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Impact of shelter on crop microclimates: a synthesis of results from wind tunnel and field experiments

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Abstract. The purpose of this paper is to synthesise data from the literature, and acquired during an extensive set of wind tunnel and field experiments, to quantify the effect of porous windbreaks on airflow, microclimates and evaporation fluxes. The paper considers flow oriented both normal (i.e. at right angles) and oblique to the windbreak, in addition to the confounding effects of topography.

The wind tunnel results confirm the validity of the turbulent mixing layer as a model for characterising the airflow around a windbreak and for predicting the locations of the quiet and wake zones. This mixing layer is initiated at the top of the windbreak and grows with distance downwind until it intersects the vegetation or surface, marking the downwind extent of the quiet zone where the maximum shelter occurs. The 3 factors that determine the growth of this mixing layer are the windbreak porosity, windbreak height and the nature of the terrain upwind. For wind that is flowing normal to a porous windbreak in the field, the latter 2 have the primary influence on the size of the sheltered zone, while windbreak porosity is the main factor determining the amount of shelter. Analyses of the effect of porosity revealed that the amount of wind shelter increases as windbreak porosity is reduced, but the downwind extent of the sheltered zone does not vary with windbreak porosity. Thus, the suggestion from older studies that low-porosity (i.e. dense) windbreaks lead to a reduced sheltered area is not supported by the wind tunnel measurements.

In the absence of shading effects, temperature and/or humidity are increased in the quiet zone, mirroring the pattern and magnitude of wind shelter. Thus, the increase in temperature and humidity is greatest where the minimum wind speed occurs, and the magnitude of the increase is smaller for more porous windbreaks.

The humidity and air (but not surface) temperatures are decreased very slightly in the wake zone, although these small changes were not significant in a field situation. Microclimate changes, therefore, occur over a much smaller distance downwind than wind shelter, and are negligible for the very porous windbreak. For example, at 20 windbreak heights downwind, the wind speed may still be 80% of its upwind value, while the air and surface temperature and humidity have returned to their upwind values after 12–15 windbreak heights. Furthermore, these changes in temperature and humidity vary with the type of land cover, surface moisture status and the temperature and humidity of the 'regional' air. Over the course of a growing season, these changes can be masked by soil and climate variability.

The turbulent scalar fluxes, i.e. evaporation and heat fluxes, also differ from the pattern of near-surface wind speeds. While significantly reduced in the quiet zone, they show a very large peak at the start of the wake zone — the location where the mixing layer intersects the surface. Thus, caution is required when extrapolating from the spatial pattern of shelter to microclimates and turbulent fluxes.

Wind flowing at angles other than normal to the windbreak has 2 effects on the pattern of wind shelter. First, for the medium and low porosity windbreaks used in the wind tunnel, the amount of wind shelter is increased slightly in the bleed flow region near the windbreak, i.e. there is an apparent reduction in windbreak porosity as the wind direction becomes more oblique to the windbreak. Second, the profile of near surface wind speeds is similar to that for flow oriented normal to the windbreak, providing that the changes in distance from the windbreak are accounted for using simple geometry. The field data agree with these results, but show an even greater influence of the windbreak structure on the pattern of wind shelter in the bleed flow region, extending from the windbreak to at least 3 windbreak heights downwind, precluding any generalisations about the flow in this region.

Additional keywords: windbreaks, National Windbreaks Program, turbulent mixing layer, turbulent scalar fluxes.

Introduction

The possibility that shelter from windbreaks may improve plant and animal productivity has been the subject of a great deal of research (e.g. reviews by Kort 1988; Bird 1998) over the last 50 years. Much of this work has focused on measurements of the airflow around barriers, including screens, artificial windbreaks and tree windbreaks, and/or monitoring plant growth and yield around windbreaks. As noted by Cleugh (1998), there have been far fewer studies quantifying the subsequent effects of wind shelter on the microclimate experienced by plants, in particular the temperature and humidity fields and the turbulent fluxes of heat, water vapour and carbon dioxide. Yet, these are some of the processes by which shelter affects plant productivity (see Cleugh 1998 for a complete discussion of this and other factors). The absence of such thorough and detailed investigations has limited the ability to identify the mechanisms by which windbreaks modify plant growth, and thus predict windbreak effects on plant productivity.

As a contribution to the National Windbreaks Program (see Cleugh et al. 2002 for description), a series of wind tunnel and field experiments have been conducted with the aim of quantifying the effect of windbreaks on airflow, microclimate and evaporation fluxes. Detailed descriptions of these experiments, and analyses of their results, are contained in the publications of Cleugh (2002a) and Judd et al. (1996). (Details about the wind-tunnel experiment and data analyses are available from the authors.) The aim of this paper is to draw on these results, and other experiments and literature as appropriate, to present a synthesis of the impact of shelter on airflow, microclimates and evaporation fluxes. The specific objectives of the experimental program, and hence this paper, are to: (i) characterise the mean and turbulent velocity and scalar fields around porous windbreaks; (ii) evaluate the current model of airflow around porous windbreaks, and hypotheses regarding microclimate changes in response to this airflow; (iii) thus, quantify the effect of windbreak porosity, upwind terrain and oblique flow on airflow, microclimates and evaporation fluxes; and (iv) develop simple parameters for predicting the spatial

patterns of wind, temperature and humidity fields downwind of porous windbreaks.

Cleugh (2002b) addresses the last of these objectives, while this paper focuses on the first 3.

Background

The main airflow regimes around a porous windbreak, for wind flowing normal (i.e. at right angles) to the windbreak, are pictured in Figure 1. Representing the currently accepted model, the key airflow zones shown in Figure 1 include the bleed flow through the porous windbreak; triangular-shaped quiet zone; and a mixing or wake layer that grows in depth with distance downwind until it intersects the surface at about 5-10 windbreak heights (H). (As noted by a reviewer, the term 'wake zone' is inappropriate. A better term would be 'high turbulence zone', which is more descriptive of the processes at play; however, the familiar wake zone is retained in this paper for consistency with earlier work.) The idealised streamlines indicate the speed-up of wind flowing over the top, and deceleration in the lee, of the windbreak. The vertical profiles of mean wind speed show the characteristic inflection, and associated large wind shear (change in wind speed with height, dU/dZ), immediately downwind of the windbreak and at a height coincident with the top of the windbreak. The reviews by Heisler and deWalle (1988) and Cleugh (1998) discuss this airflow pattern in much greater detail.

The mixing layer shown in Figure 1 is believed to closely match a classic laboratory mixing layer (Raupach *et al.* 1996; Judd *et al.* 1996), which develops when 2 fluids with different velocities are allowed to merge. This model, if valid, is potentially very useful, e.g. it can predict the position of the windbreak mixing layer, and thus the locations of the quiet zone and wake zones. The model also reveals the main factors that influence windbreak flows, e.g. windbreak height will be important because this is where the mixing layer is initiated. Furthermore, Judd *et al.* (1996) hypothesised that mixing-layer growth is increased by turbulence in the flow upwind of a windbreak and so the location and downwind extent of the quiet zone also depend on the nature of the



Figure 1. Airflow regimes for winds blowing normal to a porous windbreak (after Cleugh 1998; Judd et al. 1996).

surface and airflow upwind. The measurements described in this paper can be used to assess the validity of this turbulent mixing-layer model.

Of greater importance to the question of how shelter modifies plant growth is the interaction of this airflow with entities such as heat and water vapour (collectively referred to as scalars), which are emitted from the plant canopy and soil surfaces that extend up and downwind of the porous windbreak. (These are referred to as scalars because they are zero-order tensors, not vectors. The scalars of interest here are water vapour, heat and CO₂ and are also assumed to be passive, i.e. they do not modify the flow.) For example, by day the transport of water vapour away from a plant canopy is the evaporation flux, while the transport of CO_2 to a plant canopy is the photosynthesis rate. There have been very few studies of this interaction, yet it is the transport of these scalars, either in the mean flow or through turbulent transport that determines their concentration and thus the temperature and humidity in the near-surface air. McNaughton (1988) hypothesised that the effect of the airflow regime shown in Figure 1 might be to change scalar concentrations in the quiet and wake zones because of differences in the strength of turbulent mixing. For example, for an actively growing and transpiring crop whose surface is warmer than the air, we might expect to see the air temperature and humidity to be enhanced in the quiet zone and reduced in the wake zone. While field measurements of near surface air temperatures and humidity (e.g. Argete and Wilson 1989; studies reviewed by Cleugh 1998; McNaughton 1988) have provided empirical proof of this hypothesis, the wind tunnel results presented in this paper enable a more thorough analysis of the effect of windbreak flows on turbulent scalar fluxes and mean concentration fields.

Overview of wind-tunnel experimental methods

CSIRO's Pye Laboratory boundary-layer wind tunnel has been used for all the wind tunnel experiments described in this paper. A wind tunnel provides an ideal controlled environment in which to conduct a detailed exploration of windbreak effects, but if the results are to be applicable to real world situations then the following aspects of the wind-tunnel setup must mimic field conditions as closely as possible: (i) the model canopy must be similar to a real crop or pasture, especially its height and aerodynamic roughness; (ii) the model windbreak needs to be a close replica of a tree windbreak in terms of its drag on the airflow and its height relative to the model canopy and depth of the wind-tunnel boundary layer; and (iii) the boundary layer created in the tunnel must be similar to the atmospheric boundary layer. The Pye laboratory wind tunnel simulates a neutrally stratified boundary layer, i.e. density effects are not included. This means the results are applicable only to field conditions where the wind speeds are sufficiently strong that shear driven turbulence strongly dominates buoyancy driven turbulence.

The following description explains the main elements of the windbreak experimental setup which was designed to simulate a porous windbreak sited in a field with a plant canopy that extends up and downwind.

The working section of the wind tunnel is $17 \text{ m} \log 0.65 \text{ m}$ tall and 1.78 m wide. The floor of this working section was lined with an

aerodynamically rough surface (see below) to both mimic a plant canopy and ensure the development of a deep and fully turbulent momentum boundary layer that adequately represents the real atmospheric boundary layer. Model windbreaks were constructed from brass mesh and placed sufficiently far downwind in the tunnel working section to be within this turbulent boundary layer. For most of the experiments, and unless otherwise stated, the model windbreaks spanned the full width of the wind-tunnel working section, in the field situation this would represent a very long windbreak where effects of the windbreak ends are negligible. Three windbreaks of different porosities (β) were created using different density meshes. [The general symbol for porosity is $\beta;\,\beta_O$ for optical porosity and β_A for porosity determined using wind speeds measured upwind and in the bleed flow; β is used throughout this paper to refer to the porosity of the model windbreaks used in the wind tunnel and should not be confused with the same symbol used in Meinke et al. (2002) which is a calibration factor.]

These porosities span the range that is typical of most field windbreaks, especially those used in the National Windbreaks Program: $\beta = 0.30$ (low porosity); $\beta = 0.43$ (medium porosity); $\beta = 0.69$ (high porosity).

The results presented below use a spatial coordinate system where X denotes horizontal distance, expressed in multiples of windbreak height (H), along a transect oriented at right angles to the windbreak. Positive and negative values of X indicate distances downwind and upwind of the windbreaks, respectively. Z denotes height above either the ground surface, or the floor of the tunnel, and Y denotes lateral distance, oriented normal to X.

The surface used in the multiple windbreak experiments (see below and Judd *et al.* 1996) simulated a tall, flexible plant canopy whose height (h_c) was 47 mm, with a momentum roughness length (z_{om}) of 3.8 mm and a zero plane displacement (d) of 35 mm. The model windbreaks were 150 mm tall, so the ratio of the plant canopy and roughness length to windbreak height (H) was 0.3 and 0.025, respectively. This experiment mimicked multiple windbreaks sheltering a kiwifruit crop, however it also represents any tall, flexible, aerodynamically rough plant canopy whose height is about one-third that of the windbreak height. The topography experiment used a different surface roughness, as described later in this paper. Hot wire anemometers were used to measure the mean and turbulent velocity field in all these experiments.

The experimental configuration used to investigate airflow and scalar fields around single porous windbreaks differed again. For these experiments (Fig. 2), the surface was a model plant canopy created using 3 mm tall pegs, spaced 25 mm apart and arranged in a diamond-shaped grid. This surface simulated a rigid canopy with z_{om} of 0.12 mm and d of 1.2 mm. The model windbreaks were 40 mm tall, so the ratio of the plant canopy to windbreak height (0.075) represents a medium height windbreak (6-10 m) sited in a pasture paddock. These experiments aimed to examine the effect of a windbreak on the microclimate, i.e. temperature and humidity as well as wind speed. This meant creating a ground-based scalar source that simulates a crop or pasture that, by day, is warmed and transpires as it absorbs solar radiation. Heat was chosen as this scalar source because air and surface temperatures are relatively easy to measure accurately. It is important to realise that these air and surface temperatures and the measured heat fluxes reveal the influence of the windbreak flow on the fluxes of heat and water vapour (by day) as well as the near-surface temperature and humidity fields.

A spatially uniform heat source was created by etching copper circuit board to provide the appropriate resistance and then supplying power to the circuit board. A series of these heated panels (1.2 by 0.8 m) were constructed and placed beneath the pegboard roughness for the last 4.8 m of the tunnel working section to provide a ground level heat source at the base of the model plant canopy. The power supply to the

heated panels was controlled to maintain a constant heat flux density of 135 W/m^2 . The model windbreaks were placed 3 m downwind from the start of the heated mats (Fig. 2), and the tunnel working section was sufficiently long to allow measurements to more than 30 H downwind of the windbreak.

Surveys of the mean and turbulent wind and temperature fields, using a combination of hot wire and laser doppler anemometry and fine wire temperature sensors, were conducted around the model windbreak. For the most part, measurements were taken along a single transect oriented normal to the windbreak, positioned near the centre of the windbreak's length (Y = 150 mm or 3.75 H) starting at the most downwind station (X = 40 H) and moving upwind towards the model windbreak. A vertical profile at 28 heights, from Z = 0.23 H (9 mm) to Z = 7.5 H (300 mm) was measured at each station. As a full survey for any particular windbreak configuration took up to 2 days to complete, it was necessary to correct all data for variations in atmospheric pressure (which modifies the wind speed in the tunnel), air temperature and power input to the heated panels.

Surface temperatures were measured in 2 ways. First, an Agema infrared video camera (Model 870) and data acquisition system was used to survey the surface radiative temperatures. The camera was mounted on an automated traverse system and surveys conducted for each of the windbreak porosities and varying angles of approach flow. Calibrations enabled surface temperatures to be measured to an accuracy of $\pm 0.1^{\circ}$ C. Each Agema image was about 5 H by 5 H (0.2 by 0.2 m), with a resolution of about 0.025 H (1 mm). Second, a string of 32 thermistors (diameter 1 mm) were arrayed diagonally along the heated tunnel floor to provide a continuous time series of the floor surface temperatures.

Shelter from wind

In this section, the mean wind speed field is characterised for flow-oriented normal to long, porous windbreaks. Because a great deal is already known about airflow, the focus is on quantifying the main factors that control the reduction in wind speed and the downwind extent of this sheltered zone, viz. the windbreak porosity and the effect of upwind terrain. These results do not include the effects of buoyancy.

Wind flowing normal to a single, porous windbreak

Figures 3 and 4 summarise the main airflow features from the measurements, using hot-wire anemometers, around a medium porosity windbreak ($\beta = 0.43$) oriented normal to the flow. Note that these data are presented in Cartesian, not streamline, coordinates. [Streamlines show the instantaneous velocity field of a fluid, as tangents to streamlines are parallel to the instantaneous velocity at that point. Mass is also conserved beneath a streamline, hence zones of streamline convergence and divergence denote places where the flow is accelerating and decelerating (e.g. Fig. 1). Because streamlines are tilted in windbreak flows, all velocity components should be rotated to be parallel (u and v) and normal (w) to the streamlines rather than to the surface. This introduces new terms into the momentum and scalar conservation equations. These terms will not be large beyond 3 H, where the streamline tilt is less than 2°.]

The contour plot (Fig. 3*a*) of the mean wind speed, for an X-Z 'slice' through the flow, illustrates the expected sheltered region downwind of the windbreak at heights less than the windbreak height; accelerating flow over the top of the windbreak and the slowly readjusting mean velocity field which, even by 30 H, is still perturbed. The maximum shelter near the surface occurs at around X = 6 H, but the near-surface sheltered zone extends as far downwind as X = 30 H.

Superimposing the vertical profiles of mean wind speed, each measured at a different location up and downwind of the windbreak (Fig. 4), reveals the expected inflection in the profile which is most marked immediately in the lee (X = 0.5 H) and becomes smeared in the profiles measured further downwind. This inflected velocity profile resembles that observed in a classic mixing layer. The wind shear (dU/dZ) increases in magnitude, and the inflection becomes



Figure 2. Cross-section of the experimental wind tunnel working section, showing the configuration used in wind tunnel experiments investigating scalar transport and airflow around single windbreaks. The dashed line represents the growing momentum boundary layer.

more obvious, with decreasing porosity as illustrated in the sequence for each of the windbreak porosities (Fig. 5).

 U_{\min} occurs at about 6 H. This position of U_{\min} moves back towards the windbreak as distance from the surface increases towards the top of the windbreak. This location (defined as X_{\min}) does not vary with porosity, but will vary as the nature of the upwind terrain changes, as discussed below.

Figures 3 and 4 also show that the downwind position of the minimum wind speed (U_{min}) varies with height. Near the surface, where shelter for animals and plants is required,



Figure 3. Contour plot (X-Z slice) showing contours of (a) mean wind speed (U); (b) turbulent momentum (u'w') fluxes; (c) air temperatures and (d) heat (w'T') fluxes for the medium porosity windbreak in the wind tunnel. The vertical dashed line indicates the location of the model windbreak.

3.0 H -20 H 1 H 2.5 H 3 H 6 H 15 H 2.0 H 30 H H/Z 1.5 H 1.0 H 0.5 H 0 H 0.2 0.6 0 0.4 0.8 1.0 $U/U_{\rm o}$

Figure 4. Mean wind-speed profiles in the quiet, wake and re-equilibration zones downwind of a medium porosity windbreak. These have been normalised to the wind speed at Z = 8 H.

Figure 5 illustrates the downstream variation of shelter for each windbreak porosity, quantified using the 'shelter factor' (f_s) developed by Judd *et al.* (1996):

$$f_{\rm S} = \frac{U(X_0, Z) - U(X, Z)}{U(X_0, Z)}$$
(1)

where U is the mean horizontal wind speed, X and Z refer to the location of the measurement. The subscript 'o' indicates measurements at a location unaffected by the windbreak, typically at X = -20 H. Note that f_S moves closer to unity with increasing shelter.

Figure 5 confirms expectations that windbreak porosity determines the size of the wind speed reduction, referred to as the amount of wind shelter, for all distances downwind of the windbreak. In contrast to some of the early literature, especially Naegeli (1946), Figure 6 illustrates that for all measurement heights (Z) less than 0.5 H there is no reduction in the size of the sheltered zone with decreasing porosity. Windbreak height is thus the main factor that determines the downwind extent of the sheltered zone.

These results mean that the relative minimum wind speed (U_{\min}/U_{o}) and the wind speed in the bleed flow region $(\overline{U}_{b} = \text{wind speed averaged over all heights up to windbreak})$



Figure 5. Variation of the vertical profiles of shelter (see text for definitions) at varying distances upwind and downwind of the low-, medium- and high-porosity windbreaks: top panel, $\beta = 0.7$; middle panel, $\beta = 0.4$; lower panel, $\beta = 0.3$.



Figure 6. Mean wind speed reduction, plotted as U/U_o (= $U(X, Z)/U(X_o, Z)$), for Z = 0.25 H, 0.5 H, 1 H and three windbreak porosities.

height, at X = 0.5 H) can be related to the windbreak porosity (β), defined as the aerodynamic porosity β_A . Table 1 presents these relationships derived from the wind tunnel experiments described here and 1 from Heisler and de Walle (1988).

Effects of upwind terrain — land cover

An important result from our work is the effect that the surrounding terrain has on the windbreak flow, especially the size of the quiet zone and the location where the minimum wind speed occurs. The effects of topography are not addressed here, although some of the interactions between topography and windbreak flows are briefly described in a later section. The multiple windbreak experiments described in Judd et al. (1996) revealed the influence of the upwind land cover on shelter where the increased upwind roughness created by a multiple windbreak array increased the turbulence in the approach flow. They found that the non-local shelter was greater for repeated windbreaks compared to an isolated windbreak because the windbreaks sited upwind progressively reduced the approaching wind speed and increased the ambient turbulence. However, the local shelter, defined as the ratio of wind speeds measured downwind and upwind of a windbreak, in the lee of the most downwind windbreak was reduced because of 2 factors. First, the greater turbulence in the approach flow enhanced the growth of the turbulent mixing layer, which restricts the size of the quiet zone (Fig. 7 and discussion, below). Second, the reference wind speed immediately upwind is reduced, because of the non-local shelter effect, and thus the local shelter is reduced. The example given below also shows the effect of the surrounding vegetation on the growth of the mixing layer and thus the location of the sheltered zone. It shows that windbreak porosity and height are the sole factors determining wind speed reduction and the size of the sheltered region for a particular type of land cover and terrain.

The mixing-layer analogy

Previous research has shown that the mixing-layer model is an excellent analogy for the turbulent wake layer generated at the top of the windbreak and pictured in Figure 1. This means that a classic mixing-layer growth equation can be modified to predict the growth rate of the 'windbreak' mixing layer that exists in a turbulent boundary layer. This then enables the location and downwind extent of the quiet zone to be predicted. The model for predicting the growth of the mixing layer includes 2 terms: the first is the growth rate for a classic mixing layer, which varies with the size of the

Table 1. Expressions relating the aerodynamic porosity of a windbreak (β_A) to minimum wind speed (U_{\min}) and wind speed in the bleed flow (\overline{U}_b)

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Parameter	U_{\min}/U_{o}	Ub
Equation and origin	(a) $\frac{U_{\min}}{U_{o}} = 1.14 \beta_{A} - 0.16$ (derived from this experiment) (b) $\frac{U_{\min}}{U_{o}} = \beta_{A}$ (Heister and de Walle 1988) where U_{o} is the mean wind speed measured upwind, at the same height as U_{\min}	(a) $\frac{\overline{U_b}}{U(X_o, H)} = \beta_A$ (derived from this experiment) where $U(X_o, H)$ is the mean wind speed measured upwind, at windbreak height



Figure 7. Effect of canopy height on X_{TML} , the location where the windbreak mixing layer intersects the surface, and X_{\min} , the location of the minimum wind speed (U_{\min}). Circle symbols are measured data from Judd *et al.* (1996).

inflection in the profile at X = 1 H (Fig. 4) and thus windbreak porosity. The second term accounts for the influence of the turbulence in the flow approaching the windbreak. It is larger if the land use upwind is aerodynamically rough, for example if the land cover comprises scattered trees or forest.

Together, these 2 terms illustrate the factors that determine the size of the sheltered zone and the pattern of wind shelter. This is because the downwind limit of the quiet zone occurs where the growing turbulent mixing layer intersects the surface and X_{min} , the position of the minimum wind speed, is within this quiet zone. Factors that influence the growth of the mixing layer will therefore affect the downwind extent of the quiet zone and X_{min} . The wind tunnel results just presented clearly show that porosity determines the size of U_{min} and thus the amount of shelter. The upwind roughness and windbreak height are the main influences on the location of the quiet and wake zones and the extent of the sheltered zone downwind, for flow normal to the windbreak. This suggests that simple models can be developed to predict U_{min} , based on β , and X_{min}

A simple example is provided here to illustrate how this equation for mixing layer growth can be used to investigate the effect of changing upwind roughness, in this case the change in crop height over a growing season, on the quiet zone dimensions and location of U_{\min} . Figure 7 plots X_{TML} , the downwind distance where the turbulent mixing layer intersects the canopy, for a range of crop heights. The downwind extent of the quiet zone near the surface is equivalent to X_{TML} . Also plotted is X_{\min} , the location of the near-surface wind speed minimum, which is assumed to be 0.6 of the downwind extent of the quiet zone based on data from our wind tunnel experiment and that of Judd *et al.* (1996). The assumptions that underlie Figure 7 mean that it is valid for a medium porosity windbreak and neutral atmospheric stability only, and is intended for illustrative

purposes. Nonetheless, it reveals that the quiet zone will vary over the growing season as the upstream surface changes from bare soil to crop. For a windbreak site where the upwind fetch is a growing wheat crop, X_{min} will vary from about 8 H early, to 5 H later, in the season.

The implications of these results are that the upwind roughness, and thus the approach flow turbulence, has a strong influence on the degree of shelter experienced downwind of a windbreak. In terrain that is very rough in an aerodynamic sense (e.g. a treed landscape), the local shelter may differ from that for an identical windbreak placed in terrain that is much less rough (e.g. grazed pasture). Evidence of such a role was seen clearly in the results presented above from the multiple windbreak experiment.

Effects of shelter on microclimates

The conceptual view of windbreak flow pictured in Figure 1, with a spreading turbulent mixing layer lying above a sheltered quiet zone where turbulent exchanges are damped, leads to the hypothesis that mean temperature and humidity may be enhanced in the quiet zone and possibly reduced in the wake zone (Fig. 8). While much is known about the aerodynamic effects of windbreaks, the same is not true for the effects of windbreaks on the transport of scalars such as water vapour, heat and CO_2 . Understanding the impact of shelter on plant water use and microclimates requires knowledge of how windbreak flows modify the fluxes of these scalars, especially heat and water vapour, and their resulting mean concentrations.

One of the main objectives of the wind tunnel and field experimental program was to quantify the effects of windbreaks on the near surface air temperature and humidity, and thus the microclimate experienced by plants growing in the shelter of windbreaks. Heat was used as a passive scalar in the wind tunnel experiments as described earlier. These experiments thus measured and characterised the air and surface temperature fields, and the heat fluxes that result from the interaction of the airflow around the windbreak with



Figure 8. Hypothetical spatial variation of the concentration of a scalar, *C* emitted into a windbreak flow from a uniform ground source (see text for definitions). The mean scalar concentration is plotted as a difference from upwind (ΔC). The vertical dashed line is a windbreak.

a spatially uniform heat source. This experimental arrangement and the results are directly applicable to a field situation where the atmosphere is neutrally stratified as illustrated by the excellent results of an intercomparison with field data described below and in Cleugh (2002a). In the field, the air temperature and humidity result from the interaction between the windbreak airflow and the fluxes of heat and water vapour from the soil and plant canopy. Thus, the scalar source, by day, is the sum of the latent and sensible heat fluxes $(LE + H_A)$ that balances the available energy at the underlying soil and plant surface. [LE, the latent heat flux, is the energy equivalent of the evaporation flux. A, the available energy, is the difference between the net all-wave radiation (R_n) absorbed by the surface, and the heat conducted into the soil (G). R_n , in turn, is the sum of the net shortwave (incoming - reflected) and longwave (incoming emitted) radiation terms and represents the radiant energy available at the surface.] The air temperature and humidity measured in the field must be combined into an 'equivalent' temperature before they can be compared with the wind tunnel air temperatures. The important results from these wind tunnel and field experiments are presented to illustrate the way that a windbreak modifies the surface and near-surface microclimate.

Wind tunnel results: mean temperature fields

Figure 3a and 3c are contour plots of the complete wind speed and temperature field, measured around the medium porosity windbreak in the wind tunnel. The wind speed contour plot clearly shows the sheltered zone; the accelerating flow over the top of the windbreak; and the location of the minimum wind speed near the surface at about 6 H. The spatial pattern in air temperature (Fig. 3b) matches this wind speed plot in that air temperatures are enhanced in the quiet zone and the location of the maximum increase in temperature coincides with the location of the minimum wind speed. The displacement of streamlines as the air flows through and over the windbreak (recall Fig. 1) is the main cause of the rise in air temperature in the quiet zone. The reduction in heat transport by turbulence and the mean wind also contributes to this rise. There are also differences in the spatial pattern of the temperature and wind speed fields, e.g. the peak in temperature at all heights below windbreak height is located at about 6 H, whereas X_{\min} moves back towards the windbreak with increasing height above the surface. Similarly, the perturbation to the air temperature field is noticeably less than for the mean airflow, both in terms of the downwind extent of the temperature changes and the magnitude of these changes in response to the reduced wind speed.

Figure 9 presents a more quantitative picture of the effects of shelter on temperature and wind speed. It shows the spatial trend of the temperature and wind speed, plotted as normalised (see below) differences from upwind, at a height near the surface (Z = 0.3 H) for each windbreak porosity used in the wind tunnel experiments and for the 3-row pine windbreak used in the field measurements. These changes in wind speed and temperature have been normalised by the friction velocity, u_{*0} and its analogous temperature scale, T_{*0} , to account for differences in the heat input, upwind wind speed and surface roughness between the wind tunnel and the field. The friction velocity is an indicator of the surface roughness and thus the level of turbulence in the atmosphere. It is defined as:

$$u_{*_0} = \sqrt{u'w'} \tag{2}$$

where u' and w' are the turbulent fluctuations of the horizontal and vertical components of the velocity. The temperature scale is defined as:

$$T_{*0} = w'T' / u_{*0} \tag{3}$$

where $\overline{w'T'}$ is the turbulent sensible heat flux (= H_A/ ρC_p and ρ and C_p are the density and specific heat of air, respectively).

The resulting differences in wind speed (δU) and temperature (δT), which can now be compared between wind-tunnel and field measurements, are calculated as:

$$\delta U = \left[\frac{U(X, Z_{\rm S}) - U(X_{\rm o}, Z_{\rm S})}{u_{*_{\rm o}}}\right] \tag{4}$$

$$\delta T = \left[\frac{T(X, Z_{\rm S}) - T(X_{\rm o}, Z_{\rm S})}{T_{*_{\rm o}}}\right] \tag{5}$$

where X and Z refer to the vertical and horizontal measurement locations; Z_S refers to measurements near the surface (0.3 H in both the field and wind tunnel) and the subscript 'o' refers to measurements at a reference station far upwind (>10 H) of the windbreak.

The key features in Figures 2 and 9a are the following: (i) The spatial trend in the near-surface air temperature mirrors that for the near-surface wind speed. The air temperature reaches a peak at the same location as the wind speed minimum, so the maximum velocity deficit and air temperature increase occurs at about 6 H (Figs 2 and 9a). (ii) Discrepancies between δU and δT begin to emerge at distances downwind of 12 H. By X = 15 H, δT is close to zero, i.e. the temperature has returned to its upwind value, but wind speeds are still reduced by 30–40% ($\delta U \approx -4$). While wind shelter extends as far as 30 H downwind, this does not apply to the increase in near surface air temperature. (iii) The near-surface air temperatures in the wake zone (X>15 H) are decreased slightly below those in the approach flow, as hypothesised by McNaughton (1988); however, the magnitude is small and extends over a larger downwind distance than predicted. The surface temperatures (Fig. 10) do not follow this trend, rather, they return to upwind values between X = 12 and 15 H. (iv) The size of the temperature gain increases with increasing shelter, i.e. δT increases with



Figure 9. Spatial variation of normalised wind speed and temperature differences (δU , δT , δT_e ; see text for definitions), measured near the surface: (*a*) wind tunnel measurements for windbreaks of 3 porosities, upper curve is δT and lower curve is δU ; (*b*) from the field experiment described in Cleugh (2002*a*), upper curve is δT_e and lower curve is δU .



Figure 10. Spatial variation of surface temperature (plotted as differences from upwind temperatures) measured using the Agema infrared camera, for windbreaks of 3 porosities ($\beta = 0.3, 0.4, \text{ or } 0.7$).

decreasing porosity. The peak near-surface air temperature rise for the medium and low porosity windbreaks is about 0.7° and 1.4° , respectively. There is only a very slight temperature increase behind the high porosity windbreak (0.14° C).

The variation of δT with height does not deviate from this pattern, i.e. the maximum temperature increase occurs at 6 H at all heights. For the medium and low porosity windbreaks, this temperature rise in the quiet zone has diminished by Z = H.

Figure 10 is the matching plot of the surface temperatures, for the 3 windbreak porosities, which were measured using an Agema infrared camera and are plotted as differences from the upwind values. The peak increase in surface temperature is about 2.5, 1.8 and 0.6° C for the low, medium and high porosity windbreaks, which are linearly related to the peak increases in near-surface air temperatures (1.4, 0.7 and 0.14° for each porosity). The surface temperature reaches its peak at about X = 4.7 H, which is upwind of the air temperature maxima and wind

velocity minimum at 6 H. This location of the peak surface temperature appears to shift closer to the windbreak with decreasing porosity. The return to upwind surface temperatures occurs, like air temperature, at about X = 12 H.

Figure 11 provides some insight into the mechanisms leading to the observed antisymmetric spatial patterns of wind speed and temperature. The positions of the temperature maxima and wind speed minima coincide not only with each other, but also with the zone of maximum deceleration and hence streamline displacement (see Fig. 1) indicated by the location where the mean vertical velocity (W) passes through zero. The mean horizontal (U) and vertical velocities determine the advective heat transport, so it is not surprising that an increase in temperature occurs where these are at a minimum.

The turbulent fluxes u(u'w', w'T' and u'T'), included in Figure 11 and discussed in greater detail later, reach their respective peaks further downwind of this location, at X of about 8–10 H, marking the downwind extent of the quiet



Figure 11. Spatial variation of horizontal (*U*) and vertical (*W*) mean wind speeds; air temperature (plotted as the difference from upwind) and turbulent momentum and heat fluxes $(\overline{u'w'}, \overline{w'T'})$. Data are from wind tunnel measurements near the surface around a medium-porosity windbreak.

zone. These peaks in turbulence contribute to the re-establishment of the upwind temperatures and wind speeds. Moreover, the very large magnitude of the peak in the turbulent scalar terms (σ_T^2 , $\overline{u'T'}$ and $\overline{w'T'}$), compared to the turbulent velocity terms, explains the more rapid decrease in the near surface temperatures downwind of 10 H. The increase in air and surface temperatures in the region extending from about X = 0.5–8 H thus results from the combined reduction in advective and turbulent transport.

The similarity between the temperature and wind speed patterns in the quiet zone suggests that wind speed could be used as a predictor of near-surface air temperatures. This is confirmed by the excellent relationship between $[\log T]$ and $[\log U]$ shown in Figure 12. The line of best fit for the data in Figure 12 should not, however, be used directly to predict near surface air temperature changes in the field from wind speed measurements for 2 reasons. First, the temperatures plotted are differences from the ambient temperature measured in the wind tunnel. Second, a universal relationship needs to be normalised by u_{*0} and T_{*0} , which were about 0.55 and 0.18 in the wind tunnel experiments.

Field results: mean temperature and humidity fields

Figure 9b is a plot of the normalised equivalent temperature and wind speed differences behind a 3-row pine windbreak in the field. These data were acquired during a 6-week field campaign conducted in spring 1996 as described in Cleugh (2002a). Turbulent fluxes of water, heat and momentum, plus the available energy, air temperature and humidity, were measured simultaneously at stations up and downwind of the windbreak. As noted above, the temperature and humidity data were combined into an equivalent temperature, following the method of



Figure 12. Near-surface air temperature change *v*. wind speed. Data are from wind tunnel measurements around all 3 windbreak porosities and are plotted as $\log_{10} (T) v$. $\log_{10} (U)$. The line of best fit is:

$$\log_1 T(X, Z_{\rm S}) = -0.881 \cdot \log_{10} U(X, Z_{\rm S}) + 0.33617$$

(r² = 0.96; Z_{\rm S} \le 0.5 \text{ H}, 1 \text{ H} \le X \le 30 \text{ H}).

McNaughton (1988), to enable field and wind tunnel data to be compared. The equivalent temperature (T_e) is given by $[T_a + q_a C_p/L_v]$, where T_a and q_a are the air temperature and humidity; L_v is the latent heat of vaporisation and C_p is the specific heat of air. The uppermost plot in Figure 9b thus shows δT_e , defined as:

$$\delta T_{\rm e} = \left[\frac{T_{\rm e}(X, Z_{\rm S}) - T_{\rm e}(X_{\rm o}, Z_{\rm S})}{T_{\rm e^*o}} \right] \tag{6}$$

where the terms are as defined above and the normalising velocity scale, T_{e^*o} , is:

$$T_{e*o} = \frac{\overline{w'T'} + \overline{w'q'}}{u_{*o}} = \frac{H_{A}/\rho C_{p} + LE/\rho L_{v}}{u_{*o}}$$
(7)

The field data included in Figure 9b have been filtered to include only those data where the approach wind direction was within 15° of normal to the windbreak and wind speeds exceeded 4 m/s. The data have also been diurnally averaged, from 0900 hours to 1500 hours. Cleugh (2002a) notes that this is justified by the lack of diurnal variability in the filtered, normalised data. Figure 9b has 2 obvious features. The first is the similarity in the spatial trend of δU and δT_{e} measured in the field, and δU and δT measured in the wind tunnel. The magnitude of δU and δT_{e} is similar to δU and δT measured around the low porosity windbreak used in the wind tunnel and the position of the peaks are almost identical. The second feature is the difference between the wind tunnel and field scalar concentrations in the bleed flow region (-3 H-3 H). Part of the reason for this is that the field windbreak intercepts solar radiation, which is shed via latent and sensible heat fluxes. It thus acts as a scalar source, whereas the model windbreak in the wind tunnel did not emit heat. However, an order of magnitude estimate of this scalar flux from the windbreak suggests that δT_{e} would be about 1°C as a result of transpiration and heating from the windbreak, which is much smaller than the value interpolated between the data measured at +1 H and -3 H in Figure 9b.

The field data also show the effect of shelter on the saturation deficit and thus any direct plant physiological response to atmospheric demand. [The saturation deficit is the difference between the saturation specific humidity (the amount of water vapour contained as vapour in a parcel of air at a given temperature) and the actual specific humidity of the air. It is thus a measure of the atmospheric dryness and atmospheric demand.] McNaughton (1988) argued that the air within the sheltered quiet zone would be effectively decoupled from the regional, or overhead, air approaching the windbreak. Whether shelter reduces or enhances the saturation deficit in the quiet zone depends on the humidity of this regional air. If it has a large saturation deficit, and so is very dry, then the saturation deficit of the sheltered air in the quiet zone may be reduced. Conversely, if the regional air

has a low saturation deficit, i.e. is moist, then the saturation deficit in the quiet zone may be enhanced and atmospheric demand increased.

Evidence of this effect of shelter on atmospheric demand was found in the field measurements (Cleugh 2002*a*). During the morning, when the upwind atmospheric demand (D_0) was small, the atmospheric demand (D) in the quiet zone (from 0 to 6 H) was increased slightly. In the afternoon, when D_0 was large, D in the quiet zone tended to be decreased. In situations where the upwind air is relatively dry, the sheltered quiet zone may thus be protected from these high saturation deficits and atmospheric demand thus reduced. Conversely, if the regional air is humid, wind shelter can increase the saturation deficit in the sheltered quiet zone.

Summary

Both near-surface (all heights less than 2 H) and surface temperatures are increased in the sheltered quiet zone in the lee of a porous windbreak. The magnitude of the temperature increase varies with the windbreak porosity as expected given that streamline displacement and a reduction in advective fluxes are 2 of the main mechanisms leading to this increase.

One of the key results to emerge from both the field and wind tunnel experiments is that wind speed reductions extend much further downwind than the temperature (and humidity) changes that result from the perturbed airflow. While temperature and humidity are increased in the quiet zone, they rapidly return to their upwind values from X = 10 H to X = 12 H. Moreover, the magnitude of the increase is small for the low and medium porosity windbreaks and almost negligible for the highly porous windbreak. In contrast, near-surface wind speeds are reduced over an area extending at least 30 H downwind for all windbreak porosities, including the high porosity windbreak. This means that any direct link between plant growth and wind, e.g. through mechanical damage from leaf abrasion or sandblasting, will occur over much larger distances and thus potentially be of greater economic importance than indirect links through changes in water use or microclimates.

Turbulent fluxes of heat and water

Figure 3 includes contour plots, in the X–Z plane, of 2 of the turbulent transport terms, the momentum flux $(\overline{u'w'})$ (Fig. 3b) and the vertical heat flux $(\overline{w'T'})$ (Fig. 3d) measured in the wind tunnel for the medium porosity windbreak. As discussed above, the measurements of $\overline{w'T'}$ in the wind tunnel are analogous to the turbulent water vapour flux $(\overline{w'q'})$ which, when converted to energy or mass, is the latent heat or evaporation flux. Detailed analyses of these data have confirmed their validity and they also compare well with the field measurements presented in Cleugh (2002*a*). It is important to note that these data have not been analysed, nor presented, in streamline coordinates (see earlier note). Nonetheless, the plots do show, in a qualitative sense, several important points about the turbulent momentum, heat and water vapour fluxes in windbreak flows:

(i) The near surface turbulent scalar $(\overline{u'w'} \text{ and } \overline{w'T'})$ and velocity terms $(\overline{u'T'})$ are reduced dramatically in the quiet zone (this is also seen clearly in Fig. 11). Presenting the data in streamline coordinates would not alter this picture drastically, especially for the scalar terms.

(ii) The spreading turbulent mixing layer that was pictured schematically in Figure 1 is evident in the contour plot of $\overline{u'w'}$. Its signature is the layer of increased turbulence (i.e. increased $\overline{u'w'}$) that is initiated at the top of the windbreak and grows deeper with increasing distance downwind of the windbreak. The contour plot also shows that $\overline{u'w'}$ is not increased when this mixing layer intersects the canopy at about 8 H.

(iii) The spatial trend in the turbulent heat flux is different. The signature of the turbulent mixing layer, so clear in the turbulent momentum flux, is absent except for a zone of enhanced $\overline{w'T'}$ that extends downwind from X=8 H – 12 H. For Z<0.5 H, this term is almost twice its value upwind and is the result of the interaction between the growing turbulent mixing layer with the ground-based scalar source. This difference is not unexpected, as the windbreak obviously provides a large perturbation to the velocity field whereas the scalar (i.e. temperature) field, including the mean and turbulent fluxes, is only modified via changes in the airflow; the windbreak itself has no influence. Nonetheless, these data are among the first to show this difference so clearly.

(iv) The turbulent windbreak flow obeys classic mixing layer scaling, for both velocities and scalars, albeit a mixing layer that is embedded in a turbulent boundary layer. This is an important and significant result as it enables us to better quantify windbreak flows and their impact on scalar transport. For example, it means that expressions for predicting the growth of a windbreak mixing layer can be developed from classic mixing layer theory, as discussed before. Estimates of the turbulent diffusivities can also be derived from mixing layer theory (Cleugh 2002*b*).

Figures 9 and 11 illustrated the different spatial trends in the near-surface air temperature, compared to the mean wind speeds and some of the turbulent fluxes. The large turbulent scalar fluxes at about X = 8-10 H rapidly re-establish the upwind scalar concentrations as seen in the surface and air temperatures in the wind tunnel, and the near-surface humidity and temperatures in the field. The spatial pattern of the turbulent terms and the mean wind speed are also different. In the near-surface layers, the turbulent terms return to their upwind values within 5–10 H of the mixing layer reaching the canopy. This is in contrast to the mean wind speed, which is perturbed throughout the entire profile 30 H downwind of the windbreak. Caution is thus required in extrapolating from the pattern of near-surface wind speeds to other processes and atmospheric terms that affect plant productivity.

Effects of oblique flow on shelter and microclimates

Much of our understanding of windbreak effects on shelter is from studies where the mean flow direction is close to normal to the windbreak. On real farms, of course, the wind will often approach the windbreak at an oblique angle, i.e. not normal, to the windbreak. Note that the incidence angle is defined as α , where $\alpha = 90^{\circ}$ is normal and $\alpha = 0^{\circ}$ is parallel, to the windbreak. The shelter behind a porous windbreak may differ from that described in the previous sections if the mean wind blows at an oblique angle to the windbreak. Some important aspects of oblique flows include:

(i) There are 2 distances to consider: (a) the distance from the windbreak along a transect oriented normal to the windbreak (denoted X_N) and (b) the distance from the windbreak along a transect oriented parallel to the mean wind direction, denoted as X_S to indicate streamwise distance. This simple concept means that the reduction in wind speed typically varies with $[\cos (90 - \alpha) = \sin \alpha]$.

(ii) If the porous windbreak has width (*W*) such that W>0.1 H, the pathlength through the windbreak will be increased in oblique flow. In many cases this might decrease the effective porosity of the windbreak, especially for medium and low porosity windbreaks. Conversely, the porosity might effectively be increased for highly porous windbreaks. The changed 'effective' porosity will change the magnitude of $U_{\rm min}$ and thus the amount, and possibly the downwind extent, of shelter.

(iii) The mean flow direction can also be altered as the air flows through the windbreak as predicted by theory. An appealing explanation is that the flow takes the 'path of least resistance' through the break and thus appears to straighten such that its direction is closer to being normal to the break. Further downwind, the wind changes direction again, flowing more parallel to the break. The magnitude of this effect depends on the windbreak's porosity, width and the incidence angle of the flow.

There are relatively few studies investigating the effect of oblique flows on shelter and none that consider the effects on the microclimate and turbulent fluxes. Heisler and de Walle (1988) presented results from a range of artificial windbreak experiments showing that the relative wind speed reduction [defined as $U(X, Z)/U(X_0, Z)$, which is hereafter shortened to U/U_0 , where U is the mean horizontal wind speed at height Z and distance X from the windbreak and X_0 denotes the upwind or reference position] increases as the wind blows more obliquely towards the windbreak, i.e. as α decreases from 90°. They argue that the open area 'seen' by the wind as the flow becomes more oblique to an artificial screen increases, and that this explains the observation that relative protected distance (X_N/X_S) varies with [cos² (90 - α)] for

artificial windbreaks and $[\cos (90 - \alpha)]$ for tree windbreaks. The situation is somewhat different for tree windbreaks where the increase in pathlength means that U/U_0 is very likely to decrease with increasing obliquity.

Wang and Takle (1996) perform numerical simulations of flow oblique to windbreaks with varying porosities and widths. Despite concerns in the literature (Wilson and Mooney 1997) about the quality of their simulations, it is worth summarising their main results as there are so few studies of this kind: (i) The downwind variation in U/U_0 (Z<0.7 H, W = 0.5 H) when $\alpha = 75$ and 60° is almost indistinguishable from the pattern for normal flow, especially when Z>0.1 H. (ii) Measured along a transect oriented normal to the windbreak, the position of U_{\min} moves back towards the windbreak with increasing obliquity, and the sheltered area contracts. (iii) The magnitude of U_{\min} decreases with increasing obliquity for Z = 0.3 H, 0.5 H and 0.7 H and slightly increases at Z = 0.1 H. This behaviour depends on windbreak width and porosity. These results are for a wide windbreak (W = 0.5 H). For narrow, especially low porosity, windbreaks, increasing obliquity may lead to an increase in U_{\min} . (iv) Changes to the extent of the sheltered region vary with windbreak porosity, height and the wind direction relative to the windbreak. Only minor changes in this sheltered distance were found as α was varied from 60–90°. (v) $[X_N/X_S]$ may decrease more or less than $[\cos (90 - \alpha)]$. Importantly, this relationship varies with measurement height as well as with windbreak porosity and width. For dense, narrow windbreaks the rate is larger than $[\cos (90 - \alpha)]$ and smaller than $[\cos (90 - \alpha)]$ for porous windbreaks.

Given this lack of empirical data, a series of experiments was conducted to investigate the mean and turbulent airflow; surface and air temperatures and turbulent heat fluxes behind windbreaks oriented at varying angles to the mean wind. The first set of experiments used porous windbreaks that extended the full width of the tunnel, and so can be considered to be 2-dimensional (2D) as their length is very long in comparison to their height, oriented at 45 and 67° (medium porosity windbreak only) to the mean flow (i.e. $\alpha = 45$ and 67°). The second set of experiments measured only the surface temperature and mean wind speed around a short (20 H in length), medium-porosity windbreak oriented at 45° and normal ($\alpha = 45$ and 90°) to the mean flow. The object of the first set of measurements, using long windbreaks, was to investigate how obliquity modified the spatial pattern of the mean and turbulent flow, while the second experiment aimed to confirm an optimum windbreak length (see below). For all these experiments, the same surface roughness and scalar source was used as described previously. The wind speeds were measured using laser doppler anemometry (LDA), a method more suited to the 3-dimensional (3D) flows expected around windbreaks oriented obliquely to the flow. The surface and air temperatures were measured using the fine wire and radiative methods described above. The detailed field experiment described in Cleugh (2002a) also enabled an analysis of the interaction between a tree windbreak and oblique flow.

Wind tunnel measurements of oblique flow — mean fields

Long (2D) windbreaks. The spatial pattern in mean horizontal wind speed (U) and air temperature (T_a), for flow at an incidence angle of 45° to a medium porosity windbreak ($\alpha = 0.4$), is almost identical to that for $\alpha = 90^{\circ}$ (Fig. 13), providing U and T_a are plotted v. streamwise distance (X_S). This holds true for all heights up to 1.5 H, with the exception that the lowest transect (at Z = 0.25 H) shows slightly greater shelter from 0–4 H, in agreement with Wang and Takle's (1996) numerical results.

Although not shown here, the same is true for the spatial trend in the turbulent velocity terms and temperature at Z = 0.25 H and 0.5 H. The vertical turbulent flux of horizontal momentum $(\overline{u'w'})$ is also little changed, but the vertical turbulent scalar flux $(\overline{w'T'})$ at Z = 0.25 H and 0.5 H is slightly reduced in the case of oblique flow. It can be concluded that there are no gross changes in the spatial pattern of turbulent fluxes for flow with incidence angles of 45° , once changes in streamwise distance are accounted for.

Figure 14 illustrates the spatial variation of the near-surface ratio of U/U_0 [=(X, 0.3 H)/U(X₀, 0.3 H)] for

 $\alpha = 45$ and 67° ($\alpha = 0.43$), and $\alpha = 45^{\circ}$ for the low-porosity windbreak ($\alpha = 0.30$). These profiles are plotted *v*. *X*_S.

As expected from both Figure 13, and the numerical simulations of Wang and Takle (1996), the only obliquity effects are seen for $\alpha = 45^{\circ}$, there are no significant changes in the spatial pattern of U/U_{0} and amount of shelter when α is closer to normal.

There is a reduction in U/U_0 from X = 1 - 4 H ($\alpha = 45^\circ$) for both the medium- and low-porosity windbreaks at the lowest Z. For these porosities, therefore, the amount of shelter is increased with oblique flow. This effect increases with decreasing porosity and is most obvious near the surface (Z = 0.25 H).

For the low porosity windbreak only, the position of U_{min} moves closer to the windbreak and the magnitude of U_{min} slightly increases when $\alpha = 45^{\circ}$. This is not evident for $\alpha = 67^{\circ}$.

These results show that providing the changed streamwise distance is accounted for, which can be done using geometry, then the downwind pattern of shelter is not grossly changed by oblique flows when $90^{\circ} < \alpha < 45^{\circ}$.

Short (3D) windbreaks. Most texts suggest that windbreaks should be at least 12 H long to ensure that edge effects do not significantly reduce the efficacy of the windbreak by altering the size of the sheltered zone or the reduction in wind speed (e.g. Oke 1987). A sequence of



Figure 13. (a) Mean wind speeds and (b) air temperatures measured at 3 heights (Z = 0.25 H, 0.5 H, 1 H) around the medium-porosity windbreak, for incidence angles of 0° (normal) and 45°. The X-axis is streamwise distance (X_S).

wind-speed and surface-temperature measurements was conducted around a short, medium-porosity windbreak with length (L) of 20 H and $\alpha = 90$ and 45° to evaluate this recommendation. Measurements were made along a transect extending downwind of the windbreak and positioned midway along the windbreak's length (i.e. Y = 0.5L). The object of these measurements was to investigate the effect of a short windbreak on wind shelter and surface temperatures. Of particular interest was whether the level of wind shelter, and associated changes in surface temperature, were significantly modified by the shorter windbreak, and to what extent the limited length reduces the extent of the sheltered zone. The magnitude and spatial variation of wind speed and surface temperature are compared with those determined from equivalent measurements around the very long windbreaks, which have already been presented.

Figure 15a ($\alpha = 90^{\circ}$) and 15b ($\alpha = 45^{\circ}$) superimpose the isotachs (wind speed contours) and wind vectors on the surface temperature contour plots. For winds oriented normal to the windbreak ($\alpha = 90^{\circ}$), flow around the windbreak ends is evident in both the wind and surface-temperature plots, but this only encroaches into the sheltered zone within about 2–3 H of the windbreak ends. The isotachs in the zone extending 5 H either side of the windbreak midpoint (i.e. Y = +200 mm to -200 mm in Fig. 15a) are of a similar magnitude to those for the same porosity, long windbreak. For example, U/U_0 near the surface at X = 10 H is about 0.6 for both the 3D and 2D windbreaks. Downwind of X = 10 H, the upwind wind speeds recover slightly more quickly around the shorter windbreak, as U/U_0 at X = 15 H and 20 H is about 0.8 and 0.9.

The isotach pattern is essentially unchanged when α shifts to 45°, except there is no zone of increased shelter from 0 to 5 H seen in the oblique flow around the 2D

windbreak (Fig. 15*b*). The vectors reveal a very slight change in wind direction, both immediately upwind and downwind of the windbreak. A slight encroachment into the sheltered zone behind the leading edge (i.e. upwind) end of the windbreak is apparent, but the level of shelter from Y = +8 H to Y = -8 H is remarkably similar to that for the flow oriented normal to the windbreak.

Increased temperatures are observed where the quiet zone is expected, between 0 and 5 H, for both orientations. This warmer zone is reduced in downwind extent for the short windbreak and normal flow ($\alpha = 90^\circ$, Fig. 15*a*), but the size of the temperature rise is very similar, compared to the very long windbreak. Increased flow around the windbreak ends erodes this warm zone, especially within 1 H of the ends of the windbreak, leaving a warm core extending 2 H on either side of the midpoint of the windbreak's length.

A similar picture emerges for oblique ($\alpha = 45^{\circ}$, Fig. 15*b*) flow, but the warm core seen in Figure 15*a* appears as an elongated region running parallel to the windbreak. The peak surface temperature is still located midway along the windbreak's length, at $X_{\rm S} \approx 5$ H ($X_{\rm N} \approx 3$ H), and the size of this temperature increase is similar, if a little larger, than for the short windbreak oriented normal to the flow. The effect of obliquity is to reduce the amount of shelter, and thus surface temperature rise, at the leading edge of the windbreak. Flow around the trailing edge appears to truncate the warm zone, which otherwise extends right to the end of the windbreak.

This suggests that windbreaks of at least 20 H in length are sufficiently long that the edge effects do not significantly erode the size of the sheltered zone. It also demonstrates that the changes in surface temperature are quite similar to those found for a very long windbreak where both have flow oriented normal to the windbreak. Discrepancies between the spatial pattern in wind speed and temperature between the short and long windbreak appear within 2–3 H of the



Figure 14. Near-surface wind-speed reduction $(U/U_0 \text{ where } Z = 0.3 \text{ H})$ for windbreaks of 2 porosities (low and medium) oriented at 3 angles to the flow ($\alpha = 0^\circ, 45^\circ, 67^\circ$) and plotted *v*. streamwise distance (X_S).



Figure 15. Effect of a short windbreak (L = 20 H) on mean wind speed reduction (contour lines show $U/U_0 \times 100\%$ at Z = 0.3 H), wind direction (arrows) and surface temperatures (shading) for a medium-porosity windbreak, (a) $\alpha = 90^{\circ}$; (b) $\alpha = 45^{\circ}$.

windbreak ends. The same result was found for flow oriented 45° to the windbreak, when interpreting the results along streamlines. However, for a field windbreak, this means that the sheltered zone windward of the windbreak centreline will be contracted, as illustrated in Figure 15*b*.

Oblique flow around tree windbreaks: mean fields

The applicability of these wind tunnel results to field windbreaks can be explored using the field measurements of oblique flow around a wide ($W \approx H$), 3-row pine windbreak, described in detail in Cleugh (2002*a*). Figure 16 is a plot of the near-surface wind speed reduction (U/U_0 at Z = 0.3 H) v. the incidence angle of the wind to the windbreak, plotted as $[1 - \cos (90 - \alpha)]$, for the 1 H, 3 H, 6 H and 9 H stations.

These distances are normal to the windbreak and the data have been filtered to remove all 15-min runs where U < 4 m/s. If the changing streamwise distance with oblique flow is the sole influence on U/U_0 then a linear relationship should be observed between U/U_0 and $[\cos (90 - \alpha)]$. Seginer (1975), for example, found that $U/U_0 = m [1 - \cos (90 - \alpha)] + U/U_{0N}$, where U/U_{0N} is the value of U/U_0 for flow normal to the windbreak. Figure 16*a* shows that such a relationship also holds for the U/U_0 at 6 H, 9 H and 12 H. Figure 16*b* shows the streamwise variation in U/U_0 for broad wind directions, and thus α , classes plotted against streamwise distance, X_S . Ideally, recalling the discussion around Figure 14, we would expect all data to collapse into a single curve. While Figure 16*b* shows some scatter about the mean curve, for each α



Figure 16. (a) $U(X, 0.3 \text{ H})/U(X_0, 3 \text{ H}) v. \cos (90 - \alpha)$, where α is the incidence angle of the flow, for X = 1 H, 3 H, 6 H and 9 H (from Cleugh 2002*a*); (b) $U(X, 0.3 \text{ H})/U(X_0, 3 H) v. X_S$ (streamwise distance) for 5 classes of wind direction (α). All data have been measured using sonic anemometers, and have been filtered for U > 4 m/s.

class, it is clear that accounting for streamwise distance does account for much of the variation in U/U_0 . Some of the scatter in Figure 16*b* arises from the fact that each station has a differing length of record and the measurements span a range of upwind conditions (Cleugh 2002*a*).

The picture for sites closer to the windbreak (3 H and 1 H stations) is slightly different. U/U_0 at 3 H shows 2 different relationships with [cos (90 – α)]. For α >45°, the trend is linear but as α approaches and decreases below 45°, U/U_0 becomes constant. This means that for α <45°, the change in U/U_0 is no longer determined by changes in streamwise distance to the windbreak. At 1 H, there is no evidence of a [cos (90 – α)] relationship. The most obvious feature at 1 H is the slight increase in U/U_0 with increasing α (from 0.05 at $\alpha = 30^\circ$ to 0.1 for α <70°).

In summary, the maximum shelter will be around the 6 H position in normal flow. U/U_0 will increase (i.e. shelter decrease) linearly with $(90 - \alpha)$ at all sites downwind of, and including, 6 H as a result of the increase in streamwise distance. Closer to the windbreak, its structure will begin to influence the pattern of shelter, which may be greater or lesser than predicted from theory or wind tunnel experiments.

Interaction between topography and windbreak flows

Almost all of our understanding of windbreak flows is for flat locations, but windbreaks are just as likely to be located in undulating terrain and so it is important to consider how airflow and microclimate effects are altered when windbreak and topographic effects are combined. There are 3 key aspects to the interaction of topography with windbreak flows. First, topography modifies the airflow, creating regions where wind speeds are enhanced and regions where they are reduced. In addition, thermal circulations — driven by differential warming and cooling rates on slopes - can create local winds. Topography can also steer, and thus modify the direction of, regional winds. Intelligent siting of windbreaks therefore requires an understanding of the effect of topography on the mean and turbulent wind field. Second, the interaction between a windbreak and topography may modify the local shelter provided by the windbreak, both the level of shelter and the spatial extent of the sheltered area. Last, these changes to the mean and turbulent airflow will modify the microclimate of, and evaporation fluxes from, surfaces located near the windbreak. This paper does not consider this last issue, but it is worth noting the comments by Raupach and Finnigan (1997): that other topographic effects, such as the effect of slope and aspect on radiation receipt, water holding capacity of the soil and elevation effects on air temperature, may have a greater impact on plant microclimates and growth in hilly terrain than the effects of shelter alone. The next 2 sections describe, first, some general features of flow over hills and, second, the interaction between a windbreak and topography as revealed by a simple wind tunnel experiment.

Airflow over hills: general features

The simplest scenario is one of flow over a 2-dimensional ridge with the wind flowing normal to the long axis of this ridge and a neutrally stratified atmosphere. In flow over hills, air parcels traversing a hill respond to a pressure minimum at the crest of the hill, and pressure maxima on the windward and leeward faces of a hill. The flow responds to this pressure pattern by first decelerating as air parcels approach the pressure maximum on the windward face of the hill. There is an adverse pressure gradient, i.e. directed upstream, on this windward face. Once past the pressure maximum, air parcels then accelerate in the favourable (downstream) pressure gradient between the pressure maximum on the windward face of the hill and the pressure minimum at the crest of the hill. This leads to the common pattern of wind speed minima on the windward and leeward slopes of a hill or ridge, and a wind speed maximum, often referred to as a speedup, at the crest. Figure 17 shows typical mean wind speed profiles at key locations over a ridge for both separating and non-separating (attached) flows.

This picture is complicated slightly by 2 factors. The first is hill geometry. For axisymmetric (roughly circular section) hills, the pressure gradient at the face of the hill tends to deflect the flow around, rather than accelerate the flow over, the hill. As a result the speedup over an isolated hill tends to be less than for an elongated hill or ridge which have their long axis across the wind. Taylor and Lee (1987) give rough approximations for the magnitude of acceleration over hills of different geometries:

 $\Delta S \approx 1.6 h/L_{\rm H}$ for axisymmetric hills

 $\Delta S \approx 0.8 h/L_{\rm H}$ for 2-dimensional escarpments

 $\Delta S \approx 2.0 h/L_{\rm H}$ for 2-dimensional ridges

where $\Delta S = U/U_0 - 1$ is the fractional speedup over the hill, ridge or escarpment, *h* is the height of the hill and L_H is half the distance between 2 points (1 of the windward and 1 on the leeward side) that are halfway up the hill faces.

The second complication to the above simple picture is the response of the turbulence to the acceleration of the flow. For hills of modest slope, the flow does not become detached on the leeward side of the hill and so the flow is said to be non-separating (see below). Near the hill's surface, the production and dissipation of turbulence is essentially in local equilibrium — that is, turbulent eddies are produced by the shear in the mean flow near the surface and are strongly dissipated before they have a chance to move downwind. As a result of this, perturbations to turbulence levels correspond to areas of accelerated or decelerated flow, where there is increased or decreased shear. Thus, near the surface, on the crest of a hill, turbulence levels are higher than on the windward or leeward side of the hill where the wind speed is diminished. Over steeper hills, the picture is somewhat different. To see this first, note that over flat ground, turbulence in the flow moves the horizontal momentum of the faster moving upper air downward to be finally absorbed

by the ground surface. The mechanism for this vertical transport of momentum is the turbulent exchange of air parcels with faster moving (higher momentum) parcels moving downward and slower moving (lower momentum) parcels moving upward in the mean flow. On the lee side of a steep hill, the adverse pressure gradient can be very strong - strong enough to reverse the direction of a parcel if it does not have sufficient momentum. If enough parcels reverse direction, the mean flow on the lee side of a hill recirculates and the flow separates. Using this simple model, it is easy to see that in order for the flow to stay attached, enough momentum must be supplied to the region with an adverse pressure gradient to keep the mean flow from changing direction. If this supply of momentum is diminished, the flow will reverse more easily. Observations show that the hill slope at which flow separation takes place is about 20°. Turbulent intensities will increase dramatically in the recirculating flow leeward of the hill, where the mean wind speeds are reduced but the levels of turbulence remain similar. There may also be higher levels of turbulent production in the strong shear in the area near the top of the recirculation zone, much like that described in the lee of a windbreak near the top of the windbreak (Fig. 1 and the wind speed plot of Fig. 3).

One of the main effects of a windbreak on the mean flow is to remove momentum as a result of drag on canopy elements such as leaves and stems within the windbreak. This drag increases (roughly) as the square of the flow velocity. This being the case, 2 effects follow naturally from locating a windbreak on a hill. The first is that the amount of momentum removed from the flow will be greater at the crest of a hill than over flat ground as the wind speed is higher at the crest of a hill. Therefore, with all other things held constant, the windbreak will be more effective at the crest of a hill by virtue of the increased efficiency of momentum removal. The converse is also true - a windbreak placed in the lee of a hill, in non-separated flow, will be less effective than over flat ground. The second point is that there is a deficit of momentum in the sheltered zone in the lee of a windbreak. If this zone coincides with the adverse pressure gradient in the lee of a hill, the flow may separate. Thus, a flow that is non-separating in the absence of a windbreak may separate as a result of the reduced momentum in the lee of a windbreak placed on a steep hill. This is consistent with Finnigan and Brunet (1995), who suggested that the angle at which separation occurs for a given slope and hill geometry increases for smooth slopes and decreases for rougher, vegetated slopes. The effectiveness of a windbreak placed in separated flow in the lee of a hill depends on the windbreak height and the hill length and height. The latter determine the depth of the separated flow. A windbreak placed within the separated flow may be quite ineffective as this separated flow can be more correctly viewed as a region where there will be intermittent flow reversals, with the probability of upslope flow being greater than downslope flow. Whatever shelter is provided is thus likely to occur on alternating sides of the windbreak, if at all.



Figure 17. Typical wind speed profiles (solid lines) and streamlines over an isolated (a) low and (b) steep 2-dimensional ridge with no vegetation. The dashed lines are wind speed profiles over a flat surface.

Finally, the steering of the wind by topography needs to be considered in terms of windbreak placement. Flow tends to be preferentially channelled up or down valleys, even though the regional winds might be aligned at an oblique angle. Areas of accelerated flow can occur not only on ridge tops but also over features such as saddles, which, although lower in elevation than the surrounding ridges, can be an area of converging and accelerating flow. Windbreaks will be most effective if they are placed in regions where the topography enhances the wind speed, and oriented normal to the perturbed wind direction.

Case study of the interaction between a hill and a windbreak

A wind tunnel experiment was conducted to explore the interaction between small-scale topography and windbreaks. A 2D ridge (height, $H_h = 0.08$ m) filled the width of the wind tunnel and a model windbreak (H = 0.12 m) was mounted on the hill's crest. In contrast to other wind tunnel windbreak experiments, these model windbreaks were constructed of model trees. The roughness of the ridge itself, and the surfaces up and downwind, represented short grass. Three basic configurations were used — hill alone; windbreak alone and a windbreak sited on the hill. For this experiment, the heights of the windbreak and hill were comparable.

Figure 18 illustrates the variation of U/U_o for each of the configurations. The wind speed profile on top of the ridge, with no windbreak, shows the expected acceleration over the crest. The U/U_o profiles for trees alone, and the ridge alone, are not too different from each other in the zone immediately in the lee. Further downwind, more shelter is observed behind the porous tree windbreak than behind the solid hill. But the main result is the enhanced shelter, both in terms of the downwind extent of the sheltered zone and wind speed

reduction, found behind the combined ridge and windbreak. This increased sheltered zone is expected because the height of the 'barrier' (ridge plus trees) is now 0.2 m, more than twice the height of the hill alone. However, the reduction in U/U_o at heights greater than 0.5 H_h is greater than for the hill or tree windbreak alone, and extends upwards to 2.5 H_h. The added benefit of adding a windbreak to the hill's crest is that this is typically a zone of enhanced wind and placing a porous barrier here allows a more effective reduction in the wind speed.

These points are illustrated by the contours of wind speed reduction (U/U_0) shown in Figure 19. The contour line representing a 50% reduction in wind speed $(U/U_0 = 0.5)$ occurs below Z = 0.5 H_h for the ridge alone, and at about Z = 1 H for the tree windbreak alone. By 10 H_h (hill alone) and 12 H (windbreak alone), U/U_0 has increased above 0.5. For the combined hill plus windbreak, the zone where $U/U_0 < 0.5$ extends vertically to 2.5 H_h and beyond 12 H_h downwind (the furthest downwind measurement station). Another interesting feature is that a zone of maximum shelter extends downwind from 5 H to beyond 12 H. This is a much greater extent of downwind shelter than one would receive even with a lower-porosity windbreak than the one used here. Upwind of this zone is an area where 'jetting' through the windbreak trunk zone causes an increase in U/U_0 .

Two important points emerge about the interaction of windbreaks and topography. First, adding extra height is always advantageous in terms of increasing the downwind extent of the sheltered region. Second, the results emphasise the benefits of planting windbreaks on the crests of hills, where they protect a part of the landscape that is prone to erosive and powerful winds, and can enhance both the wind speed reduction and the downwind extent of the sheltered region.



Figure 18. Mean wind speed reduction (U/U_0) for ridge only, windbreak only, and windbreak on ridge combination. Data adapted from Finnegan *et al.* (1983).

Conclusion

Results from series of wind tunnel and field measurements have been synthesised to characterise the effect of windbreaks on shelter and microclimates, especially the spatial fields of temperature, humidity and atmospheric demand, and the turbulent fluxes of heat and water vapour from crops growing downwind of a turbulent windbreak. The important results are:

(i) The spatial pattern of wind speed follows that described in earlier studies, with a sheltered zone of reduced near-surface wind speeds that extends 5 H upwind and over 30 H downwind of the windbreak. Importantly, the spatial pattern of near-surface wind shelter does not vary with windbreak porosity, assuming a uniform vertical porosity



Figure 19. Contours of shelter $(U/U_0$, expressed as a ratio) for (*a*) ridge only (*b*) windbreak only, and (*c*) windbreak on ridge combination. Data adapted from Finnigan *et al.* (1983).

profile, leading to the conclusion that windbreak porosity determines the amount of wind shelter, while windbreak height determines the downwind extent of the sheltered zone.

(ii) Turbulence measurements clearly indicate the presence of a quiet zone underlying a growing turbulent wake layer, initiated at windbreak height, which grows in depth with distance downwind. This model enables equations to be developed for the growth of the turbulent mixing layer and thus the locations of the quiet and wake zones and the position of the minimum wind speed, which occurs towards the downwind limit of the quiet zone.

(iii) The spatial pattern of the mean scalar concentrations mimics that for wind speed, at least in the quiet zone. For a scalar with a spatially uniform ground level source, concentrations are enhanced in the quiet zone, as predicted from theory, and the maximum increase occurs at the same location as the minimum wind speed. Thus, by day we would typically expect to see enhanced air temperature and humidity in the quiet zone of a windbreak placed in a field of growing vegetation. Similarly, the size of the temperature and humidity increase varies with windbreak porosity and hence the amount of wind shelter. For example, the maximum increase in air temperatures near the surface was 0.14, 0.73, and 1.42°C for the high, medium and low porosity windbreaks, respectively, used in the wind tunnel experiments. There is a reduction in temperature and humidity in the wake zone, as predicted, but this is almost negligible and, in the wind tunnel, the reduction occurs over much greater distances than predicted from theory. As a result of the enhanced scalar fluxes in the wake zone, temperature, humidity and atmospheric demand return to their upwind values.

(iv) The difference in the spatial pattern of the near-surface wind speed compared to that for the mean scalar concentrations has twofold implications. First, microclimate effects are small in magnitude and occur over a much smaller downwind distance than wind speed reductions. Because of their dependence on surface conditions such as the type of land cover, surface moisture status and the 'regional' air mass, the effects of wind reduction on temperature and humidity are variable and may be easily masked by climate variability over a growing season. Second, the spatial extent of microclimate changes is much smaller than the downwind extent of wind shelter. This means that the direct impacts of wind on agricultural productivity through, e.g., mechanical damage (Cleugh 1998) will occur over a larger part of a paddock than indirect effects on productivity that result from these small changes in temperature and/or humidity.

(v) The turbulent scalar fluxes, i.e. evaporation and heat fluxes, also differ from the pattern of near-surface wind speeds. While significantly reduced in the quiet zone, they show a very large peak at the start of the wake zone. Thus, caution is required when extrapolating from the spatial pattern of shelter to microclimates and turbulent fluxes. Although these turbulent fluxes do not indicate the magnitude of the total flux — scalars will also be transported by the mean wind along spatial gradients — the behaviour of the scalar concentrations indicate reduced and enhanced scalar transport in the quiet and wake zones.

Wind flowing at angles other than normal to the windbreak has 2 effects on the pattern of wind shelter. First, for the medium and low porosity windbreaks used in the wind tunnel, the amount of wind shelter is increased slightly in the bleed flow region near the windbreak, i.e. there is an apparent reduction in windbreak porosity as the wind direction becomes more oblique to the windbreak. Second, the profile of near surface wind speeds is similar to that for flow oriented normal to the windbreak are accounted for using simple geometry. The field data agree with these results, but show an even greater influence of the windbreak structure on the pattern of wind shelter in the bleed flow region, extending to at least 3 windbreak heights downwind, precluding any generalisations about the flow in this region.

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