustment that is essentially complete by about $x = 10h$ from the forest edge, and (2) complex flow structures within the adjustment zone, often with a jet in the trunk space and probably an intermittent rotor above and within the upper canopy. Shinn (1971) showed that the wind speed in the trunk space follows an exponential decay function with distance from the forest edge.

Several workers reported that the Reynolds stress over the forest in the adjustment zone is not constant with height (Kruijt et al., 1995; Irvine et al., 1997). This is not a surprise because a large horizontal pressure gradient at the edge (term 5, Eq. (1); Li et al. (1990)) must be balanced a vertical divergence of the Reynolds stress (term 8, Eq. (1); Baldocchi and Meyers (1988)).

Both numerical simulations (Li et al., 1990) and observations (Fritschen, 1985; Irvine et al., 1997) show that the mean velocity vector in the adjustment zone is tilted upward at a slight angle, implying horizontal flow convergence or a positive mean vertical velocity. Partitioning of scalar fluxes observed near the edge to contributions by the open field and the forest via, for example, footprint decompositions, will be misleading, because a large mass flow component of the surface-air exchange (Lee, 1998; Lee et al., 1999) is not captured by the conventional micrometeorological methods.

It is proposed that low frequency turbulence may adjust to new surfaces more slowly than high frequency turbulence (Panofsky and Dutton, 1984). Since variations in the horizontal velocity are dominated by lower frequency or larger eddies whereas those in the vertical velocity are dominated by higher frequency or smaller eddies, one expects both the spectrum and variance of the horizontal velocity to adjust more slowly than those of the vertical velocity. This postulate is however not supported by observations (Gash, 1986; Irvine et al., 1997). It appears that the eddy structure is quickly modified in the adjustment zone through a dynamic coupling between the flow and the forest (Hill, 1989; Kaimal and Finnigan, 1994).

4.3. Forest clearings

Wind patterns in a forest clearing vary from direct throughflow to recirculations (Bergen, 1975; Weiss and Allen, 1976; Miller et al., 1991; Stacey et al., 1994; Flesch and Wilson, 1999a). The recirculation or flow reversal occurs near the ground ($z/h < 0.4$) in the form of an intermittent rotor (Bergen, 1975). It appears that the flow reversal is most likely to be detected in a small clearing, when the ambient wind speed is strong, or in a clearing surrounded by dense forests. Above the recirculation cell, the streamlines are tilted downward at an average angle of 7° (Miller et al., 1991), a feature similar to the exit flow discussed above.

In synthesizing both data of exit flow and flow in forest clearings, Flesch and Wilson (1999a) found that wind recovery downwind of the forest edge appears to increase linearly with distance from the edge, but a large scatter exists among the datasets they examined. They attribute the scatter to the differences in the pressure fields due to variations in the forest/clearing geometry. Interestingly, data of the TKE (turbulent kinetic energy) recovery seems to follow a more universal form, showing a rapid increasing trend with distance from the upwind edge, a peak value at $x/h \simeq 4 - 6$ and a decreasing trend with distance beyond $x/h > 6$. Once again the increasing trend in TKE can be viewed as evidence of the Kelvin–Helmholtz instability linked to the inflected velocity profile at the upwind forest edge.

If the clearing is large enough, thermal circulations will develop as a result of the differential cooling/heating between the clearing and the surrounding forest. Schilling (1991) used a 2D ensemble model to simulate air flow in a 4 km wide clearing bordered by coniferous forests. He found that in the daytime, air was forced out of the forest in the direction of the clearing where convergence leads to an upward motion. The pattern reverses at night. In his model domain, the clearing is bare. If the clearing is occupied by agricultural or grass plants, the combination of a number of factors —higher heat flux into the ground, lower stomatal resistance and higher albedo in the open field than at the surrounding forest —would result in a higher sensible heat flux over the clearing than over the forest in the daytime, and hence the flow will be directed from the clearing to the forest (Andre et al., 1989).

4.4. Sparse canopies

Orchards (Weiss and Allen, 1976; Baldocchi and Hutchison, 1987; Wang et al., 1992) and agroforestry systems (Alcock and Thomas, 1986; Sibbald et al., 1987) are good examples of sparse canopies. The flow field is no longer uniform in the horizontal, and the effect of individual plants is discernible. The roles of the plant elements are twofold, mechanical (generation of wakes and sheltering) and thermal (differential heating/cooling between plants and gaps). The mechanical effect was observed by Green et al. (1995) in a sparse forest (tree spacing $1h$) where large spatial variations exist in the mean velocity and velocity fluctuations between the row and gap positions. The study of Sun and Mahrt (1995) is a good example showing the thermal effect. They observed a large temperature difference between well-ventilated treetops and the less ventilated ground surface in a sparse boreal black spruce forest. When the ambient wind is weak, it is possible to observe a regular, small-scale thermal circulation pattern in such a forest whereby cold downdrafts and buoyant updrafts are locked to the trees and the gaps, respectively, leading to a significant dispersive flux (Finnigan and Raupach, 1987).

In contrast, little spatial variations were observed in the velocity fluctuations (Amiro and Davis, 1988), sensible heat flux (Lee and Black, 1993a) and carbon dioxide flux (Yang, 1998) in dense forests, suggesting that wakes and local scalar sources/sinks associated with plant elements are blended to form a flow field that is more or less uniform in the horizontal.

4.5. Flow models

Modeling studies of canopy flow face two fundamental issues, how to quantify the drag and heat exchange of individual plant elements and how to treat the closure problem. Theories addressing the first issue are well established now (Thom, 1975; Wilson and Shaw, 1977; Raupach and Shaw, 1982; Finnigan, 1985). The issue about the closure problem remains an active area of research. The problem exists because there are more unknowns than the equations that can be derived from first principles. Methods of handling the problem are called closure schemes, which add to the model a few more (empirical) parameterization equations so that the number of model equations equals the number of unknowns (Stull, 1988).

With only one exception (Patton et al., 1998), all modeling studies of the disturbed flow rely on ensemble models. Two of the simplest of these are a similarity solution with a modified wall-jet approach (Shinn, 1971) and a phenomenological analytical description (Albini, 1981) for the edge flow. Subsequent models can be sorted into groups based on their closure schemes (up to $1\frac{1}{2}$ order). Second-order schemes have been used for windbreak flow (Wilson, 1985; Wang and Takle, 1997) but not for the flow configurations considered here.

The first-order closure, also called 'K-theory', parameterizes second-order moments in relation to the appropriate first-order moments. For example, the Reynolds stress ($\overline{u'w'}$, see term 8 of Eq. (1)), which is the covariance between the longitudinal and the vertical velocity and hence a second-order moment, is related to the mean longitudinal velocity (u, a first-order moment), as

$$\overline{u'w'} = -K\frac{\partial u}{\partial z} \qquad (3)$$

where the eddy viscosity, K, is often expressed in a form similar to the following

$$K = l^2\left|\frac{\partial \vec{U}}{\partial z}\right|\frac{\partial u}{\partial z} \qquad (4)$$

where l is mixing length. To simulate the edge flow, Li et al. (1990) and Miller et al. (1991) adopted the boundary layer model of Blackadar (1962) for l for the open field, related l to the canopy structure for the forest, and interpolated between the two asymptotic forms for the transition zone. This type of model often draws criticism because model parameters are tuned to match observations (Wilson and Flesch, 1999). Another limitation is that higher-order statistics useful for wind load studies, such as TKE and velocity variances, are not computed.

A variant of the first-order closure, termed Prantdl–Kolmogorov closure, assumes

$$K = alE^{1/2} \qquad (5)$$

where a is an empirical constant and E represents TKE which is obtained by including in the model a TKE budget equation. This closure scheme was used by Schilling (1991) for flow in a large clearing, with Blackadar's lengthscale formulation without accounting for the presence of the canopy. In two recent studies, Wilson and Flesch (1999) and Wilson et al. (1998) proposed a lengthscale which is the maximum of an outer scale and an inner scale: the outer scale takes the usual inertial subrange form and the inner scale recognizes the presence of both the ground barrier and shear instability near the canopy top. Their parameterization avoids the need of interpolation in the transition zone and appears to have captured, at a conceptual level, current understanding of canopy turbulence (Raupach et al., 1996). The model is coupled with a simplified Sk_u (skewness of the horizontal velocity) budget equation to give prediction of the velocity skewness in the clearcut.

The next level in the closure hierarchy is the E–ϵ closure or $1\frac{1}{2}$ closure. The model consists of the continuity equation and equations for momentum, TKE (E) and the rate of dissipation of TKE (ϵ). The eddy diffusivity is expressed as

$$K = \frac{C_\mu C_D E^2}{\epsilon} \qquad (6)$$

and standard values are assumed for coefficients C_μ and C_D. Reasonable agreement with observations was achieved for flow in and around an isolated forest patch (Green, 1992), in a forest edge transition (Liu et al., 1996), and in a model forest on a 2D hill (Kobayashi et al., 1994). The ϵ equation, however, suffers from a fundamental uncertainty about how best to treat the dissipation mechanisms within the canopy (Wilson and Flesch, 1999; Wilson et al., 1998). Fine-tuning of the model parameters may be needed to achieve a good match with observations (Liu et al., 1996).

It has long been recognized that in the lower part of an extensive forest, the Reynolds stress (term 8, Eq. (1)) vanishes, thus reducing the flow field to a simple balance between the canopy drag (term 9, Eq. (1)) and the pressure gradient force (term 5, Eq. (1); Shinn (1971), Smith et al. (1972) and Holland (1989)). This leads to a secondary wind speed maximum and the alignment of wind direction with the geostrophic wind near the forest floor (Kondo and Akashi, 1976; Lee et al., 1994). The typical values of the pressure gradient force are very small, being 0.01 and 0.03 hPa/km for synoptic weather systems (Wallace and Hobbs, 1977) and mesoscale thermal circulations (Atkinson, 1981), respectively, and yet are sufficient to result in the secondary wind speed maximum and the wind directional shear. In comparison, the pressure gradients in the edge flow (Li et al., 1990) and in the lee of a windbreak (Wilson, 1997) are much larger (0.2–2 and 4 hPa/km, respectively). Therefore, proper account of the pressure gradient force is crucial for successful simulations of the disturbed flow (Finnigan and Brunet, 1995; Wilson and Flesch, 1999; Wilson et al., 1998).

5. Canopy flow over undulating terrain

Topographic influence on the wind pattern is of considerable interest because windthrow results from complex interactions of the topography with soil properties and stand characteristics (Mitchell, 1995). The nature of the influence depends on the scale of the topography. Over isolated small hills ($<$ 1km) with moderate slopes (≤ 0.3), wind variations are primarily associated with aerodynamic (by modifying term 5, Eq. (1)) rather than thermal forcing (term 6, Eq. (1)), and it is possible to obtain simple guidelines for estimating near-surface wind speed variations (Taylor and Lee, 1984; Taylor et al., 1987). In more complex terrain, both local circulations arising from differential heating and cooling (e.g. mountain/valley circulations) and mechanical modifications of the prevailing wind (e.g., channelling and speedup over ridge-tops) are important. Thermally-induced flow does not pose a threat to trees except the case with large slopes ($>$ 50 km) which can generate katabatic wind (also termed chinook or föhn wind) as strong as the hurricane wind (Barry, 1992).

There is a large gap between forest meteorology which, traditionally, is concerned with microscale processes (about 1 km) and mesoscale meteorology which is concerned with atmospheric phenomena at scales of several tens of km. In principle, mesoscale flow models are the desirable tool for studies of wind and turbulence in complex terrain. Some of these models are quite advanced (Pielke, 1994). For example, the NCAR mesoscale model has been adopted for operational forecasts of turbulence at the Hong Kong Airport situated in extremely complex terrain (NCAR, 1996). The cost of running such models for wind damage assessment is however prohibiting. It remains an open question whether the mesoscale models can generate accurate wind fields near the tree canopy.

An alternative and more economical method was proposed by Hannah et al. (1995) and Quine et al. (1994), who relate, in an empirical fashion, the surface wind speed to local topographic and geographic characteristics. In a similar vein, Yamazawa and Kondo (1989) proposed a two-step empirical-statistical approach to estimate the surface wind speed over complex terrain under strong wind conditions. Their first step extends the wind speed observed at a nearby reference point (e.g., an airport weather station) to the forest, accounting for the difference in the surface roughness between the two locations using the Rossby similarity (Garratt, 1977; Silversides, 1978; Tennekes, 1981). The second step corrects for the local terrain effect using a simple topographic parameter. The surface wind estimates can then be further extended, using the appropriate logarithmic/exponential functions, to the air layer within the canopy for wind load assessment. These empirical approaches do not take thermal stratification into account and hence should be limited to situations of strong winds and near-neutral stability. Uncertainties also remain about how to best estimate wind speed within the canopy as the shape of the wind profile is now dependent upon both the canopy structure and position along the slope (Finnigan and Brunet, 1995).

Only limited datasets exist for canopy flow on hills (Baldocchi and Meyers, 1988; Inglis et al., 1991; Lee and Black, 1993b; Kobayashi et al., 1994; Finnigan and Brunet, 1995). Of these, the wind tunnel study of Finnigan and Brunet (1995) provided the most comprehensive dataset to date. They identified the pressure perturbation with the topography as a key feature to explain the evolution of the canopy velocity profile on a hill. The dominant role of the pressure force is further supported by the success of a first-order closure simulation of the flow (Wilson et al., 1998).

Both the above wind tunnel and numerical simulations assume neutrally stratified flow. While this is a good simplification for wind load studies because wind damage occurs at high winds with minimum thermal stratification, it is worth mentioning that numerous practical reasons call for information on canopy flow under stable stratifications, particularly drainage flow in the forest environment. Drainage flow is a major problem in studies of nighttime forest respiration rates with the eddy correlation technique (Lee, 1998). Cold air drainage enhances the risk of frost near the bottom of a slope (Oke, 1987) and forest fire at the mid-slope (Schroeder and Buck, 1970). Models of drainage flow work reasonably well for slopes without vegetation (Brost and Wyngaard, 1978; Mahrt, 1982; Kondo and Sato, 1988). These models may perform poorly in the canopy layer because they do not include any treatment of the canopy drag (term 9, Eq. (1)) and plant–air heat exchange processes.

6. Extreme wind events

Extreme wind gusts, their force increasing in proportion to the square of the wind speed, are the low frequency events of most concern in building (Panofsky and Dutton, 1984) and wind turbine designs (Justus, 1985). One approach to determining the frequency of occurrence is to use the Weibull model distribution for wind speed, long used in wind energy assessment (Justus, 1985), as

$$p(u) = \left(\frac{k}{c}\right)\left(\frac{u}{c}\right)^{k-1}\exp\left[-\left(\frac{u}{c}\right)^{k}\right] \tag{7}$$

where u is wind speed, c is called the scale factor (units of speed) and k is the shape factor (dimensionless). The probability of speeds above a threshold u_x is given by

$$p(u > u_x) = \exp\left[-\left(\frac{u_x}{c}\right)^{k}\right] \tag{8}$$

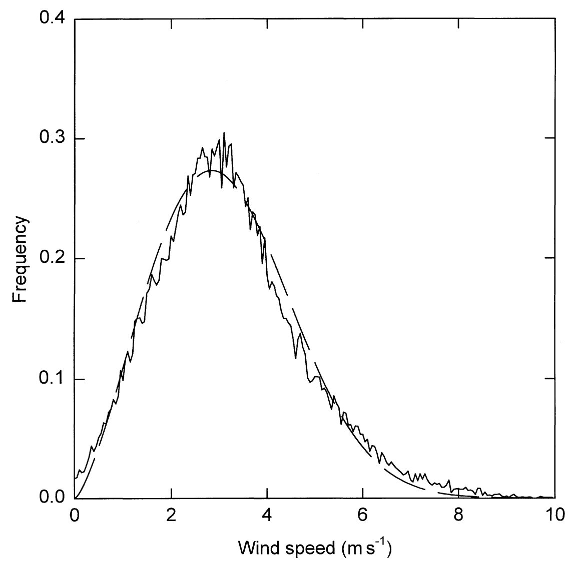

Fig. 4. Distribution of 30 min average wind speed at 23 m above the Camp Borden forest (Shaw et al., 1988; Fuentes and Wang, 1999) during 1995–1997: solid line, observation; smooth dashed line, Weibull model with $k = 2.39$ and $c = 3.56$.

Fig. 4 provides an example of the wind speed distribution observed at the Camp Borden forest on flat terrain. Justus (1985) reviewed empirical methods of inferring c and k for situations when only limited wind information is available. The predicted probability of extreme events (Eq. (8)), after proper account for the edge effect, wind attenuation in canopies, and the streamlining effect (Mayer, 1989; Monteith and Unsworth, 1990; Hedden et al., 1995), can be used with static wind load models (Blackburn et al., 1988; Peltola et al., 1997) to provide an assessment of wind risk probability (Gardiner and Quine, 1999).

A second approach of handling extreme events involves simulations of the dynamic response of trees to the wind. It is recognized that wind damage to trees is most likely to occur at high wind speeds and when the excitation frequency (frequency of wind gusts) coincides with the natural frequency of sway vibrations. The resonance is considered to occur if the spectral peak frequency of the velocity time series matches the natural frequency of sway vibrations (Mayer, 1989; Gardiner, 1994; Wood, 1995; Peltola, 1996). In reality, the spectral peak frequency is a rather ambiguous quantity because of the spectral flatness around the peak frequency. To that end, simulations of turbulent time histories via the proper orthogonal decomposition (Panofsky and Dutton, 1984; Mahrt and Frank, 1988) or wavelet functions (e.g., Katul and Vidakovic, 1998) may be more appropriate. The generated time series can then be used as a forcing function in tree dynamic response models (Wood, 1995; Kerzenmacher and Gardiner, 1998).

7. Discussion and conclusion

Kelvin–Helmholtz instability of the plane mixing flow, proposed first as a mechanism of mixing in windbreak flow (Plate, 1971; Zhuang and Wilson, 1994) and later to explain the coherent eddy structure (Raupach et al., 1996) and the canopy wave phenomena (Lee, 1997) in extensive canopies, is now a widely-accepted framework for the dynamic coupling of the vegetation with the moving air (Hill, 1989). It is possible that the mechanism is also responsible for much of the mixing in the lee of a forest edge. Because turbulence is generated locally by the instability, the Monin–Obukhov similarity strictly does not hold. The non-ideal situations discussed here introduce additional complications such as lack of a constant flux layer, the non-negligible role of the (horizontal) pressure gradient and the baroclinic forcing, and the difficulty in footprint decomposition.

The synthesis has also identified a few key issues in need of further research:

1. Wavelike motion is a common form of air motion in the forest vegetation. The mechanisms by which it interacts with the trees and its role in the evolution of the nocturnal boundary layer remain poorly understood. For example, there is observational evidence to suggest that weakening of the low level jet over the forest could be associated with enhancement of the surface drag by the turbulence generated by wave breaking, but a definite link is yet to be established.
2. We lack data of air flow in irregular (and realistic) clearcut patterns. It is comforting to note that models based on simple closure schemes work reasonably well for the disturbed flow so long as the pressure gradient force is treated properly. Information on the background pressure gradient is however not always available.
3. Windthrow hazard assessment calls for guidelines on the topographic influence on the surface wind. Documenting qualitative indicators (Justus, 1985; Mitchell, 1995) is only a small step towards this goal. More quantitative studies like those cited above, albeit empirical, should be encouraged.
4. Obviously, extreme wind events are of great concern. The illustrative treatment of this topic has identified a few important elements (wind probability, extrapolation of wind from a nearby monitoring station to the forest, the streamlining effect under strong wind, and the dynamic response of the trees to the turbulence) for future studies.

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