REPRESENTATIVE ROUGHNESS PARAMETERS FOR HOMOGENEOUS TERRAIN

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Abstract. Objective requirements are drawn up for experimental determination of the roughness of homogeneous terrain, particularly with respect to matching of the observation array to a terrain situation with given fetch, blending height and displacement length. Some fifty well-documented published experiments over various surfaces, ranging from mobile surfaces (sea, or moving sand or snow) to forests and towns, are shown to meet these criteria and are compiled. It is shown that most presently popular terrain reviews underestimate actual terrain roughness lengths by about a factor two.

1. Introduction

The surface roughness of the earth continuously serves as a momentum sink for the atmospheric flow. For description, modelling and forecasting of the behavior of average winds and turbulence at all scales, correct knowledge of the strength of this momentum sink is a necessary boundary condition. At large scales, Anthes (1980) illustrates the necessity of a good assignment of surface friction for modelling of synoptic-scale storm development, and Arya (1977) shows how surface roughness influences the development of global atmospheric circulations. At small scales, significant dependence on terrain roughness is shown by e.g., Pasquill (1974) for vertical diffusion of pollutant plumes, and by Seguin (1973) for regional evaporation modelling. Jensen (1958) demonstrates that wind tunnel modelling of wind pressure on buildings can be severely biased if an incorrect upwind roughness is used. In practical local application of wind observations we always require some information on roughness of the upwind terrain, if we want to know wind speed at a nearby place and height different from the anemometer position (Wieringa, 1976).

Now assume that a roughness estimate is required for some non-complex area, without significantly large hills or coastlines, but maybe with terrain inhomogeneities at smaller scales. The present approach to this problem is to assign roughness values to moderately homogeneous parts of the area, and then to combine these into an "effective" roughness value for the whole area. The problem of evaluating such an effective roughness has been approached in various ways (e.g., Fiedler and Panofsky, 1972; Smith and Carson, 1977; Van Dop, 1983; Wieringa, 1986; Taylor, 1987; Mason, 1988; Claussen, 1990) and will not be dealt with here. Rather, attention is focused on the first step of assigning roughness to "homogene-

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ous" sub-areas, because this step is based on "typical" roughnesses from various published lists – which at present differ too much for comfort. To give some authoritative examples, roughness lengths listed by Cook (1985) and Troen *et al.* (1987) are a factor *three* lower than those given by Davenport (1960) and Oke (1978) for the same terrain types. Whom can we trust most? The true values are not necessarily halfway.

One possible cause of such large differences between equally conscientious studies is found in the references given by their authors, which show that most of the roughness lists rely on a surprisingly small set of experimental studies, most of them older than 1960, very seldom more recent than 1970. It is probable that some of those did give biased results, due to limited technical means and to ignorance of some recently developed insights.

The purpose of this study is twofold. Primarily, objective criteria are developed for determining true values of homogeneous roughness. Particular attention is given to the matching of observation arrays and their location to the terrain situation, taking into account the upwind fetch and the existence of a transition sublayer.

Secondarily, a non-parochial search was made for all experiments over identifiable terrain, which agree with these criteria and are accessible in the open literature. Too much excellent experimental fieldwork done in the last thirty years has not found its way into application, presumably because the relevant publications are unknown to the active investigators. About fifty acceptable high-quality experiments on roughness of homogeneous terrain types, varying from smooth surfaces through low vegetation to terrain with a full cover of trees or buildings, are summarized here. A short review on experimental studies of non-homogeneous terrain is contained in a separate publication (Wieringa, 1992).

The listed homogeneous roughness averages may prove suitable as elementary building blocks for other studies on e.g., effective roughness of inhomogeneous surfaces. In this study, the averages are used as checkpoints to investigate the validity of various popular lists and classifications of landscape roughness.

2. Roughness Parameterization

Roughness describes the surface sink strength for momentum and is therefore defined for neutral thermal stratification, where the vertical flux of horizontal momentum is not augmented or decreased by buoyancy effects. Generally the interaction of the earth boundary with atmospheric flow is an integrated result of turbulent drag caused by many obstacles, from very small to very big. When compared to form drag due to pressure differences across surface obstacles, even small ones such as grass leaves or capillary water waves, the viscous no-slip interaction of air moving parallel to the surface is negligible. In other words, the surface of the earth is nearly always aerodynamically rough in the fluid dynamic sense. For the dominance of viscous surface drag we would require a smoothness of such a high degree, that in nature it has only been observed once, over coral mucus in very shallow water (Hicks *et al.*, 1974; Deacon, 1979).

Two parameters are commonly used to describe surface roughness, the roughness length z_0 and the drag coefficient $C_{D(z)}$, both to be defined below. The notation already indicates, that the value of $C_{D(z)}$ always depends on the choice of reference height z, and it will be shown that over homogeneous terrain, z_0 is height-independent for a sizeable height interval. This is a major reason for preferring z_0 as a basic descriptive roughness parameter, even though $C_{D(z)}$ is a practical working alternative for modelling purposes.

Consider a stationary, adiabatic, horizontally homogeneous planetary boundary layer (PBL) of height h, where the interaction of wind speed U and roughness results in a (turbulent) shear stress $\tau_0 \equiv \rho u_*^2$ at the surface. Here ρ is mean air density in the surface layer, which can be assumed constant with height, and u_* is the friction velocity. For such a boundary layer it was shown by Blackadar and Tennekes (1968) that an inertial sublayer (ISL) exists over a range of heights, which are well below h and not very close to the rough surface. The ISL is defined by the property, that it is both high enough to apply the velocity-defect law valid in the upper part of the PBL, and also low enough for validity of the lower-PBL wall law, so that both laws can be matched. In the adiabatic ISL, matching then results in a dimensionless wind speed gradient ϕ_M of value unity, i.e.,

$$\phi_M \equiv \frac{\kappa z}{u_*} \frac{\partial U}{\partial z} = 1 \tag{1}$$

where κ ($\approx$0.40) is the von Kármán constant. Integration of Equation (1) results in:

$$U_Z = \frac{u_*}{\kappa} \ln\left(\frac{z}{z_0}\right) . \tag{2}$$

This ISL-relation, the logarithmic wind profile, defines the roughness length z_0 as a height-independent constant. Similarity arguments (Tennekes, 1973, 1982) show that applicability of the velocity-defect law sets a lower limit to the ISL at $z \approx 20\, z_0$ to $50\, z_0$, while simultaneous use of the wall law requires an upper limit at $z \approx 0.1$ h to 0.2 h. Over homogeneous terrain, the practical extent of the ISL is from a few meters above the ground up to $z \approx 50$ to 100 m, depending on roughness and wind speed. The popular saying "z_0 is the height at which the wind speed becomes zero" is therefore true in a purely algebraic sense only, since it implies extrapolation of Equation (2) below its limit of validity.

Roughness can also be specified by way of the stress resulting from roughness interaction with the wind speed at level z. The drag coefficient

$$C_{D(Z)} \equiv (u_*/U_Z)^2 , \tag{3}$$

can be interpreted as the ratio of free-flow kinetic energy ($\approx \rho U_z^2$) to surface stress. From Equations (2) and (3) it follows at once that

$$C_{D(z)} = [\kappa/\ln(z/z_0)]^2 , \tag{4}$$

implying that $C_{D(z)}$ and z_0 are interchangeable descriptions of roughness, as long as the chosen reference level of $C_{D(z)}$ lies within the ISL.

The height-independence of z_0 is confirmed by an argument of Monin and Yaglom (1971). If we equalize U_Z in Equations (2) and (3):

$$U_Z = \frac{u_*}{\sqrt{C_{D(Z)}}} = \frac{u_*}{\kappa} (\ln z - \ln z_0) ,$$

and eliminate the ($\ln z_0$)-term between two arbitrary heights z_1 and z_2:

$$\ln z_1 - \frac{\kappa}{\sqrt{C_{D(Z_1)}}} = \ln z_2 - \frac{\kappa}{\sqrt{C_{D(Z_2)}}},$$

we can conclude that

$$z \exp[\sqrt{C_{D(z)}/\kappa^2}] = z_0 ,$$

does not depend on height. In other words, unlike the drag coefficient, the roughness length is a height-independent description of surface roughness influence on flow dynamics in the entire inertial sublayer.

In the ideal stationary case with a very long homogeneous upwind terrain fetch (say >10 km), equilibrium between turbulence production and its dissipation is achieved locally. Then the wind profile will plot on semi-logarithmic graph paper as a single straight line over the whole ISL height interval. Then also terrain roughness length values which are determined from wind profiles observed entirely within the ISL, using Equation (2), will be equal to z_0-values obtained by way of Equations (3) and (4) from observations of turbulent stress or gustiness within the ISL.

To match z_0-values obtained from turbulent u_*-observations with those obtained from profile observations, we require some certainty about the value of the Kármán constant κ, since a κ-decrease by 0.05 changes z_0 by a factor 2 on the average. Before 1970 mostly $\kappa \geqslant 0.41$ was applied. After a temporary preference during the seventies for $\kappa = 0.35$, based on the Kansas experiment alone, use agreement has shifted back towards $\kappa = 0.40 \pm 0.01$ as more experimental information became available (Wieringa, 1980; Kondo and Sato, 1982; Tennekes, 1982; Zhang *et al.*, 1988). Since κ is eliminated in the evaluation of wind speed ratios (see e.g., Equation (12)), profile-determined roughness data (i.e., the majority) are not biased by a Kármán constant uncertainty. Turbulence-derived roughness data are corrected towards $\kappa \approx 0.40$ if needed.

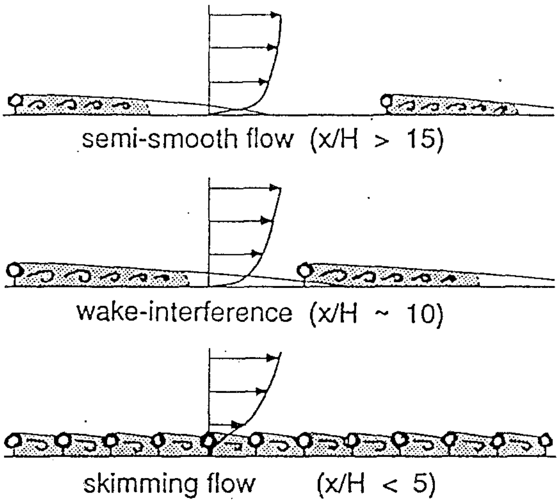

Fig. 1. Flow categories with typical density of terrain obstacles and indication of appropriate wind profile shapes (Wieringa, 1981).

3. Terrain Categories and Homogeneous Fetch Requirements

It is illuminating to subdivide the surface aspect of natural terrain according to physical flow behavior. The following four categories are widely used in surface flow investigations (see Perrier *et al.* (1972), and Figure 1):

(A) *Smooth* turbulent flow occurs over flat surfaces without any obstacles which are prominent enough to produce noticeable wakes.

(B) *Semi-smooth* turbulent flow occurs over surfaces with isolated obstacles which are sufficiently far apart, that their individual wakes are almost dissipated in the interspaces between the obstacles. In this situation the obstacle form drag and the surface friction drag of the large interspaces are approximately additive (Marshall, 1971).

(C) *Wake-interference* flow occurs, when obstacle interspaces are equal to or slightly less than typical wake lengths – of the magnitude of 5 to 15 obstacle heights H, depending on shape, porosity and distribution of the obstacles. Then the various types of drag are not simply additive, and the flow will nowhere be in equilibrium at levels $z < H$.

(D) *Skimming* flow occurs when the surface is so closely covered with high obstacles (at relative distance $D \leqslant 3H$), that flow in the interspaces between obstacles has a regime quite separate from the bulk flow above.

Homogeneous terrain is dominant in the terrain categories (A) and (D), smooth flow and skimming flow. However, in semi-smooth terrain of category (B), homogeneous fields can also be found, e.g., large areas of crops between distant shelter belts.

The purpose of this study is determination of typical roughness lengths for well-specified homogeneous terrain types. Field observations can only be generalized, if all observation levels are within an inertial sublayer of which the structure is nearly completely determined by terrain of the investigated type. Primarily this requires that the lowest observation level $z_{\min}$ should be at a height $\geqslant 20\, z_0$ above the reference surface. But at the topmost measurement level, too, wind structure should still be determined by the same homogeneous roughness, and this imposes an upper observation limit if the homogeneous fetch is finite.

If the terrain changes its surface roughness at upwind fetch distance x, the wind profile will only be related to the local terrain roughness within an internal boundary layer (IBL) of limited height $\delta(x)$, while the wind structure at higher levels is still determined by the roughness at upwind distances $>x$. This was first shown by Elliott (1958) and Brooks (1961); subsequent IBL research has recently been reviewed by Garratt (1990). It should be noted that the value of $\delta(x)$ need not be the same for average wind speed as for various turbulence parameters (see e.g., Shir, 1972). In all IBL studies the IBL-height $\delta(x)$ is defined as the highest level, where the influence of the roughness change begins to be noticeable.

Two major alternative analytical approaches exist to model $\delta(x)$. The first approach, initiated by Elliott (1958), is based on matching flow parameters upwind and downwind of the change. The second approach, pioneered by Miyake (1965), is based on an analogue with the downwind diffusion of a pollutant plume from a surface source, and has the practical advantage of relying only on downwind roughness. Experimental studies by Jackson (1976) and Walmsley (1989) show that the second approach agrees better than the first with field experiments of Kutzbach (1961), Blackadar *et al.* (1967), Bradley (1968), Angle (1975) and Peterson *et al.* (1979). Miyake's model, as presented by Panofsky and Dutton (1984) and Walmsley (1989), states that

$$\frac{x}{z_0} = \left(\frac{\delta}{z_0}\left[\ln\frac{\delta}{z_0} - 1\right] + 1\right) \Big/ \left(\kappa\frac{\sigma_w}{u_*}\right), \tag{5}$$

where σ_w = standard deviation of vertical turbulence components. If $\kappa \approx 0.40$ and $\sigma_w/u_* \approx 1.25$ (Merry and Panofsky, 1976), the value of the denominator is 0.5.

Peterson (1969a) first showed, that most of the layer below $\delta(x)$ is a transition layer where $\phi_M \neq 1$, varying between ≈ 0.5 and ≈ 1.5 according to height, fetch and type of upwind terrain change. Near-equilibrium adaptation of fluxes to the downwind roughness occurs at levels below $\approx 0.1\ \delta(x)$, so that is the largest acceptable height for observations used in determining terrain roughness correctly. Applying Miyake's Equation (5) and substituting $\delta = 10\ z$, we get

$$F \approx 2z_0\left(\frac{10z}{z_0}\left[\ln\frac{10z}{z_0} - 1\right] + 1\right), \tag{6}$$

for the fetch F which is necessary to ensure, that a particular profile level z is still

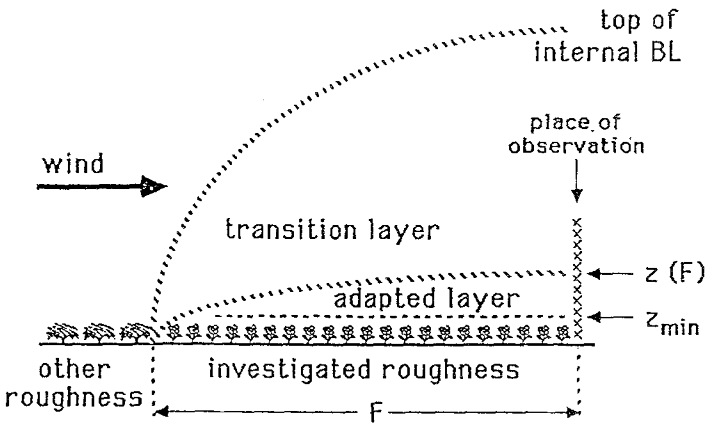

Fig. 2. Structure of the internal boundary layer downwind of a roughness change, showing at fetch distance F the observation height range (z_{min} to $z(F)$) which is acceptable for unbiased determination of the investigated nearby roughness. (Vertical distances not drawn to scale).

in the equilibrium layer fully adapted to upwind roughness z_0. The overall concept is illustrated in Figure 2.

Regarding the validity of Equation (6), it is no proof but still an indication of plausibility that the popular rule of thumb, that fetches should be ≈100 to 1 (e.g., Peterson *et al.*, 1978), agrees with it for common terrain situations with $z_0 \approx 0.1$ m. Next, for estimating heights of adapted profiles, Munro and Oke (1975) developed a similar z_0-dependent procedure, based on the model of Elliott (1958). For smooth roughness ($z_0 \leqslant 0.005$ m), their results agree with Equation (6), but for $z_0 \approx 0.1$ m the adapted height from their model is a factor 1.8 larger than the value from Equation (6). This difference fits in with experimental checks of the Elliott and Miyake IBL-approaches by Walmsley (1989), which also indicated that Miyake's model is better. Unfortunately, Walmsley's control study with field data on IBL height lacks fetches larger than ≈150 m.

Several field experiments can be added to Walmsley's list. First, Brooks (1961) shows a wind profile over concrete; the flow adapted up to $z \approx 0.5$ m after 90 m ($z_0 \approx 0.0005$ m), which fits Equation (6). Second, Deacon (1953) found his profiles to be logarithmic up to $z \leqslant 4$ m after 170 m fetch over short grass ($z_0 \approx 0.014$ m), for which Equation (6) predicts $z \leqslant 1.5$ m, but the terrain farther upwind was long grass, not strongly different. Third, Gash (1986) found that at 3.5 m above heather ($z_0 \approx 0.05$ m) about 400 m fetch was necessary for satisfactory flux adaption, which fits Equation (6) well. Gash also found shorter adaptation fetches over rough forest, as expected from Equation (6), but those observations were rather close to the treetops and are therefore quantitatively less reliable (see Section 4.1, below). At sea, Hupfer (1978) showed that at $z = 10$ m we require a fetch of several kilometers for equilibrium, again agreeing with Equation (6).

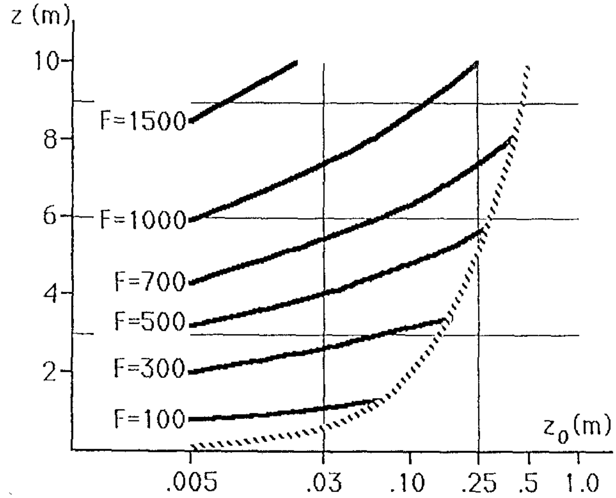

Fig. 3. Minimum fetch length F (m), necessary for adaptation of profiles with top level z to a homogeneous surface with roughness length z_0. The interrupted curve at the right corresponds to $z/z_0 = 20$.

Concluding, the model Equation (6) gives results of correct magnitude, though new field data on adaption over fetches longer than 200 m would be most welcome. The fetch modelling has been kept simple and two-dimensional on purpose. For evaluation of hundreds of field experiment publications, mostly with very scantily reported terrain information, employing three-dimensional "footprint" models (e.g., Schmid and Oke, 1989) is not a feasible approach.

To find easily the largest profile height range which can properly be used to determine a given upwind roughness, Equation (6) is given graphically in Figure 3. If roughness is determined from wind profiles, several profile levels must fit in between the highest acceptable observation level (given by the available fetch) and the lowest acceptable observation height z_{min}, which lies at $z \approx 20\ z_0$ for (semi) smooth terrain. Discussion of reliable roughness determination will be continued in Section 5, when criteria for z_{min} have been discussed for terrain with skimming flow.

Next to fetch homogeneity, we must also ascertain whether the fetch is horizontal. If the terrain slopes upward towards the mast, minor effects of hill overspeeding will be largest near the ground and the profile steepness will increase, producing inappropriately low profile-derived roughness values. Improbably low z_0-values on upslopes were observed by Rider *et al.* (1963) over grass and by Sunde (1968) on a beach. This stationary long-fetch slope effect is much more difficult to recognize that localized flow changes close to short escarpments, such as were investigated in the field e.g., by Peterson *et al.* (1976) and by Sacré (1981).

4. Roughness Properties Over Dense Surface Elements

Roughness properties for skimming flow situations must be discussed at length. One reason is that surfaces which are densely covered with slender objects, like grass, grain or trees, have skimming flow aspects at scales comparable to the height H of the objects and to their average interspace width D. On the other hand, the same surfaces look essentially smooth at the larger scale of the entire surface layer, say ≈ 50 m vertically and a few kilometers horizontally. When we use observations, which were made relatively close to such dense obstacle arrays, evaluation of roughness lengths involves two other vertical length scales, namely the transition sublayer height z_* and the displacement length d.

4.1. Transition sublayer

Many studies on wind over dense arrays of similar obstacles (say trees, houses, or wheat stalks) have been done at relatively low levels above the obstacle tops. Some investigators may have done so because of the high price of tall masts. A more scientific argument to stay low was the desire to make measurements in a layer which was completely adjusted to the vegetation, i.e., in the lower part of the IBL. In locations with short fetch, the acceptable height was then severely limited. For example, above trees with $H \approx 27$ m and $D \approx 6.6$ m, Leuning and Attiwill (1978) measured profiles to heights no more than 5 m above the treetops, explicitly arguing that their fetch was only 600 m.

The danger in using low-level observations was first shown in a good experimental study of wind over a forest by Dubov and Marunich (1973). Measuring turbulence above the trees ($H \approx 26$ m, $D \approx 9$ m) they found that only at levels between 32 and 42 m were the normalized standard deviation of wind (σ_u/u_*) and the normalized drag coefficient approximately independent of height, as expected in an inertial sublayer. Moreover they observed in the wind spectrum a significant peak at wavelengths comparable to D. In the layer below 32 m, the wind structure showed various anomalous features.

Independently of Dubov and Marunich, field experiments by Thom *et al.* (1975) and Garratt (1978) and wind tunnel studies by Raupach *et al.* (1980) led to the conclusion that a transition sublayer (TSL) exists close above the tops of dense vegetation. As summarized by Raupach and Thom (1981), mean flow varies three-dimensionally in the TSL as a function of the nearness of canopy elements, and there is a vertical so-called "dispersive" momentum transfer due to non-turbulent location-dependent motion. The wakes of the canopy tops produce rapidly-diffusing turbulence at length scales of the canopy top diameters and the interspaces. The detailed structure of this highly organized TSL flow, as investigated in field experiments by Gao *et al.* (1989), Bergström and Högström (1989) and Chen Fazu (1990), does not concern us here.

Our purpose here is good determination of momentum flux in the inertial sublayer above skimming flow. The top of the TSL, z_*, is defined as the level, where well-organized "dispersive" flux combines with the turbulent wake flux into the one-dimensional vertical flux of the inertial sublayer. This means that flow above z_* is not affected by location-dependent horizontal flow variations due to individual roughness elements below, and that there is no vertical average flow across z_*. In other words, z_* is the skimming-flow equivalent of the blending height, defined for semi-smooth or wake-interference surface-layer flow as the level above which the flow does not depend on horizontal location (Wieringa, 1976, 1986; Mason, 1988; Claussen, 1990). Therefore it seems not wrong to call z_* also "blending height".

In the TSL itself, part of the vertical transport of horizontal momentum is non-turbulent, and moreover the wind gradients are influenced by wake diffusion. Therefore z_0-values derived from TSL-profiles will be small and nonrepresentative for the total ISL stress (Raupach *et al.*, 1980). So it is not wise to obtain ISL roughness parameters from profiles with levels below z_*, even when the profiles are corrected for displacement (see Section 4.2).

For drag coefficients observed in the TSL, the net bias is uncertain. On the one hand, u_* is too small to be representative in the TSL (Raupach *et al.*, 1980), but on the other hand, U is horizontally inhomogeneous, so (u_*/U) might come out either slightly too large or much too small, depending on the location of the observation mast in the interspace wakes (De Bruin and Moore, 1985; Shaw *et al.*, 1988). The high-frequency wake spectrum peaks are smoothed significantly by many instruments, which might lead to additional underestimation of roughness derived from turbulence observations in the TSL.

4.2. Displacement length

The top of the TSL is not necessarily the same as the lower boundary of the ISL, since the fact that horizontal homogeneity of flow structure is achieved does not guarantee that the dimensionless wind gradient ϕ_M is already unity. We can approximate the deviation in ϕ_M by a height-dependent power series:

$$\frac{\mathrm{d}U}{\mathrm{d}z} = \frac{u_*}{\kappa z}\left(1 + c_1\frac{H}{z} + c_2\left(\frac{H}{z}\right)^2 + \cdots\right). \tag{7}$$

If z is still appreciably larger than H, we can neglect higher-order terms and abbreviate $c_1 H \equiv d$. Then

$$\frac{\mathrm{d}U}{\mathrm{d}z}\frac{\kappa}{u_*} \approx \frac{1}{z}\left(\frac{z+d}{z}\right) \approx \frac{1}{z-d},$$

which can be integrated to

$$U_z = \frac{u_*}{\kappa}\ln\left(\frac{z-d}{z_0}\right). \tag{8}$$

This is an empirical introduction of the displacement length d, as a correction to the logarithmic wind profile shape near the surface, which is proportional to H and negligible if $z \gg H$.

However, d has formal physical significance. Its basic role is to account for the fact that for a dense canopy of surface-covering objects (e.g., grain, trees or houses), only a small fraction of the total surface shear stress is taken up by the bottom surface at $z = 0$. The remaining stress fraction is carried by the canopy elements, with a vertical drag distribution depending on their shape. The logarithmic velocity profile has its origin at the level where the canopy-averaged surface stress acts, which is not at $z = 0$ if the canopy thickness is not negligible. Following a hypothesis of Thom (1971), Jackson (1981) proved from similarity arguments, that use of an appropriate value of d adjusts the reference level of the logarithmic profile to the effective level of mean drag on the surface canopy elements. Displacement will only be negligibly small when the obstacle density is so sparse that after individual obstacles, the separated flow is reattached to the surface before any upstream influence of the next obstacle is noticeable (Gartshore and De Croos, 1979).

Summarizing, if z_0 is the length scale for the total magnitude of shear forces acting on the surface, then d scales the vertical distribution of those forces in the surface canopy. A third profile scale, the blending height z_*, is the lowest height where horizontal location above the canopy top structure becomes unimportant for the flow structure. These three scales account for the basic influences on wind structure in the surface layer above homogeneous roughness. In non-homogeneous situations a fourth scale, called displacement thickness (Kawatani and Sadeh, 1971; Lo, 1990), can serve to describe flow over jump-like horizontal changes of surface structure.

The determination of d is needed if wind profiles must be analyzed or modelled close above the surface canopy, particularly in cases of wake-interference or skimming flow if z_0 is large. Then the lower limit of the ISL may be much higher than z_* and in an intermediate layer, defined loosely by

$$z_* < z < d + 20\, z_0$$

(approximately between $\approx 1.5\, H$ and $\approx 2.5\, H$), a wind profile analysis is still useful on condition that the displacement length is included. One advantage of the experimental use of this layer is the limitation of fetch requirements.

An unorthodox approach to evaluation of z_0 and d from wind profiles over skimming flow was recently developed by De Bruin and Moore (1985). It is based on the hypothesis of mass conservation in the sense, that in the TSL, the negative deviations of the real wind profile from the logarithmic profile Equation (2) should be compensated by the actually existing flow at levels $z < d$. Experimental application of this method requires, along with the wind profile, availability of turbulent flux data at the uppermost profile point, which must be not lower than z_*; Lo (1990) gives an alternative, using two profile levels above z_*. De Bruin and Moore discuss the additional problem that at levels $z \approx H$, effects of horizontal flow inhomogeneity must be taken into account.

The dependence of z_0 and d on size, shape, density and distribution of surface elements have been studied in the wind tunnel (e.g., Marshall, 1971, Counihan, 1971; Wooding *et al.*, 1973; Iqbal *et al.*, 1977; Hussain and Lee, 1980), by analytical investigation (e.g., Lettau, 1969; Kondo, 1971; Leonard and Federer, 1973; Businger, 1974; Seginer, 1974), by numerical modelling (e.g., Shaw and Pereira, 1982; Watanabe and Kondo, 1990), and in a great number of field projects. All studies agree that z_0/H increases very strongly with increasing obstacle density up to a maximum for surface coverage of $\approx$5% to $\approx$20%, depending on the vertical density distribution of the canopy. With further density increase, z_0/H decreases again, and also it scatters more because in fully skimming flow, the length of the bottom part of canopy elements is not very important for the magnitude of momentum exchange with upper airflow. The much-touted relation of Paeschke (1938), $z_0/H \approx 1/7$, proves to be quite unreliable, since z_0/H varies between $\approx$0.01 and $\approx$0.2 (see Garratt, 1977b). Obviously, roughness cannot be estimated well enough from H alone.

A first-order correction to the simple linear z_0/H-relation was given by Lettau (1969), estimating the effect of form drag on individual dominant roughness elements (e.g., trees). If on average per surface area A, one such element with height H and side-area S occurs, he proposed:

$$z_0 \approx 0.5\, HS/A\ . \tag{9}$$

Discussions by Businger (1974) and Seginer (1974) on this type of relation, e.g., on the change in effective drag by individual elements with increased relative sheltering, make clear that useful applicability of Equation (9) is limited to moderately inhomogeneous situations. It is certainly not a universal recipe for roughness estimation, and its application for this purpose in wake-interference conditions by e.g., Pielke (1984) produced some strange results. Even so, Equation (9) clearly shows that surface coverage and obstacle shape can be as important as H for the value of z_0.

On the other hand, d is found to depend mainly on H, and d/H increases only slowly with increasing obstacle density. For skimming flow, average d/H-estimates range from 0.64 for crops (Munro and Oke, 1973) to 0.75 for forests (Thom, 1971). Values of z_0 and d obtained from wind profiles are inversely correlated, so the result of z_0-evaluation relies heavily on the value of d used (Perrier *et al.*, 1972; Leonard and Federer, 1973; Baldocchi *et al.*, 1983). But d has to be derived from the degree of profile curvature, which is very difficult to gauge from three or four profile levels only (see Priestley, 1959). If faulty profile analysis has produced an improbably high d-value, calculated z_0-values will be improbably low (e.g., in Szeicz *et al.*, 1969).

Therefore z_0-values obtained for skimming flow from profiles with $z_{min}/H < 2$

will only be trusted if $d/H \approx 0.7 \pm 0.1$. Lower d/H-values occur only for (non-homogeneous) wake-interference flow.

5. Reliability Evaluation of Profile-Derived Roughness

The measurement of good reliable wind profiles is hard work, because the evaluation results are very sensitive to small errors. Let us define an ideal experiment. First, the observations are made on a slender mast on booms which are so much longer than the diameter of either mast or anemometer, that mast interference is negligible or can be corrected (see e.g., Gill *et al.*, 1967; Wucknitz, 1977; Wessels, 1984). Vertical spacing between anemometers is also sufficient to avoid interference. – Second, well-calibrated anemometers are used, and wind speed is averaged over at least 10 minutes. If anemometer overspeeding is large, it must be accounted for in the context of profile analysis (Busch and Kristensen, 1976). – Third, temperature gradients are measured simultaneously for diabatic profile correction, or alternatively the observations are restricted to situations when the stability is likely to be near-neutral, i.e., strong wind over rough terrain with overcast sky. – Fourth, near-adiabatic conditions at sunset or sunrise, or any other situation with bad nonstationarity aspects, are avoided (Stearns, 1971). It goes without saying that the experiment consists of several runs at least.

Altogether this is a tall order, but if a publication indicates (either explicitly or between the lines) that all these have been considered, I have estimated the observation reliability to be good. I call it acceptable, if only the second and third condition seem assured, while there is no reason to suppose that the first one was badly neglected.

5.1. Acceptable Height Range of Profiles

Next to experimental reliability there is another quality requirement, namely good matching between the observed location and the anemometer array used. There are two aspects to be considered in this matching. First, the number of measurement levels required for sufficient statistical accuracy depends on z_0, as discussed below in Section 5.2. Second, the profile height range should match the given terrain situation. The relation between topmost profile level and fetch was already given in Section 3, and if the TSL blending height z_* and displacement height d have negligible values, then the lowest acceptable observation height is $20\, z_0$. This applies above smooth terrain and for semi-smooth situations with sufficient fetch.

In order to obtain reliable ISL-z_0-values from wind profile experiments above skimming flow (without additional turbulence observations), the entire profile must be located above the TSL, i.e., $z_{\min} \geqslant z_*$. So we need primarily an advance estimate of z_* from directly available external information on the roughness elements. In field experiment publications usually the height H of obstacles is specified (N.B.: $h \equiv$ PBL-height!), even though the definition of H may vary

TABLE I

Transition layer heights above skimming flow

Author	Forest	H (m)	D (m)	z_* (m)	$(z_* - H)/D$	z_*/H
Dubov and Marunich (1973)	Pine	26	9	32	0.7	1.23
De Bruin and Moore (1985)	Pine	18.5	3.5	24.4	1.7	1.32
Bergen (1987)	Pine	27	3.7*	31	1.1	1.15
Högström *et al.* (1989)	Pine	16	5	29	2.5	1.81
Chen Fazu and Schw. (1989)	Bush	2.3	5	6.7	0.9	2.91
Viswanadham *et al.* (1990)	Tropical	35	2	39	2.0	1.11
				Average:	1.5 ± 0.7	1.6 ± 0.7

*Estimated from Eq. (11).

widely between authors (Leonard and Federer, 1973; Ford, 1976). Less often, obstacle density, element structure or interspace size D is also given.

A wind tunnel investigation of regular arrays of cylinders by Raupach *et al.* (1980) showed that at heights above $z \approx H + 1.5\,D$, the flow profile was approximately logarithmic, with a profile slope agreeing with turbulent flux measurements. Apparently the non-turbulent flux fractions of obstacle wakes are essentially diffused at those levels; also the profile scatter due to horizontal inhomogeneity was sufficiently smoothed at those heights.

A limited number of recent field experiments have made explicit checks on the applicability of ISL-relations as a function of height, and these are given in Table I. Here z_* is given as the lowest level, for which authors reported either applicability of a logarithmic profile, or height-independence of a flux quantity which should be approximately constant in the ISL. Because for most field experiments only the average obstacle density in numbers per area is given, we define here the interspace size D as the distance between centres of adjoining obstacles. If the obstacles are relatively broad (e.g., Chen Fazu and Schwerdtfeger, 1989), then the actually open interspaces may be noticeably smaller than D as defined here.

For field application, Table I validates Raupach's wind tunnel-derived criterion

$$z_* - H \approx 1.5\,D\,. \tag{10}$$

Relating z_* to displacement height d instead (e.g., Garratt, 1980; Jacobs and Van Boxel; 1988, Parlange and Brutsaert, 1989) is rather impractical, because we cannot determine d well before we know z_*. Iterative analysis is then required, and often authors do not mention whether this was actually done.

Practically, in descriptions of dense roughness arrays often even density is not specified. But it may then be possible to relate it to height for the case of plant stands, since for any height these tend to attain an optimum density for sharing of soil nutrients and solar radiation. Gorham (1979) found a highly-correlated relation, approximately linear, between H and D for naturally-developed growths of 19 different species, ranging from moss and reeds to trees. An analogous fully-linear relation is derived empirically here for pine forests, using data from various pine species with heights between 4 and 30 m: forests in Thetford (Oliver, 1971) and Brookhaven (Raynor, 1971), a German forest (Vogt and Jaeger, 1990), and 14 other forests listed by Silversides (1979). The result is:

$$D = 0.09\,H + 1.3 \pm 0.5\,\mathrm{m} \quad (n = 17, \text{ correlation coefficient } 0.79)\,. \tag{11}$$

A simplified relation, $D \approx 0.15\,H$, has an uncertainty $\pm\,0.7$ m. For mature pine forests of known height H, this gives a rough estimate of D to use in Equation (10).

However, Equation (10) does not help to estimate z_* for skimming flow over high grasses, grains and row-arranged agricultural crops, whose density cannot be defined as easily as for cylinders or trees. An indication that maybe for such vegetation we can bypass the use of D altogether, can be found in the results of Jacobs and Van Boxel (1988). Assuming $z_* \approx d + 10z_0$ for a maize crop, they obtained $d = 0.75\,H$ and $z_0 = 0.26\,(H - d)$, implying $z_* \approx 1.5\,H$. Anyhow, the last column in Table I shows that for the tabulated experiments $z_*/H \approx 1.6$ has an uncertainty similar to that of $(z_* - H)/D$. Therefore for situations where aerodynamic element density is difficult to define or unknown, a rounded-off estimate of $z_* \approx 1.5\,H$ will be used. In particular for wind profile levels over crops, this lower limit for profiles agrees reasonably well with good agrometeorological practice.

5.2. Required Number of Profile Levels

The above-given criteria show often that only a small height interval can be considered for evaluation of homogeneous roughness. Formally, the number of profile measurement levels, N, within this interval can be restricted to two. Equation (2) indicates that wind speeds at heights z_1 and z_2 are related by

$$\frac{U_2}{U_1} = \frac{\ln(z_2/z_0)}{\ln(z_1/z_0)}\,, \tag{12}$$

in which z_0 is the only unknown. However, simple calculations show that for this minimal arrangement any small wind measurement errors will result in huge errors in z_0 (Langleben, 1974), apart from the fact that biases due to stability of reference level displacement cannot even manifest themselves if $N = 2$. It is preferable to have a larger number of levels to provide redundancy checks. For the simple adiabatic case, z_0 can then be found from least-square regression of U against $\ln z_0$. However, most profiles are non-neutral and possibly displaced, and for their analysis the reader is referred to Stearns (1970), Leonard and Federer (1973), Nieuwstadt (1978) and Riou (1984).

An interesting experimental analysis of the relation of N to the accuracy of profile-derived u_* and z_0-values was made by Legg *et al.* (1980), using 10 days of data from a good six-level profile experiment over beans ($z_0 \approx 0.1$ m). Assuming $\pm\,2\%$ error in any single profile anemometer, they recalculated all profiles and found as average for the entire experiment, that N-values of 6, 5 and 4 resulted in possible u_*-changes of 8, 17 and 26% respectively. Since the error in z_0 is a factor $(\kappa U/u_*)$ larger, corresponding z_0-error percentages would have been triple! This empirical result agrees with theoretical studies (e.g., Covey, 1963; Tanner, 1963; Langleben, 1972; Foken and Skeib, 1980).

Over smooth terrain, wind gradients are small, so the observational error sensitivity is large. We can use as a working estimate that for determination of z_0 within a factor 2 we need at least 3 profile levels over high roughness ($z_0 \approx 1$ m); 4 levels for moderate roughness ($z_0 \approx 0.1$ m) and preferably 5 or more over really smooth terrain.

6. Roughness Evaluation from Turbulence Measurements

The alternative to profile determination of roughness is the use of direct observations of turbulence, using Equations (3) and (4). From about 1965 onwards, measurement of turbulent stress became more feasible through development of 3-component windmeters which were both fast-responding and weatherproof, together with increased availability of fast digital recording. In order to measure u_* with sufficient accuracy, the observed spectral bandwidth must be adequate, at least 2-Hz sampling over 10 minutes (McBean, 1972; Garratt, 1975), and instruments should be levelled to better than 1° (Wieringa, 1971). For determining z_0 by way of the standard deviation of vertical gusts, σ_w (Fiedler and Panofsky, 1972), levelling requirements are less strict.

Earlier, the only possibility to obtain z_0 from turbulence measurements was through the gustiness of the horizontal wind, using the relation

$$\sigma_u/u_z = 1.0/\ln(z/z_0)\,, \tag{13}$$

(Lumley and Panofsky, 1964). Again some bandwidth information is required, but there are no levelling requirements. Standard-deviation registrations of wind speed are feasible with present-day electronic processing, and therefore are recommended by WMO for station use (Beljaars, 1987b). A poor-man's version of this approach is statistical analysis of the gust factor, the ratio between peak gust u_{mx} and average wind speed U: if for some azimuth sector the median gust factor $\langle u_{mx}/U\rangle$ is known from $\geqslant$15 gust observations, then

$$z_0 = z\exp\left(-\frac{f_T A[1.42 + 0.3\ln(-4\times 10^3/Ut)]}{\langle u_{mx}/U\rangle - 1 + A - f_T A}\right). \tag{14}$$

Here A ($\approx$0.9) is the attenuation of u_{mx} by the anemometry, and f_T is a factor which is unity for 10-min averaging periods and increases to 1.1 for hourly averages. Ut is the average wavelength (wind run) of maximum gusts observed by the anemometer-recorder combination used, and varies usually between 50 and 100 m. For derivations and determination of the instrument constants, we refer to earlier publications (Wieringa, 1973, 1976, 1977; Oemraw, 1984). This method has the operational advantage that it allows objective determination of the roughness surrounding a single anemometer with simple registration.

For the above turbulence-derived roughness determination methods, their fetch requirements will differ from those for profile methods. Hunt (1971), Shir (1972) and Taylor (1983) conclude for various reasons that IBL-fluxes at a given height require more equilibrium fetch than is necessary for profiles. Field experience has shown that z_0-values from gustiness observed at 10 m height generally relate to an upwind fetch of several kilometers (Wieringa and Van der Veer, 1976; Wieringa, 1986), about twice the fetch expected from Equation (6). However, Beljaars (1987a) argues that relaxation of σ_u and gustiness is not so fast as for stress, because σ_u has extra input from large and slowly-dissolving eddies of sizes comparable to h. It is concluded that for eddy-correlation observations, the homogeneous fetch requirements should be larger than the fetch value given by Equation (6), and in particular we should use twice that value to estimate the upwind area influencing a gustiness-derived roughness.

The use of surface drag plates to determine surface roughness (Lynch and Bradley, 1974) is a great effort. Since these plates do not need a long fetch, they have been occasionally employed in experiments over surfaces which are smooth enough, that the plate does not need to be large to be representative.

7. Observed Roughness of Homogeneous Surfaces

Using the criteria described in the previous sections, quality checks were made for several hundred experimental field roughness determinations, for which original publications could be traced in the library. Results only, listed in other publications, were not accepted because it was necessary to know about the experimental setup and about the terrain situation explicitly, not in vague terms like "open terrain". The term "open" is one of the most abused words in small-scale meteorology: publications by renowned authors describe their location as "open" and assign a roughness length of a few centimeters, while photographs included show sagebrush or clumps of trees. In other "open" cases average wind speeds at $\approx$10 m did vary as much as 20% between subsequent masts. Only experiments with proven homogeneous fetches are tabulated here.

In the tables, the experimental arrangement is given under "*EXP*" as the number of observation levels and the type of measurement: U = average wind speed, EC = eddy-correlation stress, GU = gustiness, DP = drag-plate stress. Too-low levels, excluded explicitly by authors of an experiment themselves from their analysis of roughness, are not counted. "Experimental quality" is a subjective judgement on instrumental matters discussed in Sections 5 and 6. Given figures are sometimes rounded off within the accuracy limits of this review. All lengths are in metres except for x, which is in km.

The centre of the tables contains two quality numbers, describing the degree to which the experimental array was matched to the surface and to fetch length. In principle, both quality numbers should be larger than unity. Sufficiency of height of the lowest observation level z_{min} is checked by the similarity criterion $z_{min}/(20z_0)$ for very low roughness, by Raupach's criterion $z_{min}/(H + 1.5\,D)$ for very dense roughness, and by $z_{min}/1.5\,H$ for other cases.

Also, the required homogeneous fetch F for the highest observation level has been calculated from Equation (6) with the z_0-value given by the author, and is compared with the actual homogeneous fetch x. When the roughness of far terrain at upwind distances beyond x looks similar to the investigated roughness (e.g., Deacon, 1953), the sufficiently adapted profile might be higher than indicated by Equation (6); then x/F-values slightly smaller than 1 are not unacceptable. For roughness of terrain, which is homogeneous within the criteria used here, the notation z_{00} will be used.

For purposes of summarizing the z_{00}-results, the tables are divided into groups of comparable terrain situations. Each group consists of at least two experiments: ancient Roman law stated "one witness is no witness".

7.1. Roughness of smooth surfaces, solid and mobile

In nature, really smooth surfaces can only persist long in that state, if they consist of mobile material which acts to fill in any developing cracks or hollows. If the motion is slow, as for mud flats, wet sand surfaces or well-packed snow over flat ground, the surface behaves as a solid boundary. In Tables II and III it is seen that no really significant roughness differences appear to exist between various forms of such naturally-flattened terrain types and artificial flat hard surfaces like tarmac or concrete.

However, if the surface elements are highly mobile (water, fine sand or dry snow), then surface roughness increases with wind speed, or rather with friction velocity. For sea waves, Charnock (1955) first proposed the relation

$$z_0 \approx \beta u_*^2/g\,, \tag{15}$$

where g is the acceleration of gravity, and for fully-developed waves the experimental constant β is $\approx$0.0185 at neutral stability (Wu, 1980). If the wave age c_p/u_* (where $c_p \equiv$ phase speed of largest waves) is small, β attains significantly larger values because "young" waves are steeper (Kitaigorodskii, 1968; Maat *et al.*, 1991). The degree of influence of wave breaking on sea roughness is still the subject of research.

For small particles, above a size-dependent wind speed threshold, the particles are lifted individually and move over horizontal distances of $\approx$1 m; this is called saltation. Since the saltating particles ascend with an initial vertical speed $\approx u_*$, Owen (1964) proposed that the roughness of the saltation layer should be proportional to the height of that layer, $\approx u_*^2/(2g)$. This relation has since been shown to apply in the field, with a typical experimental uncertainty of $\pm$ 30% for the Charnock constant β of Equation (15). Over dry sand, observations of Vugts and

Cannemeijer (1981) and Walmsley (1988) correspond to $\beta \approx 0.04$, and a summary of other experiments by McEwan (1991) gives $0.01 < \beta < 0.045$, wind tunnel studies giving lower values than field data. Over fresh cold snow, Joffre (1982) observed $\beta \approx 0.015$, and from observations by Schmidt (1982) a value $\beta \approx 0.018$ can be deduced.

It then appears that all mobile surfaces behave according to the Charnock model (Chamberlain, 1983), with variations caused by differences in surface structure. Available β-values for snowdrifts agree roughly with those of sea waves, while for saltating sand, β is a factor 2 or 3 larger. Since grain-borne shear stress is proportional to particle mass (Owen, 1964), this β-difference might be modelled by assuming that β is proportional to the density of the saltating particles. At the saltation threshold speed (typically 6 to 8 m/s), the roughness length suddenly jumps from the low smooth-surface value to the higher "Charnock value". For dry sand the increase is typically from 10^{-4} to 10^{-3} m, an order of magnitude. For snow, the z_0-increase at some saltation threshold speed is hardly noticeable – partly because the initial roughness of snow at low wind speed is quite variable, partly since for snow, β is not large and the z_0-difference therefore small, and also because the snow saltation threshold wind speed depends on its cohesion and therefore varies with the thermal history (Kondo and Yamazawa, 1986; Schmidt, 1986; Kind, 1990). Old settled snow surfaces do not saltate at all, except maybe in gales.

For mobile surfaces the roughness variation can be summarized by citing Garratt (1977a), who analyzed 17 good experiments over open water. He shows that $C_{D(10)} \approx 0.0013$ to 0.0015 for $U_{10} \approx 6$ to 12 m/s, i.e., $z_0 \approx 0.0001$ to 0.0003 m at moderate wind speeds, increasing in gales to $z_0 \approx 0.001$ m or more. Roughness of approximately similar magnitude is found over "dry" (i.e., fresh cold) snow at all wind speeds, and also over dry sand for wind speeds below $\approx$8 m/s; at higher wind speeds, sand roughness becomes twice as large.

For ice fields, smoothness is not a necessary characteristic, certainly not at sea. There, at floe edges, ridges of several metres occur and will boost roughness up to values around $z_0 \approx 0.05$ m, necessitating drag-partitioning analysis like for other semi-smooth situations (Arya, 1975). Table II only lists four illustrative examples of observations over ice, because several good reviews on roughness of various ice surfaces were published recently (Overland, 1985; Belitz *et al.*, 1987; Guest and Davidson, 1991). It appears that roughness of ice fields, if ridging is not very dominating, is comparable in magnitude to the roughness of earth which is fallow and not quite smooth.

7.2. Roughness of low vegetation in "open" terrain situations

Considering the existing large amount of published agrometeorological field research on turbulent exchange over crops, the size of Table V is quite disappointing. The reason is that most such experiments take place on rather sheltered farmland: fetch requirements for homogeneity (see Figure 3) are therefore usually not met.

TABLE II

Roughness of snow and ice fields

Surface type Reference author(s)	z_{00} d	H $z_{min}/(20z_0)$	EXP x/F	Stability x (km)	Exp. quality Far terrain
Snowy flat ice Smith (1972)	$(2 \pm 1) \times 10^{-4}$ –	– 750	2 EC 1.5	Yes ≈ 1	Good Rather flat ice
Snowy flat ice Langleben (1972)	$(6 \pm 4) \times 10^{-4}$ –	– 21	5 U ≈ 1	Yes 0.5–1.5	Excellent Rather flat ice
Snowy flat ice Joffre (1982)	$(3.4 \pm 2) \times 10^{-4}$ –	– 150	5 U 0.9	Yes 2.0	Good Ice floe edge
Flat snow field King (1990)[1]	$(1 \pm 0.1 \times 10)^{-4}$ –	– 2300	3 EC ∞	Yes ∞	Good Antarctic plain
Flat snow field	0.0001–0.0007				
Small ice floes Smith *et al.* (1970)	$(8 \pm 4) \times 10^{-3}$ –	– 24	EC ≈ 9	Yes ≈ 5	Good Sea with ice
Snowy rough ice Holmgren (1971)	$(9 \pm 4) \times 10^{-4}$ –	– 28	5 U ∞	Yes ∞	Good Arctic plain
Flat rafted ice Langleben (1972)	$(3 \pm 1.5) \times 10^{-3}$ –	– 4	5 U ≈ 2	Yes ≈ 1.5	Excellent Rough ice
Rough rafted ice Macklin (1983)[2]	$(7 \pm 4) \times 10^{-3}$ –	– 5	4 U $\geqslant 1$	Yes $\geqslant 1$	Good Similar ice
Rough ice field[3]	0.001–0.012				

[1]For background information see King *et al.* (1989).
[2]With good list of other sea ice roughness experiments.
[3]Prevailing meridional winds in the Antarctic generate elongated sastrugi (Jackson and Carroll, 1978), presenting to the occasional flow across their orientation a roughness length of ≈ 0.05 m.

In particular, it proved impossible to find articles on roughness of rice fields with sufficient fetch; maybe such fields do not exist.

A major uncertainty in roughness of crops relates to their customary arrangement in rows. First, the width of the rows is often not documented but might have a significant influence on acceptable z_{min}-levels (see Section 5.1) and on the overall roughness itself. Second, the crop roughness value will depend on the wind direction. An interesting example is given for a vineyard ($H \approx 1.5$ m, $D \approx 1.8$ m) by Riou *et al.* (1987), unfortunately with a too-low z_{min}-level of 1.6 m. Their analysis resulted in $z_0 \approx 0.55$ m, $d \approx 0$ for flow parallel to the rows, and in $z_0 \approx 0.2$ m, $d \approx 0.75$ m for cross-row flow. This disagrees with the vineyard investigation by Weiss and Allen (1976) - again with a too-low z_{min}-level, and bad fetch – which concludes that the roughness is greater cross-row than down-row; but both investigations agree that there is, and should be, a difference. As long as crop investigations do not differentiate according to wind direction, it is mandatory to be severe on the z_{min}-requirement, and this is one more reason why so few investigations were accepted.

TABLE III

Roughness of featureless land

Surface type Reference author(s)	z_{00} d	H $z_{min}/(20z_0)$	EXP x/F	Stability x (km)	Exp. quality Far terrain
Concrete Sheppard (1947)	$(3 \pm 1) \times 10^{-4}$ –	– ∞	DP Good	? >0.1	Acceptable Plains
Just-flooded marsh Jensen (1954)	1.8×10^{-4} –	– 140	3 U ≈5	$U > 8$ ≈2	Acceptable Sea
Runway tarmac Bradley (1968)[4]	$(3 \pm 2) \times 10^{-4}$ –	– ∞	DP Good	Yes 0.15	Good 1m scrub
Flat desert Tetzlaff (1974)	$(3 \pm 1) \times 10^{-4}$ –	– 15	6 U ≈9	Yes ≈11	Good Mountains
Tidal mud flats Hessler (1987)	$(4 \pm 3) \times 10^{-4}$ –	– 375	4 U 2.0	Yes 7.5	Good Coastline
Very flat surface	0.0002–0.0005				
Harrowed field Seginer (1975)	0.003 ± 0.001 –	– 7	9 U 0.8	Yes >0.5	Acceptable Ploughed field
Almost no grass Hicks (1976)	0.0012 ± 0.0001 –	– 21	6 U 1.4	Yes ≈5	Excellent Wangara
Fallow ground	0.001–0.004				

[4]Reanalysed by Nemoto (1972) and Rao *et al.* (1974).

In some cases the experiments were given the benefit of the doubt, e.g., if the investigated field itself had insufficient fetch, but the terrain farther upwind was covered with similar crops without obstacles. For the experiment of Dobesch (1976), it was clear (from other parts of his study) that the author was well aware of the severity of fetch requirements, though these were not actually quantified for the sugar-beet data listed here. The grass roughness determinations by Deacon (1953) also had insufficient fetch but still are included, both because his analysis shows explicitly that at his site the resulting error was probably small, and also because in all other respects his experiment was outstanding in quality.

Crops do grow. For the investigations covering a full growth cycle (e.g., Dobesch, 1976; Azevedo and Verma, 1986; Jacobs and Van Boxel, 1988), Table V gives results for the crop at maximum height, except for the soybean data by Baldocchi *et al.* (1983) where the wind profile no longer fulfilled the z_{min}-criterion over the full-grown crop. The roughness of very young crops is rather comparable to that of short grass (Table IV). For low crops, $d/H \approx 0.55 \pm 0.1$, because their surface coverage S/A is generally less than that of high crops.

Regarding representative roughness (and displacement height) of slender vegetation, an additional important question is whether these parameters depend on wind speed. In particular, for grains and grasses with flexible stalks, it was found that lower roughness values occur for higher wind speeds (e.g., Long *et al.*, 1964;

TABLE IV

Roughness of grassland with homogeneous fetch

Surface type Reference author(s)	z_{00} d	H $z_{min}/(1.5\,H)$	EXP x/F	Stability x (km)	Exp. quality Far terrain
Short grass Deacon (1953)	0.002–0.017 –	0.04 17	3 U (0.3)	Yes 0.17	Good Long grass
Short grass Blackadar *et al.* (1976)	0.01 –	0.03 8.0	5 U 3.0	Yes >3	Good Marsh
Tundra Harper & Wiseman (1977)	0.021 ± 0.007 –	0.05? 4.0	3 U ∞	Yes Selected	Acceptable Tundra
Short grass Saugier & Ripley (1978)	0.011 ± 0.002 –	0.025 2.5	7 U 1.0	Yes ≈2	Good Downslope
Short grass, moss	0.008–0.03				
Long grass Pasquill (1950)	0.03–0.07 0.08	0.03–0.15 1.1	5 U + DP >0.8	yes >0.15	Excellent Well-exposed
Long grass Deacon (1953)	0.04–0.06 0.25	0.6 1.1	4 U (0.6)	Yes ≈1	Good Plains
Wheat stubble Bradley (1971)	0.024 0.1	0.18 1.0	5 U + DP 24	Yes >2.4	Excellent Kansas
Heather Wallace *et al.* (1984)	0.02 (0.3)	0.44 3.0	EC ≈1.0	Yes <0.25	Good Moorland
Heather Gash (1986)	0.028 (0.2)	0.25 9.3	EC 1.0	Yes 0.4	Good Forest
Long grass, heather	0.02–0.06				

Takeda, 1966; Perrier *et al.*, 1972; Tajchman, 1981), though the effects are often not significant (Munro and Oke, 1973). When using independent profile and turbulence observations, Baldocchi *et al.* (1983) found no such U-dependence and suggest that it can be introduced spuriously, if the interdependence of z_0, d and u_* is disregarded in profile analysis. Since moreover all observed changes occurred only for low wind speeds, we can disregard this phenomenon in the context of our study if we restrict ourselves to roughness data at moderate or high wind speeds – which is anyway the situation when good parameterization of wind is of large practical importance.

7.3. Roughness of Forests and "Regular" Towns

Flows over high dense obstacle arrays are always displaced. For some experiments the displacement height of profiles was estimated, either by the author or by myself. Such estimated d-values are bracketed in the tables. In case of turbulence observations, the measurement does not furnish a value for d at all. The required fetches F in Table VI (and V) have been calculated by substituting $(z_{max} - d)$ into Equation (6).

TABLE V

Roughness of full-grown crops with homogeneous fetch

Surface type Reference author(s)	z_{00} d	H $z_{min}/(1.5\,H)$	EXP x/F	Stability x (km)	Exp. quality Far terrain
Sugarbeets Dobesch (1976)	0.07 ± 0.02 0.16	0.3 1.7	4 U OK	Yes OK	Good OK
Soybeans Hicks & Wesely (1981)	0.05 ≈0.9	0.85–1.05 1.6	3 U + EC >0.7	Yes >0.5	Good Similar crops
Young soybeans Baldocchi *et al.* (1983)	0.06 0.5	0.8 1.0	6 U ≈ 1	Yes ≈ 0.2	Good Soybeans
Cotton (wind across rows) Kustas *et al.* (1989)	0.07 ± 0.04 0.3	0.5 1.6	5 U 2.6	Yes 0.8	Good Crops & fallow
Low mature crops	0.04–0.09				
Sorghum Azevedo & Verma (1986)	0.12 0.8	1.2 0.8	<7 U ≈0.8	Yes ≈0.2	Excellent[5] Similar crops
Wheat Munro & Oke (1973)	0.16 0.65	1.0 1.0	6 U 0.2–1.0	Yes 0.1	Good Similar crops
Maize Hicks & Wesley (1981)	0.17 1.5	2.6 1.3	(3 U+)EC 0.9	Yes >0.5	Good Similar crops
Maize Jacobs & Van Boxel (1988)	0.13 (0.75*H*)	2.3 1.0	<10 U + EC >0.2	Yes >0.12	Excellent Similar crops
High mature crops ("grain")	0.12–0.18				

[5]Profile analysis and mass conservation analysis (De Bruin and Moore, 1985) agree in results.

It is most unfortunate that for deciduous forests the few published experiments either show no measurements above the transition layer, or do not give much roughness information. An experiment by Anderson *et al.* (1986) in an oak-hickory forest at 28 m height lists turbulence-values, from which it can be estimated that $0.6\text{ m} \leq z_0 < 1\text{ m}$, but from that article z_* cannot be obtained. Some studies (Dolman, 1986; Shaw *et al.*, 1988) have indicated that possibly there is no significant difference in roughness for deciduous trees with and without leaves. The reason seems to be that the wind can penetrate deeper into a bare-branched forest, resulting in the same net amount of drag as for the same forest in summer, with leaves. (See also the discussion on vineyards, Section 7.2). New high-level experiments above season-dependent forests (or z_0-evaluation from existing observations) would be most welcome.

Homogeneous cities do not exist, strictly speaking, but some town areas show little variation in building height and a rather regular distribution of buildings, so that well above the roofs the wind flow should not vary much horizontally. Unfortunately many very high masts in cities are broad objects, often Eiffel-tower-

TABLE VI

Roughness of high vegetation with homogeneous fetch

Surface type Reference author(s)	z_{00} d	H $z_{min}/(H + 1.5\,D)$	EXP x/F	Stability x (km)	Exp. quality Far terrain
Bushland Mayer (1975)[6]	0.36 ± 0.03 2.4	3 ≈1.1	3 U >3.2	Yes >10	Acceptable Namibia
Bushland Chen Fazu & Schwerdtfeger (1989)	0.43 1.8	2.3 ≈1	EC(+6 U) 22	Yes >12	Good Australia
Continuous bushland	0.35–0.45				
Pine forest Leonard & Federer (1973)	1.0 ± 0.3 9.6	11.8 ± 1.6 1.0	5 U OK?	Yes OK?	Excellent Forest
Pine forest (Thetford) Thompson (1979)	1.0 12	15.4 0.9	5 U 1.4	Yes 3	Good Forest
Pine forest (Thetford) De Bruin & Moore (1985)	1.3 12.7	18.5 1.0	EC + 6 U 2.2	Yes 3	Excellent Forest
Pine forest Gash (1986)	≈1.0[7] (7)	10 ± 1 1.1	GU 1.5	Yes 0.7	Good Heather
Pine forest Bergen (1987)	1.65 17	27 1.0	6 GU >0.7	Yes >1	Good Forest
Mature pine forest	0.8–1.6				
Tropical forest Pinker & Holland (1988)	1.8 ± 0.3 27	20–35 <1	3 U 2.9	Yes 4	Acceptable Hilly forest
Tropical forest Viswanadham *et al.* (1990)	2.2 ± 0.1 31	≈35 1.0	5 U **0.2**	Yes 0.3	Good Forest
Tropical forest	1.7–2.3				

[6]For background information see Walk and Wieringa (1988).
[7]Derived from σ_u/U.

shaped (e.g., Leningrad (Borisenko, 1977), or Tokyo Tower), and quite unsuitable as wind observation platforms except at the very top. Regarding the observations of Jensen (1958), a visit to Vor Frelsers church showed that its surroundings and his experimental setup were acceptable.

Publications on wind flow in cities almost never allow an estimate of D, and are quite vague in their fetch descriptions. Therefore in Table VII the $z_{min}/(1.5\,H)$-criterion was used to estimate the acceptability of observation heights, and sufficiency of fetch has been decided subjectively. For the latter purpose either a published overview photograph, a small-scale map, or a rather detailed description of the surroundings was required.

TABLE VII

Roughness of rather homogeneously built-up areas

Surface type Reference author(s)	z_{00} d	H $z_{min}/(1.5\,H)$	EXP observation level(s)	Stability	Exp quality Far terrain
Scattered low buildings Shiotani (1962)	0.5 ± 0.2 –	5 ± 3? ≈ 1?	4 U 9, 18, 31, 48	Yes	Acceptable Tokyo suburb
Low buildings and trees Duchêne-Marullaz (1979)	≈ 0.7 –	≈ 8 ≈ 1	7 U 15, 20, 30, 40, 50, 55, 60	Yes	Acceptable Nantes
Regular dense low houses Steyn (1982)[8]	0.5 ± 0.1 (3.5)	≈ 5 ≈ 3	EC 30	Yes	Good Vancouver
Dense low building ("suburb")	0.4–0.7				
"Regular" city buildings Jensen (1958)[9]	≈ 1.3 (12)	(15–20) ≈ 1.0	3 U 34, 50, 74	No	Acceptable København
"Regular" city buildings Brook (1972)	1.1 ± 0.4 ?	≈ 10 ≈ 1.8	GU 28	No	Acceptable Melbourne
"Regular" city buildings Karlsson (1986)	1.0 ± 0.6 (8)	8–18 ≈ 1.5	4 U 35, 50, 75, 100	Estimate	Acceptable Uppsala
"Regular" city buildings Yersel and Goble (1986)[10]	0.9 ± 0.3 ?	≈ 10 ≈ 3	EC 30	Yes	Good Worcester
Regularly-built town	0.7–1.5				

[8]For background information see Kalanda *et al.* (1980).
[9]For background information see Jensen (1968).
[10]With short list of other city roughness references.

7.4. Summary of Results

The acceptable published evaluations of homogeneous field roughness (notation: z_{00}), listed in Tables II–VII, are summarized in Table VIII. On purpose, group uncertainties have been given as rounded-off ranges of the summarized experiments, because the small size of most groups does not allow the determination of reliable standard deviations.

For open water and similar mobile surfaces, where Charnock's relation is generally used, a typical value for $U_{10} \approx 10$ m/s has been entered, referring for refer-

TABLE VIII

Roughness lengths of homogeneous surface types (z_{00})

Surface type	Roughness length (m)	Number of references
Sea, loose sand and snow	≈0.0002 (U-dependent)	17
Concrete, flat desert, tidal flat	0.0002–0.0005	5
Flat snow field	0.0001–0.0007	4
Rough ice field	0.001–0.012	4
Fallow ground	0.001–0.004	2
Short grass and moss	0.008–0.03	4
Long grass and heather	0.02–0.06	5
Low mature agricultural crops	0.04–0.09	4
High mature crops ("grain")	0.12–0.18	4
Continuous bushland	0.35–0.45	2
Mature pine forest	0.8–1.6	5
Tropical forest	1.7–2.3	2
Dense low buildings ("suburb")	0.4–0.7	3
Regularly-built large town	0.7–1.5	4

ences to Garratt (1977a). Displacement heights for rough terrain are not listed for two reasons. First, the d-parameter is independent of z_0 (though correlated to it), and e.g., two dense forests can have the same z_0-value even though their trunk lengths and their d-values differ by, say, a factor two. Second, d-values are of little interest to modellers at meso- or macro-scale, who just require z_0 (or a z_0-derived C_D-value at their surface-layer reference height) to parameterize turbulent stress in the ISL. However, d-values are often needed for experimental analysis or for local-scale profile modelling, and in skimming flow a good working estimate for this purpose is $d \approx 0.7\,H$ (see Brutsaert, 1975a, and Section 4.2).

7.5. Discussion of the Selection Process

The working principles of this study act in opposite directions. The basic purpose, to make public, as much as possible, existing valuable work on terrain roughness of the past thirty years, was achieved by compiling many hundreds of relevant publications in various languages, both experimental and methodological. It is unlikely that much from the open literature was missed. The second principle was quality selection of the experimental results and this meant that only part of these actually could be used in this survey. Many projects failed to meet the objective criteria, particularly the necessary matching of the observation height interval with fetch and canopy height.

Moreover, frequently the reason for disregarding some study was poor reporting of observation methods and of circumstantial information, making it impossible even to guess if criteria of instrumentation, exposure, fetch, etc. had been met. A glaring example must be mentioned here, because it has been misapplied much too often. For agricultural roughness, most reviews quote only Szeicz *et al.* (1969). Whoever reads that study will discover that it is impossible to find observation heights, actual fetches, reported growth stages and the analysis method. Reported displacement heights of potatoes and lucerne are suspiciously high. Added "results" of Tanner and Pelton (1960) prove to be untraceable "personal communications". Forget these studies.

Three psycho-sociological barriers in the publication process block the provision of sufficient experimental information. First, the author has to devote some valuable writing time and space to reporting on seemingly obvious issues, which are not really exciting new scientific results, such as instrumental exposure or the surroundings of the site. He (or she) ought to realize that readers cannot judge the scope and reliability of the results without such documentation. For wind experiment reporting requirements, a checklist is given elsewhere (Wieringa, 1983).

Second, journal reviewers and editors (or report-publishing institutes) should encourage some inclusion of experimental background information, instead of suppressing it in the name of conciseness. And finally, the reader should look actively for the presence of reliability information, and not blithely use the "results" in cases where such information is omitted.

8. Homogeneous and Heterogeneous Landscape Roughness

In meteorological analysis at a scale larger than several hundreds of meters, it is common to encounter terrain changes; then we cannot gauge the overall terrain roughness by way of Table VIII. Therefore we must look briefly at the effect of heterogeneity on roughness, before we can check any roughness reviews which are meant to be used for estimating mesoscale roughness. Moreover it is worthwhile to investigate whether for averaging of landscape roughness some optimal integration area size does exist.

8.1. Roughness of Heterogeneous Terrain

Inhomogeneous terrain consists of surface patches with different roughness, some of which are scattered obstacles such as shelterbelts or houses. Horizontal averaging of roughness across the landscape for determination of average stress then only becomes feasible at and above the blending height z_*, where flow retardation influences from different terrain patches have merged (Wieringa, 1976). At this height, horizontal homogeneity is approached, as in small-scale flow at the top of the TSL above large roughness (Section 4.1).

The analogy between heterogeneous terrain and the TSL can be extended to the value of the blending height. For the TSL, z_* was shown to depend on the height H of the roughness elements as well as on the average dimension D of element interspaces. For heterogeneous terrain, Wieringa (1976) estimated $z_* \approx 2\,H_{\max}$ ($\approx$60 m for typical landscapes), and this estimate was supported in an experimental analysis by Beljaars (1982) for Cabauw, where $H \approx 15$ m. Mason (1988) then derived in a more general analysis that $z_* \sim 0.01\,D$, without committing himself to a precise ratio. Claussen (1990) then showed that the sum of mean flow deviations from local equilibrium and from horizontal homogeneity should be minimal at some height which can be approximated by

$$z_* \approx 0.7\langle z_0 \rangle (D/\langle z_0 \rangle)^{0.8}, \tag{18}$$

and computed $z_* \approx 50 \pm 10$ m for some typical cases of inhomogeneous terrain. His full formulas are given in a subsequent paper (Claussen, 1991).

The landscape-averaged stress then relates to an effective roughness z_{0e} (Fiedler and Panofsky, 1972), obtainable from area-weighted averaging of drag coefficients $C_{D(z*)}$ (Wieringa, 1986; Mason, 1988). Since this includes a squaring operation, it follows mathematically from Schwarz's inequality that z_{0e} of heterogeneous terrain exceeds the area-weighted arithmetical average $\langle z_{00} \rangle$ of the z_{00}-values of individual patches, because relatively rough patches contribute more than their area fraction to the integral effective roughness.

This bias towards greater roughness is due to the fact that it is easier to charge the atmosphere with turbulence than to discharge it by dissipation. Average flow at the blending height is relatively fast over rough surfaces and slow over rough surfaces, because complete equilibrium is only established at scales >10 km (Jensen, 1978). But since the stress at z_* adapts faster to the underlying surface than the average flow, the increases of $C_{D(z*)}$ over rougher patches dominate its decreases over smoother patches, leading to an overall increase in average drag at height z_* (Claussen, 1990; Wood and Mason, 1990).

For the case of long rough patches that are parallel to the flow, retardation over their extra roughness generates downward flow over their centre and outward flow at the sides, as was shown in the wind tunnel by Vermeulen (1986) and Cermak and Edling (1987). Due to this circulation above the roughness-islands, the net downward flux of horizontal momentum is augmented. Again this implies that the large-scale effective roughness generated by the entire landscape exceeds the area-weighted arithmetical z_{00}-average.

Concluding, in semi-smooth and wake-interference terrain situations, the effective values of heterogeneous roughness, z_{0e}, will never be less than $\langle z_{00} \rangle$ and certainly significantly higher than the z_{00}-values of the open patches. In other words, Table VIII contains for any type of surface cover the *smallest* value of effective roughness, which can occur in terrain situations where such cover is dominant; usually the terrain will be rougher due to heterogeneity. This is clearly shown by data from high masts with inhomogeneous upwind roughness (Beljaars, 1982; Wieringa, 1992). For example, farmland with mature low crops (Table VIII: $z_{00} \approx 0.05$ m) can never have an effective roughness z_{0e} smaller than 0.05 m, and it is more likely that for such farmland z_{0e} is at least 0.1 m, due to drag by occasional trees or hedges.

Paradoxically, *measured* wind gradients over smoother parts of heterogeneous terrain are often less than the appropriate values (Peterson, 1969a; Andreopoulos and Wood, 1982). One possible cause is that after a roughness change the stress overshoots the downwind equilibrium value temporarily, so in experiments over semi-smooth sites with rather short fetches to rougher surroundings, the gradient will be too small (Peterson 1969b, 1971). Brooks (1961) already indicated that the resulting underestimation of shear stress could easily attain a factor two. Another possible cause of underestimation of z_0 in field projects over very rough surfaces is use of profile levels below z_*, too close to the roughness elements. A possible third cause of underestimation of roughness may be that most profiles are measured in daytime, and the necessary corrections for slight instability are sometimes not applied. Large anemometer overspeeding also diminishes measured wind gradients. Concluding, when low-level wind profiles are measured over "homogeneous" sections of heterogeneous terrain, all possible bias sources act concurrently to give a low estimate of the roughness length of the "homogeneous" surface.

8.2. Optimal roughness averaging area for landscapes

General meteorological boundary-layer modelling requires an average roughness, integrated over a domain which is well-matched to the modelling height z. It is practical to choose the latter somewhere within the inertial sublayer, where the wall law still applies and flow structure is reasonably well correlated to surface-based observations in the majority of atmospheric conditions. Experimentally it has been shown that stability-dependent upward wind speed transformation becomes increasingly unreliable above $\approx$50 m to 80 m (Cats, 1979; Holtslag, 1984). This is related to the phase change of the average diurnal course of wind speed at 60 to 80 m (Wieringa, 1989). For model description of the ISL, it is therefore sensible to choose a reference level which is not far above its lower boundary, i.e., close to the blending height z_*, which was shown above to be at $\approx$50 m in heterogeneous terrain. The matching upwind fetch is several kilometers (Pasquill, 1972; Schmid and Oke, 1990). In this context it is interesting to note that the Coriolis force ought to have no significant influence on wind flow in the ISL, and that in diffusion experiments the Coriolis-caused crosswind effects do become important for horizontal travel distances larger than $\approx$5 km (Snyder, 1972).

A "local" integration scale of several kilometers squared seems then to be the optimal integration domain for routinely averaging effective roughness of landscapes (Carson, 1986; Wieringa, 1986; Mason, 1988; Agterberg and Wieringa, 1989; Claussen, 1990). In roughness integration at this scale, the effects of small escarpments and hollows must be included. Such small-scale effects are implicitly accounted for in turbulence- and gustiness-derived z_0-data, which is one more reason why such data give the effective roughness rather than the small-scale roughness (Wieringa, 1983; Beljaars, 1987a).

Occurrence of mountains, coastal cliffs or long large slopes makes the flow multidimensional, adding other means of vertical transport of momentum besides boundary-layer turbulence. Though mesoscale flow over orography will sometimes appear logarithmic up to large heights (Kustas and Brutsaert, 1986; Grant and

Mason, 1990), its parameterization with only effective roughness seems too simplistic. More complicated modelling may then be needed to account for e.g., pressure drag, or local flows (Emeis, 1990; Tieleman, 1992).

9. Verification of Roughness Classifications

Handbooks on the atmospheric boundary layer usually contain some list of "typical" roughness. Most such lists describe their entries with a single word, usually some form of surface cover, and assign to it a roughness length from a single reference. It may be supposed that such lists are meant to indicate the roughness of small-scale homogeneous patches of terrain, and so they can be checked against the ranges in Table VIII for reliability. Often, the z_0-values in such lists prove to be rather low, but occasionally also much too high. For example, Baumgartner (1956) published a single averaged-profile pine forest roughness ($H \approx 6$ m, $d = 0.1$ m (!!), giving $z_0 = 2.9$ m), and this strange result was copied literally into several handbook-lists. A quite reliable list of small-scale roughness values is given by Oke (1978), as will be shown below.

For larger-scale modelling of atmospheric flow, it is necessary to assign roughness values to large terrain areas from e.g., map evaluation. For guidance of such estimates, it is usual to range landscape types into several roughness classes. Direct verification of these classifications against Table VIII is only feasible for approximately homogeneous terrain, i.e., in cases of smooth or skimming flow. So the extreme ends of a good roughness classification, describing either open or very rough terrain, should agree with Table VIII.

For intermediate heterogeneous landscapes, with semi-smooth or wake-interference flow, two approaches exist. Some classifications, such as Cook (1985), are designed for use in modelling procedures which deal explicitly and separately with obstacles and roughness changes; then roughness of e.g., crops should simply comply with Table VIII. The approach of other classifications is to assign a single (effective) roughness estimate to the entire landscape. It was shown above that classification values for intermediate roughness then ought to exceed relevant z_{00}-values. In both approaches, roughness lengths listed in handbook reviews should be greater than those given in Table VIII.

In Table IX four well-known roughness classifications are summarized in order to investigate their validity. These four have been selected, because they all deal with the full range from low to high roughness, and they all have sufficient subdivisions to allow comparison. Davenport's (1960) classification was originally not specified in terms of roughness lengths, but instead as (height-dependent!) exponents p of the power "law" profile, related to z_0 by

$$p \approx 1/\ln[(\sqrt{z_1 z_2})/z_0], \tag{19}$$

where z_1 and z_2 are the two compared wind profile levels (Panofsky, 1977). For

TABLE IX

Roughness in classifications ($z_{0(\mathrm{class})}$) and homogeneous z_{00}. ($\approx$ implies that terrain correspondence is only approximate).

Terrain	z_{00}	Oke (1978)	Davenport (1960)	ESDU Smedman (1978)	Cook (1985)	Troen (1987)
Flat snow	0.0003	≈0.0002	–	0.0001	0.003	0.001
Flat land	0.0003	0.0003	0.006	0.005	0.003	0.0003
Fallow ground	0.002	0.001–0.01	0.015	–	0.01	0.005
Smooth, $\langle z_{0(\mathrm{class})}/z_{00}\rangle$		1.1	14	(1)	8	2.2
Short grass	0.013	0.003–0.01	0.015	0.008	0.01	0.01
Long grass	0.034	0.04–0.10	0.04	0.02–0.05	0.01	–
Cropped farmland	≈0.1	0.04–0.20	0.11	0.05–0.10	0.03	0.05
Semi-smooth, $\langle z_{0(\mathrm{class})}/z_{00}\rangle$		1.1	1.1	0.7	0.5	0.6
Pine forest	1.2	1.0–6.0	0.8	0.4	0.3	0.30
Low suburb	0.6	–	≈1.3	0.4–0.6	0.3	0.40
Regular town	1.1	–	≈1.3	0.6–0.9	0.8	–
Skimming, $\langle z_{0(\mathrm{class})}/z_{00}\rangle$		(2)	1.3	0.6	0.5	0.5

translation of Davenport's p-classes into z_0-values, the height interval 10–80 m was chosen because his context was engineering application, where wind is adapted upward from station observation height to a typical height of wind-sensitive construction works. The so-called "ESDU" classification was never submitted for review to a journal, but is widely used in wind engineering, and is accessible because of its inclusion as an appendix by Smedman-Högström and Högström (1978). A slightly improved version of the classification by Troen *et al.* (1987) is used in the European Windatlas. A fifth non-classified review, the roughness list of Oke (1978), is also included in the comparison.

In Table IX, the left-hand column contains z_{00}-values averaged from Table VIII. To ease comparison with the classifications, the z_{00}-values for low and high crops from Table V are lumped together as "cropped farmland". Any z_0-ranges are averaged logarithmically. For each flow category the average of the roughness ratios, $\langle z_{0(\mathrm{class})}/z_{00}\rangle$, is given.

From Table IX two conclusions can be drawn. First, apart from Oke (1987) and ESDU, all classifications exaggerate z_0-values for smooth terrain. The most likely reason for this is that the classifiers used field data with fetches that were too short (good profile observations over smooth surfaces require more than a kilometer of fetch!). Second, both for semi-smooth terrain and for forests and small towns the classification of Davenport (1960) and the z_0-list of Oke (1978) give plausible results, while Cook (1985), Troen *et al.* (1987) and "ESDU" underestimate roughness by about a factor two. The limited available high-mast information for heterogeneous terrain (Wieringa, 1992) gives similar conclusions for the same classifications. Davenport's classification appears to be reasonably reliable, if lower

z_0-values are assigned to its lowest classes, which was done in updated versions (Wieringa, 1977, 1992).

The latter three classifications are not the only ones to underestimate roughness; actually, most such reviews (e.g., Biétry *et al.*, 1987; Stull, 1988) assign too-low z_0-values to moderately rough and very rough terrain. Unfortunately, many classifications rely on references which are too few and too old, and often they are just copied from some previous handbook. It is interesting to see that results of such uncritical borrowing do not just show random uncertainty, but rather have a systematic underestimation bias. Likely reasons for such a bias were discussed earlier in Section 8.

10. Conclusions

The main result of this study is contained in Table VIII, a summary of roughness lengths z_{00} of homogeneous natural surfaces, for application in various analyses of vertical turbulent transfer of horizontal momentum. It should be noted that for calculating vertical exchange of temperature (or other scalars) the appropriate thermal roughness lengths can be smaller than z_{00} by orders of magnitude (Brutsaert, 1975b; Beljaars and Holtslag, 1991).

The z_{00}-summary is based on very hard facts, some four dozen quality-selected experimental projects, deserving recognition and further use by others instead of library oblivion. In the course of proving the quality of these projects, some criteria have been developed which may be useful for judging the value of other projects, as well as for planning new experiments. For homogeneity, it seems advisable to use a fetch-to-observation height ratio of $\sim$100, dependent on local roughness (Figure 3). Also, observations made below the blending height $z_*(\sim 1.5H)$ are not acceptable for roughness determination, not even when displacement height corrections are used.

The tabulated roughness values have been used to verify some popular classifications and lists of roughness of arbitrary non-complex terrain. The conclusion is that the small-scale list of Oke (1978) and the local-scale classification of Davenport (1960) are reliable, provided that in the latter, its lowest two roughness classes are adjusted (Wieringa 1986, 1992). Other reviews underestimate most roughness lengths by approximately a factor two.

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