CHAPTER 4

Environmental Factors Affecting Dust Emission by Wind Erosion

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ABSTRACT

Factors which influence the production of dust in the Sahara may be described in terms of aerodynamic and soil physical characteristics. Work on the threshold wind velocity for soil erosion, on total soil movement by wind, and on dust production is reviewed. An extremely important factor in desert environments is the aerodynamic partitioning of wind stress by nonerodible elements and erodible soil. The factors influencing dust production are presence of nonerodible elements such as vegetation and rocks, vegetative residue, surface roughness, aggregate structure of surface soil, soil moisture, soil mineralogy and texture, and wind stress.

4.1 INTRODUCTION

Dust production is affected by a wide variety of factors both natural and man-made. I will attempt to summarize some of the physical factors involved in dust emission but will not assign them to natural or man-made causes since in most cases they could be caused by either. I will also emphasize that part of the total transported soil material which can travel far from its place of origin. In order to travel long distances, dust must have a small settling velocity compared to the root mean square vertical velocity fluctuations of the supporting air. Fine dust, i.e. that having a potential for long-distance travel for almost all soil eroding winds, is generally smaller than about 20 μm (0.02 mm) in diameter.

Threshold velocity is one of the most important parameters of wind erosion since it is that parameter, along with the frequency distribution of wind speed, that determines the frequency of wind erosion of soil. The movement of soil by the wind is the other important characteristic of wind erosion that I will discuss. Soil movement is divided into the horizontal flux of sand (which describes most soil mass movement) and vertical flux of dust smaller than 0.02 mm (which describes a large part of permanent soil loss). I will discuss separately the aerodynamic and soil factors as they relate to threshold velocities and soil movement.

4.2 AERODYNAMIC FACTORS IN FINE DUST PRODUCTION

4.2.1 Threshold velocities for soil movement

In general, threshold velocities for soil movement are those velocities in which aerodynamic forces are sufficient to dislodge particles from the soil and initiate movement. Of course, this velocity is dependent on both the aerodynamic forces and the forces holding the particle in the soil. Theoretical studies have considered simple soil systems and idealized particles. Such studies typically equate moments due to lift, drag, and weight.

Experimental studies of threshold velocities for simple soil systems consisting of beds of loose, monodisperse, and similar particles are reported by Bagnold (1941), Ishihara and Iwagaki (1952), Chepil (1951), and Greeley *et al.* (1973). For the sake of illustration, the results* of Chepil (1951) are shown in Figure 4.1. It is to be noted that there is a minimum friction velocity that will produce motion in particles of diameter of about 100 μm. Particles larger than 100 μm require greater wind speed, presumably because of their greater weight, while smaller particles are lower in the boundary layer and require larger pressure fluctuations to initiate their

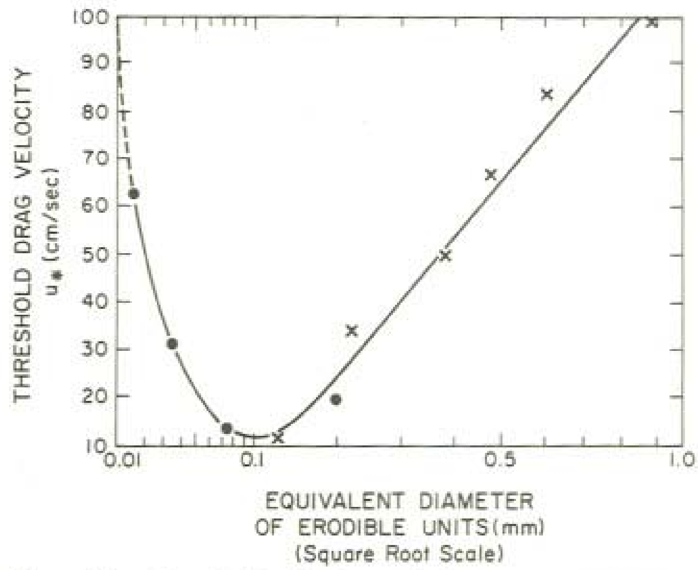


Figure 4.1 Threshold friction velocity *vs.* monodisperse particle size (after Chepil, 1951)

*Wind speeds are reported in terms of friction velocity $u_* = c_D^{1/2} u_1$ where u_1 is the mean wind at height 1, usually within a metre or two of the surface, and c_D is the drag coefficient for that height and that particular surface. For neutrally stratified air, the mean wind profile near the ground takes the form $u_2 - u_1 = (u_*/K) \ln (Z_2/Z_1)$, where u_2 and u_1 are mean wind speeds at heights Z_2 and Z_1 and K is von Kármán's constant.

movement. The main conclusion to be drawn from such data is that if loose particles are available in the soil, 100 μm particles require the lowest velocities for initiation of movement. Once particle movement has begun, momentum for downward soil movement is more effectively delivered by the soil particles colliding with the surface than by aerodynamic transfer. Natural soils, however, are rarely characterized by monodisperse particles and are rarely found in perfectly loose beds. Nonetheless, erosion is often initiated by a small quantity of loose 100 μm particles on a soil surface.

The effect of nonerodible roughness elements such as bushes or pebbles is to absorb some of the momentum being transported to the ground by the wind and thus to decrease the momentum felt by individual particles. Marshall (1971) and Lyles and Allison (1976) have made studies where the increase of threshold wind velocity due to nonerodible roughness could be calculated. Marshall measured partitioning of the momentum of roughness elements and the floor. This study was concerned mostly with the effects that the large-scale roughness elements such as bushes and boulders have in absorbing wind force. His roughness elements were cylinders having a diameter-to-height ratio of 0.5 to 5, as well as hemispheres. Lyles and Allison were concerned with the effect of standing stubble on threshold

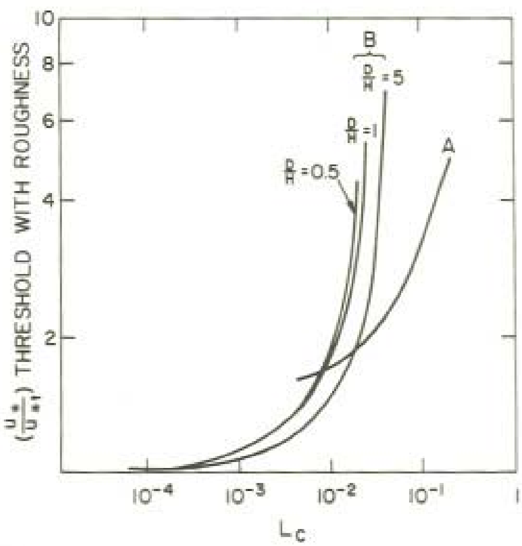


Figure 4.2 Ratio of threshold velocity with nonerodible elements present to that with none present. L_c is the ratio of silhouette area of one nonerodible element to the area of the floor divided by the number of nonerodible elements on the floor. A is the data of Lyles and Allison (1976), B is the Marshall data for elements with diameter to height ratios (D/H) of 0.5, 1, and 5

velocity and so in general were interested in much smaller diameter-to-height ratios. I have expressed the fractional increase of threshold velocity as a function of L_c which is a nondimensional ratio of silhouette area of a roughness element to the floor area occupied by one roughness element. The results are shown in Figure 4.2. In this figure u_*/u_{*t} is the ratio of threshold velocity with roughness elements to that without roughness elements, and D/H is the diameter-to-height ratio of the roughness elements. Both studies show that the presence of nonerodible elements is extremely important in the prevention of erosion.

4.2.2 The relationship of fine particles to coarse particles suspended by wind erosion

Since both fine and coarse particles are present in the soil, the emission rates of both size ranges of particles as well as the length of time they remain in the air must be considered. Figure 4.3 shows size distributions reported by Gillette and Walker (1977) of particulate mass in parent soil, air-borne particles in the first 1.3 cm above the ground, and at 1 m, for two soils at two different wind speeds. The size distribution of the particles in the first 1.3 cm for all cases is highly similar to the size distribution of the loose particles in the parent soil smaller than 0.4 mm. The size distribution very close to the ground thus reflects the availability of particles from the soil. On the linear scale used in these plots, the proportion of mass smaller than 0.02 mm is very small. At 1 m, soil I (a sand soil) shows that for both wind speeds the proportion of mass in particles greater than 0.04 mm is reduced so that a mode of particle size smaller than 0.02 mm (between 0.002 and 0.02 mm) is now important. Gillette and Goodwin (1974) showed that this change of proportion with height is due to the large sedimentation velocities of particles greater than 0.02 mm. Similarly for soil II (a loamy sand soil) the proportion of particles greater than 0.02 mm is greatly reduced from near the ground to 1.5 m, although in this case the proportion of fine to coarse particles increases with wind speed.

The enormous change in the proportion of fine to coarse particles reflects the much greater heights and distances that fine particles may be transported compared to coarse particles. A size distribution of air-borne dust, generated by wind erosion in Texas and western New Mexico and collected by aircraft in southwestern Missouri (Gillette, in preparation), is shown in Figure 4.4. The distance to the sampling location from the eroding areas, indicated in the figure, is about 800 km. The outline of the dust cloud from satellite photographs is shown, along with wind speed vectors. The size distribution, which also exhibits evidence of sulphur-containing particles ($d < 0.002$ mm), shows the dominance of particles smaller than 0.02 mm in dust carried great distances.

An explanation for the diminishing proportion of coarse to fine particles with increasing distance or height can be found in the probability distribution of the turbulent vertical air velocity. The distribution has a mean of zero, and near the ground is gaussian, with a standard deviation approximately equal to the friction velocity of the air (Lumley and Panofsky, 1964), a speed which is not referenced to a specific height above ground. Since particles having a finite settling (downward) velocity must be supported by the upward fluctuations of the wind, a particle must have a favourable ratio of upward to downward motions in order to remain suspended.

By adding the settling velocity to the vertical air fluctuations, effective particle fluctuations result. The ratio of upward to downward motions of a particle carried in air having a normal vertical velocity distribution with zero mean and standard deviation u_* is shown in Figure 4.5. For reference, sedimentation velocities versus size (after Bagnold, 1941) are plotted in the figure. For a sedimentation velocity $V_{sed} = 0.4\, u_*$ the ratio of upward to downward motions is 0.5; in other words, for every upward motion there are two downward motions. The probability that such a particle will rise very high is thus rather small. Particles smaller than 0.02 mm are sufficiently small that their sedimentation velocities are less than 0.1 u_* for virtually all eroding winds (Gillette *et al.*, 1974).

4.2.3 Movement of soil particles as a function of wind speed

The flux of particles through a surface of unit width and infinite height that is mutually perpendicular to the ground and to the wind may be represented as 'q'. In simple terms we may think of q as the quantity of particles leaving an area of interest per unit width perpendicular to the wind. Several investigators have considered q as a function of wind speed greater than the threshold wind speed. Bagnold (1941) used dimensional arguments along with the assumption that all the momentum is transferred to the surface by collision of the air-borne sand grains (having positive momentum) with the surface in a jumping movement he called 'saltation'. His expression for q as a function of friction velocity of the wind is $q \propto u_*^3$. Other authors also give q which tends to be a function of u_* to the third power [Hsu, 1971; Kawamura, 1951; O'Brien and Rindlaub, 1936 (as reported by Horikawa and Shen, 1960); Belly, 1964]. A plot of the present authors' values for q *vs.* u_* for several different soils having a similar threshold shows a relationship $q \propto u_*^2\,(u_* - u_{*\mathrm{thresh}})$ which tends to $q \propto u_*^3$ for large u_* (Gillette, 1974). The movement of air-borne soil particles is, however, dependent on soil type, since as seen in Section 4.2.1, non-erodible aggregates or roughness elements absorb varying fractions of the horizontal momentum delivered by the wind. In that case the dependence of q on u_* is altered greatly since only a fraction of the energy of the wind is being used to move soil particles. Also the agreement of soil movement as $q \propto u_*^3$ is probably due to the dominance in the size distributions of air-borne particles of sand-sized material. The derivation of Bagnold depends on air-borne sand grains striking the surface in the saltation mode of movement. If the bulk of the material did not move in saltation it is doubtful that the $q \propto u_*^3$ relationship would hold. Since, however, sand-sized grains have the lowest threshold velocity and make up the largest fraction of air-borne soil particles where erosion takes place, the approximation $q \propto u_*^3$ for large u_* seems to be good for loose soils having few nonerodible roughness elements.

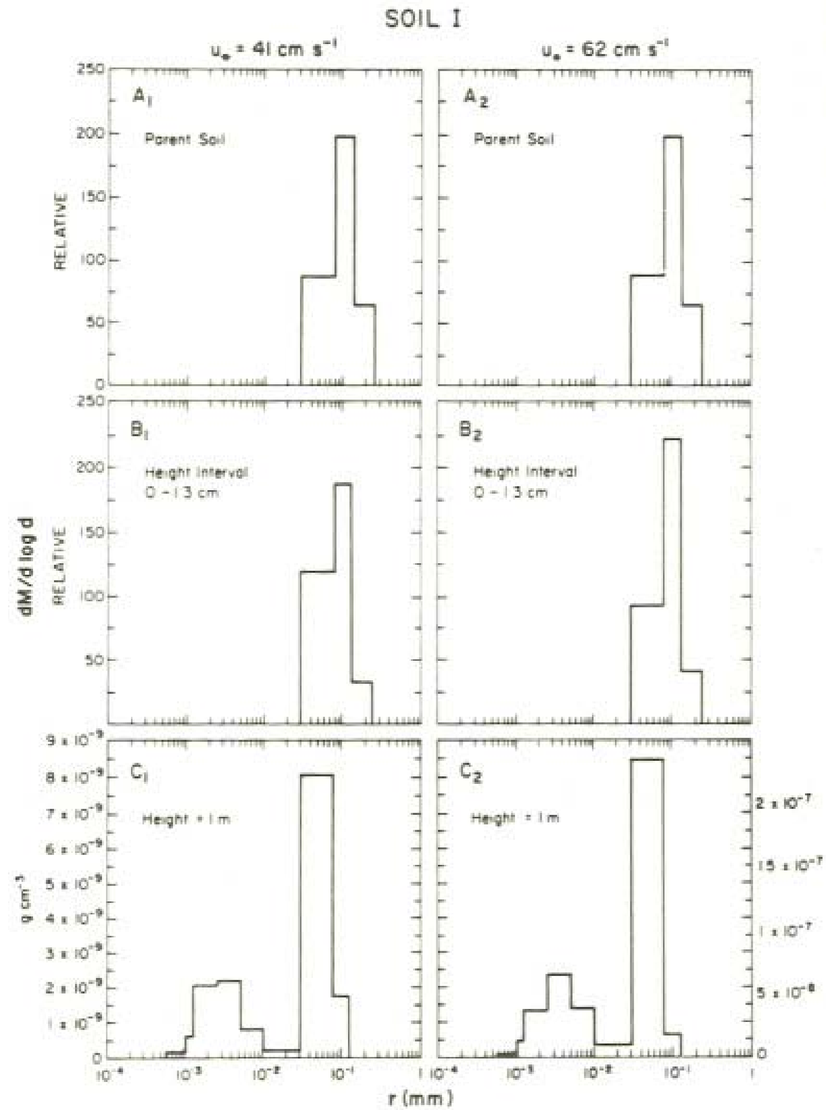
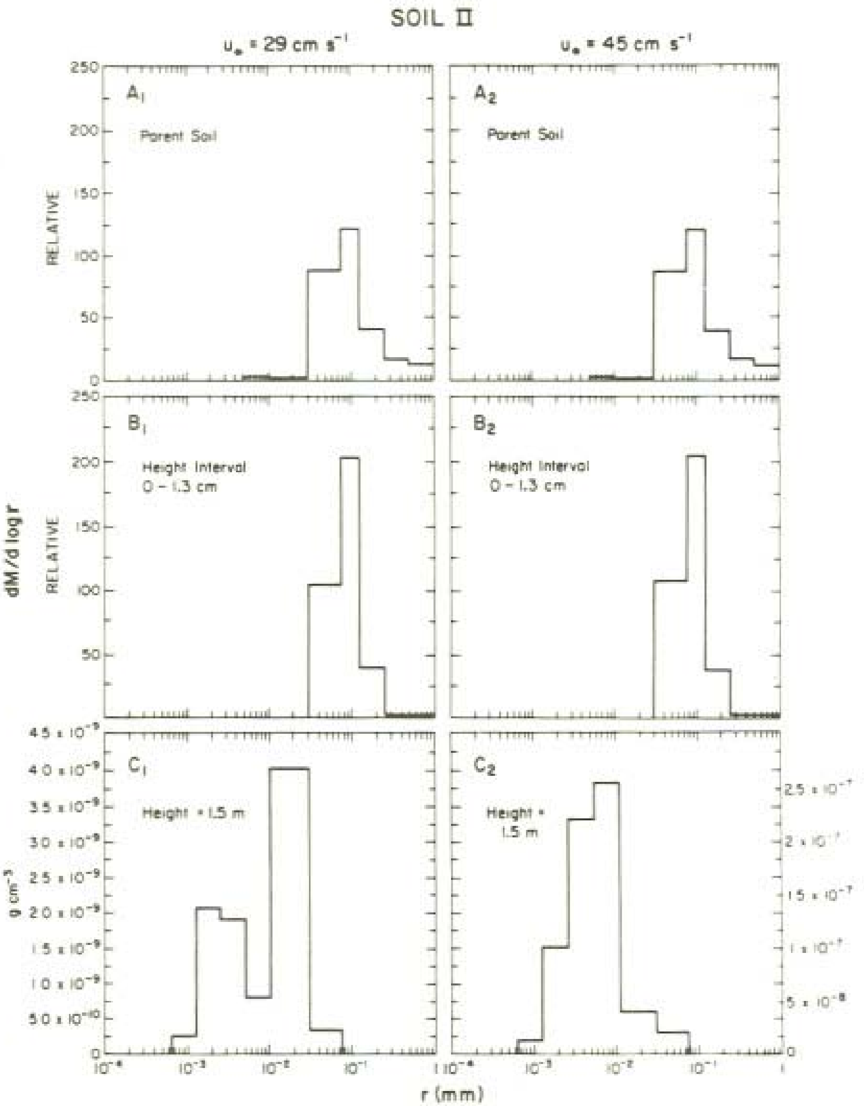


Figure 4.3 Size distributions of soil particles eroded from two soils. Soil I has a surface texture of fine sand and soil II has a surface texture of loamy fine sand. For each soil and each friction velocity 1 and 2, A is the relative size distribution of the dry parent soil, determined by wet sieving; B is the relative size distribution of



particles moving between the heights 0–1.3 cm above ground, and C is the mass size distribution for air-borne particles at 1 m above ground (after Gillette and Walker, 1977)

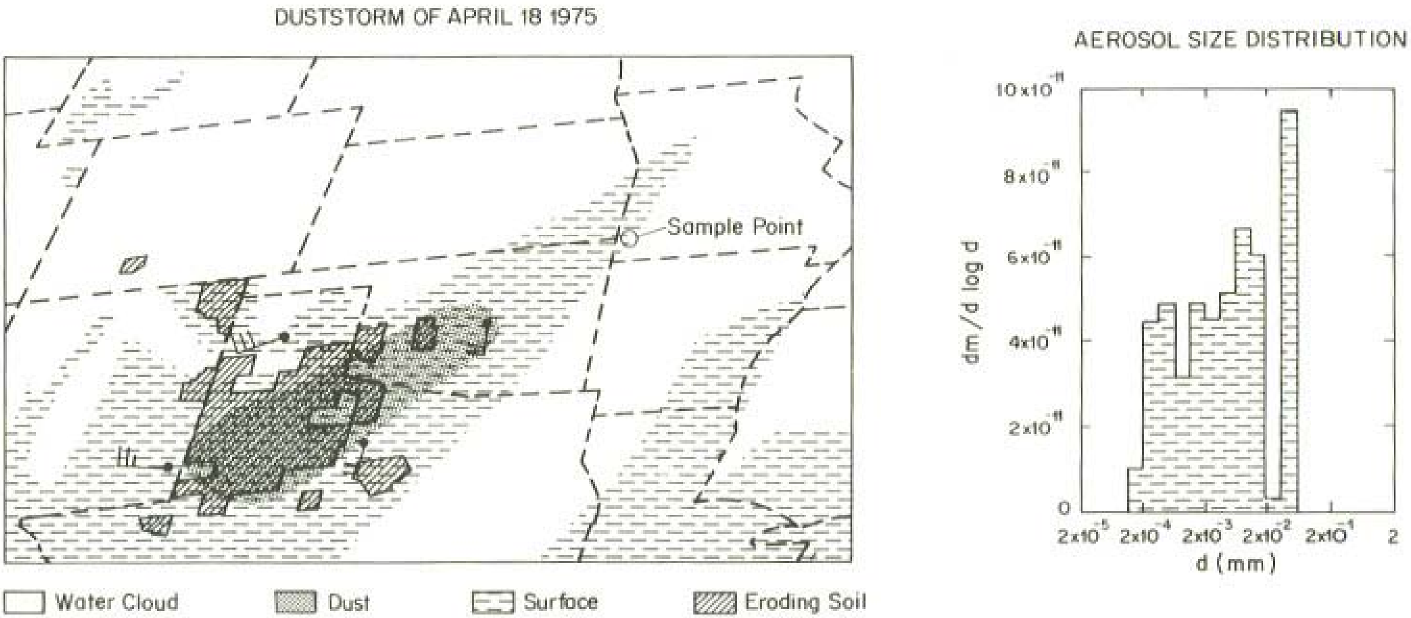


Figure 4.4 Size distribution of soil wind erosion particles carried a great distance. The dust was collected by aircraft. The production area of the dust is shown as the area of eroding soil in the figure (after Gillette, in preparation)

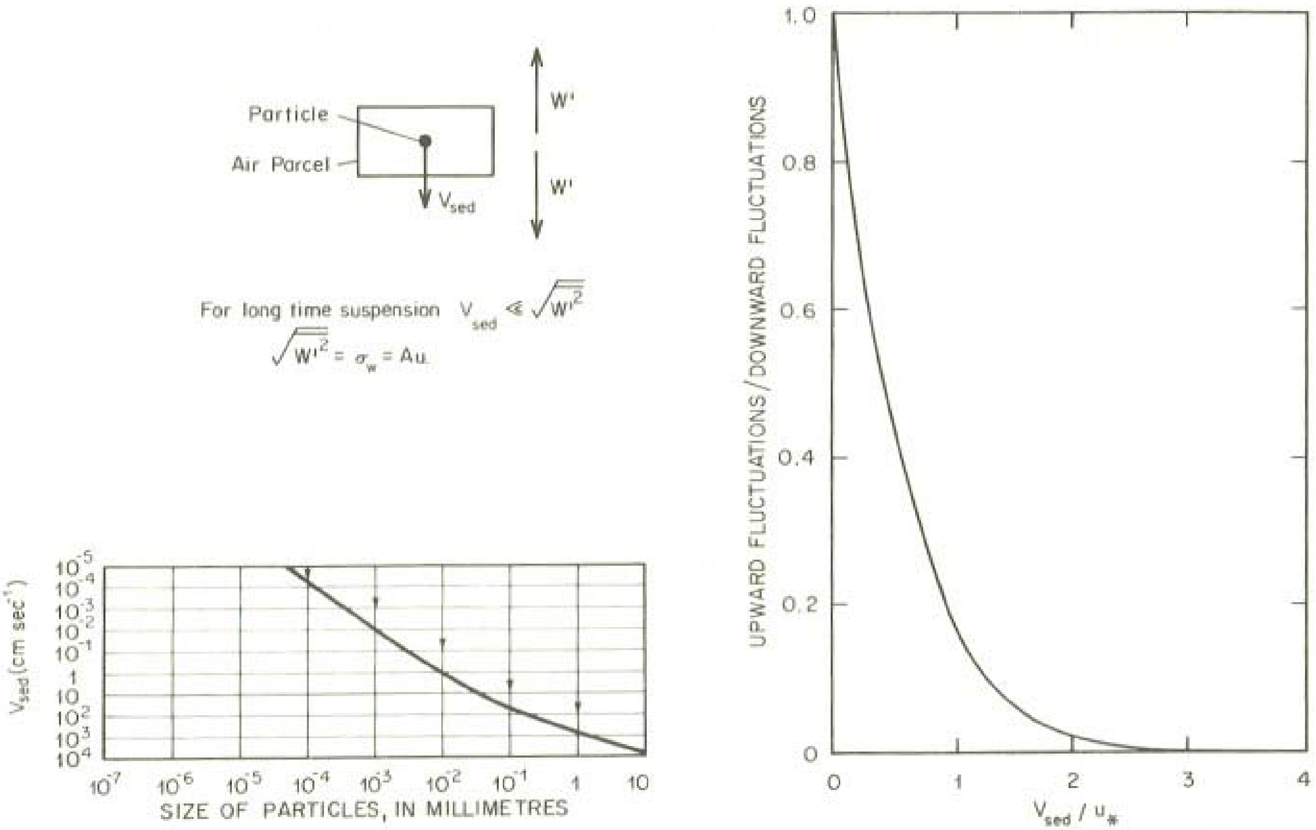


Figure 4.5 Sedimentation velocity V_{sed} compared to vertical velocity fluctuation w'; upward motions divided by downward motions for a particle having sedimentation velocity V_{sed} in air having vertical velocity fluctions with mean = 0, standard deviation u_*; sedimentation velocity *vs.* particle size (this part of figure after Bagnold, 1941); (the entire figure after Gillette, submitted to transactions off ASAE)

4.3 SOIL FACTORS IN DUST PRODUCTION

4.3.1 Soil factors affecting threshold velocity

Since threshold velocity is the minimum velocity at which the aerodynamic lift and drag is equal to the soil forces holding particles together, soil binding should be expected to have a major effect on threshold velocity. That is, a soil having greater coherence forces would require larger wind forces to set it in motion. Van der Waal's forces have been considered by Iversen *et al.* (1976) but little is known about the stronger effects of clay aggregation and water bonding. Chepil (1957) discussed erodibility in terms of soil moisture; he concluded that soil moistures near the wilting point of plants (i.e. approximately soil moisture content at 15 atmospheres of negative pressure) are ineffective in holding soil particles together. Chepil's correction for the presence of water will be given in the next section 4.3.2.1. Smalley (1970) stated that erodibility depends on cohesiveness of the soil, which may be measured in terms of its tensile strength. In Smalley's model of a simple soil, tensile strength is related to the packing density, the coordination number of the particles, and the interparticle bond strength. The tensile strength was shown in the simple soil system to be inversely proportional to the cube of the particle diameter, which suggests that very fine soils are less erodible than coarse soils; this is indeed observed.

Bisal and Ferguson (1970) investigated the effect of non-erodible aggregates on threshold wind velocity (in this case wind velocity at 30.5 cm above the soil surface). Their empirical relationship was

$$\ln V_T = 6.0438 + 0.02332\, C$$

where V_T is the threshold velocity in cm s^{-1} at 30.5 cm, and C is the percentage of soil mass in aggregates larger than 1 mm in a sample.

The classification by the US Department of Agriculture of soil texture classes into wind erosion groups also shows that soils having surface aggregates around 100 μm are more erodible than soils having aggregates of larger size. If the same wind erosion forces act on different soils, the more erodible soil will have a lower threshold velocity since the maximum soil transport at a given speed will be determined by aerodynamic factors. Thus more soil erosion implies a lower threshold velocity and more frequent erosion events. Table 4.1 (after Lyles, 1976, using data from Hayes, 1972) shows the magnitude of erosion for surface soil textural classes. Referring the textual classes to wind erosion, one notices that coarse soils are the most easily eroded, followed by clay soils (due to their tendency to form aggregates of sand size), calcareous loamy soils, and finally the soils of wind erosion groups 5, 6, and 7 (fine soils which form stable large aggregates). The erodibility of calcareous soils is quite interesting, showing chemical effects in the erodibility of soils.

TABLE 4.1 Descriptions of Wind Erodibility Groups (WEG)[a] (After Lyles, 1976)

WEG	Predominant soil textural class	Dry soil aggregates >0.84 mm	Soil erodibility 'I'
		Percent	Metric tons/ha/yr
1	Very fine, fine, and medium sands; dune sands	1	696
2	Loamy sands; loamy fine sands	10	301
3	Very fine sandy loams; fine sandy loams; sandy loams	25	193
4	Clays; silty clays; non-calcareous clay loams and silty clay loams with more than 35 per cent clay content	25	193
4L	Calcareous loams and silt loams; calcareous clay loams and silty clay loams with less than 35 per cent clay content	25	193
5	Noncalcareous loams and silty loams with less than 20 per cent clay content; sandy clay loams; sandy clay	40	126
6	Noncalcareous loams and silt loams with more than 20 per cent clay content; non-calcereous clay loams with less than 35 per cent clay content	45	108
7	Silts; noncalcareous silty clay loams with less than 35 per cent clay content	50	85

[a]Data from Hayes, 1972.

4.3.2 The effect of soil characteristics on the movement of soil

4.3.2.1 Total soil movement as affected by soil characteristics

Chepil and Woodruff (1959) related the total soil mass moved by a wind tunnel (χ) which produced winds of a given u_* (78 cm s^{-1}) to the average depth of furrows (the ridge roughness R), vegetative residue V, and an erodibility index I based on the percentage of soil mass in aggregates greater than 0.84 mm.

$$\chi = 400 \frac{I}{(RV)^{1.26}}$$

where R is measured in inches, V is measured in lb/acre, and I is given in Table 4.2. The present author modified a formula (Gillette, submitted to the ASAE) from Chepil (1957) to give the total soil flux 'q' through a surface of unit width that is mutually perpendicular to the wind and to the soil surface.

$$q = 0.16 \frac{LX}{16500} \frac{u_*^3}{78} \ (\mathrm{g\ cm^{-1}\ s^{-1}})$$

$$\text{for } \frac{LX}{16500} \leqslant 1$$

$$= 0.16 \frac{u_*^3}{78}$$

$$\text{for } \frac{LX}{16500} > 1$$

In these formulas L is the length of eroding soil parallel to the wind (in feet) and u_* is the corrected friction velocity (cm s^{-1}). X is defined above and the correction for the friction velocity for moisture content is

$$\rho u_*^2 = (\rho u_*^2)_{\text{observed}} - 6\left(\frac{W}{W'}\right)^2$$

where W is the amount of water held by the soil and W' is the amount of water held by the same soil at 15 atmospheres of negative pressure (Chepil, 1956).

The present author (submitted to ASAE) compared measurements of total soil flux to those calculated from the above formula. The calculated values were compared to observed fluxes in two categories: steady erosion and intermittent erosion. Since intermittent erosion does not meet the assumptions of the Bagnold (1941) formula (in which momentum is transported at the lower boundary by saltating sand grains), intermittent erosion was not expected to show agreement between observed and predicted mass fluxes. For steady erosion, agreement was good:

$$q_{\text{obs}} = 1.0005\, q_{\text{p}}^{1.08} (\text{corr. coeff.} = 0.84)$$

For intermittent erosion, however, agreement was poor:

$$q_{\text{obs}} = 0.58\, q_{\text{p}}^{1.49} (\text{corr. coeff.} = 0.70)$$

4.3.2.2 The vertical flux of particles which may be transported great distances as affected by soil characteristics

Since particles which may be carried great distances are a part of the total air-borne soil particulate suspension, we must consider that the factors affecting total soil movement also affect fine particulate emission. Consequently it is con-

TABLE 4.2 Soil Erodibility Index, I, Based on Percentage of Soil Fractions Greater than 0.84 mm in Diameter as Determined by Dry Sieving (after Chepil and Woodruff, 1959)

Percentage of soil fractions greater than 0.84 mm:	Units									
Tens	9	8	7	6	5	4	3	2	1	0
9										0.02
8	0.02	0.03	0.03	0.04	0.05	0.06	0.07	0.08	0.09	0.10
7	0.12	0.14	0.16	0.20	0.25	0.30	0.35	0.40	0.45	0.5
6	0.55	0.60	0.65	0.70	0.75	0.80	0.85	0.90	0.95	1.0
5	1.1	1.2	1.3	1.4	1.6	1.8	2.0	2.2	2.5	2.8
4	3.1	3.4	3.7	4.0	4.3	4.9	4.9	5.2	5.6	6.0
3	6.5	7.0	7.5	8.0	8.5	9.0	9.5	10.5	11.0	12.0
2	13	14	15	16	17	18	19	20	21	23
1	25	27	30	33	36	39	43	17	51	55
0	70	80	100	120	150	175	280	450	1000	—

DUST GENERATION

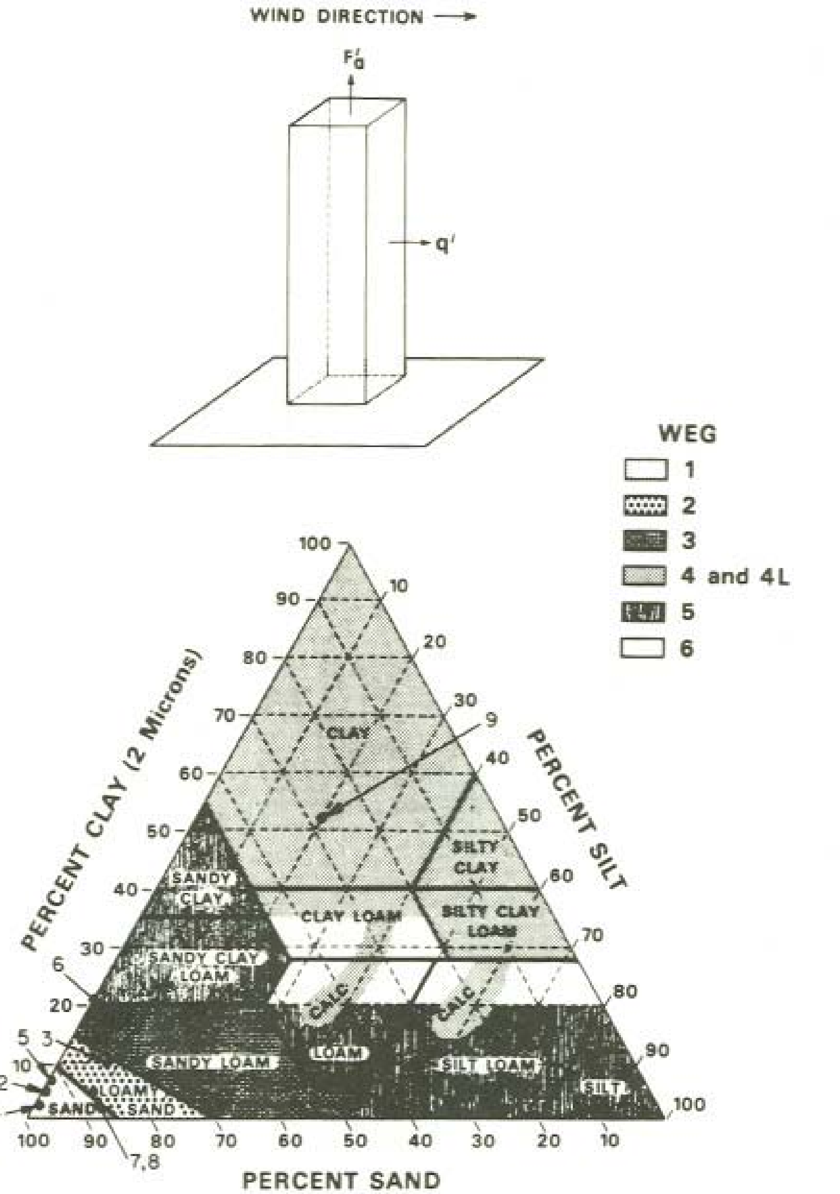
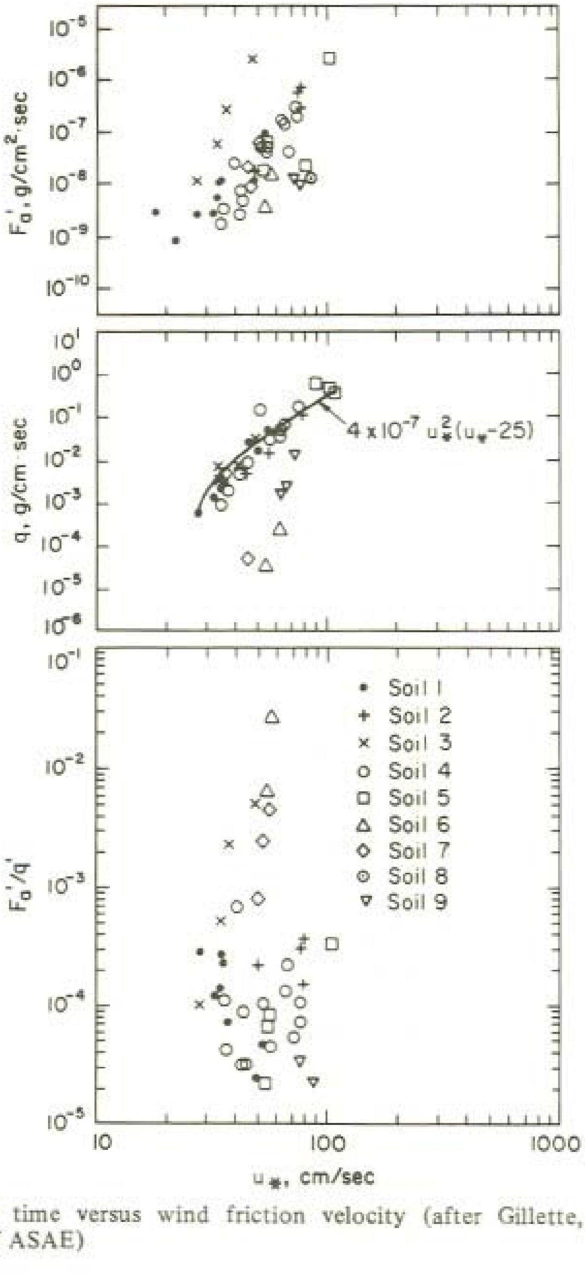


Figure 4.6 Textures of sampled soils; total soil movement *vs.* wind friction velocity; vertical flux of particles smaller than 0.02 mm *vs.* wind friction velocity; ratio of vertical flux of particles smaller than 0.02 mm to total soil movement per



unit area per time versus wind friction velocity (after Gillette, submitted to transactions of ASAE) venient to speak of the ratio of fine particles to total particles or of the ratio of the vertical flux of particles $d < 0.02$ mm (seen in a previous section to include most of the particles having the potential for long-range transport) to the total flux q. Ratios of the vertical flux of fine particles, F_a' to the horizontal flux of all air-borne soil particles*, q', were given by Gillette (submitted) and are shown in Figure 4.6. Nine different soils are shown in the soil textural diagram. Also included in the figure is a schematic diagram of F_a' and q' as well as F_a' *vs.* u_*, and q *vs.* u_*. Physical characteristics for the nine soils are given in Table 4.3. Soils 1, 2, 4, and 5 are sandy; soils 3, 6, 7, and 8 are loamy; and soil 9 is clay. The sandy soils show a fairly uniform trend of increasing F_a' with u_*, probably due to the uniformity of the dry aggregate structure of these soils. F_a' *vs.* u_* for the loamy soils shows a greater scatter, probably due to their widely different dry aggregate structures; the percentage of their mass in aggregates < 0.84 mm varied from 38% to 89%, which, in turn, affected q. Chepil and Woodruff (1963) showed that a greater percentage of mass in aggregates larger than 0.84 mm correlated with decreased wind erosion. Soil 9 was a clay soil whose eroding particles consisted of strongly layered pellets formed by shrinkage of a montmorillinitic clay upon drying from a mud state. These clay pellets were extremely resistant to breakage by hand when compared to the aggregates of the loamy soils and the sandy soils. This soil had a $u_{*\text{thresh}}$ of about 65 cm s^{-1} probably due to the large size of the aggregates (only 9.3% of the loose surface pellets were smaller than 0.84 mm). This high $u_{*\text{thresh}}$ and breakage-resistant aggregates help to explain the low values of F_a' for this clay soil. The largest values of F_a' for a given u_* are those for soil 3, a loamy soil with very little surface coherence. In general, the increase of fine particulate production (F_a') with wind speed is greater than the linear increase of wind speed.

Values of F_a'/q' for sandy soils show great scatter but little evidence of trend with wind friction velocity u_*. For loamy soils there is a great increase with u_*, which is tantamount to an increase with wind speed in the proportion of fine ($d < 0.02$ mm) to total soil movement. The value of F_a'/q' for the clay soil is very low, showing the effect of the organized structure of the clay minerals on drying and shrinking in resisting breakage into fine particles. In general, soils having finer textures produced more fine dust for unit soil movement except for that fine-textured soil whose mineral components formed small aggregates hard enough to resist impact breakage at mean wind speeds up to 25 m s^{-1}.

4.3.3 Moisture and wind history of the soil

The moisture and wind history of the soil, including freezing and thawing and drought accompanied by frequent strong winds, influences the breakage of soil

*q' differs from q in that q' is the flux of particles into a surface 1 cm wide, perpendicular to the wind and ground with vertical dimensions from the surface to 76 cm. q is the flux into a surface 1 cm wide, mutually perpendicular to the wind and ground with vertical dimensions from the ground to infinity. Practically speaking, however, most air-borne soil mass is confined to heights less than 50 cm and the approximation $q' = q/76$ is good. The dimensions of q' are g/cm^2 sec whereas the dimensions of q are g/cm sec.

TABLE 4.3 Soil Parameters

Soil erosion parameter	Soil 1	Soil 2	Soil 3	Soil 4	Soil 5	Soil 6	Soil 7	Soil 8	Soil 9
Soil moisture, %	0.52 ± 0.61	0.99	1.29 ± 0.20	0.41 ± 0.10	0.52 ± 0.13	0.75	0.6	0.6	6.6
Cloddiness, % dry									
Aggregate <0.84 mm	95.0 ± 2.78	98.9	89.1 ± 5.0	95.9 ± 1.6	98.8 ± 0.6	36.7	60.0	53.0	9.3
Vegetative residue, g/m^2	26.67	8.25	3.67	91.6	19.1	3.5	161.0	39.0	2.9
Ridge roughness, cm	2.5	2.5	2.5	3.7	2.5	5.0	5.0	22.5	2.5
Erosion fetch, km	1.6	0.8	1.6	1.6	0.2	0.5	0.5	0.5	0.1
Soil Texture									
% Sand, $r > 25\ \mu m$	96.0	95.5	81.5	96.8	93.1	77.7	88.0	88.0	28.0
% Silt, $1 < r < 25\ \mu m$	0.5	1.0	8.5	1.4	1.0	3.3	3.2	3.2	20.0
% Clay, $r < 1\ \mu m$	3.5	3.5	10.0	1.8	5.9	19.0	8.8	8.8	52.0

aggregates into small erodible units which change the initial wind speed for soil movement ($u_{*\mathrm{thresh}}$) as well as the horizontal soil flux (q) as a function of wind speed. Although it is not certain that the ratio F_a'/q' changes with weathering, it may change for certain soils.

4.4 CONCLUSIONS

Both aerodynamic and soil factors must be considered in studying dust emission by the wind. For the Sahara I must stress the importance of non-erodible elements and soil aggregation although other important factors have also been discussed. It is quite possible that human activities have a greater effect on soil surface condition and non-erodible elements than do other environmental factors.

The threshold wind velocity for wind erosion and movement of soil are affected by soil textures, wind speed, mineralogy, soil moisture, history, soil roughness, vegetative residue, and the presence of vegetation and other nonerodible elements such as rocks.

Fine particles are produced by the sandblasting effect of saltation, which acts to disaggregate fine particles on the surfaces of larger particles and by 'splashing' of the saltating particle into a reservoir of fine particles. Fine-textured soils tend to emit more fine particles upon unit soil movement unless the mineralogical structure of the soil aggregates is highly resistant to breakage. Total soil movement measurements show good agreement with values computed from formulae derived from ARS studies and this author's data.

4.5 ACKNOWLEDGEMENTS

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APPENDIX: OTHER DATA WHICH MAY BE OF INTEREST TO SAHARA DUST RESEARCH

I would like to quote two papers which might be of interest to the Sahara dust project.

1. Mass-visibility relationships

The first paper (Patterson and Gillette, 1977) deals with estimating dust concentration from visibility observations. Comparing our measurements with those of other workers, we found some discrepancies. Visibility, V (expressed in km), and mass concentration of dust, M (expressed in g m^{-3}), are related by

$$MV^{\gamma} = C$$

where C is a dimensional constant and γ is a nondimensional constant. The results of several authors for γ and C were given in Patterson and Gillette's Table 2, summarized below:

Study	C (g m^{-3} km for $\gamma = 1$)	γ
General Urban (Charlson, 1969)	1.8×10^{-3}	1
Bertrand *et al.* (1974)	1.4×10^{-3}	1.05
NCAR (Total mass)	2.3×10^{-2}	1.07
(SP $r < 20\ \mu$m)	1.7×10^{-2}	1.08
(SP $r < 10\ \mu$m)	1.3×10^{-2}	0.95
Chepil and Woodruff (1957)		
All data	5.6×10^{-2}	1.25
No local erosion	2.0×10^{-2}	–

The value of γ is reasonably close to 1 for all the studies except those of Chepil and Woodruff. The study concluded:

'Based on these data there is no single value that is generally appropriate for relating mass and visibility of soil-derived aerosols, and that can thus serve as a predictor of mass concentration when the visibility is known. In the presence of local erosion, our data and Chepil and Woodruff's data indicate that the value of C can vary between approximately 20×10^{-2} and 1.0×10^{-1} depending on soil conditions, visibility, and wind speed; the higher values for C are apparently characteristic of drought conditions. Under conditions of no local erosion, but still in the general source region for the dust generation, the value of $C = 2.0 \times 10^{-2}$ g m^{-3} km appears to be appropriate, while the Bertrand value of $C = 1.4 \times 10^{-3}$ g m^{-3} km may be appropriate for measurements made several thousand kilometers from the dust source.

Because of the variability in C, we feel that when the reduction in visibility is due to air-borne dust, mass concentration–visibility relations alone must be used carefully; if the fractional size distribution is known, however, the mass-visibility relations may be calculated using theoretical values of C.'

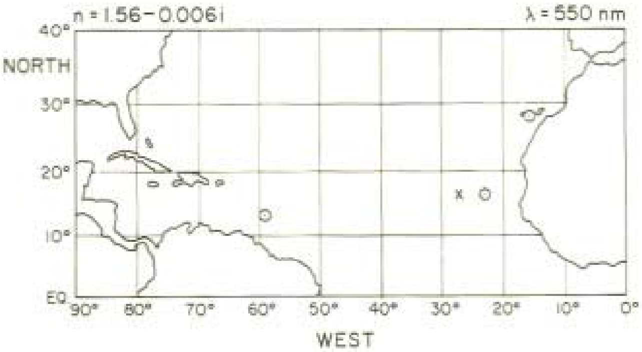

Figure 4.7 Index of refraction for Saharan aerosols over the Atlantic. Crosses and circles indicate locations for particle collections

2. Index of refraction for Saharan aerosols over the Atlantic

The second paper reports values for particle collections obtained at the circled and crossed locations in Figure 4.7 (Patterson, Gillette and Stockton, accepted manuscript). The samples were obtained by stations established by M. L. Jackson and Dale Gillette, by Joseph Prospero, and by Ruprecht Jaenicke and Lothar Schütz on the Research Vessel *Meteor.* The mean value for the index of refraction at all stations was $1.56 - 0.006\,i$.

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