Roughness-dependent geographical interpolation of surface wind speed averages

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SUMMARY
A two-layer boundary layer model is presented for regional assessment of seasonal-average surface wind climate from station data. In the surface layer the wind measurements are objectively corrected for imperfect station exposure, using gustiness-derived roughness values for each azimuth sector. This produces at 60 m a mesoscale wind speed, $U_m$, representative for a $5 \times 5 \text{ km}^2$ area block. The regional averaging of $U_m$ in the Ekman layer is done by geostrophic similarity methods, referring to a mesoscale block roughness obtained from maps through terrain classification. All wind data for the flat Netherlands are analysed, using a 30-year series of annual windness indices for time-span adjustment of non-simultaneous data series. The evaluation leads to objectively interpolated seasonal surface wind maps, supplemented by wind distributions and extreme design wind speeds. Possible extension of the methodology to complex terrain is discussed.

1. INTRODUCTION

The analysis of wind climate from station measurements has long been hampered by the absence of a systematic method to account for small-scale terrain effects. Any study of shelter effects will show that the presence of near-by buildings or trees—even at the WMO-authorized distance of 10 obstacle heights or more—can decrease local wind speed by 10% to 25%, depending on the situation. Therefore data from (partially) sheltered stations will not represent the regional wind adequately.

Traditionally there were two attitudes to this problem: the first being that wind speed data are inherently of low quality. Many synoptic meteorologists for this reason use only the azimuth component of station wind data, e.g. to recognize positions of fronts, relying exclusively on pressure fields for quantitative flow speed information. The second classical attitude is to ignore the problem and to accept the original measured wind speeds. This has reduced the value of many potentially useful investigations. For example, the much-quoted study by Richards et al. (1966) on land–water wind differences is difficult to interpret quantitatively, because from neither the paper nor its references is the degree of sheltering of the used land station traceable.

The first necessary step towards quantification of small-scale wind–terrain interaction is proper parametrization of the terrain aspect. Originally this was done empirically by way of the so-called power law wind profile, which suffers from the absence of physical interpretability. A major advance was the matching theory by Blackadar and Tennekes (1968). They showed that the logarithmic wind profile was not (as thought before) a feature only of the lower few metres over homogeneous terrain, but, rather, was a consistent description of any surface layer wind field up to heights of 30 to 100 m. Consequently, the roughness length, $z_0$, is the optimal parameter for specifying terrain effects on wind, better than e.g. the geostrophic drag coefficient, which depends on stability and pressure gradient as well as on terrain (Wieringa 1981).

In the seventies, various systematic studies investigated wind–terrain relations. Some investigators applied Blackadar and Tennekes’ geostrophic similarity relations and averaged the wind over very large horizontal distances (Garratt 1977; Petersen et al. 1981). Others used the surface layer part of the similarity relations, the logarithmic wind profile (Tennekes 1973), and worked on a smaller scale (Simiu 1973; Wieringa 1976; Smith and Carson 1977; Duchêne-Marullaz 1977; Bietry et al. 1978). The application of

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these similarity laws requires a certain degree of horizontal homogeneity at the considered scale, so for regions with many steep elevation variations special approaches were necessary (e.g. Murakami and Komine 1983).

In this study a two-layer model will be presented, which was developed for terrain-dependent wind climate study in the Netherlands. This model also requires differentiation of the roughness information. For the surface layer we need to evaluate the azimuth-dependent roughness at the wind stations, as 'seen' by the station anemometer. For the entire planetary boundary layer (PBL) we require an 'effective roughness' (Fiedler and Panofsky 1972) at a larger scale. The similarity laws are then applied separately for each layer.

It is shown that the resulting modelling of seasonally averaged surface wind climate is sufficiently accurate for practical application in non-mountainous terrain, as in the Netherlands. Other developed applications of roughness-corrected wind climate analysis are regional evaluation of wind frequency distributions and estimation of extreme design wind speeds. Possibilities for extending the method to more complex terrain are discussed.

2. DETERMINATION OF SURFACE ROUGHNESS AROUND WIND STATIONS

Sheltered station wind measurements, however precise, are inaccurate in the sense that they represent the wind climate only at the point where the station anemometer is located. Such an anemometer 'sees' in different directions terrain of varying roughness, caused by obstacles, vegetation, etc. The relevant descriptive parameter, the roughness length, $z_o$, has traditionally been obtained from measurements at several levels on a mast—however, weather stations with suitable masts are a rarity. Even if a mast is available, it is nearly impossible to obtain representative values of $z_o$ from short-mast profiles if the surroundings are not homogeneous, because then the standard flux-profile relations no longer apply (Peterson and Busch 1978; Wieringa 1981; Beljaars et al. 1983).

Since the turbulence level, $\sigma_u/U$, increases with increasing roughness, $z_o$ can also be derived from the (single-level) station wind measurements themselves by way of a gustiness analysis (Wieringa 1973). Such an analysis is already practicable for a station if its routine wind data contain maximum gusts, $u_{\text{max}}$, as well as the average wind speed $U$ and the direction at the time of the gust occurrence (Wieringa 1976). If at the station anemometer height, $z_s$, for some upwind azimuth sector the median value of the observed gust factor $\langle G \rangle = \langle u_{\text{max}}/U \rangle$ is known, then a gustiness-derived sector roughness length $z_{og}$ can be estimated from

$$z_{og(\text{sector})} = z_s \exp \left( - \frac{Af_T \left( 1.42 + 0.3 \ln(-4 + 10^3/U_t) \right)}{(G)_{\text{sector}}} - 1 + A - f_T A \right),$$

Here $A$ ($\sim 0.9$) is the attenuation of $u_{\text{max}}$ by the anemometry, and $f_T$ is a factor which is unity for 10-minute averaging periods and increases to 1.1 for hourly averages. $U_t$ is the average wavelength (wind run) of maximum gusts observed by the given station combination of anemometer and recorder, and varies usually between 50 and 100 m. Roughness lengths and other length-dimension parameters, such as $U_t$, are given in metres unless other units are specified. For derivations and for ways to determine $A$, $f_T$ and $U_t$ from instrumentation specifications we refer to previous publications (Wieringa 1973, 1976, 1977, 1980, 1983; Oemraw 1984b).

Beljaars (1982) has shown experimentally, that a roughness length $z_{og}$ derived from measured gustiness at $\sim 10$ m is a good descriptor for average wind profiles above twice the height of surrounding obstacles, in his case for a layer above 20 m. Brooks (1961) and Dyer (1963) have already shown that the adaptation of stress to surface features may
require fetches of >100 layer heights—which would mean here that $z_{os}$ values would represent an upwind terrain fetch well over 2 km. This agrees with our own experience on distances over which terrain features are still noticeable in $z_{os}$ values (Wieringa and Rijksoort 1983). For instance, if on going upwind we first encounter 2 km of open field and then a forest, the $z_{os}$ value will significantly exceed open-field roughness. But if we look along a runway of 3 to 4 km length towards a town, then the observed $z_{os}$ will be a runway roughness. These figures agree with the experience of Caton (1977) that surface gustiness is determined by an upwind fetch of ~3 km.

Such experimental results fit in with the notion that in the PBL $c_u$ is largely determined by large eddies with horizontal dimensions of several times the PBL height, resulting in long adjustment fetches for the horizontal streamwise component of turbulence (Panofsky et al. 1982). As a result, the roughness value obtained by a gustiness analysis is an area-integrated parameter. It takes into account both nearby and far-off obstacles, vegetation etc. in the upwind direction within a distance of ~3 km over a sector of 20° to 30° width.

There may be significant differences between summer and winter terrain roughness, e.g. due to seasonal changes in sparse deciduous vegetation, to crops, or to terrain-smoothing by snow. In the Dutch situation a separate roughness analysis is routinely made for the leaf-growing season (May–October) and for the winter season.

Analysing gustiness-derived $z_{os}$ values for many hundreds of azimuth sectors in all kinds of landscapes (Wieringa and Van der Veer 1976) gave an excellent opportunity to gauge the typical roughness values of various land-use classifications. It was found (Wieringa 1981) that the majority of published roughness classifications do not adequately cater for the very common situations with inhomogeneous rough terrain. A probable reason is that many published field roughness data are based on measurements on short masts situated in untypically homogeneous surroundings. In addition, field mast evaluations are biased towards too low $z_{os}$ values if they use a von Kármán constant of 0·35 instead of 0·4. Classifications based on such data will hardly feature roughness lengths greater than 0·1 m, even though larger-scale studies have shown that effective roughness commonly has values of 0·1 to 0·5 m (e.g. Fiedler and Panofsky 1972; Wamsler and Müller 1977).

Good agreement was found between gustiness $z_{os}$ values and the terrain classification of Davenport (1960), which for rough terrain is based on data from sufficiently high masts. Since this Davenport classification was originally specified by way of power law exponents and thus is not compatible with similarity analysis, it was rewritten in abbreviated form (Table 1) in terms of roughness length (Wieringa 1977, 1980).

<table>
<thead>
<tr>
<th>Class</th>
<th>Terrain description</th>
<th>$z_{os}$ (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Open sea, fetch at least 5 km</td>
<td>0·0002</td>
</tr>
<tr>
<td>2</td>
<td>Mud flats, snow; no vegetation, no obstacles</td>
<td>0·005</td>
</tr>
<tr>
<td>3</td>
<td>Open flat terrain; grass, few isolated obstacles</td>
<td>0·03</td>
</tr>
<tr>
<td>4</td>
<td>Low crops; occasional large obstacles, $x/H &gt; 20$</td>
<td>0·10</td>
</tr>
<tr>
<td>5</td>
<td>High crops; scattered obstacles, $15 &lt; x/H &lt; 20$</td>
<td>0·25</td>
</tr>
<tr>
<td>6</td>
<td>Parkland, bushes; numerous obstacles, $x/H \sim 10$</td>
<td>0·5</td>
</tr>
<tr>
<td>7</td>
<td>Regular large obstacle coverage (suburb, forest)</td>
<td>1·0</td>
</tr>
<tr>
<td>8</td>
<td>City centre with high- and low-rise buildings</td>
<td>?</td>
</tr>
</tbody>
</table>

Adapted from Davenport (1960) in $z_{os}$-form by Wieringa (1980)

Notes: Here $x$ is a typical upwind obstacle distance and $H$ the height of the corresponding major obstacles. For class 7, the applicability of logarithmic exposure correction modelling is doubtful, and requires at any rate the substitution of $(z - d)$ for $z$ in the formulae, where $d$ is the displacement length ($d \sim 0·7 \times$ average obstacle height; see Brutsaert (1975)). Class 8 is analytically intractable and can better be modelled in a wind-tunnel.
The actual estimation of terrain roughness at locations without anemometers by way of Davenport's classification has proved viable. Class assignment is unlikely to be more than one class wrong, implying that $z_0$ is known within a factor of two. This corresponds to an exposure uncertainty of $\pm 6\%$ in surface wind speed (see section 3), not much worse than the net uncertainty resulting from the much more troublesome method of estimating and evaluating multiple upwind roughness changes (see e.g. Petersen et al. 1985). Gustiness-derived roughness is still better, but gustiness data are not available in every country.

3. LOCAL-SCALE EXPOSURE CORRECTION OF STATION WIND MEASUREMENTS

(a) Methodology description

The consequence of terrain roughness variation with azimuth is that the relation between station-measured wind and regional wind climate becomes also azimuth-dependent. As a result, the original station wind measurements are not sufficiently representative of the regional wind climate, unless we remove exclusively local terrain-induced wind influences—e.g. wind shelter produced by an observer’s house in one direction and by trees in some other direction. Our final aim is a regional area-representative wind speed.

Sheltering obstacles and patches of uncommonly high roughness are sinks for momentum, so they will produce in the downwind direction a region of limited extent where the stress is relatively high, and wind speed relatively low. However, these changes become less significant with increasing height, and model calculations on the wakes of patchy roughness (e.g. Smith 1967; Duijm 1983) indicate that the relative stress anomalies decrease to $\sim 10\%$ of their surface values at heights between 50 and 100 m, depending on the situation. In that layer, subsequent roughness change effects will not be recognizable individually, and an overall stress profile will exist representing the roughness of a large area. Munn and Reimer (1968) observed in a non-homogeneous location, that at 10 m the turbulence varied with azimuth according to local obstructions, while at 60 m the levels of turbulence seemed to be independent of wind direction.

From an analogy between upward diffusion of pollution and momentum loss, Pasquill (1972) estimates theoretically that over typical country the wind structure at 50 to 100 m represents the interaction of large-scale wind with an upwind surface region of $\sim 5$ km length. Experimentally, Pretel (1984) concludes from mast data that wind profiles measured in the lowest 80 m are influenced by obstacles to a maximum distance of 5 km. This matches well with the height-fetch ratio of order 100 needed for stress adaptation, as mentioned in section 2.

Consequently, our working assumption is that retardation plumes of roughness patches in a $5 \times 5$ km$^2$ block merge at 60 m height, where we need only consider their joint retardation influence on the regional wind. At this height (called ‘blending height’ by Wieringa (1976)) the wind will not vary significantly over the block, in spite of roughness variation. We call the wind speed at that height the mesowind, $U_m$.

The mesowinds can be determined from the station-measured wind speed $U_s$, observed at a height $z_s$, by way of applying the logarithmic wind profile between $z_s$ and 60 m as follows:

$$U_m = U_s \ln(60/z_0)/(\ln(z_s/z_0))^{-1}. \quad (2)$$

The effective upwind roughness length used is best obtained from gustiness analysis of strong winds from the considered azimuth sector. A working alternative is the use of the Davenport terrain classification (Table 1).
From a given value of the mesowind we derive a working value for the surface wind, namely the wind speed $U_p$ at 10 m over open country with $z_o = 0.03$ m—which is for practical purposes the largest average surface wind which is observed over land for a given pressure field. We call $U_p$ the potential wind speed. Logarithmic profile calculation similar to Eq. (2) shows that $U_p = U_m/1.31$. The transformation of $U_s$ into $U_p$ by this procedure is named 'exposure correction', because it converts measured winds with varying upwind roughness into corresponding wind speeds over hypothetical nearby open terrain.

When the surface boundary layer is not neutrally stable, the actual wind profile above 30 m can deviate markedly from the logarithmic profile, Eq. (2). This does not induce significant errors into the exposure correction, because the deviation in the upward transformation is compensated by a similar deviation in the downward transformation from $U_m$ towards $U_p$. However, such compensation operates only if the upward and downward transformation height intervals are comparable, which makes exposure correction less reliable if $z_s > 25$ m.

The concepts of mesowind and potential wind are illustrated in Fig. 1. Here it is also indicated that the logarithmic relationship, Eq. (2), between $U_s$ and $U_m$ is valid only in the matching layer (Tennekes 1973). This implies the restriction that the observation height, $z_o$, should exceed $20z_o$. Over dense regular roughness with average height $H$ (forests, homogeneously built-up areas) the profile reference level is raised above ground level by a displacement height $d \sim 0.7H$ (Brutsaert 1975). At heights less than $(d + 20z_o)$, close to obstacles and roughness elements, the wind field is too inhomogeneous for modelling at this scale.

![Diagram](image_url)

**Figure 1.** Examples of surface layer wind profile curves (m s$^{-1}$) over various terrain situations with roughness length $z_o$ and displacement height $d$, when at a nearby meteorological station the measured wind corresponds to a potential wind speed $U_p = 10$ m s$^{-1}$ (with mesowind $U_m = 13.1$ m s$^{-1}$). Interrupted profile curves indicate the height range where microscale wind variations make average wind estimates highly unreliable.
(b) Validation and practical application

The reliability of exposure correction has been checked in various field experiments by comparing wind stations within a few kilometres of each other. The most extensive check was made at Cabauw (Wieringa 1980), comparing a 10 m mast 300 m downwind from the edge of an extensive wooded and built-up area, with a 1800 m distant 10 m mast with 2 km open fetch for the same wind direction. The gustiness-estimated ratio of the wind speed at these sites was 1.09, and actual observations showed a ratio of 1.10±0.02 for \( U \geq 3 \text{ m s}^{-1} \), even for clear days and nights.

Other validation checks were made. At Vlissingen, potential wind data from a 24 m mast at the edge of the town were shown to be practically equivalent to potential wind data from an unsheltered 10 m high mast at the water edge, 400 m distant (Wieringa 1983). Also for the stations De Bilt and Soesterberg, 7 km apart in a large wooded region, uncorrected strong wind observations differed by up to 40% for some sectors, while potential wind speeds were found to be comparable within 10% and to have similar climatological statistics (Wieringa 1977; Rijkoort and Wieringa 1983).

Application has not been restricted to Dutch stations. For Heathrow airport, a gustiness-estimated 10% retardation of wind speed for a sheltered sector was found to tally with local experience (Wieringa 1980). Christoffer and Jurksch (1985) have applied gustiness analysis to data from 200 German stations, and give interesting illustrations how this can show the effects of changes in station surroundings such as building activity or tree growth, and how corrections can be made for the effects of station relocation on climatological series.

The above-mentioned validations pertain to approximately level country, and it remains to discuss the effect of minor elevation changes around the station. For example, a hill-top surface wind will be stronger than would be implied by the mesowind, but its gustiness will also be lower than would be compatible with the surface roughness (e.g. Frenkiel 1963), and this relative decrease in gustiness results in an exposure correction modification in the required sense. For example, a gustiness analysis for 10 m wind masts located on \( \sim 4 \text{ m} \) dikes indicates an estimated velocity increment of 5 to 8% at anemometer level, depending on the dike slope (Wieringa and Van der Veer 1976). This is a plausible speed-up value in view of present knowledge (Bowen 1983; Baker 1985).

We conclude that, for climatological purposes the surface wind changes caused by minor terrain elevation variations should be reasonably well accounted for by an exposure correction using a gustiness-derived roughness. Working, instead, with Davenport class estimates of roughness Logue (1975) allowed for minor local elevation effects by, e.g., decreasing the class number by one for over-exposed stations, with satisfactory climatological results.

The general conclusion is, that for wind speeds \( \geq 3 \text{ m s}^{-1} \) and observation heights \( < 25 \text{ m} \) a roughness-dependent exposure correction gives reliable estimates of average wind over specified roughness in the neighbourhood. This implies that the application of Eqs. (1) and (2) results in a block-representative mesowind estimate from station observations, even when the station is significantly better or worse exposed than its surrounding region (within reasonable limits).

Therefore, we correct for terrain roughness influences on our station wind speed values by azimuth-dependent transformations to open terrain, using the effective roughness obtained by gustiness analysis of each azimuth sector. The nomogram in Fig. 2 is a graphical version of Fig. 1 for practical use in the transformation of wind speeds between different heights and upwind roughnesses, representing the \( U_p \) extension of Eq. (2):
Figure 2. Nomogram for finding potential wind speed $U_p$ (10 m above open terrain with $z_o = 0.03$ m) from observed wind speed $U$ at height $z$ above nearby terrain with specified roughness length $z_o$ ($z_o$ value indicated in parentheses). Conversely, when $U_p$ is known the nomogram allows estimation of wind speeds in the neighbourhood at arbitrary moderate heights above arbitrary terrain with given $z_o$. Broken lines at the left-hand side indicate that below $z \sim 20z_o$ the transformation is of doubtful validity (Wieringa 1977, 1980).

$$U_p = 0.76 U \ln(60/z_o)/\ln(z_o/z_o)^{-1}. $$

The nomogram has a non-dimensional wind speed scale and a logarithmic height scale, and profile lines for seven Davenport roughness classes.

The potential wind speeds, which are obtained after exposure correction, are really surface equivalents of mesowind values referring to a 5x5 km$^2$ block at a height of 60 m. For wind climate mapping in this context we have converted the hourly station wind speeds to potential wind speeds, and then averaged these over seasons and over the year.

4. MESOSCALE ROUGHNESS MAPPING

We have to consider now the regional interpolation of the obtained seasonal averages of the local mesowind. Stations are typically at least 30 km apart, and much can happen to the wind between them. Whether simple interpolation is possible depends on the regional terrain homogeneity. Any overall transformation in the landscape, such as change from a generally hilly and wooded region to a flat region dominated by meadowland, implies that potential wind data obtained at one side of the change are not well applicable at the other side. On the other hand, in a region where the overall aspect of the terrain varies little, we may expect that potential wind climate data from a local station will be representative over a large area.

At this point it is necessary to introduce well-specified scale definitions. We will speak about local scale when discussing the wind field and the roughness within the area over which an integrated roughness evaluation is provided by a station gustiness analysis. From previous sections, it follows that this scale typically refers to an area of about 5x5 km$^2$. The next larger scale is obtained by averaging horizontally over the largest distance to which a gust analysis of roughness applies, which means that we use areas of about 5x5 km$^2$ as units (‘block’). The variation of block wind averages over a larger region will be called mesoscale variation of wind climate.

We wish to determine the friction-caused retardation of mesoscale wind, and at this horizontal scale, boundary layer analysis should cover the entire PBL. Analogous to the surface layer formalism, where the local stress is determined from the average wind at 10 m and surface roughness, we can determine the mesoscale stress value from $U_m$ if we have a corresponding mesoscale roughness, $z_{om}$. Fiedler and Panofsky (1972) call this an
effective roughness: "the roughness length which homogeneous terrain would have in order to produce the correct space-average downward flux of momentum near the ground, with a given wind near the ground." We have to derive $z_{om}$ from available information on the various local roughness situations occurring across a block.

Essentially one wishes to average turbulent shear stress, or equivalently the square of the corresponding friction velocity, $u_*$, across the block. This is directly related to wind speed at a reference level $z_r$ and to the roughness length, by the relation

$$(u_*/U_{*r})^2 = [0.41/\ln(z_r/z_0)]^2 = C_d(z_r)$$

where $C_d(z_r)$ is the drag coefficient. Obviously $C_d$ is the parameter of which an appropriate average should be obtained, not the roughness length. Therefore we will calculate block-averages of $C_d$, but for consistent presentation we will show corresponding roughness lengths, $z_{om}$.

Various studies (e.g. Wooding et al. 1973) have invoked the principle of drag partition, which essentially means that within an array of roughness elements the various drag coefficients should be additive. Such additivity has been assumed here too in averaging $C_d$ values corresponding to roughness values occurring across a block. An interesting feature of $C_d$ averaging is that larger roughnesses in a block are weighted more heavily, due to the squaring operation. This implicit bias is in accordance with the experience gained in roughness step change analysis (see Jackson 1976). There often the largest roughness is used as a representative value, because adaptations to rough–smooth changes require much longer fetches than smooth–rough changes.

The drag coefficient reference level for these calculations was taken as 10 m, in accordance with general usage. With hindsight it might have been better to take $z_r = 60$ m because $U_m$ also refers to that level. However, the resulting difference is that average $z_{om}$ values calculated with $z_r = 10$ m are larger than those which would have been obtained with $z_r = 60$ m by about 5 to 10%, reaching 20% if the roughness variation in the block is very great. In comparison with other uncertainties in the procedure this roughness bias is not grave.

Information on the mesoscale roughness variation across the Netherlands was produced by analysis of topographical maps of scale 1:100 000. Available maps allowed us to distinguish 11 classes of land use, varying from sea to high woods and low-rise or high-rise built-up areas. To each class a characteristic roughness length was assigned in accordance with the Davenport roughness classification (Table 2). For each $5 \times 5 \text{ km}^2$ we

<table>
<thead>
<tr>
<th>Land-use category</th>
<th>$C_d(10) \times 10^3$</th>
<th>$z_0$ (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sea (minimal fetch 5 km)</td>
<td>1.5</td>
<td>0.0002</td>
</tr>
<tr>
<td>Small lake, mud flats</td>
<td>3</td>
<td>0.006</td>
</tr>
<tr>
<td>Marshland</td>
<td>5</td>
<td>0.03</td>
</tr>
<tr>
<td>Pasture</td>
<td>7</td>
<td>0.07</td>
</tr>
<tr>
<td>Dunes, heath</td>
<td>8</td>
<td>0.10</td>
</tr>
<tr>
<td>Agriculture</td>
<td>10</td>
<td>0.17</td>
</tr>
<tr>
<td>Road, canal (in Dutch landscape, tree-lined)</td>
<td>12</td>
<td>0.24</td>
</tr>
<tr>
<td>Orchards, bushland</td>
<td>16</td>
<td>0.35</td>
</tr>
<tr>
<td>Forest</td>
<td>25</td>
<td>0.75</td>
</tr>
<tr>
<td>Residential built-up area ($H \leq 10$ m)</td>
<td>35</td>
<td>1.12</td>
</tr>
<tr>
<td>City centre (high-rise buildings)</td>
<td>50</td>
<td>1.6</td>
</tr>
</tbody>
</table>

Based on classification by Davenport (1960) and Wieringa (1980).
evaluated land-use classes covering at least 10% of the block area. Then composite block roughness lengths were calculated by adding the drag coefficients of these classes, weighted by area, and recomputing from the total drag coefficient a surface roughness length $z_{ob}$, representative for the total block.

On a local scale such an evaluation of roughness due to vegetation and building would suffice, but at mesoscale the presence of hills also becomes important for the interaction between wind and terrain. At this scale the average elevation of the terrain above sea level is not important—a high plain has not necessarily a larger mesoscale roughness than a low plain. However, recurrent elevation variations become important additional flow modifiers when they exceed typical obstacle heights, say 10 m.

For evaluation of elevation-variability effects in the context of this analysis we determined per block the largest terrain elevation difference, $dH$. Figure 3 shows the distribution of this elevation variability parameter over the country. For the sake of completeness the figure also contains isolines of block-averaged absolute elevation above sea level. Elevation variability appears to be a small factor over 90% of the Netherlands.

The resulting $dH$ values were converted into ‘elevation variability roughness lengths’, $z_{ob}$, by a dimensional relation used in more mountainous countries (Smith and Carson 1977):

$$z_{ob} = 0.2 (dH)^2 / L.$$  

(3)

A discussion of the validity of the Smith–Carson formula is given by Deardorff et al. (1984). For our flat country, with $dH < 60$ m nearly everywhere, substitution of $L = 5$ km provided sufficiently plausible results of the same magnitude as obtained by others (e.g. Wamser and Müller 1977; Guyot and Seguin 1978; see also Carson 1986). In section 8 below, a possible alternative method for handling moderate orography is discussed.

Finally for each block the surface roughness, $z_{ob}$, and the elevation variability parameter, $z_{ob}$, were compounded as drag coefficients. A summary map of the resulting mesoscale roughness, $z_{ob}$, is given in Fig. 4. One should be aware that this map was made to evaluate shear stress averaged over a $5 \times 5$ km$^2$ area from the wind speed at 60 m. In this it differs from similar maps by Smith and Carson (1977) and Van Dop (1983), which were made with coarser grids ($\geq 10$ km) and were used for single-layer calculations over the entire PBL.

5. Transformation between mesowind and macrowind

The basic hypothesis for the exposure correction analysis in section 3 is that at the average top of the surface layer, at about 60 m, the wind $U_m$ does not vary on the local scale but only on the mesoscale. In the same way we can expect that at the top, $h$, of the PBL the wind varies only on a still larger scale, the macroscale. The wind speed at height $h$, when inferred from surface wind measurements, will be called the macrowind, $S_h$. Since $S_h$ contains implicitly the effects of mesoscale thermal and topographical gradients, it is not necessarily equal to the synoptic geostrophic wind speed, $G_h$, which is derived from macroscale pressure gradients. Elsewhere (Petersen et al. 1985), $S_h$ is called the effective geostrophic wind.

A basic relation between $U_m$ and wind at geostrophic level is given by PBL similarity theory (Blackadar and Tennekes 1968), valid for a horizontally homogeneous PBL. In such a boundary layer the near-surface layer is characterized by a wall law, with the turbulent friction velocity, $u_*$, and the roughness length, $z_o$, as variables. The wind at the top of the PBL has veered in direction with respect to the direction of $u_*$ through a
certain angle, the magnitude of which is related to the rotation of the earth and is accounted for by the Coriolis parameter, \( f \). At the latitude of the Netherlands \( f \approx 1.1 \times 10^{-4} \). The two components of the wind at the top of the PBL, \( U_h \) in the \( u \) direction and \( V_h \) in the orthogonal horizontal direction, are then, according to similarity theory, given by

\[
U_h = \left( \frac{u_*}{\kappa} \right) \left( \ln \left( \frac{u_*}{z_0 f} \right) - A \right)
\]

(4)

\[
|V_h| = \left( \frac{u_*}{\kappa} \right) B.
\]

(5)

The relations contain three empirical stability-dependent constants. For the purpose of evaluating seasonal average wind we assume neutral stability, and take the values of Deacon (1973), namely \( \kappa = 0.41 \) for the von Kármán constant, and \( A = 1.9, B = 4.5 \) (the lower limit of Deacon’s uncertainty range) for the other two constants.

In Eq. (5) \( V_h \) changes sign with the hemisphere, but that is not relevant here since
we restrict our interest to the macrowind

$$S_h = (U_h^2 + V_h^2)^{1/2}. \quad (6)$$

The $U$ component of the wind at height $z$ in the PBL is given by

$$U_z = U_h + (u_*/\kappa)\{\ln(zf/u_*) + A\} \quad (7)$$

and substitution into this relation of $z = h$ gives us the PBL height

$$h = u_*/(f e^A). \quad (8)$$

It is interesting that this formal interdependence of the $h-(u_*/f)$ ratio and the choice of the constant $A$ is not more often invoked.

A major result of similarity theory is the validity of the simple logarithmic wind profile near the surface over a height interval of 0.1 to 0.2$h$ (Tennekes 1973). In this layer the $V$ component can be neglected. We have used this already in the potential wind transformation formula, Eq. (2).
We will here describe the PBL wind situation based on assumed knowledge of the mesoswind $U_m$ and the mesoroughness length $z_{om}$. Since the relevant height of 60 m is situated in the logarithmic surface layer, we may derive the mesoscale friction velocity, $u_{*m}$, by way of the logarithmic profile formula

$$u_{*m} = \kappa U_m \left( \ln(60/z_{om}) \right)^{-1}. \quad (9)$$

Substitution of $u_{*m}$ and $z_{om}$ into Eqs. (4), (5), (6) and (8) gives us the $S_h$ value at specified $h$ corresponding to given values of $U_p$ and $z_{om}$.

This calculation can be inverted. When $S_h$ and $z_{om}$ are known, then we can solve for the unknown $U_h$, $V_h$ and $u_{*m}$ from the three equations (4), (5) and (6). We can then derive $U_m$ by substituting $z = 60$ m in Eq. (7). Hence $U_p = 0.76U_m$ is known when macrowind and mesoroughness are known.

The application of these similarity formulae requires formally a homogeneous PBL with neutral stability. The homogeneity requirement relates both to the terrain and the horizontal pressure gradient. The latter implies that strong horizontal temperature gradients are not admissible, because these give rise to a change of the pressure gradient with height (baroclinicity). The method is therefore unlikely to be applicable on the coastline, both because of the strong changes in surface aspect and because of the possibility of land–sea temperature differences. On the other hand, sufficient homogeneity is quite likely inland over a 5 x 5 km$^2$ block, if we have characterized the surface by a mesoscale roughness parameter.

In any single hour the stability is very seldom neutral over the full height of the PBL. This implies that the similarity approach used, with constant values of $A$ and $B$, is not applicable to the calculation of the wind at more than a few decametres above the surface in any single hour—instead, we then must use stability-dependent similarity profile formulae (see discussion, section 8).

However, our purpose here is the evaluation of seasonal wind averages, and for that case it is probable that an average near-neutral stability applies. Moreover the problem becomes less severe, because a double transformation is made from the surface to height $h$ and back. On the same grounds as for exposure correction (see section 3) we can then assume that systematic profile deviations during the upward transformation are compensated by equal deviations during the downward transformation over the same height interval. If we transform up and down along the same wind profile curve, we must necessarily reach again the same profile point.

6. Mesoroughness-dependent interpolation of potential wind

For the description of the wind climate of the Netherlands an analysis was made of long data series of hourly averaged wind from about 40 meteorological stations (Wieringa and Rijksoort 1983). Below we explain how the geographical distribution of seasonally averaged wind speeds across the country was evaluated from these data.

The previous sections have provided us with tools to interpolate the station wind measurements systematically and objectively into a map of seasonal-average wind speed. Both the terrain variations and the macroscale climate differences can be accounted for. The latter can be due to various reasons, e.g. the distance to the sea, or preferential tracks of depressions. The method is executed in five steps, which are summarized in the flow-chart, Fig. 5.

(a) Station wind evaluation and time-span adjustment of series

Our analysis started with a check of the data quality and an evaluation of the
roughness around the various anemometer locations (Wieringa 1983). For ten major stations this information was compiled into a station history report (e.g. Oemraw 1982, 1984a). At the end of this time-consuming task, all used hourly averaged wind data were considered accurate to within 5%, and exposure corrections per sector were known. The effective local roughness length, \( z_0 \), required for this was obtained by a gustiness analysis for the majority of stations, sometimes supplemented by Davenport roughness class estimation per azimuth sector.

The result of this first analysis step was a mesowind, \( U_m \), at the top of the surface layer, at 60 m, representative for a 5×5 km\(^2\) block around the station. For practical application at heights of 10 to 20 m the mesowind was transformed to a potential wind speed \( U_p \), referring to 10 m height over hypothetical open terrain. Such \( U_p - U_m \) pairs were obtained for each hour.

As a second step full year and seasonal averages of mesoscale wind speed were calculated, represented by their surface equivalent \( U_p \). Two-month seasons were used, because the three-month astronomical seasons are too inhomogeneous with respect to wind-climate. For instance, the 'official' autumn (21 September to 22 December) contains both a quiet September and a very windy November. Gaskell and Morris (1979) show that the general circulation in western Europe can be divided into six reasonably homogeneous two-month periods. For these seasons and for the year we have therefore computed average potential wind speeds. In addition, the seasonal shape of the wind speed distribution at the station was summarized by fitting a Weibull distribution over the medium speed range (4−16 m s\(^{-1}\)) and evaluating the distribution shape parameter, \( k \).

Some years are windier than others, and this makes regional comparison of station series difficult if the series are short (say, three to six years) and not observed in the same period. Therefore we determined for each year between 1951 and 1980 for each station the windiness ratio, i.e. the annual average potential wind speed \( U_{\text{year}} \) at the station,
normalized through division by the long-term average at the same station. Windiness ratios of many stations were then averaged per year—this was done in steps, first using only series longer than 20 years, then adding series of at least 12 years to increase the statistical mass. We used 381 years of data from 20 stations for this analysis.

The results are the regional windiness ratios in Table 3. The standard deviation of the mean, \( \sigma/\sqrt{n} \), was 0.016 for the first decennium and 0.010 for the period 1961 to 1980, due to varying numbers of usable series, therefore the three-figure accuracy of Table 3 is warranted. It was possible to obtain for each station from the annual \( U_p \) averages an overall equivalent 30-year average by correcting each year according to this table. For series lengths of at least five years the time-span adjustment error becomes \( \leq 0.1 \text{m s}^{-1} \).

(b) Macrowind analysis and derivation of wind climate map

The third step was the calculation of the seasonal average macrowind, \( S_h \), at the top, \( h \), of the PBL above selected basic wind stations. The relevant roughness for this computation is the mesoroughness, \( z_{om} \), of the station surroundings. Because often stations were eccentrically located within a block, we used not only \( z_{om} \) of the 5 x 5 km\(^2\) block in which the station was situated, but also \( z_{om} \) of the adjacent blocks, all in all, therefore, over an area about 8 km radius. In this block-averaging, double weight was given to roughness in the south-westerly direction, because that is the preferred wind direction at all seasons in our country. Computation of \( S_h \) was done using the similarity equations (4)–(6).

The fourth step was the determination of the smoothed macrowind field in the season, or year, over the Netherlands. This serves to describe the large-scale wind climate differences across the country. From the smoothed field we can obtain an \( S_h \) value for every 5 x 5 km\(^2\) block in the country, irrespective of whether it contains a wind station.

The fifth step was the computation of block values of the local seasonal average mesoscale wind \( U_m \) (and the corresponding \( U_p \)) from \( S_h \) and the block value of \( z_{om} \), using again the similarity formulae. This fifth step (as well as the third step) is an application of the similarity drag law, requiring no discussion beyond the comment given in section 5. However, we need to look closer at the fourth step, the geographical smoothing of the macrowind.

If any station is to provide a reliable macrowind value, two criteria must be satisfied. Primarily, the surroundings must be reasonably homogeneous within 5 to 8 km distance; therefore coastline stations were not admissible as basic stations. Secondly, a reasonably

<table>
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homogeneous distribution of stations across the country was aimed at, in order that the final smoothing result would not be influenced unduly by regions with high station density. In all 16 basic stations were used, the locations of which can be found in Figs. 6 and 7.

The macrowind appears to vary quite regularly across the country. In particular it was found that in the summertime seasons (May/June, July/August and September/October) the horizontal gradient of _S_ shows little geographical variation. A least-squares fitted plane, sloping up towards the coast in a WNW direction, approaches nearly everywhere the seasonal _S_ averages of the individual stations to within 0.35 m s⁻¹ (e.g. Fig. 6(a)). This results in an acceptable _U_p uncertainty of ±0.2 m s⁻¹.

The three wintertime seasons show a less homogeneous picture. When single-plane

![Figure 6](image-url)

Figure 6. Results of various alternative approaches to smoothing seasonal macrowind averages from 16 stations by least-squares plane fitting, δ = r.m.s. value of deviations between _S_ values and local _S_ field estimates at these stations. For comment on the four illustrated cases (a)–(d) see text.
smoothing of the macrowind was attempted, it proved not possible to approach the individual station macrowind values nearer than \( \pm 0.5 \text{ m s}^{-1} \) (e.g. Fig. 6(b)). This would entail \( U_p \) uncertainties which were clearly larger than the uncertainties in determining the original \( U_p \) values of the various stations. An acceptable result was obtained by smoothing separately a coastal zone in the west of about 50 km width. In this coastal zone the winter gradient of \( S_h \) proved to be large, approximately 1 m s\(^{-1}\) over 20 km. On the other hand, the inland \( S_h \) gradient in winter was only 1 m s\(^{-1}\) over 100 km. With this double-plane smoothing it again proved possible to approach the seasonal \( S_h \) values of almost all stations to better than 0.4 m s\(^{-1}\) (e.g. Fig. 6(c)). Obviously, in a country with less regular geographical variations, smoothing has to be done by application of polynomial or spline functions (e.g. Delaunay 1984).

The difference between summer and winter in the overall macrowind field structure may be related to the behaviour of depressions. In winter the preferred depression tracks are just north of the Netherlands along the coast, following the relatively warm North Sea water to the German Bight. It occurs much less often, certainly in the windy November season, that winter depressions cross our country towards central Europe. Therefore strong winds in winter are much more probable on the west coast than inland. On the other hand, this behaviour does not entail very strong winds in the north-east, since the winds near the depression centres are relatively weak and these centres often pass close to the northern coast line. In this respect it is acceptable that the smoothed \( S_h \) isolines hardly bend along with the coast, even though that would be expected intuitively.

Figure 7 shows the annual macrowind field, obtained by averaging the \( S_h \) fields of the six seasons. The \( S_h \) isolines still show the coastal effect of the winter seasons. However, due to the uniformity of the summer fields the net result is so regular that single-plane smoothing of the annual \( S_h \) values approaches the individual station values within accuracy limits. Incidentally, the annual-average station values of \( S_h \) deviate from the 16-station average of 10.7 m s\(^{-1}\) by as much as \( \pm 0.9 \text{ m s}^{-1} \). It may be concluded that over the Netherlands the assumption of a constant macrowind is not justified—certainly not in wintertime (e.g. Fig. 6(d)), but also not in the summer.

The surface wind counterpart of Fig. 7 is Fig. 8, a detailed map of annual-average potential wind speed \( U_p \). This has been derived over land by computing \( U_m \) for each block from the smoothed macrowind and the block mesoroughness of Fig. 4. For the six seasons, separate \( U_p \) maps have been published by Wieringa and Rijkhoort (1983).

Inspection of Fig. 8 will show that the speed class 6.2-7.0 m s\(^{-1}\) has been omitted—this is intentional. At the coast, too, we want to compare normalized wind speeds at 10 m, but over the sea it is more relevant to use an open water reference roughness, \( z_o = 0.0002 \text{ m}, \) for normalization. This implies that at sea \( U_p = U_m/1.17, \) while on the coast \( U_p = U_m/1.31, \) representing the fact that a given coastal pressure gradient generates stronger winds over sea than over the flattest land. Due to the predominance of westerly winds, the \( U_p \) climate (average, diurnal variation, annual variation, etc.) at four stations on \( \sim 1 \text{ km long west-facing piers} \) already differs little from the wind climate on light-vessels 50 km out at sea.

7. Validation of annual wind map, and supplementary results

From the map of annual-average potential wind speed, Fig. 8, estimates of \( U_p \) have been made at the location of 34 wind stations with at least three years of data; locations are indicated in Fig. 7. On comparing model estimates with measured averages the first interesting conclusion was that there is no strong indication that model performance in the coastal zone is unreliable. At the coast the data from Fig. 8 agree quite closely with
station averages, if the coastal stations are situated a few hundred metres inland. Only if the anemometer mast is actually sited over open water, has the station a markedly higher wind speed than the map indicates. Such open-water stations have not been used for checks and are omitted in Fig. 7.

At 31 out of 34 stations, $U_p$(model)/$U_p$(station) showed random variations with a mean of 0.994 and a standard deviation of 0.033. At three stations only (Beek, De Bilt, Zierikzee), separately marked in Fig. 7, annual $U_p$ values are about 0.5 m s$^{-1}$ below the map values, not compatible with a normal distribution of deviations. Of the remaining 31 stations, i.e. 16 basic stations and 15 additional control stations, 29 have their wind speeds represented by Fig. 8 within ±0.25 m s$^{-1}$. This deviation is consistent with an uncertainty of ±0.15 m s$^{-1}$ in the original averages and an uncertainty of ±0.2 m s$^{-1}$ in smoothing $S_b$.

Two of the sub-standard stations, De Bilt and Zierikzee, are badly sheltered, with obstacles ≤10 obstacle heights away in the whole westerly sector. Obviously a minimum station location quality is required for reasonable model agreement.

The third sub-standard station, Beek airport, is located in the far south-west of the Netherlands, where elevation variations are ~100 m (see Fig. 3) so that the applicability of the Smith–Carson roughness formula, Eq. (3), is problematic. Beek itself is situated near the edge of a table-shaped hill, and katabatic effects on the wind climate occur at this elevation (Dawe 1982). Since the nearby regional Belgian and German stations are so sheltered that they are unsuitable for model performance checks (Wieringa 1983), the map values for the far south-west must be regarded as unchecked and only indicative.

For the remainder of the country the reliability of the result is quite satisfactory. For any location not in the south-east, we can model the hourly average speed and its annual distribution (see below) with the same accuracy as if a wind station had been
located there for two years. Thus the simple procedure used to average the macrowind has practical validity within the borders of the Netherlands. A subdivision of Fig. 8 into zones of 0-5 m s\(^{-1}\) is justified, but interpolation to much higher accuracy is not advisable.

So far we have looked mainly at modelling of season-average potential wind speeds. However, if we know the distribution shape at any location, e.g. the Weibull parameter \(k\), then we can derive the local wind speed distribution if we have also a (modelled) estimate of the seasonal average (Hennessey 1977). From Weibull analysis of \(U_p\) at the 16 basic stations we have determined the geographical distribution of \(k\) with the purpose of interpolation (see Fig. 5); it proved to be very simple: \(k = 2.2\) at sea; \(k = 2.0\) at the coastline; and inland, \(k = 1.75 \pm 0.1\) from the coast to the German–Belgian border (Wieringa and Rijkoort 1983). At the 15 control stations shown in Fig. 7 these modelled distributions approach the actual distributions obtained from local observations sufficiently closely, that average wind energy can typically be estimated within 10%.

Other applications of exposure correction to wind analysis have been developed. Cats (1980) has shown that potential wind speeds show interesting features of synoptic-scale atmospheric flow, because local surface-generated noise has been eliminated. Consequently analysis of small-scale divergence effects from the wind field becomes feasible.

Extreme-value distributions of potential wind speeds were found to behave much less erratically than the distributions of originally observed extreme values. An additional important discovery of Rijkoort (1983) is that daytime \(U_p\) distributions are exactly Weibull-shaped. Conversely, nighttime distributions approach daytime distributions only at high speeds; at low speeds they curve away in a fashion which can be described by a stability parameter.

Since \(U_p\) distributions are representative on the mesoscale, the act of applying exposure correction evidently makes it possible to model wind speed distribution parameters regionally instead of station by station (Rijkoort and Wieringa 1983). Separate distribution evaluation for day and night and for individual azimuth sectors, followed by joint regional statistical analysis of station distribution parameters, provides extreme-value curves with a much more plausible behaviour than a Gumbel analysis of the original observations (Fig. 9). Consequently it was admissible to publish maps of design wind speeds for return periods up to 1000 years (Wieringa and Rijkoort 1983).

8. CURRENT MODEL DEVELOPMENTS AND POSSIBLE EXTENSION TO COMPLEX TERRAIN

The modelling approach presented here has two uncommon features: first the use of different roughness parameters for the surface layer and the Ekman layer; and second, the use of seasonal surface wind averages, with the subsequent assumption of neutral PBL stability. The second feature has the practical advantage that the computing requirements are very modest. However, the applicability of the model results is limited to the average surface layer wind climate at heights <30 m.

In regard to the PBL formalism used, comparison with other related models (e.g. Simiu 1973; Garratt 1977; Smith and Carson 1977) is difficult, since they are all essentially single-layer PBL models. In regard to the application to wind climatology, a different model with the same purpose was developed at Risø for Denmark (Petersen et al. 1981), based on 3-hourly station pressure data. A geostrophic wind, \(G_a\), could be derived from these for ~90% of the time and was assumed to be constant across Denmark. Using the PBL stability data from the 123 m mast at Risø, surface wind speeds over standard
roughness were derived from the $G_n$ values by way of the similarity equations (4) and (5). These winds were then analysed for each azimuth sector, and the compound distributions were checked against surface wind data.

A European collaboration is now underway to combine the useful features of this Risø model with the Dutch model described above, which can describe macroscale wind variations because it can accommodate surface wind data as input. Working with (3-) hourly surface wind averages, it becomes necessary to incorporate stability in the calculations if the results are required to make any sense at heights greater than 30 m (Holtslag 1984). However, the necessary information on stability at wind stations can nowadays be estimated from routine station measurements (Holtslag and Van Ulden 1983; Van Ulden and Holtslag 1985). It is planned to use this joint European model—which ought to be called the Troen model in acknowledgement of its chief implementor—to evaluate the wind climate in all non-orographic regions of the European Community (Petersen et al. 1985).

Still, we obviously cannot handle really complex terrain with such a model unless we can provide an extension. An interesting idea was developed by Delaunay (1984) in an analysis of the wind climate in Bretagne (western France), where terrain elevation changes are comparable to those in the south-east of the Netherlands. His basic concept
evaluates surface wind input according to the Dutch model, with addition of an azimuth-
dependent correction factor for overspeeding or sheltering caused by hills around the
station. An assumption that the mesowind $U_m$ at 60 m (i.e. the exposure-corrected station
wind) is suitable for non-divergent mass conservation flow modelling (Sherman 1978)
makes it possible to calculate the ratio between station-derived $U_m$ values and model-
derived $U_m$ values over hypothetical flat terrain. Delaunay’s preliminary results appear
promising.

In fact, this concept is comparable to the $U_p$ concept introduced in section 3, only
now we have a ‘potential mesowind’. Essentially this should make it possible to handle
moderately complex terrain, where elevation changes are very much less than the PBL
height. For handling larger orography with significant buoyancy effects, it is evidently
necessary to use full elaborate PBL models (e.g. Melgarejo 1982; Deardorff et al. 1984;
Hunt et al. 1984), though also for those models application of exposure correction to
wind data input might prove useful. Nevertheless the simple model used by Delaunay,
with its relatively limited computing requirements, may be useful for climatological
purposes in terrain of modest complexity, such as the south-eastern Netherlands.

9. CONCLUSION

The application of representative mesoscale wind information, obtained by exposure
correction of station wind data, in a similarity drag model has given rise to some
useful climatological wind analysis methods. For non-complex terrain a relatively simple
approach has produced an unusually detailed map of surface wind speed, taking into
account terrain features in an objective manner instead of by educated-guess inter-
polation. Moreover, mesoscale evaluation of wind speed distributions proved to be
possible; these are applicable, for example, to the regional determination of design wind
speeds. Possibilities of modifying the same methods for slightly complex terrain appear
to be present.

The reliability of the derived wind climate is sufficient for many engineering appli-
cations. For instance, in order to evaluate available wind energy in the Netherlands, the
model results have been applied in a wind energy handbook (Vermeulen et al. 1984).
There the concept of potential wind speed as a small-scale representative parameter has
proved useful, as well as understandable by the non-meteorologist. Consequently, the
handbook on the Dutch wind climate (Wieringa and Rijkoort 1983) contains system-
atically potential wind data instead of data ‘as measured’.

The evaluation of surface roughness from station-measured gustiness is com-
plemented by the alternative of roughness classification according to Davenport (1960). The
classification has also been applied to supply a mesoscale terrain roughness estimate.
This completes the set of meteorological tools required to produce a reliable large-scale
wind climatology from the usual imperfect wind station observations.

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REFERENCES


Carson, D. J. 1986 ‘Issues concerning the evaluation of effective surface roughness of heterogeneous terrain’. To appear in proceedings of conference on parametrization of land-surface characteristics and use of satellite data in climate models, and first results of ISLSCP. Rome, December 1985


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