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As a publication of the Canadian Meteorological and Oceanographic Society, the Climatological Bulletin provides a medium of information on climatology. The editorial board gives special encouragement to the submission of manuscripts on applied climatology, e.g., agriculture, climatic change and variability, climate impact assessment, data bases, energy, environment, forestry, health, measurement, recreation, and transportation. Several formats are provided, including the formally reviewed "Articles", and the less formal "Notes" section. The latter consists of shorter contributions, such as research notes, surveys, overviews, and book reviews. Submissions from students through their professors are encouraged. News items are welcome, and are placed in a separate "News" section.

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Contributors should submit manuscripts to Stewart J. Cohen, Editor, Climatological Bulletin, Canadian Climate Centre, 4905 Dufferin St., Downsview, Ontario, Canada, M3H 5T4. All manuscripts should be typed double spaced on one side of good quality white paper, 28 cm × 21.5 cm or its nearest equivalent. The abstract, list of references, tables, and a list of figure captions should be typed double spaced on separate sheets. The total length of research manuscripts should not exceed 5,000 words, exclusive of illustrative material. Comments, reviews, opinions, and news items should not exceed 1,500 words. Furnish an original and three copies if possible, in the order listed below.

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Publication de la Société canadienne de météorologie et d'océanographie, le Bulletin climatologique offre un moyen d'information sur la climatologie. Le comité de rédaction encourage en particulier la soumission de manuscrits sur la climatologie appliquée, comme l'agriculture, les changements et la variabilité du climat, la prospective climatologique, les bases de données, l'énergie, l'environnement, la sylviculture, la santé, les mesures, les loisirs et les transports. Cette publication compte plusieurs catégories d'articles, dont les "Articles", officiellement évalués, et la partie plus libre des "Notes", qui se compose d'articles plus courts, comme les notes de recherche, les études, les vues d'ensemble et les critiques de livres. On incite les étudiants à présenter des articles par l'intermédiaire de leurs professeurs. Nous faisons bon accueil aux informations, qu'on publie à part dans la partie des "Nouvelles".

Les auteurs, y compris les étudiants, peuvent choisir de soumettre leurs manuscrits à l'appréciation officielle ou aux "Notes". Ils doivent l'indiquer sur la lettre d'accompagnement du manuscrit. Les articles de recherche sont indépendamment soumis à l'examen d'au moins deux appréciateurs anonymes. Le rédacteur en chef examine les "Notes" conjointement avec le comité de rédaction. On accepte les articles soit en français, soit en anglais. Il faut envoyer un résumé, de préférence en français et en anglais.

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Foreword

Applied climatology is a field dominated by the needs of the user, rather than the researcher. The dissemination of climatic information depends, in part, on the ability of the climatologist to demonstrate how this information can be used for planning and design purposes. Such "marketing" efforts should lead to more widespread use of data bases and models by public and private agencies, and will ultimately lead to a higher level of environmental awareness among decision makers in all fields of endeavour.

In situations where data are not available by direct measurement, and must be estimated by using models, a step by step or "cookbook" approach may be needed in order to facilitate the successful application of these models to the problem at hand. Taylor and Lee provide such detailed documentation in their paper on estimating wind speed variations due to topographic features.

This issue also includes a review paper by Pulwarty and Cohen on CO_2-induced climatic change and its effects on world food production. Pulwarty is a student from York University who wrote an essay for a course I taught at the Department of Geography. I encouraged him to submit it to the Bulletin, where it was independently reviewed. During the Bulletin's years as a McGill University publication, students contributed close to one-fourth of all articles which appeared from 1967 to 1982.

Stewart J. Cohen

Simple guidelines for estimating wind speed variations due to small scale topographic features

P.A. Taylor and R.J. Lee

Hills, escarpments and valleys can give rise to significant local variations in near surface wind speed. For moderate to strong wind speeds ($>6ms^{-1}$), near neutral thermal stratification and horizontal terrain length scales of order 1 km or less, these variations will be primarily associated with aerodynamic rather than thermal effects. Under these circumstances and for moderate terrain slopes (≤ 0.3) Hunt (1980) notes that the maximum of the fractional speedup, ΔS, above the summit of low hills or other terrain features, is approximately double that predicted by inviscid irrotational flow theory. We use this result as a basis for proposing guidelines for estimating near surface wind speed variations. These estimates compare well with field data, wind tunnel and numerical model predictions. There are many applied engineering and field oriented activities where wind speed estimates in complex terrain are needed. Some examples and a procedure for applying the guidelines are given.

Les collines, les escarpements et les vallées peuvent causer des variations locales d'importance dans la vitesse du vent près de la surface. Pour les vitesses du vent modérées à fortes ($>6ms^{-1}$), avec stratification presque neutre et pour les échelles de longueur de terrain de près de 1 km ou moins, ces variations seront associées avec les forces aérodynamiques plutôt que les effets thermiques. Dans ces conditions et pour les pentes modérées (≤ 0.3), Hunt (1980) fait remarquer que le maximum d'augmentation fractionelle de vitesse, ΔS, au dessus des sommets des collines basses ou d'autres caractéristiques du terrain, s'approche du double de celui donné par la théorie des fluides inviscides et sans rotation.

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Nous utilisons ce résultat comme base pour proposer les lignes de conduite pour estimer les variations du vent près de la surface. Ces estimations se comparent bien avec les données observées et avec les prévisions obtenues par soufflerie et avec des modeles numériques. Il y a beaucoup d'activités d'extérieur et de génie appliqué où les estimations du vent en terrain complexe sont nécessaires. Nous présentons quelques exemples et aussi un procédé pour l'application des lignes de conduite.

1 INTRODUCTION

There are basically three characteristics of terrain whose spatial variation can cause variations in the near-surface wind field. These are

(1) surface roughness;
(2) surface thermal and/or moisture properties; and
(3) surface elevation.

On a 'local' scale, over horizontal distances of about 10m to 10km the effects of abrupt changes in surface roughness or thermal properties and the wind speed changes within the resulting 'internal boundary layer' are fairly well understood – see for example Hunt and Simpson (1982). Surface elevation changes can give rise to flow changes via thermal effects – slope and valley winds – but can also have a direct aerodynamic effect on the flow. At horizontal scales of about 1-2 km in moderate to high wind speed (say $>6 ms^{-1}$) and without strong thermal stratification the aerodynamic effects will usually dominate and it is these which will be discussed in the present paper. For a discussion of the relative importance of thermal and aerodynamic effects of topography see Taylor and Gent (1980).

The simplest example is the 'speed-up' observed as the wind blows over a hill. There is however, still some confusion in the literature regarding the cause of this speed-up. Davidson *et al* (1964), for example, tend to give the impression that wind speeds over hills are higher because an anemometer on a hill is at a greater absolute height above sea level and is thus higher in the boundary layer. While this might be true in the extreme case of a steep pinnacle (Fig. 1a), it would not apply for a 2D ridge or low 3D hill (Fig. 1b), where, as a first approximation, the whole boundary-layer is simply displaced up and over the hill and the flow acceleration is basically due to the air having to travel faster through the partial constriction caused by the terrain.

The concept is built into the following standard definitions which are based on wind speeds at a given height above the local terrain.

The 'normalized wind speed', sometimes referred to as 'amplification factor' is defined as

$$A(x,y,\Delta z) = \frac{U(x,y,\Delta z)}{U_0(\Delta z)} \tag{1}$$

where $\Delta z = z - z_s(x,y)$ is the vertical height above the local terrain of elevation z

$= z_s$. $U(x,y,\Delta z)$ is the mean wind speed at any point and $U_0(\Delta z)$ is the undisturbed or upstream wind profile – see Fig. 1b. It will also be convenient to introduce the velocity perturbation,

$$\Delta U = U(x,y,\Delta z) - U_0(\Delta z) \tag{2}$$

and define the 'fractional speed-up ratio',

$$\Delta S(x,y,\Delta z) = \frac{\Delta U(x,y,\Delta z)}{U_0(\Delta z)} = A - 1. \tag{3}$$

Hunt (1980) [see also Hunt and Simpson (1982)] makes use of this notation in deriving his 'rules of thumb' for estimating maximum speed-up above simple hills. These rules will form the basis for our guidelines.

The present paper will not discuss changes to the turbulence occurring in flow over hills and other terrain features. Some estimates are, however, presented by Hunt (1980) and additional details are given by Bitter *et al.* (1981). In general turbulent intensities (σ_u / U etc.) will tend to be significantly reduced above hilltops as a result of increased wind speeds but only small changes in σ_u. Bradley (1980) and Taylor and Teunissen (1983) discuss some observations. Our primary concern will be with near-neutral thermal stratification but some discussion of non-neutral conditions will be given in Section 5.

2 NEUTRALLY STRATIFIED FLOW OVER LOW HILLS – RULES OF THUMB

Although there have been a number of alternative theoretical modelling studies of neutrally stratified flow over isolated low hills, the work of Jackson and Hunt (1975), Mason and Sykes (1979), Jackson (1979) and Hunt (1980) is probably the most useful basis for providing simple guidelines for estimates of 'speed-up' on hills. The same work has also provided the foundation for a detailed model of flow over 'real terrain' designated MS3DJH [Walmsley *et al* (1982), Taylor *et al* (1983)]. This model has been used to provide some of the results given later.

Within the atmospheric boundary-layer, the most rapid changes of wind speed with height normally occur near the ground. In the ideal case of a horizontally homogeneous infinite flat plain, the velocity profile, ignoring any changes in direction with height, can be well-represented by the logarithmic form

$$U_0(z) = \frac{u_*}{\kappa} \ln \frac{z}{z_0} \tag{4}$$

where κ (=0.4) is von Karman's constant and u_* is referred to as the friction velocity [see for example Oke (1978)]. While it was originally derived for the relatively shallow 'constant flux layer', the log-law has been shown to often apply quite well, in near-neutral conditions, up to heights of order 150m [see Panofsky (1973)] and it is convenient to assume this form for the upstream profile in the analysis. We should however remark that there will be occasions when the profile departs significantly from equation (4), especially for heights of order L. and that these departures could have a significant impact on the observed speed-up near the surface.

For rough surfaces (say $z_0 > 0.1$m) it is often necessary to introduce a 'zero plane displacement', d, and replace 'z' by 'z−d' in equation (4) in order to match observed profiles to the logarithmic form. Oke (1978, p.97, 98) gives a discussion of this and provides an estimate for d as $d \simeq 2/3\ h_e$ where h_e is the height of the roughness elements (e.g. the height of a vegetation stand or an average building height). We will not include this in the following analysis but it may be necessary to consider it if velocity measurements are being extrapolated from one height to another. Note that the surface roughness length, z_0, will initially be assumed to be the same on the hill as in the surrounding terrain.

Jackson and Hunt (1975) argue that the effects of turbulent momentum transfer are only significant, in the flow over low hills, within an 'inner layer' of thickness, ℓ, which they show to satisfy the equation

$$\frac{\ell}{L} \ln(\ell/z_0) = 2\kappa^2 \tag{5}$$

Here L, the length scale of the hill, is defined as the distance from the hilltop to the upstream point where $z_s = h/2$, if h is the height of the hill above the surrounding terrain (see Fig. 1b). We will use this specific definition of L throughout the paper to avoid some of the uncertainties present in other work. The relationship between ℓ and L is shown in Fig. 2.

Above this inner layer is an 'outer layer' where the flow is modified by the pressure field in essentially the same way as the inviscid irrotational flow of an ideal fluid past an obstacle. Jackson and Hunt (1975) show that the pressure perturbations, Δp, within the outer-layer flow will scale with the velocity $U_0(L)$ so that

$$\frac{1}{\rho} \Delta p = O\left[\frac{h}{L} U_0^2(L)\right] \tag{6}$$

where O [ ] indicates the order of magnitude of the quantity in brackets.

Although they argue that this result cannot be 'arrived at without detailed analysis', it is nevertheless what one might suspect intuitively. Velocity perturbations, ΔU, in inviscid irrotational flow would be $O[(h/L)\ U_0(L)]$.

The outer-layer pressure field is also assumed to act on the flow in the inner layer but here, because the disturbed velocities are lower $[O(U_0(\ell))]$, the same pressure perturbations will produce *larger* values of the amplification factor or fractional speed-up than would be predicted by inviscid irrotational flow

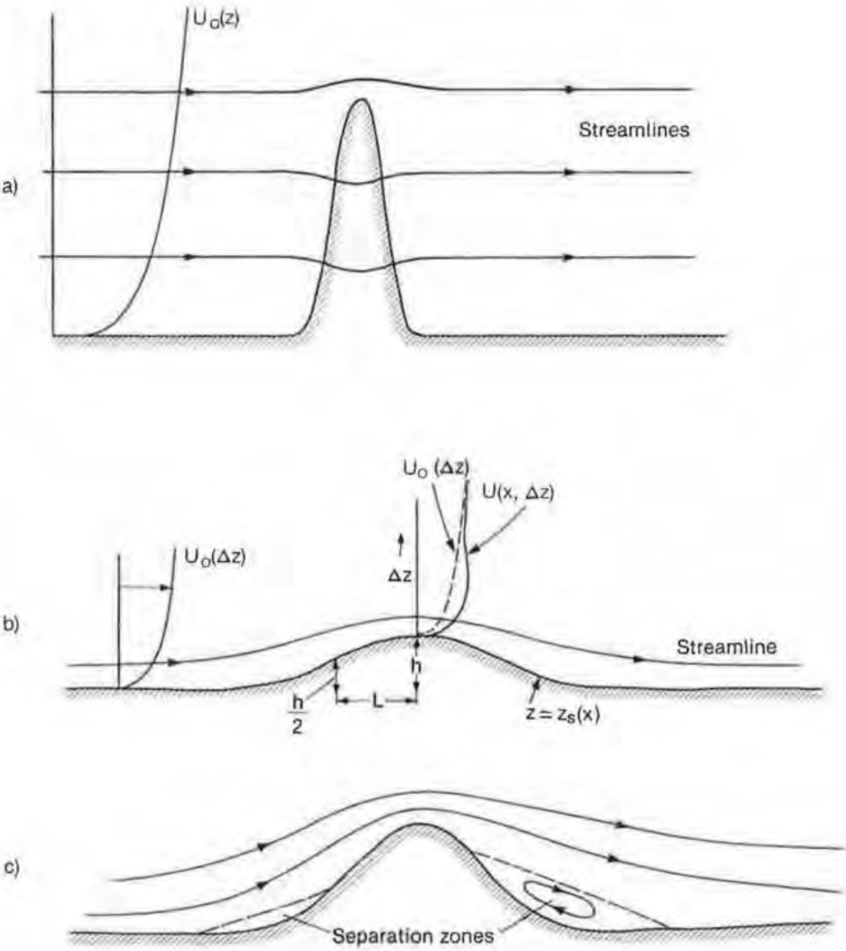

FIGURE 1 (a) Speed-up due to increased height in the boundary-layer.
(b) Aerodynamic speed-up over a low hill, without separation.
(c) Flow with separation.

over the hill with velocity $U_0(\ell)$. In fact near the outer edge of the inner layer, where frictional effects are only just effective but where shear in the incident flow causes $U_0(\ell)$ to be less than $U_0(L)$, it can be argued that

$$\Delta U = O\left(\frac{h}{L}\ \frac{U_0^2(L)}{U_0(\ell)}\right) \tag{7}$$

and

$$\Delta S = O\left(\frac{h}{L}\ \frac{U_0^2(L)}{U_0^2(\ell)}\right) \tag{8}$$

This can be compared to $\Delta S = O(h/L)$ for inviscid irrotational flow. At heights within the inner layer with $\Delta z < \ell$ both $U_0(\Delta z)$ and the perturbed flow above the hilltop will have essentially logarithmic profiles and ΔS should be roughly constant, although as we note in Section 4 there will be some variation. The ratio of the fractional speed-up ratio maxima over hilltops in boundary-layer and irrotational flow is thus $U_0^2(L)/U_0^2(\ell)$. For typical values of L/z_0 in the range 10^3 to 10^5 the value of this ratio is close to 2.

The arguments leading to this result are somewhat tenuous and, indeed, Jackson (1979) and Hunt (1980) offer slight variations on the same theme. The essential point, however, is that given by Hunt (1980) who argues that, for atmospheric boundary-layer flow over hills with low slope, the maximum value of ΔS can be well approximated by

$$\Delta S_{max} = 2\left(\frac{h}{L}\right)\sigma_{max} \tag{9}$$

where $(h/L)\,\sigma_{max}$ is the maximum fractional speed-up ratio predicted for potential flow over the same hill. It will occur near the surface at the hilltop.

For some simple mathematically-specified two-dimensional terrain features, of the form $z_s = hf(x)$, Hunt (1980) provided some of the σ_{max} values reproduced in Table 1. We have added some additional cases here. Note that for the simple 2D hills or ridges, $\sigma_{max}, \simeq 1$, while for a ramp or escarpment $\sigma_{max}, \simeq$ 0.3—0.5. These results and additional studies of flow over three-dimensional axisymmetric hills lead us to propose the following simple 'rules of thumb'.

$$\begin{aligned} \Delta S_{max} &= 2\ h/L \text{ for 2D ridges (or valleys, with h negative)} \\ &= 0.8\ h/L \text{ for 2D escarpments} \\ &= 1.6\ h/L \text{ for 3D axisymmetric hills.} \end{aligned} \tag{10}$$

These estimates should only be used for h/L values up to about 0.5 (max slopes ~ 0.35). When the slopes get too steep non-linear effects become important and the flow may well separate (Fig. 1c). $\Delta S_{max} \simeq 1.2$ is probably an upper bound for this type of aerodynamically-controlled terrain-induced speed-up and values of $\Delta S > 1$ should be treated with caution.

The 2D cases are for flow perpendicular to the feature. For flow at an angle, the equations still hold approximately, but with L adjusted (according to its definition) to take account of the angle of incidence, i.e. $L = L_0/\mathrm{Cos}\,\theta$ where L_0 is the value at normal incidence and θ is the angle between the surface flow direction and the normal. An alternative approach (see Jackson (1979) or Bradley (1983) is to consider separately the terrain influence on flow normal and parallel to the hill. The $\mathrm{Cos}\theta$ factor will again apply to first order. Note that there may be significant (~20°) changes in wind direction over 2D terrain features for non-normal flow. Taylor (1977b) gives some examples of model computations while Bradley (1983) gives a few results from his field study at

TABLE 1: Fractional speed-up maxima for potential flow over analytically-specified terrain. $\sigma_{max}=\Delta U_{max}/U_0(h/L)$.

Basic Terrain Type	Hill Shape	Analytic Form, f=	Max Slope/ (h/L)	σ_{max}
2D Ridges	Bell-shaped	$1/(1+(x/L)^2)$	0.65	1.0
	Gaussian	$\exp[-(x/L)^2 \ln 2]$	0.71	1.13
	Cosine Squared	$\cos^2(\pi x/4L), x<2L$ $0 \quad ,x\geq 2L$	0.79	0.92
2D Escarpments	1/2 Bell	$1/(1+(x/L)^2), x<0$ $1 \quad ,x>0$	0.65	0.5
	Tanh Ramp*	$1/2[1+\tanh(x/L)]$	0.5	0.29
3D Circular Hills	Bell-shaped	$1/(1+(r/L)^2)$	0.65	0.79
	Cosine Squared	$\cos^2(\pi r/4L), r<2L$ $0 \quad ,r\geq 2L$	0.79	0.73
Rolling Terrain	2D Sinusoidal	$(1+\cos \pi x/2L)/2$	0.79	0.795
	3D Sinusoidal	$(1+\cos \pi x/2L \cos \pi y/2L)/2$	0.79	0.555

* definition of L is only approximate in this case.

Bungendore. Bradley's field data show minimal direction changes. The wind direction above a two dimensional ridge should be deflected towards the normal, while for a valley bottom it would be deflected along the valley.

For the 'rolling terrain' cases the values of σ_{max} given in Table 1 indicate the increase in wind speed on the tops of the ridges or peaks compared to the basic flow velocity. In practice it may be easier to use these results (x2) to indicate the range of velocities occurring between hilltop and valley bottom. We will discuss this further in Section 3.

Equations (10) do not include any allowance for the relative roughness of the surface and are intended to be applied for L/z_0 in the range 10^3 to 10^5. In general the variations with roughness length are quite small with the lower values of L/z_0 corresponding to slight increases (10-20%) in ΔS_{max}. Jackson and Hunt (1975, p.946) give some examples of variation in ΔS for different values of ℓ/z_0. Their cases are all for relatively smooth surfaces.

3 FIELD OBSERVATIONS AND NUMERICAL MODEL STUDIES

Several recent field studies have provided data against which the proposed rules may be tested. Mason and Sykes (1979) made observations of wind speed at a height of 2m on and around Brent Knoll in Somerset, England. This isolated, mainly grass covered, 137m high hill is almost axially symmetric ($L\sim250$m) and Mason and Sykes analyzed their data on this basis. Their hilltop value of ΔS was about 1.3 while the estimate based on 1.6 h/L gives only 0.9. It is quite a steep hill with a shape tending to maximize near-surface speedup.

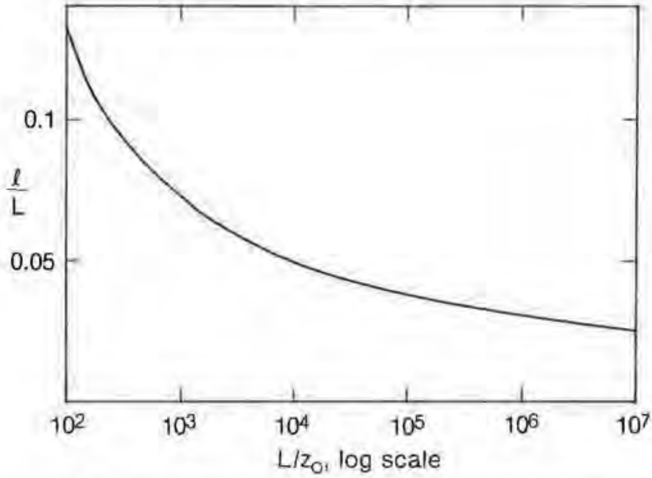

FIGURE 2 The inner layer thickness, ℓ, as a function of L/z_0.

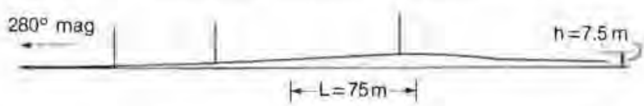

FIGURE 3 Cross-section of Bungendore ridge, without vertical exaggeration showing h, L and mast locations – reproduced from Bradley (1983).

Bradley's (1983) work on a low, nearly two-dimensional ridge near Bungendore (NSW, Australia) gave $\Delta S \simeq 0.15$ at $\Delta z \simeq \ell \simeq 3$m for winds normal to the ridge but with variations from 0.1 to 0.3 depending on stability and surface roughness conditions. Values of h and L are 7.5m and 75m respectively and our 2D ridge estimate of ΔS would be 0.2. However Bradley's ridge (see Fig. 3) is in fact part escarpment since there is a net change in elevation and on this basis an estimate of $\Delta S_{max} = 0.5(2.0+0.8)h/L = 0.14$ provides a good match to the data. This study was for a feature with low slopes in good accord with the theoretical assumptions. Taylor (1977b) gives some numerical model results for planetary boundary layer flow above 2D Gaussian ridges with the geostrophic wind direction at various angles (α_u) to the normal to ridge. Within his model the angle between the surface and geostrophic wind directions is 18.7° (for $Ro = |U_g|/fz_0 = 10^7$) and so the angle between the surface wind and the normal to the ridge is $\theta = \alpha_u + 18.7°$. His results are given in Table 2 together with estimates based on $\Delta S = 2\,h\,\mathrm{Cos}\theta/L_0$. Our rule of thumb values appear to underestimate both the value of ΔS at normal incidence and the reductions due to non-normal incidence but the overall agreement is adequate. Bradley's (1983) data also suggest that non-normal angles of incidence lead to larger reductions in ΔS than the simple $\mathrm{Cos}\theta$ rule predicts.

TABLE 2 Flow at an angle to 2D Gaussian Hills

Results from Taylor (1977b)			Rule-of-Thumb Values	
α_0	$\tau_s^{max}/\rho\lvert U_0\rvert^2$	ΔS	Θ	ΔS
−67.5	1.68×10^{-3}	0.26	−48.8	0.32
−45	2.27×10^{-3}	0.47	−26.3	0.43
−22.5	2.64×10^{-3}	0.58	− 3.8	0.48
0	2.54×10^{-3}	0.55	18.7	0.45
22.5	2.05×10^{-3}	0.39	41.2	0.36
45	1.48×10^{-3}	0.18	63.7	0.21

Taylor's (1977) computations are for a gaussian hill $z_s = a \exp(-(x/\beta)^2)$ with $a = 4000\ z_0$ and $\beta = 2 \times 10^4 z_0$ ($L = 1.666 \times 10^4 z_0$). ΔS is calculated from the surface shear stress values (τ_s^{max}) given and represents the limiting surface value at the hilltop. (Upstream shear stress has $u_* = 0.0325\ \lvert U_0\rvert$).

TABLE 3 Data* on hilltop amplification factors from the 1981 Kettles Hill Experiment

Run No.	Average Wind Direction	Hilltop Amplification Factor+ 3m	6m	10m
5a	257°	1.33	1.30	1.31
5b	255°	1.23	1.23	1.25
6a	220°	1.61	1.55	1.57
7a	262°	1.17	1.15	1.18
7b	260°	1.20	1.19	1.21
8a	263°	1.26	1.22	1.26
9a	251°	1.22	1.26	1.31
9b	261°	1.26	1.22	1.23

* Corrected for Tower 1 perturbation (Tower 1 = 95% of upstream flow for directions 250°-263°, 99% for 220°)

\+ Based on measurements with cup anemometers at 3m and 6m and Gill UVW propeller anemometers at 10m.

The work on Kettles Hill (Fig. 4) reported in Taylor *et al* (1983b), included measurements upwind and on the hilltop at 3m, 6m and 10m. Some data from that experiment are given in Table 3. Amplification factors, $A = 1 + \Delta S$, for the three levels were approximately the same but there were pronounced variations with incident wind direction, as is to be expected since the hill is distinctly asymmetric. The data obtained were mostly for directions (degrees true) between 250° and 260° but on one day the direction was approximately 220°. Values of L for these directions are estimated from the contour map to be 520m for 260°, 440m for 250° and 330m for 220° while we take h = 100m. Ground cover was low grass, mud and stubble with an estimated surface roughness length, $z_0 = 0.01$m. In applying equations (10) to Kettles Hill we can subjectively interpolate between the 2D ridge and axisymmetric hill values. For the 220° case a factor 1.8 would appear logical, while for the 250° and 260° cases

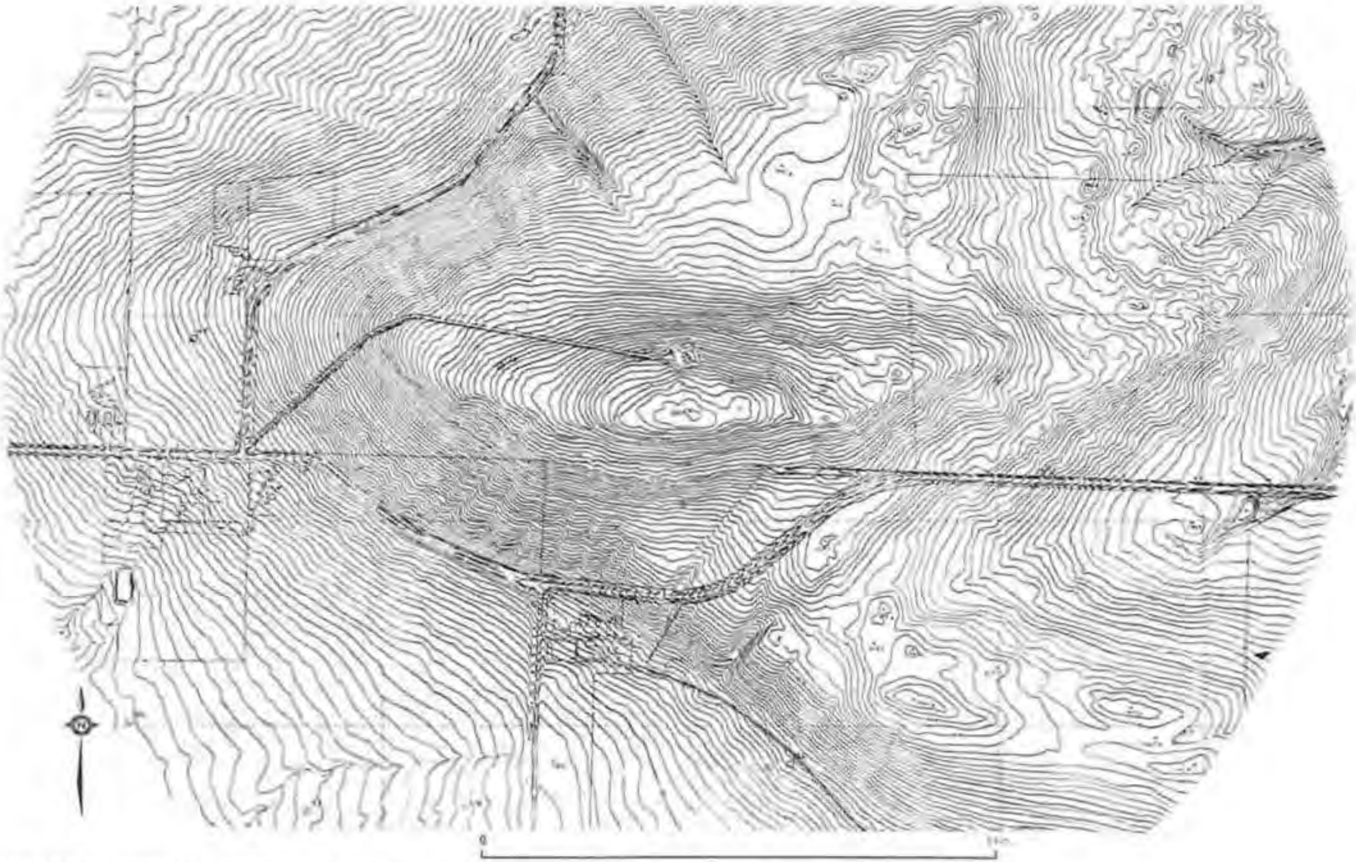

FIGURE 4   Kettles Hill contour map, contour interval 1m.

TABLE 4 Fractional speed-up ratio data and estimates for Askervein from the 1982 Experiment, h=116m.

Run No.	Wind Direction	ΔS measured at Δz=10m at HT	L(m)	ΔS_{max} Estimates 1.6 h/L	1.8 h/L
1.22	180°	0.61	280	0.66	0.75
1.23a	230°	0.78	220		0.95
1.23b	245°	0.61	240		0.87
2.25	120°	0.34	790* (650)	0.24 (0.29)	
2.27	165°	0.62	380	0.49	0.55
2.29b	235°	0.65	220		0.95
2.01a	165°	0.53	380	0.49	0.55
2.01b	155°	0.40	530	0.35	
2.02	200°	0.82	210		0.99

* Distance to upstream contour is anomalously large in this case, value in brackets is more representative.

the axisymmetric hill formula is probably appropriate. Our estimates for ΔS_{max} would then be 0.31 for 260°, 0.36 for 250° and 0.55 for 220°. These are a little high in comparison with the values in Table 3 (approx. 0.24, 0.31 and 0.57) for the three directions considered but still provide useful guidance. During the experiment, hilltop winds from a standard U2A anemometer were compared at half-hour intervals with those recorded at Pincher Creek airport, about 12km to the west of the hill. Fractional speed-up values based on these data were grouped into two classes based on airport wind directions, which were sometimes about 10° different from directions measured near the hill. For 245° ± 10° the average value of ΔS was 0.40 while for 220° ± 10° it was 0.64.

During September and October 1982 a team of Canadian, British, Danish, and German scientists and technicians made detailed measurements of near-surface wind and turbulence in the flow over Askervein on the island of South Uist, Scotland – see Fig. 5. This is a relatively isolated hill with rather steeper slopes than Kettles Hill. Its height is 126m above sea level, about 116m above the upstream terrain. Surface cover was heather, flat rock, coarse grass and peat bog with an estimated surface roughness length of 0.02-0.05m. The main emphasis of the 1982 experiment was on a detailed resolution of spatial variations in the 10m mean wind field. The data, which are described in a report by Taylor and Teunissen (1983), include winds from an upwind reference site and a hilltop location (HT in Fig.5b). Table 4 lists computed speed-up values for selected two-hour runs together with wind directions, corresponding values of L and our simple estimates for ΔS. The L values are the upstream distances from HT to the 68m contour. In one case, run 2.25, this is anomalously large due to the location of HT relative to the hill, but in general the value given is characteristic of points on the ridge between HT and CP. As with Kettles Hill, Askervein is an asymmetric hill and we have used coefficients of 1.6 or 1.8 to reflect the different character of the hill for different wind directions. In general the estimates seem reasonable although for runs with wind directions from 200° to 245°

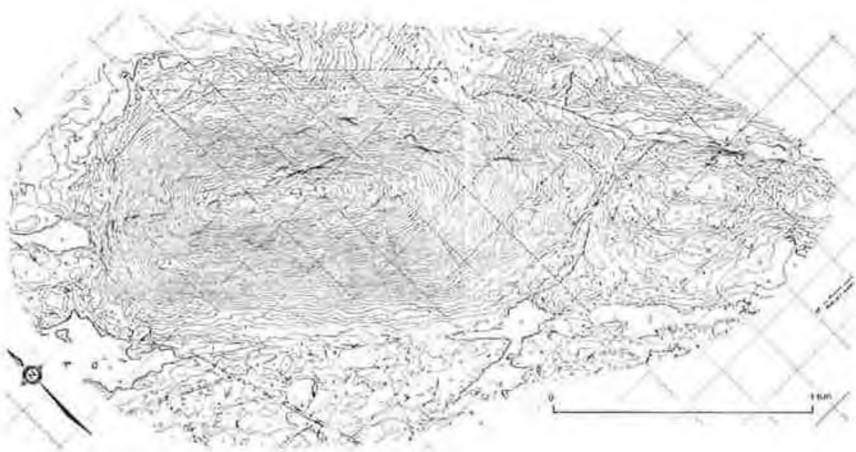

FIGURE 5 Askervein: Photograph from SW and detailed contour map. Contour interval is 2m although some lines with 1m spacing are also shown.

the observed ΔS values at Δz=10m are significantly lower than our estimates. All the data are for moderate to strong winds with near-neutral stability. It is clear that the values for L in the case of asymmetric hills can be strongly dependent on wind direction and that wind direction can be an important parameter affecting the magnitude of ΔS. Mean wind data from a total of thirty-five 10m towers were available from the Askervein experiment in addition to profiles up to 50m at the reference site and on the hilltop. Wind speed variations over the hill are closely linked to the topography with the maximum at the hilltop.

The field data discussed so far were all from studies conducted on hills with relatively smooth surfaces. It is, after all, much easier to make representative wind speed measurements with 10m or shorter towers in these cases, whereas for wooded hills one needs to go to much greater heights to get above the roughness elements. Bradley (1980), however, made a series of measurements from a 100m tower on top of a wooded 170m hill (Black Mountain) near Canberra, Australia. The hill is only slightly asymmetric and Bradley gives a value of 275m for L. It is thus a rather steep hill with slopes of order 0.4. The roughness length was estimated as $z_0 \simeq 1$m. Bradley found an average $\Delta S = 1.07$ on the hilltop for Δz<28m which he compared with Jackson and Hunt's (1975) estimate of $\Delta S_{max} = 2\ h/L = 1.24$. Allowing for the fact that the hill is almost axially symmetric suggests however that we should use the estimate $\Delta S_{max} = 1.6\ h/L$ which gives $\Delta S_{max} = 0.99$, in good agreement with the observation considering that it is such a steep hill.

We are not aware of any really suitable field observations with which to test the results for escarpments but a range of $\Delta S_{max} \sim 0.7 - 1.0\ (h/L)$ is compatible with numerical model computations [e.g. Taylor (1977a)] and wind-tunnel measurements [Bowen and Lindley (1977)]. The model results indicate a substantial dependence on the specific shape of the escarpment while for the sharp-edged escarpments used in the wind-tunnel studies there tend to be strong variations with height close to the surface.

For flow across river or other valleys we can use $\Delta S_{max} = 2\ h/L$ to estimate wind speed reductions but with 'max' interpreted as 'extreme' and h taken as negative. Once again there is a scarcity of suitable field data but Taylor (1977a) presents numerical computations for 2D Gaussian valleys ($z_s = h \exp(-(x/\beta)^2)$ where h (negative) is the depth of the valley and β is a measure of its width). He finds $\Delta S_{max} \simeq 2.45\ h/\beta = 2.04\ h/L$, for flow normal to the valley. For flow at an angle to the valley the appropriate adjustment in the choice of L should be made. Recent studies by Mason and King (1984) do, however, suggest that 3D effects may be important for valley flows at non-normal incidence. Note that the flow across valleys will be likely to separate for maximum slopes greater than about 0.3 ($h/L \simeq -0.5$) or even less, leaving essentially stagnant or very light winds in the valley. One should therefore be very cautious about applying these estimates in reverse, i.e. estimating wind speeds in the surrounding area from observations made in valleys. Wind directions in valleys, even in the absence of thermally-driven circulations of flow separation, may easily deviate by 20°-30° from flow in the surrounding area [see Taylor (1977b)].

For the case of terrain which really has no flat areas and consists of a continuous sequence of 'rolling hills' we can use equation (9) applied to both hilltop and valley bottom locations and, for simple, regular, sinusoidal terrain use the results given in Table 1 for σ_{max} to give the estimates

$$\frac{U_H - U_v}{\frac{1}{2}(U_H + U_v)} = 3.1 \frac{h}{L} \quad \text{for flow normal to 2D rolling hills}$$

$$= 2.2 \frac{h}{L} \quad \text{for flow over 3D isotropic rolling hills} \qquad (11)$$

Here U_H and U_v are the hilltop and valley bottom winds at a height $\Delta z < \ell$. $(U_H+U_v)/2$ is an estimate of the horizontally averaged wind at Δz, h is the vertical height of the hilltop above the valley or hollow bottom and L is ¼ of the distance between ridge crests or hilltops.

If either U_H or U_v can be measured then equation (11) gives an estimate of the other one. If an estimate of $0.5(U_H+U_v)$ can be made, then this, coupled with equation (11), can give an estimate of both U_H and U_v.

Equations (11) have been corroborated by MS3DJH model calculations of flow over the sinusoidal terrain specified in Table 1. Taylor *et al* (1983a) give results for the 2D case. There are also some field observations available from the recent paper by Mason and King (1984) which describes a study of flow over a succession of 2D ridges and valleys in South Wales. The value of h is approximately 200m while we estimate L as 400m. Flow separation does occur for flow across the valleys and the near surface valley bottom wind U_v is near zero. Thus $(U_H-U_v)/0.5(U_H+U_v) \sim 2$ while our estimate is 3.1 h/L=1.6. Separation thus appears to occur at a slightly less steep slope than our simple linear estimate would predict although this may be partly due to the slightly irregular cross-section of the valleys studied.

For 3D rolling hills some typical MS3DJH model results are shown in Fig. 6. Normalized wind speed and wind direction perturbations are for a height of 2m above the terrain, which has h=100m, L=300m and z_0=0.03m. The upstream flow is parallel to the x-axis. Note that there can be quite significant changes in wind direction at different locations in the terrain (±20° relative to average). The normalized wind speeds vary from about 0.63 in the hollows to 1.37 on the peaks relative to the 'undisturbed' or average value. This is in good agreement with equation (11).

4 SPATIAL VARIATIONS

The simple rules of thumb presented in equations (10) apply only to estimates of near-surface wind speeds on hill, ridge or escarpment summits or at the bottom of valleys or hollows. The surface roughness should be reasonably uniform or

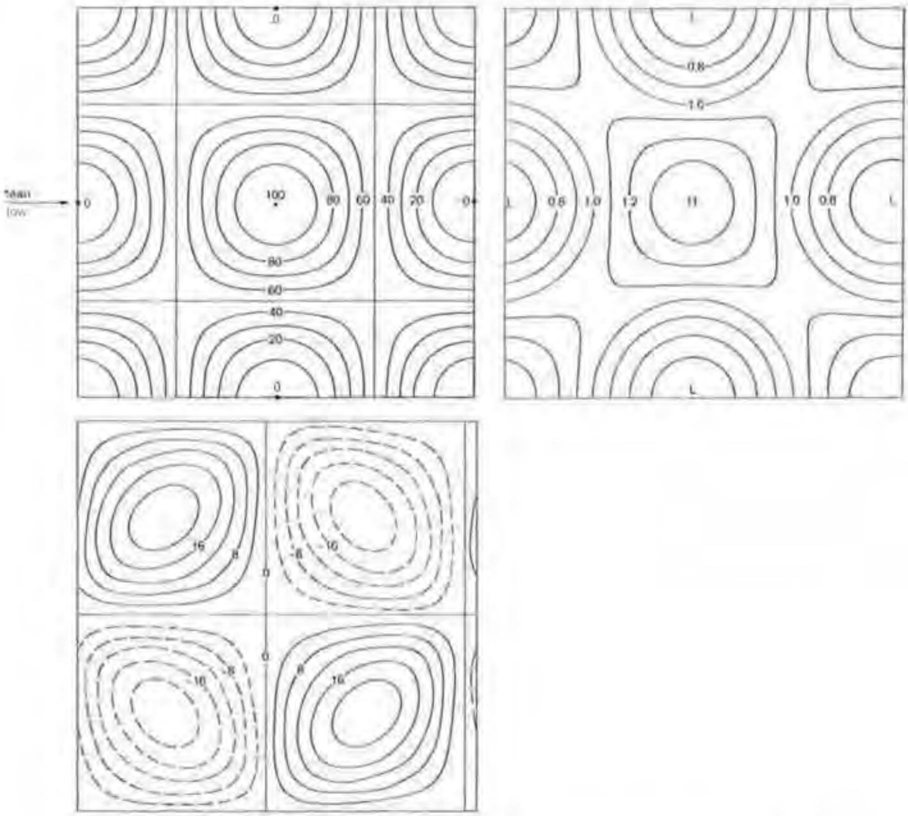

FIGURE 6 (a) Terrain, (b) normalized wind speed and (c) wind direction perturbations for flow over 3D rolling terrain predicted by MS3DJH/3.1. Undisturbed flow is from the left – see text for terrain details. Domain size is 4L x 4L. Contour intervals are (a) 10m, (b) 0.1, (c) 4°. Negative contours are dashed in part (c).

additional estimates made of the effect of roughness change. The terrain feature should be isolated and the reference wind speed U_0 measured sufficiently far from the hill for its influence to be minimal. A distance of at least 4L upwind or to the side of a 3D hill is recommended. Immediately upwind and in the lee of most hills there will be reductions in the wind speed (see for example Jensen (1983) or Taylor *et al* (1983a). Measuring U_0 in this area could lead to underestimates of hilltop wind speed. It is rather hard to estimate wind speeds at other positions, e.g. mid-slope, on hills or other terrain features, since the detailed shape of the hill will strongly influence the spatial variations. The maximum will however usually occur at, or very close to, the hilltop. The upstream minimum will typically correspond to a decrease in wind speed by about ¼ of the increase at the hilltop (i.e. $\Delta S \simeq -0.5\ h/L$ for a 2D ridge) and will occur close to the upstream base of the hill. The minimum will be most pronounced in cases where

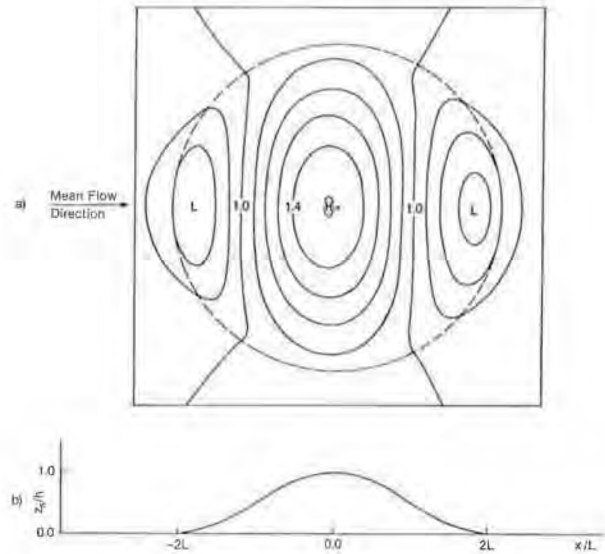

FIGURE 7 Normalized wind speed at 2m height above a cosine squared hill. Undisturbed flow is from the left. Topographic cross section also shown. h=100m, L=300m, z_0=0.03m. Contour interval 0.1. x marks the hilltop while the dashed line is at r=2L.

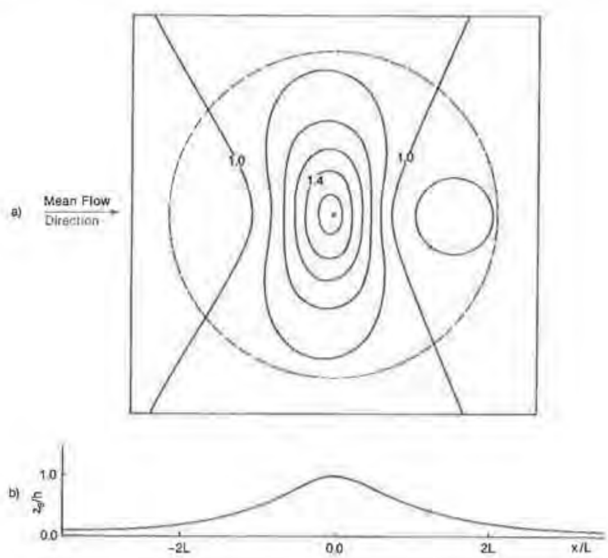

FIGURE 8 Same as Fig. 7 but for bell-shaped hill.

the hill rises abruptly in an otherwise flat plain and could be absent in cases where there is a gradual transition. Simple linear interpolation between this point and the hilltop would probably be as good as anything for estimates at intermediate points on the upwind slope. Wind speeds in the lee of the hill are more difficult still, especially for steep hills where flow separation and high turbulence levels are likely. As guidance in assessing spatial variations of near surface wind speed, Figs. 7 and 8 give results for two idealized isolated hills. These were computed using a computer model, MS3DJH/3.1 (see Taylor et al (1983a) for details), which can also be used with real terrain input. The hills are defined by

$$z_s = hf(R)$$

where

$$R^2 = \left(\frac{x}{L}\right)^2 + \left(\frac{y}{L}\right)^2.$$

The first is a 'cosine-squared' hill with

$$f = \begin{cases} \mathrm{Cos}^2 \frac{(\pi R)}{4} & R < 2 \\ 0 & R \geq 2 \end{cases} \tag{12}$$

The second has

$$f = 1 / (1 + R^2) \tag{13}$$

and is an axially symmetric bell-shaped hill. Both forms maintain the standard definition of L as the distance from the hilltop to the point where $f = \frac{1}{2}$. The contour plots are of normalized wind speed at Δz=2m for a case with L=300m, h=100m and z_0=0.03m. Only the central portion (5L×5L) of the computational domain (total 10L × 10L) is shown and the dashed circle is at r = 2L. Our rule-of-thumb would predict a maximum $\Delta S_{max} = 0.53$ or A = 1.53, which is consistent with the numerical computations for the 2m level (A = 1.49 at the hilltop for the cosine squared hill and A = 1.59 for the bell-shaped one). In both cases the area of increased wind speed extends down the lateral sides of the hills but the upstream and lee minima are much more pronounced for the cosine-squared hill, with extreme values of A = 0.71 and 0.67 respectively. Corresponding values for the bell-shaped hill are 0.93 and 0.87.

In the derivation of equation (9) it was assumed that the fractional speed-up, ΔS, would be approximately constant with height throughout the inner layer of depth, ℓ, defined by equation (5). The inner layer may be quite shallow; for example, with L=300m, and z_0=0.03m we would have ℓ=15.4m. In practice ΔS will usually be maximum at the surface but may decay significantly

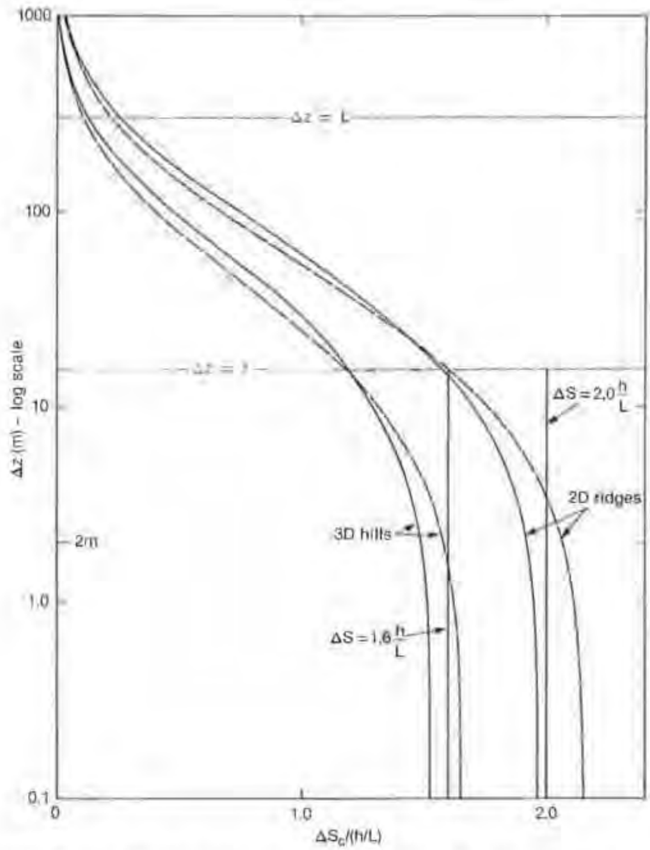

FIGURE 9 Profiles of fractional speed-up ratio, ΔS, above the crest of two and three-dimensional idealized hills.

———— Cosine squared hill.
……… Bell shaped hill.

through the inner layer, depending on the detailed shape of the terrain. On the basis of model computations and field studies we find that, as a general rule, ΔS will be approximately constant for $\Delta z \leq 0.2\ell$ and then decay more or less exponentially, like the inviscid flow solution, up to about 3L at which height it will approach zero. This is illustrated in Fig. 9 which shows the ΔS values above the hilltop for the circular cosine squared and bell-shaped hill computations discussed earlier and for two-dimensional versions of the same hill shape. The decay with height can be roughly approximated by

$$\Delta S(0,0,\Delta z) = \Delta S(0,0,0)\, e^{-A\Delta z/L} \tag{14}$$

where $A = 4$ for the 3D hills and $A = 3$ for the 2D hills. For 2D escarpments, we suggest $A = 2.5$.

Note that for the Kettles Hill field data, for which $\ell \sim 22$m for most of the data but is only ~ 15m for the 220° direction, the observed amplification factors (and hence fractional speed-up values) were essentially the same for $\Delta z =$ 3m, 6m and 10m. On the other hand at Askervein there was about a 50% variation in observed values of ΔS between the surface and the 10m level for some wind directions. The detailed shape of the hill concerned can have a major impact on the spatial changes in ΔS. Bowen (1983) gives a discussion of height variations in ΔU for simple 2D hill shapes.

5 THERMAL STRATIFICATION

Before discussing the application of the rules-of-thumb we should note that they have been developed for conditions of neutral thermal stability. They should work fairly well for near-neutral and unstable conditions but with moderate or strong stable stratification there may be substantial departures from the predicted values. The following discussion is intended to give a brief overview of some aspects of stratified flow over hills without attempting to provide quantitative guidelines. There are several mechanisms by which thermal stratification can affect the flow over complex terrain. Since we are considering local scales and moderate to strong winds we will ignore slope wind effects, but these will become important at low wind speeds or larger scales (say >5km). Let us consider the gravity-wave mechanism which will be primarily active in the outer layer. If a temperature sounding is available and the potential temperature gradient, $\partial\theta/\partial z$, is reasonably constant over a height range comparable with the length scale of the hill (L) then we can define a Froude number,

$$F_L = U_0(L)/NL \tag{15}$$

where the Brunt-Vaisala stratification parameter,

$$N^2 = \frac{g}{\theta}\frac{\partial\theta}{\partial z} \tag{16}$$

Hunt (1980), based on the work of Brighton (1977) and others, suggests that the near-surface flow is only affected by outer-layer stratification effects when $F_L<2$. For typical values, $U_0 = 6\text{ms}^{-1}$, $L = 300$m, this corresponds to $N>10^{-2}\text{s}^{-1}$ and $\partial\theta/\partial z>3°$C/km [$\partial\theta/\partial z>-7°$C/km]. Stratification effects, including a tendency to upstream blocking, upstream and downstream separation and the generation of lee waves are most pronounced for $F_L < 1$ — see Hunt (1980) or Smith (1979) for more details. There is an extensive literature on inviscid, stably-stratified flow over hills, especially two-dimensional flow over ridges, but work on stratified boundary-layer flow over hills is rather limited and mostly quite recent. Simple guidelines are not readily available but one should be aware that, to quote Hunt (1980), the 'speed-up over two-dimensional hills can be con-

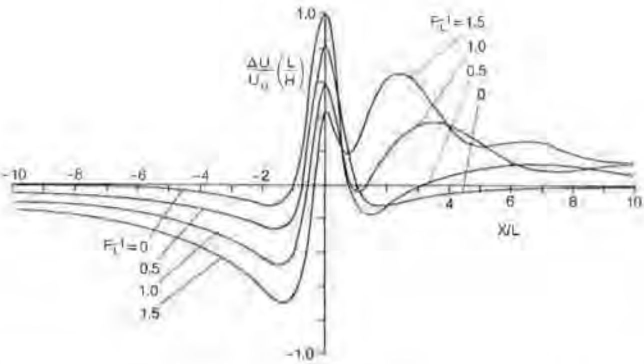

FIGURE 10 Surface velocity perturbations in stably stratified inviscid linearized flow over bell-shaped hills with small slopes. It is assumed that the upwind velocity and density gradients are uniform. The shape of the hills is $f=1/(1+(x/L)^2)$ (redrawn from Hunt (1981) see also Hunt *et al.* (1984)).

siderably greater in stably than in neutrally stratified flows'. One factor causing this can be the presence of an elevated inversion. If this is located at a height $< L$ and is sufficiently strong it could tend to restrict vertical motion at that level and cause inhanced acceleration of the flow as it passes through the 'gap' between the top of the hill and the inversion base. In other cases it reduces speed-up. Carruthers and Choularton (1982) have recently studied this type of flow over two-dimensional ridges. They show considerable variations in the behaviour of ΔS depending upon the stratification in and above the inversion and the height and thickness of the inversion layer.

For a deep, uniformly-stratified, surface-based stable layer Hunt (1981) shows relatively small departures from the neutrally-stratified results for the near-surface fractional speed-up ration at the hilltop, ΔS_{max}, provided that $F_L > 1$. However, as the stratification increases still further (i.e. F_L decreases), there is a pronounced change in the flow pattern as the strongest winds start to occur in the lee of the hill, as indicated in Fig. 10. These computations are for inviscid flow over 2D bell-shaped hills as indicated in the figure caption, but we would expect qualitatively similar behaviour in the boundary-layer, and for other hill shapes. Walmsley's recent calculations (private communication) have confirmed this.

For 3D hills an additional effect of strong stable stratification is to encourage air to flow around rather than over the hill. This is discussed in some detail by Hunt and Snyder (1980) and by Strimaitis *et al.* (1982) in terms of a critical streamline height, H_{CRIT}, defined such that all streamlines originating from below that level pass around the hill, while those starting at $z > H_{crit}$ can go over the hill. Hunt and Snyder (1980) find, from their laboratory studies that

$$H_{CRIT} \simeq h(1 - F_h)$$

where F_h is a Froude number $(U(h)/Nh)$ based on hill height.

An additional factor associated with non-neutral thermal stratification in the near-surface layer is the effect it has on the upstream profile $U_0(\Delta z)$. Bradley (1983) discusses this and finds that as a result of these effects ΔS can be reduced (by up to about 30%), relative to the neutral case, for unstable stratification and increased (up to 50%) with stable stratification. Theory and observation are in good agreement for unstable stratification while for stable stratification the observed changes in ΔS appear to be significantly less than Bradley's theoretical predictions.

6 ROUGHNESS-CHANGE EFFECTS

Boundary-layer flows above changes in surface roughness have been widely studied and have recently been reviewed by Hunt and Simpson (1982). We will consider only the simplest case of an abrupt, step change in surface roughness (from z_{01} to z_0) across a line normal to the flow. Three-dimensional effects associated with roughness changes are relatively minor and most changes are fairly abrupt (water to land etc.) so this is not too restrictive a limitation. If we are concerned with the turbulence structure within the internal boundary-layer downstream of a roughness change (see Fig. 11) or require especially accurate velocity predictions then a relatively sophisticated turbulence model will be needed. For rough estimates, however, we can do quite well with a very simple model based on that proposed by Elliott (1958). The essential features are that flow modifications are confined to an internal boundary-layer of depth δ_I (this notation is used to avoid confusion with h and ℓ already used) within which the velocity profiles is assumed to be of the form $U \propto \ln(z/z_0)$. Above this layer $U = (u_*/\kappa)\ln(z/z_{01})$ and, if we know the depth δ_I, we can construct the profiles as indicated in Fig. 12. All we need to know are the roughness lengths z_0, z_{01}, and δ_I as a function of x, plus our upstream measurement U_0 at height Δz_m. There are various formulations for δ_I – see for example Jackson (1976) or Hunt and Simpson (1982) – but for present purposes Elliott's simplest form

$$\frac{\delta_I}{z_0} = 0.75 \left(\frac{x}{z_0}\right)^{0.8} \tag{17}$$

will suffice. For typical value of x and z_0 this gives $\delta_I \simeq 0.1$ x. Note that z_0 is the downstream value of the roughness length. The velocity perturbation at height, Δz_m and distance, x, downstream from the roughness change is then ΔU_R which can be measured from a graphically constructed profile (on log-linear paper) similar to that sketched in Fig. 12. Alternately it can be calculated as:

$$\Delta U_R = U_2(x, \Delta z_m) - U_0(\Delta z_m)$$

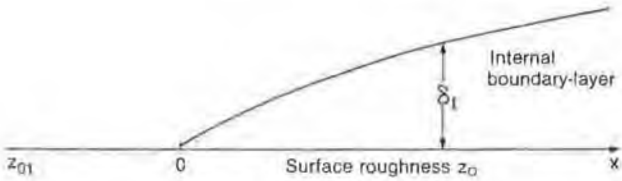

FIGURE 11 Internal boundary-layer due to roughness change.

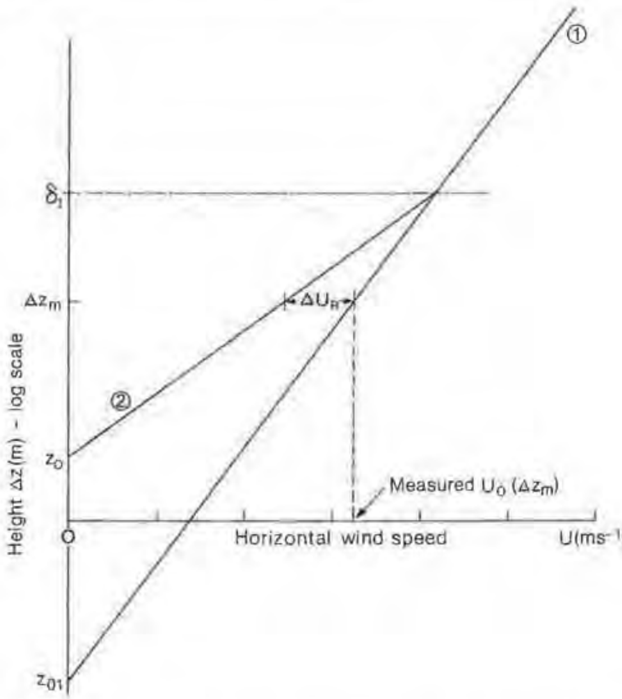

FIGURE 12 Construction to determine velocity perturbation, ΔU_R, due to a roughness change. (1) Upstream profile, (2) Profile in internal boundary layer. Smooth to rough case shown, but technique can also be applied for rough to smooth transitions.

where U_2 the velocity at $(x, \Delta z_m)$ within the internal boundary layer, is given as

$$U_2(\Delta z_m) = \frac{\ln \frac{\Delta z_m}{z_0}}{\ln \frac{\delta_1}{z_0}} U_0(\delta_1)$$

$$= \frac{\ln \frac{\Delta z_m}{z_0}}{\ln \frac{\delta_1}{z_0}} \cdot \frac{\ln \frac{\delta_1}{z_{01}}}{\ln \frac{\Delta z_m}{z_{01}}} U_0(\Delta z_m)$$

The velocity perturbation is thus

$$\Delta U_R \quad = \quad \frac{\ln \frac{\Delta z_m}{z_0}}{\ln \frac{\delta_1}{z_0}} \cdot \frac{\ln \frac{\delta_1}{z_{01}}}{\ln \frac{\Delta z_m}{z_{01}}} - 1 \quad U_0(\Delta z_m) \tag{18}$$

which will be negative for flow from a smooth to a rough surface and positive for the rough to smooth case.

7 PROCEDURES

Well, so much for the 'theory'; how can we apply it? We will consider first a situation that meets the requirements for the application of equations (10) and discuss some potential real terrain difficulties in the next section.

Let us suppose that we wish to estimate the wind speed near the top of a roughly circular hill of radius ~ 1 km in otherwise relatively flat terrain, and that we are located sufficiently far from the hill, maybe 3-4 km, to make a representative measurement of the wind speed and direction at a height Δz_m, in the undisturbed flow. With a hand held anemometer or portable anemograph this would usually be at 2m unless the terrain were wooded, in which case it would be essential to make measurements well above the canopy. This could possibly be achieved with a commercially available kite anemometer. The steps needed are simply:

(1) Measure mean wind speed, $U_0(\Delta z_m)$, and direction at a representative point - bear in mind that the estimates you are about to make are most appropriate for $U > 6ms^{-1}$ and should not be used for $U < 3ms^{-1}$, when thermal effects would probably dominate. If hilltop values are required at a height other than Δz_m it is probably best first to estimate U_0 at that height, by interpolation or extrapolation based on a logarithmic profile, and then proceed as if that were a measured value. This would require a good estimate of z_0 and, if

TABLE 5: Surface roughness lengths for typical surfaces.

Surface	Comments	z_0(m)
Ice	Smooth	$\sim 10^{-5}$
Water	Windspeed Dependent	10^{-5}-10^{-3}
Snow	Assumed Smooth	$\sim 10^{-3}$
Sand, desert	Dependent on grain size and presence of dunes or ripples	$\sim 3\times10^{-4}$
Bare Soil	Higher values if ploughed	10^{-3}-10^{-2}
Grass	0.2 – 0.1m high	0.003-0.01
	0.25 – 1.0m high	0.04-0.10
Agricultural Crops	Can be windspeed dependent	0.04-0.20
Typical Rural Area	Farmland with isolated trees & buildings	0.01-0.10
Orchards	May be seasonally dependent	0.5-1.0
Forests	—	1.0-6.0
Suburban Areas or Small Towns	Low Housing, Trees, etc.	0.10-2.0
City Centres	Buildings 10-50m high	1-10

Notes: (1) Based in part on data given by Oke (1978).
(2) For $z_0 \geq 0.1$m one should also use displacement heights for displaying the profile in logarithmic form.
(3) As a rough guide $z_0 \simeq 1/30$ to $1/10$ of the size of the roughness elements.
(4) Values for urban areas are somewhat speculative and the log law may not hold in some circumstances.

appropriate, the displacement height, d. A power law profile (see for example Counihan (1975)) could be used if preferred.

(2) Determine h and L for the hill, either from a detailed contour map or from visual estimates. Note that it is the value of L for the particular wind direction that is required, recalling that L is 'the distance from the hilltop to the upwind point where $z_s = h/2$'. The rules are intended for 0.1 km $\lesssim L \lesssim 2$ km and $h/L \lesssim 0.5$, not for major mountain ranges.

(3) Assess the value of the surface roughness length z_0, both on the hill and in the upwind terrain. Values for typical surface covers are given in Table 5. If hilltop and upwind values differ substantially (say be a factor 5 or greater) roughness change effects should be estimated (step 3a) and added to the terrain-induced wind speed perturbations. From the estimates of L and z_0 we can determine ℓ, either iteratively from equation (5), graphically from Fig. 2 or, very roughly, as $\ell = 0.05L$. If the measurement height, $\Delta z_m > 0.2\,\ell$, we will need to either reduce Δz_m or use equation (14) or Fig. 9 to adjust the estimate of ΔS following completion of step 6.

(3a) If there is a step change in surface roughness between the measurement point and the point of interest, construct a diagram similar to that shown in Fig. 12 or use equation (18) to determine ΔU_R, the velocity perturbation at height Δz_m. Input parameters required are

z_0, the hill roughness, z_{01}, the upstream roughness and x, the distance from the roughness change to the point of interest. Profile (1) can be drawn based on U=0 at $\Delta z = z_{01}$ and the measurement of U_0 at Δz_m. Then δ_I is calculated from equation (17) and, if $\Delta z_m < \delta_I$, profile (2) can be constructed with U=0 at $\Delta z = z_0$. ΔU_R is then determined as the velocity difference between the two profiles at $\Delta z = \Delta z_m$. Fig. 12 is for a step change from a smooth to a rough surface and ΔU_R should be taken as negative. The construction for rough to smooth cases is similar with ΔU_R positive. If there is a roughness change upstream of the measurement point, and it is suspected that roughness change effects could affect wind speeds at the measurement point and the point of interest differently, thus this procedure can be applied to both profiles to obtain an appropriate ΔU_R.

(4) If the flow is stably stratified, i.e. $\partial\theta/\partial z > 0$, make an estimate of N, (see equn.(16)) the Brunt-Vaisala frequency for $0 < z < L$ and calculate $F_L = U/NL$. If $F_L < 1$ the flow over the hill will be strongly influenced by stability effects, so these procedures are inappropriate. If $F_L > 2$ or thermal stratification is neutral or unstable, equations (10) should apply. For $1 < F_L < 2$, wind speed maxima on the hilltop will probably be slightly lower than those predicted by equations (10) but upstream profile considerations may offset this.

(5) Apply the appropriate form of equations (10) to determine ΔS at the hilltop. For a circular hill this would be $\Delta S = 1.6\ h/L$. If $\Delta S > 1.0$ treat the estimate with caution since there will probably be flow separation behind the hill. Bear in mind that the recommended range of application of equations (10) is for $h/L < 0.5$.

(6) The estimate of the wind speed at the hilltop, at the same height above the local terrain, Δz, is now.

$$U_H(\Delta z_m) = (1 + \Delta S)\, U_0\, (\Delta z_m) + \Delta U_R \qquad (19)$$

Fig. 13 displays these steps in a flowchart format which could be used for field operations.

8 POTENTIAL PROBLEMS

The first difficulty is likely to be with the terrain. Isolated hills in otherwise infinite flat plains are the exception rather than the rule. The case of generally 'rolling terrain' was discussed in section 3 while in other cases the best solution is probably to try to make the measurement of the undisturbed flow in as good a location as possible. Then, if necessary, one can make a subjective correction, based perhaps on the discussion in Section 4, to give an estimate of $U_0(\Delta z)$.

We should stress that the guidelines given here consider only the effects of the dominant terrain features in the area. Smaller scale humps, creek

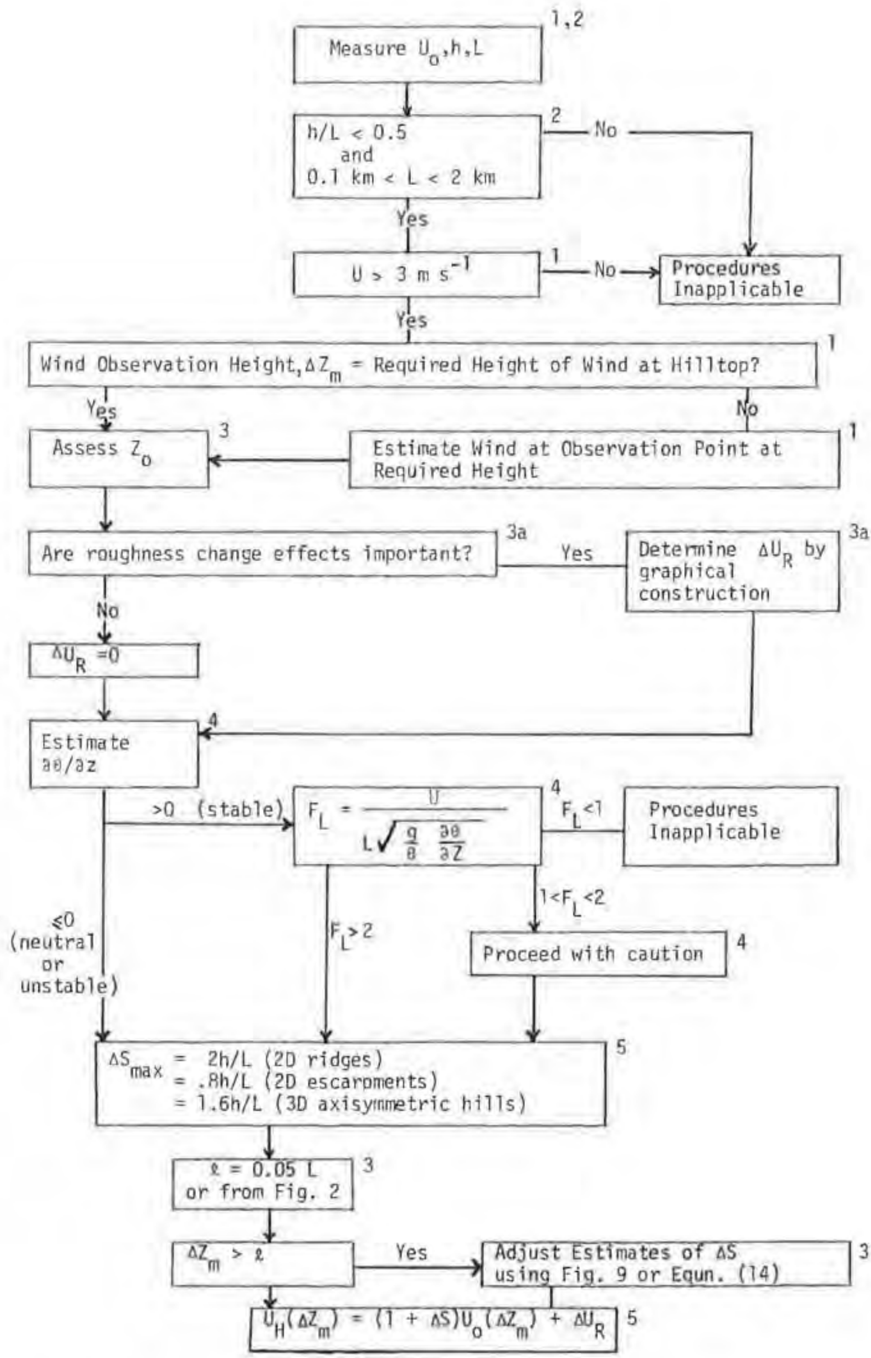

FIGURE 13 Flow chart for wind speed estimates.

valleys or other features superimposed on this terrain will also affect the flow, but on a more local scale. One must keep well clear of such features, as well as clumps of trees, buildings or anything else which would disturb the local wind field when measuring $U_0(\Delta z)$.

Determining the thermal stability is easier said than done. If a dropsonde or AIRsonde release can be arranged and monitored then a direct measurement of $\partial\theta/\partial z$ in the lowest levels can be made. If it is not available then we must beware of applying the rules under clear sky, night time conditions with light ($<6ms^{-1}$) winds. One could try to make an estimate of the appropriate Pasquill (1974) stability class, widely used in diffusion studies, and use the standard values of $\partial\theta/\partial z$ for those classes [0.02°C m^{-1} for class F(=6 if the numerical categories are used) and 0.035°Cm^{-1} for class F(7)]. These are, however, rather strong gradients which correspond to light wind conditions with very low values of F_L. ($F_L \simeq 0.4$ if $\partial\theta/\partial z = 0.02$°C$m^{-1}$, $U=3ms^{-1}$ and $L=300m$) and the range we are most concerned with falls within Pasquill's rather broad, near neutral, class D (4 and 5).

Deciding how to interpolate between circular and 2D hills and sometimes between hills and escarpments can be difficult. If the terrain feature under consideration does not fit into, or between, the categories, and if the situation has been anticipated far enough in advance, one solution would be to run the MS3DJH/3.1 model, either on idealized terrain with the same general characteristics or in a real terrain mode. The latter approach is also recommended if more detailed or accurate predictions are required for a particular hill.

In summary, users should be warned that estimating near-surface winds in complex terrain is a difficult and uncertain process and that the more complex the terrain, the larger the errors and uncertainties are likely to be. The subject is, however, an area of active research at the present time and improvements in our understanding and ability to estimate wind speed variations is to be expected in the future.

9 PRACTICAL APPLICATIONS

The procedures set out in Section 5 above, with due regard to the limitations enunciated in Section 6, have potential applications in a wide spectrum of applied engineering and "field" oriented activities. A far from exhaustive list could include: polutant dispersal, crop and forest spraying, hang gliding and kite flying, ski jumping, estimating snow drift, soil erosion, spreading of forest fires, heat loss and ventilation of buildings in complex terrain, sailing on lakes surrounded by complex terrain – and estimating the surface wind stress and circulation within such lakes. We have in fact been consulted on techniques for estimating wind speeds for most of the activities listed above and continue to be surprised at the range of activities where wind is an important factor. Some details of two sample applications are given below. This paper is based on a report prepared for the Department of National Defence in respect of the first of these.

(a) Drop-Zone Winds

Airdrop operations, both recreational and military, involve the parachuting of personnel and equipment into a target area known as a drop-zone. Of prime consideration in executing the "drop" is the wind, both aloft and near the surface.

While personnel can, to varying degrees, compensate for adverse winds, there is no means to steer equipment load drops during their descent. Navigational procedures are thus used to adjust for wind drift and form part of established routines during military drops.

However, the most critical wind is near the surface. Equipment and personnel can suffer injury and damage if their impact is too hard, as a result of high horizontal wind speed. Maximum allowable winds by day are 13 knots ($6.7ms^{-1}$) and by night, 9 knots ($4.6ms^{-1}$). For certain equipment drops surface winds up to 18 knots ($9.3ms^{-1}$) can be tolerated, but more usually, 13 knots is the upper limit.

When a large number of parachutists are involved in a drop, or many pieces of equipment must be delivered into a drop-zone, then an area up to several kilometers in length may be required. Although flat drop-zones are normally selected, instances arise where undulating terrain is encountered either by choice or necessity. On this scale, rolling terrain, as described in previous sections, can cause winds to exceed safe values. Normally, winds are manually measured in the drop-zone. The person doing so should be fully aware of the hazards involved in selecting a location to measure winds. Equipped with topographical maps, an estimate of thermal stratification and an anemometer, the procedures described could help a drop-zone official prevent an underestimation of winds and thus prevent injury and damage to personnel and equipment.

(b) Wind Engineering Applications

For wind engineering applications the instantaneous measurement of a wind speed would usually be replaced by climatological average or extreme wind data of the area. Such data are usually obtained from airports or other sites with relatively uniform and unobstructed terrain and the techniques described in this paper could be used to provide estimates of the adjustments that should be made to account for terrain-induced windspeed modification. This could be an important factor for estimates of wind loads on hilltop structures such as microwave towers (Force $\propto U^2$) or of the available energy for wind turbines (Power $\propto U^3$). In these cases the heights of primary interest may be somewhat greater than ℓ and the height adjustments discussed in Section 4 should be applied. Also if the climatological data is from a site a long distance (say $>$ 10km) from the point of application the effects of surface roughness differences may need to be taken into account.

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Possible effects of CO_2-induced climate change on the world food system: a review

Roger S. Pulwarty and Stewart J. Cohen

Global surface warming as a result of a doubling of atmospheric carbon dioxide is likely to reach an equilibrium of 3°C ± 1.5°C. This means that the probability of change in patterns of rainfall, wind circulation and temperature distribution undoubtedly will increase. A review of predicted climatic scenarios and the possible effects of these on the global food system is presented. Given the uncertainties of current climate forecasts for specific regions and future changes in global politics and economics, it is premature to predict the magnitude or location of changes in food production. However, work done now to reduce vulnerability to any climatic change could only prove to be beneficial to humanity.

La quantité d'acide carbonique ayant doublé dans l'atmospherè, le réchauffement de la surface du globe se stabilisera probablement autour de 3°C ± 1.5°C. Alors, la probabilité d'un changement dans la tendance des pluies, la circulation du vent, et la distribution de la température, augmentera. On passe en revue les scénarios climatiques et leurs influences possibles sur le circuit alimentaire mondial. Étant donné l'incertitude des prévisions du climat de régions particulières et les changements futurs dans la politique et l'economique mondiales, il est trop tôt pour prévoir l'importance ou l'emplacement des changements en production d'aliments. Néanmoins, le travail effectué maintenant pour réduire la vulnérabilité à tout changement climatique ne peut être que favorable à l'humanité.

1 INTRODUCTION

Carbon dioxide increase in the atmosphere is largely due to anthropogenically induced factors such as the burning of fossil fuels, deforestation and changing

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land use. Of all the gases contributing to the so-called "greenhouse effect", it is the most abundant. There is now almost universal agreement that the increasing carbon dioxide (hereafter CO_2) will cause a general warming of the earth's surface (Manabe and Wetherald, 1975; Walts, 1980). Recent estimates of the increase give a projected doubling (to 660 ppm) of atmospheric CO_2 by the year 2065 and a resultant $3°C \pm 1.5°C$ global temperature increase (NRC (U.S.) 1982; Herbert, 1983). Complications arise with changes caused by the increase. Variations in precipitation, atmospheric circulation patterns and seasonal temperatures are expected. The possible changes in climatic distribution and intensity will have impact on energy requirements, food production, fisheries, water resources, forestry, land use, transportation, tourism, health and human well being (Bach et al., 1980). The magnitude of temperature increase is however uncertain because of difficulties in modelling various feedback mechanisms, such as cloud cover (Hansen et al., 1980). In addition, a CO_2-induced warming could possibly be offset by a natural cooling trend.

Although there is some doubt about the magnitude and significance of a warming trend, its possible consequence, particularly with regard to the world food system, is so great that the possibilities must be entertained seriously. No other human activity is so influenced by weather and climate as the productivity of terrestrial biological resources, including food, feed, fiber, energy crops, range and forage crops and forests (Cooper, 1982). The prospect of climatic change from increasing atmospheric CO_2 must therefore bring into concern the "long term ability" of humans to feed themselves.

There are two categories of biological response to increased atmospheric CO_2. The first is a direct biological effect on photosynthesis and plant growth. The second is response on a spatial scale to climatic change (Baker and Lambert 1980). In regard to the first response, results indicating an increase in the net productivity of plants because of changes in photosynthetic and transpiration rates, and an increase in water use efficiency, have been well documented by many researchers (Strain, 1977; Ajtay et al., 1979; Baker and Lambert, 1980; Patterson and Flint, 1980; Wittwer, 1980; Rosenberg, 1981). The focus of our discussion will therefore centre on the second of the two responses. Direct reference to present models and predicted scenarios will be employed in order to obtain some idea of (i) variations in climatic distribution and effects on agriculture and natural vegetation and (ii) the possible impact on the global food system. The word "scenario" will be used as a deliberate distinction from a "prediction". A scenario can only provide a guide to patterns of change which could accompany a global warming (Wigley et al., 1980).

2 THE GLOBAL WARMING

The implications of an induced warming effect can only be presently assessed through the study of past climate and climate models. Paleoclimatic evidence, when viewed together with an understanding of physical processes, suggest that surface warming at high latitudes will be two to five times the global mean warming (Lamb, 1977; CLIMAP, 1976). Models predicting larger sensitivity at high latitudes relates this to feedback mechanisms (e.g. snow-ice albedo) and increased atmospheric stability, reducing circulation and therefore increasing surface warming. The models also predict accompanying regional variations in climate. These will be of great human significance. Precipitation patterns and temperature changes determine the location of deserts, fertile areas, marginal lands and the general soil moisture content and carrying capacity (Wigley et al., 1980). The beneficial effects associated with a warming trend will include an increased growing season, both in terms of heat and length.

Precipitation is dependent on both temperature and the global circulation pattern that transports water vapour, which together with regional factors determine the type of precipitation (rain or snow). Higher temperatures also result in higher open water evaporation rates (Kellogg, 1978). There must be, as Kellogg (1978) points out, a relationship between these circulation patterns and the large scale heat balance since they are both functions of the atmospheric "heat engine" in action. Indeed, it is expected that a mean decrease in the Pole-to-Equator temperature difference (gradient) as a result of greater increases in polar temperatures, could cause a decrease in windspeed across the globe (WMO, 1981).

3 IMPACTS ON AGRICULTURE

Plant Productivity

The effects on agriculture from a potential CO_2 induced climate change can best be estimated, if in turn, the available estimates for temperature and precipitation changes are reliable. However, since all of these requirements cannot be satisfied, it is presently possible to predict only the orders of magnitude of some potential effects (Jäger 1983).

Lieth (1976) and Box (1975) have, among others, developed procedures to test the correlation between environmental impact variation and vegetation functions. Their models predict the net primary production from average annual temperature and amount of precipitation. Table 1 demonstrates that the global net primary productivity may change as much as 5% with temperature changes of 1°C accompanied by a 6% change in rainfall. About one-third of this 5% is due to temperature changes and the remaining two-thirds due to rainfall changes (Lieth, 1976). It is precipitation which largely determines whether vegetation will thrive and whether or not agricultural practices are feasible (Kellogg, 1978). The changes and causes, according to Lieth (1976), differ for different latitudinal belts. For example, an 8% productivity change predicted in the 60°-70°N belt is mostly due to temperature, whereas the 5% change in the 0°-10°N belt is almost entirely due to precipitation. Lieth (1976) confirms that the significance of temperature effects tends to decline below 40° latitude.

It is reasonable to assume that these changes will effect not only net

TABLE 1: Change in NPP caused by changes in temperature and/or precipitation calculated from the "Miami Model" (Lieth, 1976)

Latitude Belt	$10^6 km^2$ Area Land	SMM NPP t/km²	−1°C Δ%	−6% Ppt Δ%	−1°C −6% Ppt Δ%	+1°C +6% Ppt Δ%
70-60N	13.3	387	8.0	0.3	8.0	11.0
60-50	14.7	645	5.8	1.2	7.0	7.0
50-40	16.5	804	2.8	3.2	6.0	5.6
40-30	15.6	860	0.7	4.4	5.1	1.5
30-20	15.2	553	1.1	5.0	6.1	4.9
20-10	11.4	894	0.3	4.7	7.5	7.0
10-0	10.1	1823	0.1	5.2	5.2	4.5
0-10	10.5	2024	0.8	2.3	2.5	2.2
10-20	9.5	1275	0.8	4.8	5.2	6.0
20-30	9.3	832	1.2	1.6	2.6	3.9
30-40S	4.1	856	0.6	4.2	4.8	5.6
Total Land	149.4	124.7 x 10^9t	122.8 x 10^9t	120.5 x 10^9t	118.45 x 10^9t	130.6 x 10^9t
%Δ			−1.6	−3.4	−5	+4.8

SMM = Standard Miami Model
NPP = Annual net primary productivity

primary productivity but also the borderlines among vegetation types or biomes. These boundaries will not only be affected by temperature and precipitation but by changes in soil moisture content as well (Kellogg, 1981). The migration of major agricultural crop zones across climatic gradients is therefore highly probable. The Manabe-Wetherald Model (1980) predicts that the zone between 37°-40°N will experience a reduction in precipitation and an increase in evapotranspiration. Kellogg and Schware (1981) thus force an adjustment of agricultural practices, which may mostly be beneficial. However, since every crop does not respond uniformly to climatic factors such as growing season temperature and length, as well as soil conditions and water availability, perspectives on this complex issue remain varied (Gribbin, 1983). Brinkmann (1980) suggests that the term '"growing season" provides a meaningful indicator relating macro-scale climatic variations to their potential impact on the micro-climate. For agricultural purposes, however, the indicator has to be more specific in order to incorporate different plant responses to temperature (Kellogg, 1978).

Kellogg (1978) makes an empirical deduction about growing season lengths in relation to mean temperatures. He describes a scenario for a 1°C rise in mean summertime temperature, corresponding to about a 10-day longer growing season. By A.D. 2000, a mean global rise of 1°C might mean a 2°C rise at 60° latitude. A model produced by the Goddard Institute of Space Studies projects much larger temperature increases. At 60°N, 100°W, these are 7.7°C in January and 3.1°C in July (GISS, 1984). This would approximately correspond to a 40-day longer growing season.

FIGURE 1 Projected shift of the American corn belt, based on a temperature increase of 3°C evenly distributed over the year, and no change in precipitation. The solid black line represents the current corn belt. Source: U.S. Dept. of Energy (1983).

An increase in photosynthetic rate, because of greater atmospheric CO_2 concentration, is likely to be enhanced by higher temperatures. In addition, a longer growing season could allow multicropping in some areas (e.g. planting of both winter and spring wheat varieties in the same year). Wittwer (1980) suggests that global warming could open up parts of the USSR for cultivation, and improve agricultural productivity in other subarctic regions of Canada and Scandinavia (Kellogg and Schware, 1981). However some present areas of productivity in these countries (e.g. Kazakstan USSR) could become drier (Bach, 1978).

Considerable attention has been given and concern has arisen for the North American grain belt's future productivity. Summer temperatures may become too high for optimal production of maize and soybeans forcing their cultivation northwards to areas of less fertile soil regimes. It has in fact been estimated that for each 1°C rise, the corn belt would shift approximately 175 km northeast (Rosenberg, 1982). Fig. 1 shows the simplest projection based on a scenario consisting of an increase in growing degree days only, with no change in precipitation (U.S. Dept. of Energy, 1983). The American corn belt could move partly into Canada and into an area mostly occupied by the Great Lakes. Similar dislocations could affect most of the major food and oil seed crops and commercial forest species. Many countries with marginal precipitation may experience shifts in rainfall patterns which would prove detrimental or beneficial to food production and natural vegetation, depending on location (Hall et al., 1979).

Productivity Of Northern Soils

Given the prediction that a 1°C rise will facilitate 10 extra growing days, it seems evident that the greatest restriction on northern agriculture (i.e. the length of growing season) may to some extent be relieved. According to Flohn (1980) the shores of the Arctic Ocean were forested 2.4 million years ago and could be so again. Northern warming to the extent suggested by Flohn would create temperatures suitable for farming in some areas.

The restrictions placed on this apparently substantial benefit lies in the limitation of precipitation and in the lack of suitable soil. Even though Flohn (1982) postulates some increase in precipitation in what will be former polar desert regions, whether or not this will be adequate remains to be seen (Cooper, 1982) or more preferably remains to be forecast. The soils however are of most critical concern. Most minerals and organic matter has been leached away by podzolization (Jenny, 1941, 1980). Agricultural production is severely limited in Alaska, Northeastern Canada, some parts of Siberia and large parts of Scandinavia, because of thin soils and rugged glaciated topography (Cooper, 1982). While the clay belt (particularly the Peace River Valley in Alberta, and parts of Siberia) may appear to benefit from the changes, poor surface drainage in these areas, especially in the northern parts, will require that regional surface drainage action be undertaken (Stewart, 1984). It also appears that while a warming trend may be beneficial to northern areas, the actual productivity of these seem limited and the net global productivity may well be offset by decreases in production in the south, because of droughts and floods (Stewart, 1984).

Water Resources

Water table levels vary according to rain and snow input, and the outflow of evaporation, transpiration, and runoff. A change in precipitation or a change in watershed losses to the atmosphere will alter runoff available for human use. Stream runoff, according to Cooper (1982), may be the most sensitive of the economically important responses to CO_2-induced climatic change. Climate models are limited in their ability to determine runoff response of specific river basins. Despite this and other shortcomings, at the very least, they can provide some indication of change in the processes involved (Cooper, 1982).

Model simulation by Manabe, Wetherald and Stouffer (1981) suggest that precipitation increases are likely to occur in subtropical and tropical areas. These could result in increased flooding of low lying areas, particularly as a result of heavier, summer monsoon rains over India and south-east Asia. Present problems with precipitation minus evaporation deficits (P-E) between 15° and 35° latitude can be expected to be alleviated somewhat. It is expected, however, that problems with summer dryness would increase between 35° to 50° latitude, with the greatest soil moisture deficits being in the latitude belt 37°-47°N. The projection of dryness in mid-latitudes is of particular concern to North America. Growing areas in the southwestern U.S.A. will experience a decrease in streamflow from their principal water sources. Meanwhile, in the northwestern U.S.A., the Columbia river basin could well receive an increased water supply even though it already possesses sufficient streamflow (Cooper, 1982).

Agriculture and population centres in water deficient areas can be negatively affected by decreased streamflow, and increased climatically induced water demand. Idso (1982) counters this by arguing that future circulation patterns will not only help existing agriculture around the world, but could also lead to enhanced productivity in relatively infertile and arid lands, which now lie idle, thereby greatly increasing the areas of the globe available for cultivation. He also notes that as CO_2 concentrations increase, growth rates and yield of plants could actually double, with water use being cut to a fraction of its present value. Semi-arid lands not now suitable for cultivation could be brought into profitable production with the additional water supply expected. What Idso does not make clear however is the location of possible sources of nutrients for sustaining plant life, nor the scale and regional distribution of these "beneficial effects".

Present agricultural and silvicultural practices in most countries do not allow optimum yields and regeneration rates. Alkalinity, salinity, water logging, domination of weeds, diseases and pests are a few of the reasons cited for this (Kovda, 1981). Even if new types of irrigation methods are introduced, Kovda (1981) believes that the development of new plant varieties, fertilizers and reasonable crop rotation, must be coincident with any new developments. The major problem with this however is that most countries likely to be affected cannot afford to implement these adaptive strategies. More will be said on this later. While some of the adaptive measures may nevertheless be covered by human intervention, the natural vegetation, a major sink for atmospheric CO_2 (WMO, 1981), will have to fend for itself. This has been ignored in our great uncontrolled experiment, in which mankind plays an important part. There is an urgent need to study the climatic and economic sensitivity of the world's major water supplies. Reassessment of laws and policies regulating water allocation will be necessary if climatic change increases the demand deficit. An example, cited by Cooper (1982), would be the diversion of water to urban areas from agricultural lands.

4 RESPONSE OF NON-CULTIVATED ECOSYSTEMS TO CO_2

The American Association for the Advancement of Science in 1980 convened a panel (Revelle, 1980) consisting of an impressive collection of scientists from many disciplines. In an attempt to provide a worldwide research agenda, partly to assess risks and benefits to the world food system from a CO_2-induced climatic change, the panel outlined some of the possible responses in the areas of non-cultivated ecosystems. While the impact of the research agenda has been somewhat limited, the information provided about such a poorly studied area is still useful.

The fact that most large scale agricultural crops are managed as mono-clutures means that all healthy individuals in a stand will have a fairly uniform response to CO_2 increase. Systems that are not cultivated have numerous competitive species, which will respond differently to this change. Early growing, cool season plants are generally more responsive than those plants with their primary growth in summer. Seedling emergence is also a function of CO_2 concentration. The differential responses can lead to changes in species competition in forests and rangelands (Revelle, 1980). The degree of human intervention is apparently the distinguishing factor between cultivated and non-cultivated ecosystems, given that the actual biological processes in these two environments do not vary significantly. The low level of manipulation means, in natural resource systems, less opportunity for economic benefit from climatic change than in agriculture. The AAAS panel correctly concluded that present management practices are insufficient to optimise favourable changes which may occur, or to minimize adverse consequences of unanticipated changes, which could prove costly (Revelle, 1980; Loomis and Gerakis, 1975).

About a quarter of the U.S. is still forested. According to the AAAS panel, these range from small areas of intensive high yield silviculture to economically unproductive woodlands. The larger forests in developing countries around the world are already under severe stress from cultural practices such as overcutting and the pastoral lifestyle. These may succumb to the stealthily encroaching desert condition. Climate change may serve to exacerbate this type of stress and may also affect mid-latitude North America, where reduced soil moisture contents are expected. The ability of both theoretical and applied ecologists to predict consequences to forest ecosystems from a CO_2-induced climatic trend seems as yet to be inadequate. While it is known that both population dynamics and rates of ecosystem processes will be affected, the lack of proper model synthesis and basic research are mainstay problems in the making of accurate predictions (Revelle, 1980). Full evaluation of climatic change will require increased emphasis on "spatio-geographic" modelling in contrast to point scale modelling, including the adequate incorporation of anthropogenic factors (Kellogg, 1978). Kellogg and Schware (1981) point out that while the distribution of biomes are climate dependent, human intervention in the next 50 to 100 years will have greater ecological significance in most parts of the world.

5 IMPACTS ON THE GLOBAL FOOD SYSTEM

Several aspects of the global food system can be directly or indirectly affected by a CO_2-induced climatic change. One of the major indirect effects occurs through changes in land use. In many regions of the world, ecosystem dynamics are greatly influenced by land use practices, through clearing, reforestation, conversion of grazing land to agriculture, farm abandonment, and grazing and utilization practices (Revelle, 1980). Provided there are social and economic incentives, areas once limited in agricultural capability could be cleared and used. The con-

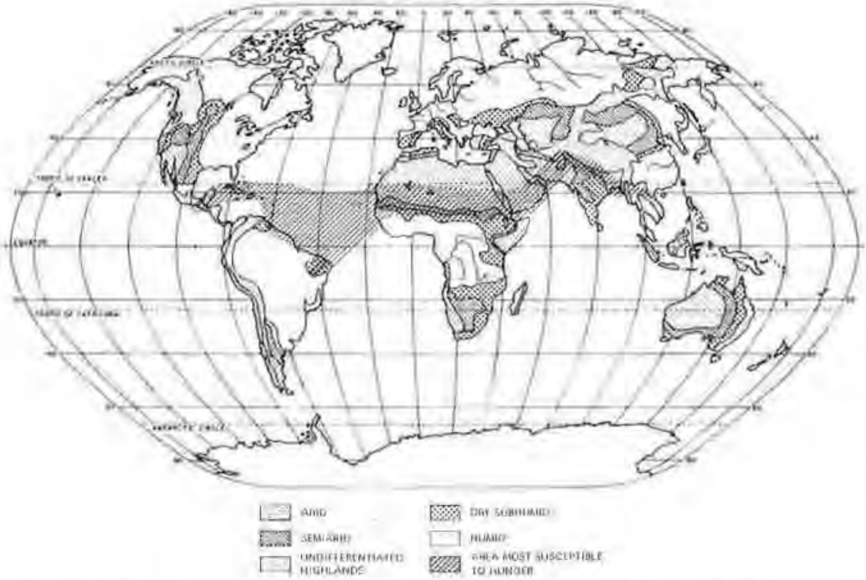

FIGURE 2 Region most susceptible to significant hunger in the future because of the interaction of climate and societal factors. Adapted from Dando (1980). Climate regions base map adapted from Hall et al. (1979).

verse condition, according to the AAAS (1980) panel, could result in extensive abandonment of agricultural land. The effect in marginal agricultural regions would be more pronounced, as was the case of drought in the Sahel 1968-1972 when over 100,000 people lost their lives (Croat, 1972; Rogers 1977). Agricultural expansion could occur at the expense of natural ecosystems, even though the expansion may not prove to be extremely productive. Changing rates of land use are indicators of the complex interplay of climatic, economic, social and demographic factors. More research will be needed in predicting future landscape patterns resulting from the interplay of climate and social forces in specific regions (Kellogg and Schware, 1981).

Already in our past, population shifts have been at least partly due to adverse climate. Kellogg (1981) illustrates this point using fairly recent examples, such as the abandonment of farms in the Great Plains of Nebraska, Kansas, Texas and Oklahoma during the dry periods of 1890's, 1910 and the 1930's, and the Irish Potato famine of 1845-1851 (which was in fact the main reason for Irish emigration to the U.S. and Britain).

The developing nations (of this era) are probably especially vulnerable to the pressures of climatic variation because of population increase. An estimated three-quarters of the world's population will be in developing countries by the year 2000 (Mitchell, 1977). Future famines, according to Dando (1980), will last longer and cover larger areas than ever before. They will not only be the result of an area's vulnerability to climatic change, but will directly depend on the characteristics of human societies at different levels of development, i.e. societal resilience (WMO, 1980). It seems obvious, as Dando points out, that the world's future food problems will occur in areas of low "societal resilience" i.e. the developing nations (Fig. 2). It is necessary that changes in traditional behaviour occur to some extent, in order to avoid further reduction of the land's carrying capacity and greater dependence on food and financial aid from relief agencies (Walsh, 1984). In this way, it is hoped that increased poverty and starvation will not perpetuate themselves as deterministic factors in culture and tradition (Lewis, 1966; Sneden, 1970).

Regarding direct effects on the global food systems, it is evident that some aspects of the projected biological and climatic change, as a result of a CO_2-enriched atmosphere, could be beneficial. The relationships between climate and productivity for a number of major food crops are known. However, every food crop responds differently to a given climatic change. Present studies are severely limited in terms of their ability to assess resilience and rigidity of ecosystems against anthropogenically caused climate variations. This seems to be most significant in subtropical marginal areas (Oram, 1982). Frequent and severe pest and disease outbreaks are extremely important to food production in a changing climate. Pests are presently responsible for up to 25% of agricultural losses. A temperature increase may make pest control even more difficult than it already is, since plants may be exposed to pests not previously encountered, and a warmer, longer season will increase the probability of major outbreaks. The same may be true for plant diseases e.g. stripe rust on winter wheat. A warmer earth could mean an increase in latitudinal distribution of pests and diseases to which specific crop plants have no resistance mechanism of their own (Kellogg, 1981). Even with our most hardy crop strains, the results of agricultural technology are far more vulnerable to climatic variations and change than the natural ecosystems they replaced. Our deliberate damage or elimination of natural ecosystems are rapidly decreasing stocks of wild strains and disease resistant crops. In a world where 95% of human nutrition is derived from less than 30 different kinds of plants, and just three crops (wheat, rice and maize) account for over 75% of our cereal production (Kellogg, 1981), this realization seems to and should be one of the most disturbing.

The attempt here, it must be noted, is not to present a case for an imminent doomsday scenario, but to make evident the fact that the beneficial effects of atmospheric CO_2 on plant productivity and growing season length do not necessarily provide a linear correlation with world food production. Regional shifts in climate could reduce food production and increase hunger in some nations facing unfavourable social and economic conditions (Cooper, 1982). However, given the uncertainty of current climate model forecasts and dominant socio-economic factors, it is premature to predict the magnitude or location of changes in productivity, although at least a better idea of the direction of change is available.

At current food production levels, supplies are adequate to feed the world's population even in lean years (Reutlinger, 1978), but even if world productivity does increase, it appears likely to happen in areas already well provided for. The probable benefits of a warmer world must be weighed against "the present practical achievements of the world food production and distribution system which falls (too) far short of what is in theory, possible" (Gribbin, 1983). It would seem idealistic and naive to assume that the world's poor and disadvantaged will not still suffer. Malnutrition and starvation, it seems, has less to do with the overall scarcity of food and more to do with its grossly uneven distribution and unfair exploitation practices globally and within nations (Bigman and Reutlinger, 1979; Murdoch, 1980). The scientists of the International Institute for Applied Systems Analysis's (IIASA) Food and Agricultural program provide a harsh accurate reason, if not solution for this. They insist that "we are part of a highly complex robust system which is malfunctioning and which is extremely resistant to change, even to improvement" (cited in Cooper, 1982). The existing distribution of income and wealth among individuals and nations prevents the free market economy from regulating the capricious uses of food (e.g. grain dumping) at the same time that people are starving (Bigman and Reutlinger, 1979). These are the same people who are resigned to face the consequence of droughts and floods, and who, to a great extent, are not responsible for the causes of catastrophe. Insurance against hunger will require significant financial assistance (Bigman and Reutilinger, 1979), but given the lack of an effective global policy making process (Glantz, 1979), this adaptive strategy seems more suited to developed nations than to developing ones.

A real and immediate problem concerns risk perception and decision making (Revelle, 1980). This involves the way in which people of different cultures perceive scientific and other information, and make decisions regarding adaptation or other responses to climatic change. Concern in this area lies not only with lay people, but more importantly, with the "experts" who provide information on climatic changes, their effects and societal response, and with the policy makers at all levels of society who decide how to act upon the information (Meyer-Abich, 1980). Given the 50- to 100-year time lag between the present and the expected appearance of the effects of atmospheric CO_2 increase (NRC, 1982), responses are not seen as being immediately required.

6 CONCLUSION

The authors seem to assume a generally pessimistic view of the CO_2 problem, for it is a view supported by most researchers cited here. This type of speculation can only result when little else in the form of alternatives is being offered. Idso (1980) attempts to counter the arguments, but even now his questions of doubt are being seriously challenged (Crane, 1981; NRC, 1982).

Howeve1, one must always consider that models may be proven to be wrong or marginally relevant. Other severe weather occurrences, anthropo-

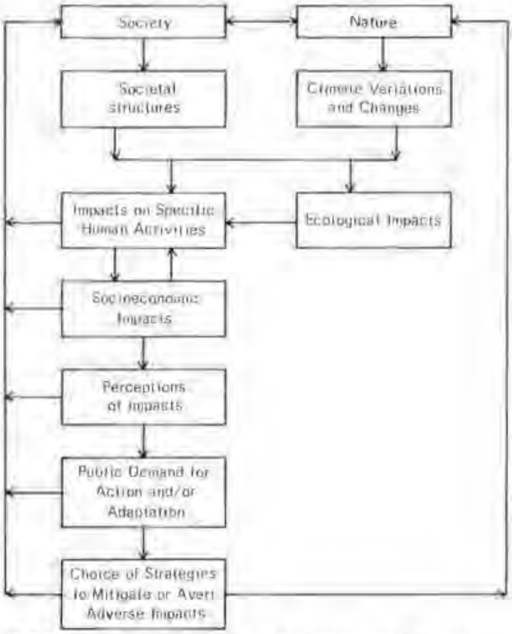

FIGURE 3 The interconnected components that are involved in climate impact studies (see Kellogg and Schware, 1981).

genically or naturally induced, may impede the predicted warning or change present forecasts, on a regional-specific basis. Much inter-disciplinary work is required. Many of the present recommendations to provide definitions of environmental and societal consequences are weakened by a lack of a sense of immediacy and proper identification and scaling of priorities. These fail in technique to transcend traditional disciplinary boundaries of the climate-society system (Chen, 1981). Work done now to counteract a perceived threat (in this case) could only prove beneficial to humanity, should the threat not materialize to the extent predicted. Vulnerability to any climatic change could in any event be decreased.

Studies must involve the physical, biological, sociopolitical and economic systems under which we function (Fig. 3) and new means of integrating research results for use in policy making decisions must be discovered. Above all, these studies must assume a human dimension. Predictive models which include "vulnerability" and "resilience" factors will question the very way of life of societies (Timmerman, 1981). The values of parameters of such models are not universally standard. The successful institution of change, can only be brought about through education, oriented toward the perspective of the people concerned. This seems to be true regardless of "level of development." Modelling of this type requires much greater efforts at interdisciplinary communication for the purposes of gathering and interpreting data, and it may provide the beginnings of a sufficient solution. It is necessary therefore that the problems be understood before data are gathered (Eddy, 1983) and be continuously reassessed during data gathering. Perhaps, contrary to Dylan[1], we do need a weatherman just to known which way the wind blows.

The CO_2 problem, and its effects on world agriculture and vegetation, appears to be increasingly imminent. The final answer lies with nature, but our ability to cope depends on the integrity of policy makers and the honesty of those who implement the policies. Swaminathan (1982) cautions that we know less about human perception and behaviour than about natural processes. In this case, the authors must conclude that we know little of either.

ACKNOWLEDGEMENTS

We would like to thank the referees for their comments and suggestions.

[1]Dylan, Bob (1966) Rainy Day Women No. 1235. Dwarf Music, New York.

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News & Comments

COMMENTS ON "*DES ERREURS SYSTÉMATIQUES DANS LES DONÉES CANADIENNES DE LA DURÉE D'ENSOLEILLEMENT.*"

Reading the paper of Hufty et Theriault (1983) has left me also "avec stupéfaction" that such major systematic errors in sunshine records could have been overlooked by our national meteorological service. For the benefit of those who may not have read the article, the authors pointed out discrepancies in many sunshine recorder records due to the "horizon effect" (shading by objects on the horizon), and to shading by the projecting ends of the Casella equinoctial sunshine cards.

Since the article was published, over a year ago, I have been searching each issue of the Bulletin for a rebuttal to, or at least an explanation of, the criticisms. The fact that no reply has appeared, at least up to now, has prompted me to offer a few comments, and pose a few questions, on the subject.

To begin with, the equinoctial card shading effect is well documented in the A.E.S. publication: SUNSHINE – Manual of Standard Procedures for Obtaining Sunshine Data (1974). Quoting from this manual (p.17) – "At the times of the equinoxes, when the straight cards are used and the sun rises very nearly in the east and sets very nearly in the west, the edges of the card will shade the sphere resulting in a loss of record on a clear day. To overcome this observers are instructed to bend or trim the overlapping ends of the equinoctial cards. The ends of the short and long curved summer and winter cards need not be bent or trimmed."

In view of the above, we must conclude that this type of error is due either to negligence on the part of the observers, or to shading that remains even after the edges of the card are bent over. If due to the former, perhaps an easy solution would be to make sure that the cards are trimmed to the correct length to begin with.

With regard to the horizon effect, one should keep in mind that shading by hills is a *real* effect, and thus should be corrected for only with caution. Shading by trees, buildings, etc. may vary with time, and requires constant vigilance on the parts of both weather observers and users of weather data.

I sincerely hope that these observations will serve some purpose in helping to solicit a reply, to the comments of Hufty et Theriault, from the A.E.S., so that we may once again use their published data with complete confidence.

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FRIENDS OF CLIMATOLOGY MEETING

The 1984 Friends meeting, arranged by Scott Munro, was held on April 13-14 at the University of Toronto's Erindale College, Mississauga, Ontario. Ken Hare of Toronto gave a brief history of this "anti-organization", from its founding in 1968 to the present. Once a year (most years), climatologists and others with an interest in climate, from Southern Ontario and Southern Quebec, gather to discuss various topics. Hare's address, "Canadian Climatology in Perspective", was very broad, touching on climatology as a "non-discipline", and the important new effort in studying impacts. He noted that few people are actually doing climate impact studies, but that governments are providing seed money, so hopefully the research effort will expand.

The remainder of the meeting included some excellent reviews of Atmospheric Dynamics (Charles Lin, Toronto), Radiation Modelling (John Davies, McMaster), Carbon Cycle Modelling (Kaz Higuchi, AES Downsview), and a geophysicist's view of the ice ages (Richard Peltier, Toronto). In addition, there were presentations on research activities at Windsor (Marie Sanderson) and Guelph (Terry Gillespie, John Wilson, Marty Heckenheimel, Rob Place), and a brief presentation by Lin Hanzhong from China on agrometeorological problems in China.

The 1985 meeting will be held at Queen's University in Kingston, Ontario, probably in late spring. Contact Professor McCaughey at the Geography Department for details.

The 1984 AAG Annual Meeting took place on April 22-25 in Washington, D.C. There were 13 climatology sessions, and similar to the 1983 program, climate modelling was a popular topic, as was applied climatology and synoptic climatology. Several papers in these areas were also presented in poster sessions. In addition, there was a large number of sessions on water resources, hazards, energy, medical geography, biogeography, and geomorphology, which included papers of interest to climatologists. Other climatology sessions included presentations on water budget, air pollution and radiation balance.

The featured speaker, invited by the specialty group, was Donald Gilman of the U.S. National Weather Service, who provided an overview on the methods and difficulties of extended forecasts. Considerable attention has been given to winter temperature forecasts, because of the significant impacts of recent severe winters on the transportation and energy sectors in the U.S. In general, forecasters use lag correlations between fall vs winter geopotential height anomalies, and fall vs winter air and sea surface temperature anomalies. Forecasters also look for teleconnections between sea surface temperature anomalies or height anomalies, and temperatures over the continent, so that if heights in the Asian Arctic could be predicted, for example, a forecast could be given for temperature in Southern British Columbia. However, there are very few statistically significant teleconnections and correlations, so forecasters also look for precedents in past records, particularly the exceptional cases, such as previous El Niño years. Finally, the four forecasters in Gilman's group issue a joint forecast by combining the methods of statistical and historical analyses. Skill appears to be higher for temperature than precipitation. There are also spatial differences with the forecasts for western and southeastern sectors showing higher skill than those in the central region of the U.S.

The climatology specialty group elected John Oliver of Indiana State as Chairman. An executive committee was formed, consisting of Anthony Brazel of Arizona State, James Burt of Wisconsin, and Cort Willmott of Delaware. The 1985 AAG meeting in Detroit will include several sessions on climate impact assessment and climate-society interactions. These are being organized by William Riebsame of Colorado and Diana Liverman of Wisconsin.