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An analytical representation for the Pasquill-Gifford-Turner σ_z curves for elevated sources

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RESUMEN

Representaciones gráficas de esparcimiento vertical de la pluma σ_z requiere de digitalización, en la implementación de cálculos numéricos del esparcimiento de la pluma. La representación gráfica actual para estas curvas (desarrolladas a partir de las curvas originales de Pasquill-Gifford-Turner) se centra en la determinación de regresiones de mejor ajuste junto con otra evidencia empírica. Aquí se propone para la cuantificación de las curvas de PGT en condiciones neutras y estables, un modelo conceptual, validado observacionalmente, y basado en la teoría de mezclado turbulento.

ABSTRACT

Graphical representations for the vertical plume spread parameter, σ_z, require digitisation for implementation in numerical calculations of plume spread. Current algebraic representation for these curves (developed from the original Pasquill-Gifford-Turner curves) centres on determining a "best-fit" regression together with other empirical evidence. Here an observationally validated conceptual model based on turbulent mixing theory is proposed for the quantification of the PGT curves for neutral and stable conditions.

1. Empirical (algebraic) representations for σ_z

Although the Pasquill-Gifford-Turner (PGT) curves for the lateral and vertical plume spread rate parameters, σ_y and σ_z, have been much criticized, they are still in common regulatory use (see e.g., Irwin, 1983). The original curves (Pasquill, 1961; as modified by Gifford, 1961 and Turner, 1967) were constructed graphically and were useful for hand calculations. However with the advent of digital computers it has become necessary to represent these observational curves with algebraic formulae. Much effort has been placed into "best-fit" representation, and to the extension of the range of applicability of the original curves to take into account varying topography/roughness length, release height etc. (e.g., Smith, 1972; Briggs, 1973; Irwin, 1979a) in order to satisfy the regulatory agencies' requirements for a "working model".

However relatively few attempts (e.g., Högström, 1964) have been made to relate these curves to a conceptually derived model; as will be undertaken in this paper by relating the vertical spread rate parameter, σ_z, to eddy diffusion coefficients using assumptions compatible with the limitations inherent in the original data sets.

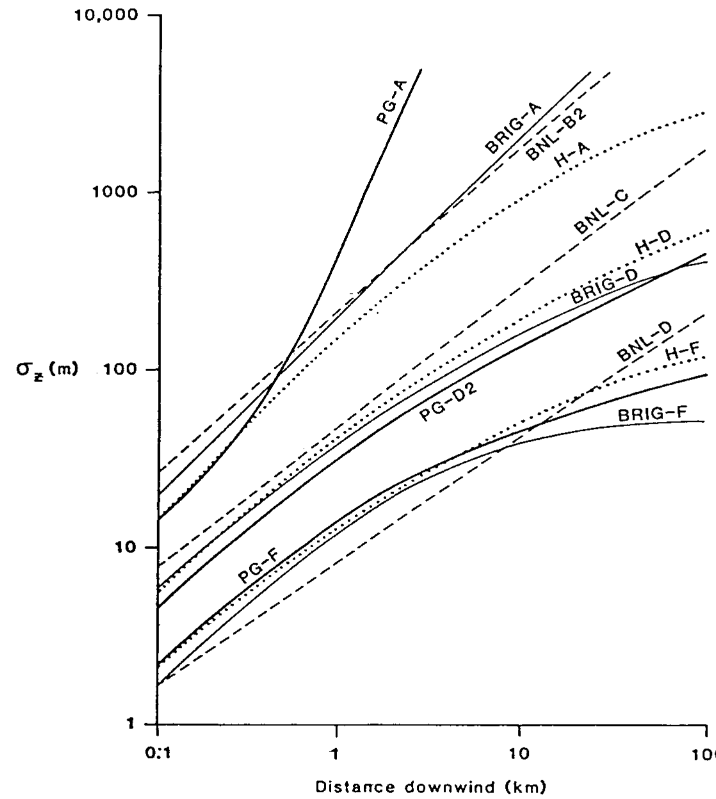

Fig. 1. Curves for σ_z for PG stability classes based on Pasquill (1961) and Gifford (1961) ($z_0 = 0.03$ m) (redrawn from Turner, 1970) plus best-fit curves from equations (1-4). [PG = Pasquill-Gifford; B = Briggs; BNL = Brookhaven National Laboratory; H = Hosker].

The original σ_z curves (Pasquill, 1961; Gifford, 1961) (Fig. 1) were for non-buoyant ground level releases, for distances to only 1 km downwind and for a surface roughness of 0.03 m. Representations for these curves have been attempted by use of a power law expression:

$$\sigma_z = cx^d \tag{1}$$

(e.g., Weber, 1982, p117; Pasquill and Smith, 1983, p338). The values of the coefficients were first estimated specifically for surface-based releases, and modified by Smith (1972) to take into account different roughness lengths (Table 1). F. B. Smith's (1972) graphical method may also be quantified:

Table 1. Values of constants for equation (1) (with x in km) (after Pasquill and Smith, 1983)

Stability Category	Coefficient c			Cofficient d		
$z_o =$	1 cm	10 cm	1 m	1 cm	10 cm	1 m
A	0.102	0.140	0.190	0.94	0.90	0.83
B	0.062	0.080	0.110	0.89	0.85	0.77
C	0.043	0.056	0.077	0.85	0.80	0.72
D	0.029	0.038	0.050	0.81	0.76	0.68
E	0.017	0.023	0.031	0.78	0.73	0.65
F	0.009	0.012	0.017	0.72	0.67	0.58

a well-known analytical "best-fit" expression is that of Hosker (1973):

$$\sigma_z = \frac{ax^b}{1+cx^d}F(z_o,x) \tag{2}$$

(although better algebraic fits are possible: F. B. Smith (*p.c.*, 1985)). In equation (2) the values of a, b, c, d (Table 2) depend on stability and $F(z_o,x)$ is a function of roughness length:

$$F(z_o,x) = \begin{cases} \ln[fx^g(1+(hx^j)^{-1})], & z_o \geq 0.1 \text{ m} \\ \ln[fx^g(1+hx^j)^{-1}], & z_o < 0.1 \text{ m} \end{cases} \tag{3}$$

and f, g, h, j (Table 3) are functions of roughness length only.

Table 2. Coefficients a, b, c, d for use in equation (2)

PG stability category	a	b	c	d
A	0.112	1.06	5.38×10^{-4}	0.815
B	0.130	0.950	6.52×10^{-4}	0.750
C	0.112	0.920	9.05×10^{-4}	0.718
D	0.098	0.889	1.35×10^{-3}	0.688
E	0.0609	0.895	1.96×10^{-3}	0.684
F	0.0638	0.783	1.36×10^{-3}	0.672

Table 3. Values of coefficients f, g, h, j for equation (3)

Roughness length (m)	f	g	h	j
0.01	1.56	0.0480	6.25×10^{-4}	0.45
0.04	2.02	0.0269	7.76×10^{-4}	0.37
0.1	2.72	0	0	0
0.4	5.16	-0.098	18.6	-0.225
1.0	7.37	-0.0957	4.29×10^{3}	-0.60
4.0	11.7	-0.128	4.59×10^{4}	-0.78

However, it should be noted (NRPB, 1979) that this numerical scheme can only be used for a specific subset of possible values for roughness length; although these values are not stated explicitly.

In order to estimate the corresponding coefficient values for elevated releases, M. E. Smith (1968) analysed the 108 m (non-buoyant) release of the Brookhaven National Laboratory (USA) (BNL) data set to derive coefficients valid in the range 0.1 to 10 km for 4 stability classes (Table 4). Many other workers have attempted to derive coefficients from various data sources - a more detailed assessment of the historical development of these differing approaches is to be found in the excellent review of Gifford (1976).

Table 4. Values of constants for equation (1) (with x in meters)(after Hanna *et al.*, 1982)

Stability category		Parameter	
Brookhaven	Equivalent PG	c	d
B2	B	0.41	0.91
B1	C	0.33	0.86
C	D	0.22	0.78
D	F	0.06	0.71

Furthermore, several attempts at rationalisation have been made. Briggs (1973) proposed a series of algebraic interpolation formulae based on a wide variety of data sources (including the BNL data) containing surface and elevated sources with a range of initial buoyancies:

$$\sigma_z = b_1 x (1 + b_2 x)^c \tag{4}$$

The coefficient values were derived for both rural and urban terrain and are given in Table 5. It is worth noting that, in neutral conditions, from equation (4) for small values of $x, \sigma_z \ \alpha \ x$; and at large distances $\sigma_z \ \alpha \ x^{1/2}$ - a range of exponent values encompassing the values given in Tables 1, 2 and 4.

Table 5. Parameter values for equation (4) (after Briggs, 1973)

PG stability category	a) Open country conditions			b) Urban conditions			
	b_1	b_2	c	b_1	b_2	c	
A	0.20	0	-	A-B	0.24	0.001	0.5
B	0.12	0	-				
C	0.08	0.0002	-0.5	C	0.20	0	-
D	0.06	0.0015	-0.5	D	0.14	0.0003	-0.5
E	0.03	0.0003	-1	E-F	0.08	0.00015	-0.5
F	0.016	0.0003	-1				

In order to provide a baseline against which to test theoretically derived value of σ_z (Section 2 *et seq.*), these algebraic representations of the various graphical methods have been plotted in Fig. 1 along with the original PG curves. The power law expression, equation (1), as evaluated by F. B. Smith (see Pasquill and Smith, 1983, p338) is noted to underestimate (by 10%) at 0.1 and 10 km and to overestimate (by 5%) at $x = 1$ km. Furthermore, comparison of Hosker's curves with M. E. Smith's (1968) (despite being for a completely different data set) reveal strong similarities and only differ significantly at distances of $\gtrsim 5$ km. Briggs' (1973) values (equation (4)) tend to underestimate (as compared to Hosker's values) at a similar distance range. It should also be noted that the curves of F. B. Smith relate to near-surface releases, whereas M. E. Smith's, Hosker's and Briggs' relate to higher level (~ 100 m) releases.

2. σ_z in terms of eddy diffusion coefficients

A relationship for the plume standard deviation, σ_z, can be derived from equation 6.102 of Kao (1984) by noting that for the term $exp[-az^{2+m-n}/b(2+m-n)^2x]$ to be compatible with the normally assumed 'Gaussian plume model':

$$\sigma_z^2 = \frac{b(2+m-n)^2 x\, z^{n-m}}{2a} \tag{5}$$

when the profiles of wind, $U(z)$, and eddy diffusivity, $K(z)$, are given by power law profiles:

$$K(z) = bz^n \tag{6}$$

and

$$U(z) = az^m \tag{7}$$

and where x is the distance downstream from the source and z is the height above the surface. [The corresponding equation for the σ_y component is not discussed here]. It should be noted that this general result (equation (5)) is *not* restricted to ground level sources.

In the special case of $m = n$, this gives

$$\sigma_z^2 = (2b/a)x \equiv (2K/U)x \tag{8}$$

which is equivalent to the Fickian diffusion approach in which, additionally, $m = n = 0$ (viz. $K(z)$ = constant = K and $U(z)$ = constant = U). Hence, in this special case, $\sigma_z \; \alpha \; x^{1/2}$. However, such a functional dependence is not borne out in any of the empirical curves discussed above except at large distances downstream (F. B. Smith, *p.c.*, 1985).

A more realistic approximation to the K and U profiles requires the velocity profile to depend upon both atmospheric stability and surface roughness length. This can be accomplished in several ways: using the logarithmic profile or including such dependency in the coefficient a and/or m of equation (7) (see e.g., Irwin, 1979b). The parameter a must incorporate a reference height; for example it is common to refer values to those at a height of 10 m such that

$$U(z)/U_{10} = (z/10)^m \tag{9}$$

However for the theory to be developed here, the reference value will be taken as the geostrophic velocity, U_g, which is assumed to occur at the top of the boundary layer of depth, h, such that

$$U(z)/U_g = (z/h)^m \tag{10}$$

This is the "top down" profiling method of Moran *et al.* (1985). In a theoretical discussion (as here) these two power law profiles should have identical slopes such that the exponent m in equations (9) and (10) must be identical. However, it should be noted that if a single theoretical curve is fitted to a set of real data it is usually only possible to ensure that the values of U and $\frac{dU}{dz}$ are in agreement at a single height, z_1, which becomes a factor in determining the value of m i.e., m is a function of this reference height, z_1 (e.g., Gryning *et al.*, 1983) as well as of the choice between top down and bottom up profiling method.

Furthermore it is commonly assumed that K can be given by

$$K = K_o\ f(R_i) \tag{11}$$

where R_i is the Richardson number and K_o is the neutral value of the eddy diffusion coefficient given by e.g., the "law of the wall" relationship (Mellor and Yamada, 1974)

$$K_o = ku_* z \tag{12}$$

(where k is Von Karman's constant, u_* is the friction velocity and z the height), then equation (5) can be rewritten as

$$\sigma_z^2 = ku_*(1+m)^2 xz^{1-m} h^m f(R_i)/(2U_g) \tag{13}$$

Furthermore, since u_* (which is a measure of surface roughness and therefore related to the roughness length, z_0) is proportional to the wind speed U, the value of σ_z becomes a function of distance, height and stability class, but only a relatively weak function of wind speed - via the mixing depth, h. This independence of wind speed is one of the assumptions embodied within the Pasquill-Gifford-Turner curves. However it should be noted that under several stability-typing schemes, there is an implicit dependence on wind speed insofar as the Pasquill-Gifford (PG) stability class is itself a function of wind speed. Such classification schemes incorporate both the variables (wind speed, temperature gradient) which can be alternatively synopsised in the Richardson number, R_i.

3. Pasquill-Gifford-Turner curves

Although equation (13) is the main conclusion of this analysis, it is considered useful to evaluate this general formula for the Pasquill-Gifford-Turner (PGT) σ_z curves for specific stability cases. This can be done by substituting appropriate expressions for the boundary layer depth, h, and the Richardson number, R_i. Furthermore, for any given stability, the effect of varying roughness lengths can be evaluated either directly in terms of different values of u_* or by employing an (analytic) relationship between u_* and the roughness length, z_0 (e.g., based on the geostrophic drag law: Jensen, 1978; Tennekes, 1982, p46).

As two examples of the applicability of the analysis undertaken in Section 2, in Section 3.1, equation (13) will be considered for neutral (PG class D) stability in which $R_i = 0$ and $f(R_i) = 1$ and in Section 3.2 stable cases ($f(R_i) < 1$) will be discussed.

3.1 Neutral stability case (PG class D)

In the case of a neutrally stable boundary layer, further relationships are known which can be used to simplify the expression for σ_z (equation (13)). The neutral boundary layer depth can be related to the surface friction velocity, u_* (e.g., Shir, 1973):

$$h = u_*/4f \tag{14}$$

where f is the Coriolis parameter. Secondly the value of the Richardson number is unity and $f(R_i)$ is also unity (i.e., $K = K_o$). With these substitutions, equation (13) is rewritten as

$$\sigma_z^2 = ku_*^{1+m}(1+m)^2 xz^{1-m}/(2U_g(4f)^m) \tag{15}$$

Furthermore, values of the exponent, m, are known as a function of z_o, for specified height ranges (e.g., Irwin, 1979b). For a typical surface roughness length of $\sim$ 0.1m, m is frequently taken as 0.14. With this value and a value of $f = 10^{-4}s^{-1}, \sigma_z^2$ can be written as

$$\sigma_z^2 = 1.943ku_*^{1.14}z^{0.86}x/U_g \tag{16}$$

For neutral conditions, the value of the friction velocity, u_*, is related to the roughness length z_o, and the geostrophic velocity, U_g, by

$$u_* = \frac{0.5U_g}{\ln\,(U_g/(z_0 f))} \tag{17}$$

(Jensen, 1978).

Values of σ_z, calculated using equation (16) for $z_o = 0.1$ m are compared in Table 6 with the range of values derived from the analytic-fit curves of Fig. 1 (*viz.* derived from equations (1-4)). It can be seen that there is excellent agreement in the downwind range of $\sim$ (500-5000 m).

However, in contrast to Fickian diffusion where $\sigma_z\ \alpha\ z^o x^{1/2}$, equation (16) leads to the result that

$$\sigma_z \propto z^{0.43}x^{1/2} \tag{18}$$

At large distances, z is approximately constant and σ_z is, as anticipated, proportional to $x^{1/2}$. However, the value of σ_z is now additionally a function of z: *viz.* the stack height for non-buoyant plumes described by the Pasquill-Gifford curves. [This height dependency should also permit future investigation (beyond the scope of this present paper) of differential diffusion rates of the upper and lower portions of the plume as a consequence of the non-constant value for K].

Table 6. Comparison for neutral stability (D) from Fig. 1 of empirical values of σ_z (m) (z_0 = 0.1 m) and calculated values (equation (16)) for reference heights of 50 and 100 m for two wind speeds.

z(m) =	100	500	1000	5000	10000	100,000
i) Ranges for empirical values from Fig. 1 (z_0=0.1 m)	5-8	15-30	30-40	100-200	150-250*	480-700*
ii) Calculated values U_g=2 m s^{-1}(z=50 m) (u_* = 0.0819 m s^{-1})	8.1	18.0	25.5	56.9	80.5	254.6
(z=100 m)	10.8	24.3	34.3	76.7	108.5	343.0
U_g=6 m s^{-1}(z=50m) u_*=0.2255 m s^{-1})	8.3	18.5	26.2	58.5	82.8	261.8
(z=100 m)	11.2	24.9	35.3	78.9	111.6	352.8

* The upper values here neglect the much larger estimates of the BNL curves.

3.2 Stable cases (PG class E, F, G)

In terms of the eddy diffusion coefficient formulation proposed here for the evaluation of σ_z, the inclusion of a non-neutral atmosphere is only possible where σ_z scales with the friction velocity. Hence the approach is likely to be useful for the stable categories, E, F, G; but not useful for the unstable categories when the dominant diffusion mechanism is convection. In such unstable situations, scaling appears to be with the convective velocity (e.g., Pasquill, 1985); consequently the unstable case willl not be discussed here.

For stable cases, the value of the exponent m, the depth of the surface boundary layer, h, and the functional form of the Richardson number, $f(R_i)$, will be different from the neutral case. The depth, h, of the stably stratified boundary layer can be represented by

$$h = 0.4(u_* L/f)^{1/2} \tag{19}$$

(see e.g., Zilitinkevich, 1972, 1975; Hanna, 1982) where L is the Monin-Obukhov length scale. One particular parametrisation of the stability function in terms of the Richardson number (Henderson-Sellers, 1982) gives K as

$$K = K_o(1 + 37R_i^2)^{-1} \tag{20}$$

The atmospheric boundary layer observations by Ueda *et al.* (1981), upon which this formula is based, show that turbulence is not exterminated totally at a critical Richardson number (of $\sim \frac{1}{4}$); although the value is severely damped. The continued existence of turbulence at Richardson numbers in excess of $\frac{1}{4}$ is also supported by observations in aquatic boundary layers by e.g., Thorpe (1969), Mortimer (1974).

Writing the appropriate stable value of the exponent m as m', equation (13) becomes

$$\sigma_z^2 = k u_*^{1+m'/2}(1+m')^2 x z^{1-m'} 0.4^{m'} L^{m'/2}(1+37R_i^2)^{-1}/(2U_g f^{m'/2}) \tag{21}$$

Results for stable cases can either be derived from direct evaluation of equation (21) or can be expressed by the ratio of σ_z in stable cases to the value of σ_z in neutral conditions. In this latter case, dividing equation (21) by equation (15) gives

$$\frac{\sigma_z(stable)}{\sigma_z(neutral)} = \left[\frac{u_*^{(1+m'/2)}}{u_{*n}^{(1+m)}}\left(\frac{1+m'}{1+m}\right)^2 z^{m-m'} 0.4^{m'} L^{m'/2}(1+37R_i^2)^{-1} 4^m f^{m-m'/2}\right]^{1/2} \tag{22}$$

where m is the neutral value of the exponent and subscript n denotes neutral. For stable conditions, equation (17) must be replaced by a geostrophic drag law for non-neutral conditions e.g.,

$$\frac{U_g}{u_*} = \frac{1}{k}\left[\left\{\ln\left(\frac{u_*}{f z_0}\right) - A\right\}^2 + B^2\right]^{\frac{1}{2}} \tag{23}$$

(e.g., Tennekes, 1982, p46) which gives an implicit relationship between u_* and z_o for known U_g. The constants A and B in equation (23) are empirical functions of stability. Typically $\frac{u_{*n}}{u_*} = O(0.6)$.

As an indication of the potential usefulness of equation (22), the ratio $\sigma_z(stable)/\sigma_z(neutral)$ will be evaluated using typical values for the parameter values (which themselves may often be the state variables of a larger air pollution model). Taking $L = O(20\text{m})$ (Hanna, 1982), and $m' = 0.54$ (stability class: F), $m = 0.16$ (for a $z_o = 0.1$ m) (e.g., Irwin, 1979b) together with a corresponding value for the Richardson number of ~ 1.25 (Sedefian and Bennett, 1980), gives a value for this ratio of

$$\sigma_z(stable)/\sigma_z(neutral) = 0.396\frac{u_*^{0.635}}{u_{*n}^{0.580}} z^{-0.19} \tag{24}$$

Values derived from equation (22) for a typical value of $u_{*n} = 0.2255$ (corresponding to $U_g = 6$ m s^{-1} and $z_o = 0.1$ m s^{-1} as before) [and hence assuming $u_* = 0.1353$] and for three different heights are compared in Table 7, for two distances downstream (1 km and 10 km), with the power law representation of the PGT curves using coefficient values of F. B. Smith and M. E. Smith (the latter specifically for elevated sources in the distance range of 0.1 to 10 km) as given in Pasquill and Smith (1983, p338). It is seen that the agreement with M. E. Smith's values are good for the high level releases, whilst the surface-release based coefficient values of F. B. Smith, which are much larger, are underestimated by the near-surface calculation ($z = 10$ m) which gives a ratio value of 0.17 in comparison with Smith's value of 0.26 for the same roughness length.

Table 7. Ratios of σ_z in stable (class F) conditions of σ_z in neutral conditions (in part, after Pasquill and Smith, 1983)

	SURFACE RELEASES	ELEVATED RELEASES
	F. B. Smith	M. E. Smith
Empirical calculations (equation (1))		
x=1 km	0.32	0.12
x=10 km	0.26	0.11
Calculated ratios from equation (22)		
z=10 m	0.17	
z=100 m		0.11
z=200 m		0.10

4. Discussion

It is, perhaps, worth noting that the validity of the above argument relies on the representation of U and K as power law profiles. However such functional forms are monotonic increasing with height. Since observations suggest that such values will reach a maximum at some height in the atmosphere, z_{max}, and thereafter decrease, several functional forms for K have been proposed to permit this phenomenon to be modelled more accurately. For example, Shir (1973) proposed that

$$K = 0.4u_* z\ exp(-4fz/u_*) = 0.4u_* z\ exp(-z/h) \tag{25}$$

In contrast to the law of the wall which gives a zero value for K_O at the interface and a value monotonically increasing with height, equation (25) gives a maximum at a height above the interface of $z = h$: a height which varies with wind speed (see also figure 9 of Kundu, 1980 and figure 8.3 of Hanna *et al.*, 1982). [This functional form is also supported by theoretical work in the aquatic boundary layer by Henderson-Sellers (1985)].

It is also interesting to note that a truncated version of this formula (equation (25)) has already gained some acceptance (see e.g., Nieuwstadt, 1980; Pasquill and Smith, 1983):

$$K = ku_* z(1 - z/h) \tag{26}$$

The correspondence with equation (25) is most easily seen by undertaking a series expansion for the exponential:

$$K = ku_* \sum_{i=1}^{\infty} (1/h)^{i-1} \frac{(-1)^{i-1} z^i}{(i-1)!} \tag{27}$$

and then rewriting equation (26) in a similar format as

$$K = ku_*(z - z^2/h)$$

$$= ku_* \sum_{i=1}^{2} (1/h)^{i-1} \frac{(-1)^{i-1} z^i}{(i-1)!} \tag{28}$$

The disadvantage with this truncated version is that it is symmetrical about a maximum at $z = h/2$, with a zero value at $z = h$. However many observations and models (e.g., Brost and Wyngaard, 1978; Hanna *et al.*, 1982) suggest that there is likely to be a strong asymmetry (skewness) in the vertical profile of the eddy diffusion coefficient, K. Furthermore, the extent of the mathematical reanalysis that introduction of such modifications (equations (25) or (26)) would require in order to derive an analytical solution to the diffusion equation so that a relationship between σ and z, x etc. analogous to equation (16) can be derived is beyond the scope of this present paper and will potentially provide detail in excess of the degree of accuracy inherent in the original PGT curves. Qualitatively it can be seen that since these modified forms for K describe a smaller value at greater height than the law of the wall equation (equation (12)), the moderation of values of σ in e.g., Table 6 are likely to result in decreases (probably marginal).

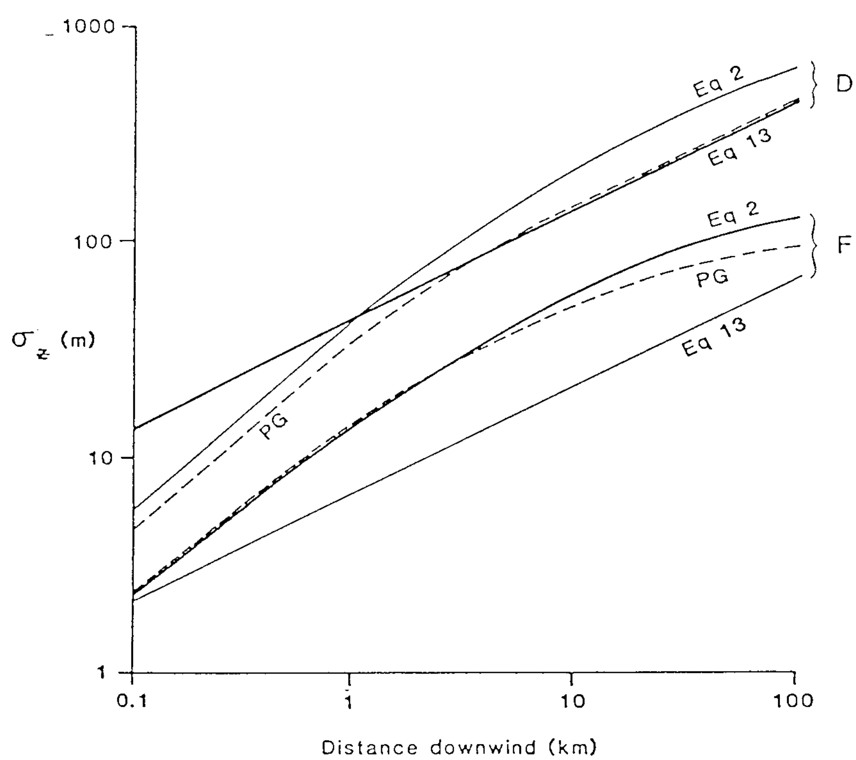

Fig. 2. Curves for σ_z for PG stability classes D and F (approximate Richardson numbers following Sedefian and Bennet, 1980) derived from the analytical expression derived here equation (13) plus Hosker's best-fit curves (equation (2)) and the original PG curves.

5. Conclusions

Graphical representations for the vertical plume spread parameter, σ_z, required for digitisation in numerical calculations of plume spread must be superceded by analytical expressions. Current algebraic representation for these curves (developed from the original Pasquill-Gifford curves) centres on determining a "best-fit" regression together with other empirical evidence. An observationally validated conceptual model has been developed here for the quantification of the PG curves based on turbulent mixing theory - leading to a new algebraic equation (equation (13)) in which the stability class is expressed in terms of the Richardson number. For neutral values (PG stability class D $\equiv R_i = 0$), σ_z can be deduced using equation (15) and stable values for σ_z given either by equation (21) or, as a ratio, by equation (22). For ease of visual comparison, the curves thus derived (for PG classes D and F) are shown in Fig. 2 (for $z = 100$ m and $U_g = 6$ m s^{-1}) for Richardson numbers equated to Pasquill-Gifford stability classes D and F following the scheme of Sedefian and Bennett (1980). (Also plotted in this figure for comparison are the best fit curves of Hosker (1973): equation (2) and the original PG curves). It should be noted that for stable atmospheric conditions, this new approach is not appropriate for heights above the boundary layer; neither does it show the lower growth rates distances in the range 1-10 km, as in for example the Briggs' curves for open country (equation (4) and Table 5). This is related to the fact that such curves implicitly contained an assumption regarding the depth of the boundary layer, which for stable cases is assumed to be < 800 m (see, for instance, Table 2 of NRPB, 1979). Hence, although there are more variables within this new conceptual approach, this added flexibility, derived by consideration of the turbulent mixing processes in the atmosphere and not by simple curve fitting to specific data sets, should render equation (13) appropriate for a wide variety of cases of atmospheric dispersion.

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