CHAPTER 2

The Importance of Mineral Dust as an Atmospheric Constituent

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ABSTRACT

The importance of mineral dust in the atmosphere is twofold: Dust deposition and possible effects on climate.

Depending on the conditions and dimensions of desert areas and meteorological transport processes, dust depositions along the fringe areas of the source region can be of geological, pedological and ecological importance. The loess deposits during the glacial age are the classical example. Present day deserts seem to be much less efficient sources for loess formation.

The mineral dust in the atmosphere affects both the radiation of cloud free air and the optical properties of the clouds. The interaction is complex but in general one can expect the mineral dust to affect primarily the absorption of aerosols and clouds for short wave radiation. Large scale natural or anthropogenic changes of dust production which have undoubtedly occurred should therefore be considered with respect to their effect on climate.

This workshop is concerned with the production and transport of dust from the area of the southern and western Sahara. This concern is based primarily on ecological reasons. But once the dust has become air-borne it is part of the atmospheric system and is itself of interest. I will discuss here the question of its importance for atmospheric sciences.

We have gaseous and particulate trace substances in the atmosphere. Both can be of local, regional and global importance. The aerosols influence cloud formation, radiation and air electricity. Changes in their concentration may induce variations of the climate by affecting the global radiation budget. There are aerosols from different sources and the question to be discussed here is the role of the dust component of the tropospheric aerosol in this context.

Aerosols in the troposphere have residence times of the order of a week (see Martell and Moore, 1974). This is long enough for considerable spread within both hemispheres with their strong zonal circulation, but it is too short for an exchange between the two hemispheres across the equator which is of the order of one year.

The aerosols of the two hemispheres remain, therefore, almost completely separated with some large scale mutual leakage resulting from the Indian Monsoon system. Since there is no substantial exchange between the hemispheres in the Atlantic sector, Saharan dust remains confined to the northern hemisphere for all practical purposes.

If we disregard the anthropogenic sources we can distinguish three dominant natural aerosol production mechanisms listed in the order of their quantitative importance:

1. Gasformed particles, produced within the atmosphere by oxidation and reaction of a variety of trace gases.
2. Sea spray.
3. Mineral dust.

Table 2.1 gives published estimates of their source strength, indicating the degree of uncertainty. The uncertainty is particularly large for mineral dust, because these figures were derived by rather dubious methods. For some strange reason, however, they seem to be in fairly good agreement with recent estimates based on fairly comprehensive data sets on atmospheric dust concentrations published over the last years. Covering essential parts of the world, these data became available by routine application of such analytical methods as neutron activation, absorption spectroscopy, and others, capable of determining simultaneously in clean air aerosol samples a large number of elements. The results show that a considerable number of these elements, in particular Fe and Al, occurred in ratios quite similar to those in crustal or soil material indicating that their dominant source was mineral dust. By applying appropriate factors it is, therefore, possible to derive fairly reliable figures for total mineral dust concentrations from these analyses. Figure 2.1 gives a survey based on such observations, with the most prominant authors indicated. We will come back to this Figure later. On the basis of such concentration data covering

TABLE 2.1 Global Estimates of the Major Natural Aerosol Sources in the Troposphere in 10^6 tons yr^{-1}

Source	Strength
Particles formed in the troposphere by oxidation of H_2S, NO_x, NH_3, HC and other trace gases	300–1100
Sea spray	1000
Mineral dust	100–500
Other sources (forest fires, volcanic, *etc.*)	3–150

Figures from the SMIC-Report (1971)(Table 8.1) and other sources.

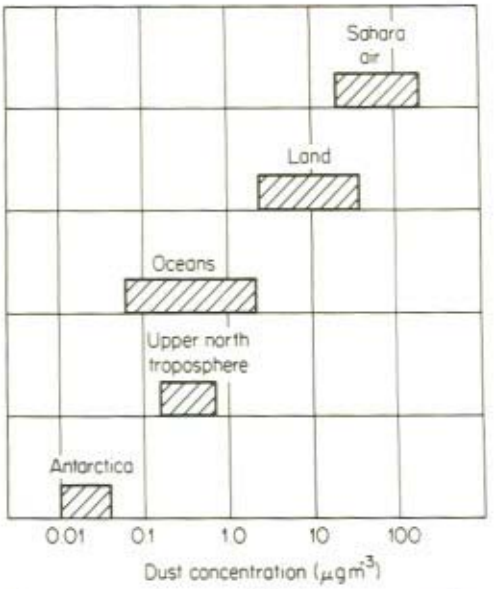


Figure 2.1 Survey of the concentration ranges of mineral dust in the troposphere based on numerous studies by – among others – the following authors: Blifford, Chesselet, Duce, Ferguson, Hoffman, Gillette, Goldberg, Griffin, Jaenicke, Prospero, Rahn, Schütz, Winchester, Zoller

most essential areas of the world and also of meteorological considerations about the vertical distribution within the troposphere, the global tropospheric dust burden can be estimated. With the knowledge of the average residence time for aerosols, we can then derive at figures for the total source strength. Table 2.2 gives such figures separate for both hemispheres for reasons indicated earlier together with our guess for reasonable uncertainty factors.

We think that the basis for these estimates is more reliable than that for the previous estimates. But it becomes equally clear that despite this fact the uncertainty is estimated to be even larger, which we feel is more realistic. We see that even without the Sahara source the northern hemisphere has a higher dust load than the southern one as was to be expected from the larger land areas. Included in Table 2.2 is the range of recent estimates of the Saharan dust production by Prospero and Carlson (1972) and Jaenicke and Schütz (1977). Clearly, the Sahara desert seems to play an important role in the northern hemispheric dust cycle, providing about half of it. The addition of the Saharan source results in a range for the global source strength of 260 to 400×10^6 tons yr^{-1} (disregarding the uncertainty of non Saharan production) which agrees well with the older estimates in

TABLE 2.2 Estimates of the Global Tropospheric Dust Cycle

Part of troposphere	Dust burden 10^6 tons	Source strength 10^6 tons yr^{-1}
Northern hemisphere[a]	3.0[b]	150[c]
Southern hemisphere[a]	1.0[b]	50[c]
Whole troposphere[a]	4.0[b]	200[c]
Sahara plume	1.2–4.0	60–200[d]
Total troposphere (plus Sahara)[e]	5.2–8.0	260–400
Total troposphere (plus Sahara)[f]	3.2–12.0	130–800

[a]Estimates disregarding any special production in deserts, particularly the Sahara area and its plume in the north Atlantic trade winds.
[b]Estimated uncertainty factor about ±2.
[c]The source strength was calculated from the dust burden assuming an aerosol residence time of 1 week. Estimated uncertainty factor about ± 3.
[d]Range given by Prospero and Carlson (1972) and Jaenicke and Schültz (1977)
[e]Without applying the uncertainty factors *b* and *c*.
[f]After application of the uncertainty factors *b* and *c*.

Table 2.1. Applying our uncertainty factors gives the possible range of 130 to 800×10^6 tons yr^{-1}.

But the actual uncertainty is still larger because there are other desert areas besides the Sahara which may act as huge individual sources. There are yet, at least to my knowledge, no data on the frequency and the magnitude of these sources. Figure 2.2 shows a map indicating areas covered with dunes which are potential sources for wind blown dust. Observation of dust haze over the oceans seems to indicate, however, that the transport of Saharan dust over the subtropical Atlantic is the most spectacular phenomenon of this type. But we should consider these facts to get a realistic feeling of the lack of reliable information in this field and for the uncertainty of the figures in Table 2.2.

It is perhaps of interest to compare the flux of the eolian transport of Sahara dust with other fluxes out of this desert. The transport of sand by moving dunes across the West coast of the Sahara by N to NE winds occurs apparently only along a front of about 80 km length south of Cape Blanc. Flux estimates give 0.15×10^6 tons yr^{-1}, with particle sizes peaking around 100 μm radius. Transport of mineral dust out of the Sahara by the river Niger is 60×10^6 tons yr^{-1}. These data are compiled in Table 2.3. It is clear that the long range eolian dust transport out of the West Sahara in Table 2.2 is the most important one, at least much more important than the other eolian transport mechanisms. This may of course be different for other arid zones and deserts.

Figure 2.2 shows also the known loess deposits. Loess is an eolian deposit and is

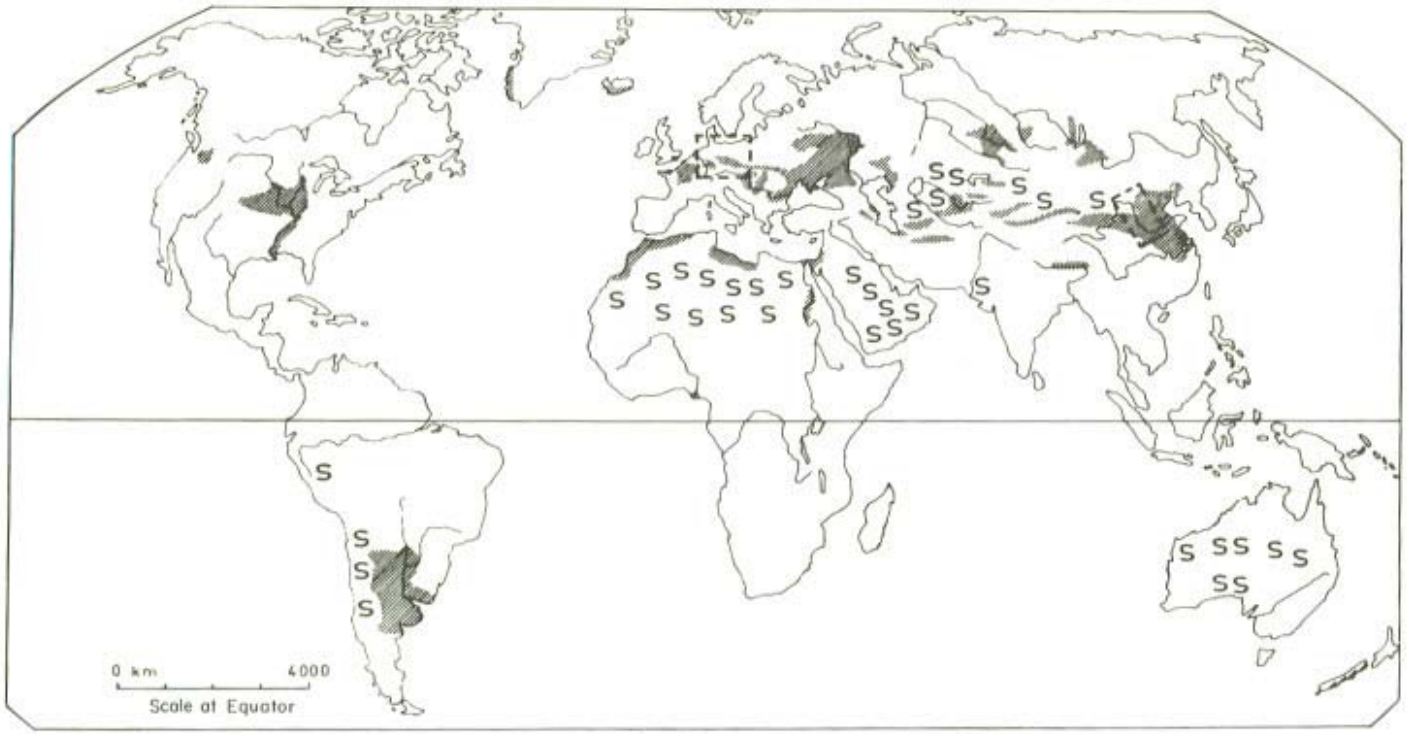


Figure 2.2 Distribution of loess on Earth (according to Scheidig, 1934) as well as desert areas with sand dunes (according Boyko, 1967). From Smalley and Vita-Finzi, 1968, Section of Central Europe, reproduced in Figure 2.6 indicated by a dotted square (after Füchtbauer and Müller, 1970)

Legend: Shaded areas = Proved loess deposits S = Deserts with sand dunes

TABLE 2.3 Comparison of various transport rates of Saharan material across the western coastline

Mode of transport	Flux in 10^6 tons yr^{-1}
Moving sand dunes	0.15[a]
Saltating sand	10–20[a]
River Niger	60[a]
Eolian transport	60–200[b]

[a] Sarntheim and Walger (1974).
[b] Table 2.2.

primarily made up of particles in the range 5 to 50 μm radius. Figure 2.3 shows cumulative mass distributions (size is expressed as diameter) of 8 different types of loess from various parts of the world showing a fairly uniform distribution. This is not too surprising because the fractionation process between loess and soil particles is always the same depending to some degree on the distance between source and deposition area. In desert dust storms even much coarser material becomes air-borne or is jumping along the surface but settles down again at much smaller distances. In

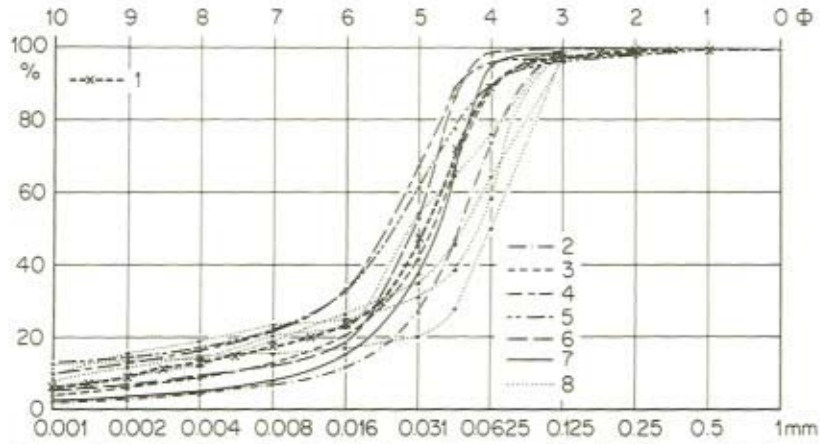


Figure 2.3 Particle size distributions (diameters!) of different kinds of loess (according to Swineford and Frye, 1955, and Teruggi, 1957) as well as recent eolian dust (according to Swineford and Frye, 1945) (after Füchtbauer and Müller, 1970)

Legend: 1. =Dust 1939, Meade, Kansas, U.S.A. 2. = Torino, Italy.
3. =St. Vallier, France. 4. = Achenheim, France.
5. =Wiesbaden, West Germany. 6. = Liege, Belgium.
7. =Thomas Co., Kansas, U.S.A. 8. = Loess from the Argentine

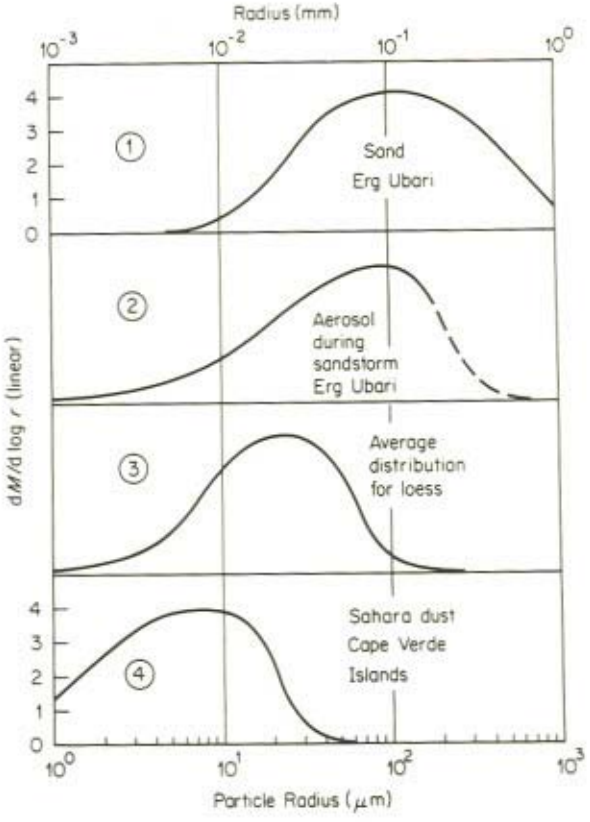


Figure 2.4 Comparison of different idealized mass distributions based on the following sources:

(1) Schütz and Jaenicke (1974), sand from the Libyan desert, Erg Ubari
(2) Same source and location as (1) but aerosol during sand storm
(3) Füchtbauer and Müller (1970), average of 8 loess distributions from various continents, according to Figure 2.3
(4) Jaenicke and Schütz (1977), average mass distribution over the Cape Verde Islands

Figure 2.4 we plotted linear mass distributions of sand and sand storm aerosols from the Libyan desert, loess (according to Figure 2.3) and air-borne dust in an idealized fashion to show the essential size ranges. Figure 2.5 demonstrates schematically how the major fractions are derived from the original source by successive stages of a winnowing process controlled by the interaction between air-borne transport and gravitational settling until Stage 4. For atmospheric travel-

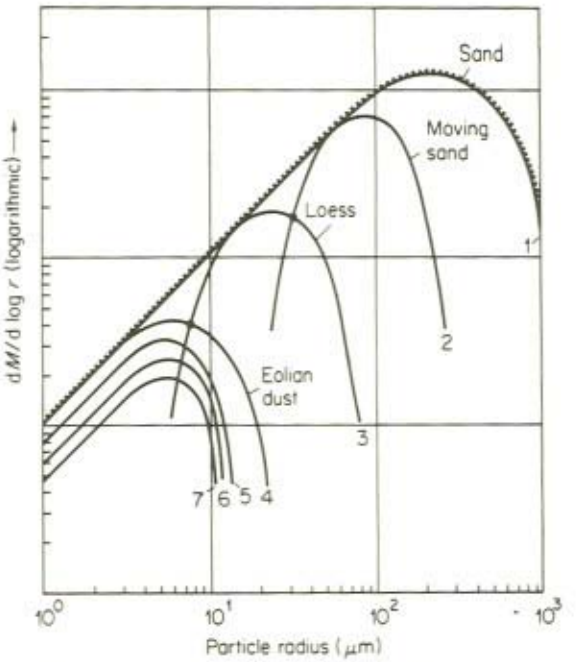
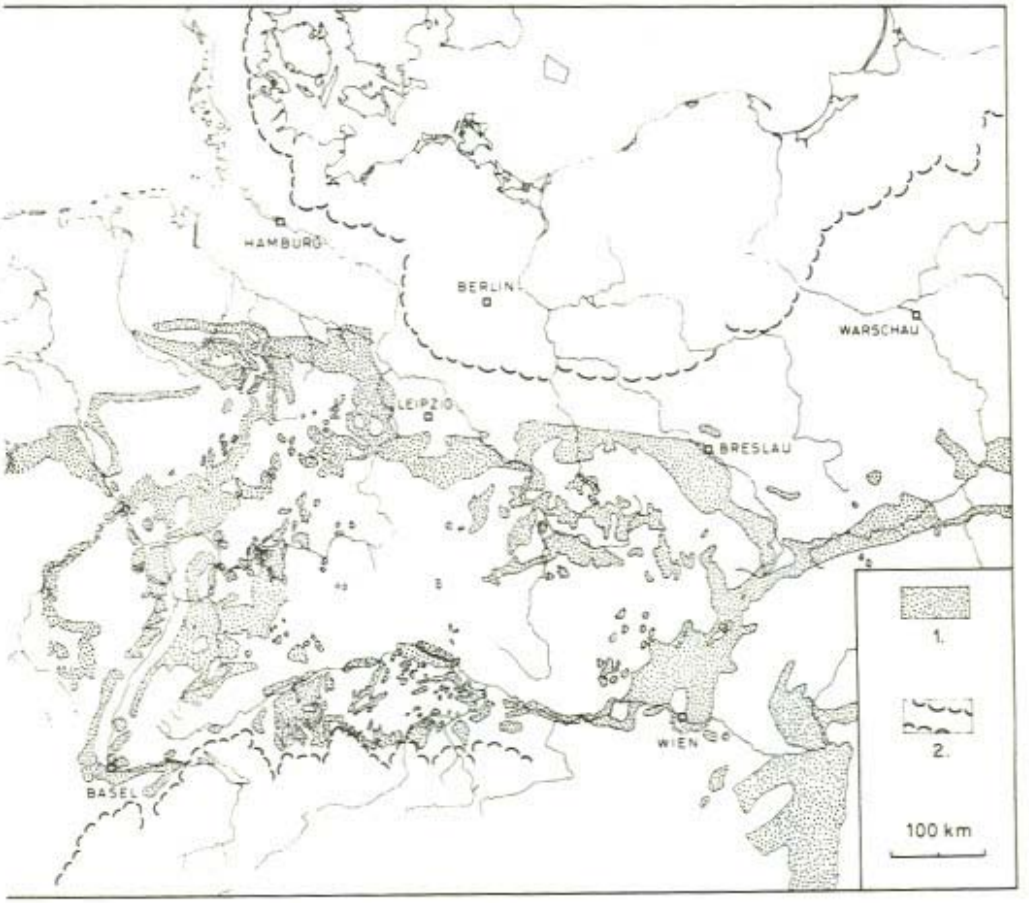


Figure 2.5 Sand fractionation processes by wind, schematic. The original sand distribution is fractionated into major fraction 2, 3, and 4 as a function of distance from the source. Curves 5, 6, and 7 depict the change in concentration due to both wet and dry removal from the atmosphere. The sum of curve 2, 3, and 4 should be equal to the original curve 1

ling times comparable to the average aerosol residence time in the troposphere removal by precipitation becomes important besides fallout by sedimentation. The size distribution of the individual fractions will, of course, vary with the environmental parameters, but Figure 2.3 suggests that the differences may not be very dramatic. In Figure 2.5 a smooth, continuous distribution was assumed for the original sand. In reality this may not be the case because of size dependent efficiencies of the various physical and chemical disintegration processes constantly active in the sand. In fact there are indications of bimodal distributions in the case of fraction 2, Figure 2.4, at some places, but in principle one should expect single mode distributions if the sand distributions are sufficiently smooth. Figures 2.4 and 2.5 demonstrate that there is and should be considerable overlapping in the tails of distribution curves.

It is clear that the loess fraction becomes air-borne only in fairly strong winds. The large majority of all loess deposits in Figure 2.2 were formed during or at the



ure 2.6 Distribution of loess in Central Europe, according to Scheidig, 1934 (after :htbauer and Müller, 1970)
;end: 1. Shaded area = Loess, thickness more than 0.5 m.
2. Limit of Würm-glaciation

end of the last glaciation period in distances up to several hundred kilometers from the border lines of the ice sheets. This is well demonstrated by Figure 2.6 which shows that all the loess deposits in Central Europe are located between the ice shields of the last glaciation. The huge amounts of material of these deposits were most likely produced by physical weathering processes associated with glaciers. But there is sufficient evidence that deserts also can act as source areas for loess, though apparently at a much smaller scale. One example is the Negev desert in Israel (Smalley and Vita-Finzi, 1968). Apparently in most deserts the weathering rate just is not sufficient to produce deposits along their fringe areas.

The large areas covered with glacial loess deposits imply that at those times the

atmosphere in such latitudes must have carried a considerable dust burden, most likely due to stormy and dry weather conditions. This conclusion is in fact supported by high dust concentrations in Greenland ice layers during the end of the last glaciation. It is also implied by the actual absence of loess formation at present times in these latitudes. An intriguing question is if and to what degree such dusty atmospheres induced feed processes on the climate development during the ice age.

The processes by which dust becomes air-borne are complex. Structure, moisture content and other soil parameters determine a threshold wind speed at which the dust is picked up by the air. Once this threshold is passed the vertical flux of dust increases rapidly with wind speed (see e.g. Gillette, 1974). A considerable fraction of the material, however, remains near the surface and is soon redeposited again. It would, of course, be unreasonable to include the flux of such material in figures for the global dust cycle, as given e.g. in Table 2.2. Dust considered in this respect should remain sufficiently long in the air to become of importance for meteorology. It is suggested that a one day residence time would represent a reasonable lower limit for this purpose.

The dust component in atmospheric aerosols is highly variable as shown in Figure 2.1. The reasons for this are the rather short residence times of aerosols together with the pronounced variability of the dust sources with time and space. The data show that on a global scale the average dust concentrations at fixed locations usually fluctuate with time by about one order of magnitude. They also show that average dust concentrations range more than 4 orders of magnitude with values of 100 $\mu g/m^3$ in air coming from the Sahara and values as low as 10^{-2} $\mu g/m^3$ at the South Pole (for comparison: The normal range of sea salt concentrations over the oceans is 1 to 10 $\mu g/m^3$, i.e. only about one order of magnitude). Inside of desert dust storms the visibility can be less than 100 meters implying concentrations of the order of 10^4 $\mu g/m^3$.

We have shortly reviewed production, redeposition and atmospheric transport and distribution of mineral dust. Besides the geological, pedological and ecological effects of soil erosion and dust fallout the atmospheric dust may also have importance for the radiation budget of the atmosphere and hence for the climate if large scale changes of land surfaces have occurred or do occur, natural or man made. There are two ways in which the aerosol can influence the radiation budget: In a direct way by changing the radiation fluxes in a cloud free atmosphere, by scattering and absorption of radiation, and in an indirect way by modifying the optical properties, particularly the albedo of clouds (see SMIC Report, 1971).

The direct influence on the short wave radiation is primarily affected by the aerosol particles in the size range from about 0.1 to 2 μm radius. Most of the particles in this range are formed by gas reactions and under normal conditions the fraction of mineral particles in this size range is small. But we know that the mineral particles have, among other constituents, an iron content of about 5%, mostly present in the form of oxides which are good absorbers for radiation. In clean atmospheres the gasformed particles show little absorption, and here changes

in the concentration of dust may result in marked changes of the absorption characteristic of the aerosol even if the number concentration is not much affected.

The optical properties of clouds are influenced by the aerosol in two ways. The number concentration, particularly those larger than about 0.1 μm, controls the number concentration of cloud droplets during cloud formation because these particles act as cloud condensation nuclei. The number concentration of cloud droplets in turn affects the albedo of the clouds in the sense that more but smaller cloud droplets result in higher albedos. Due to multiple scatter of light on the cloud droplets within the cloud a certain fraction of light becomes absorbed in the cloud resulting in a decrease of the cloud albedo. The amount of absorbing material in cloud droplets therefore controls also the cloud albedo. Under normal atmospheric conditions the mineral dust will hardly influence the number concentration of cloud condensation nuclei and thus the cloud droplet concentration. But in clean air condition the mineral dust may be important for the absorption characteristic of the cloud droplets due to the content of iron oxides and other mineral compounds, just as it may be important for the absorption characteristics of the aerosols themselves.

Details of the radiation effects of aerosols are rather complex but in a simplifying way we may say that the gasformed particles affect primarily the aerosol number concentration whereas under clean air conditions the mineral dust component may affect the absorption characteristic of the visible light both of aerosols and clouds resulting in changes of the Earth albedo. Increase of the mineral dust content of the atmosphere should therefore under most conditions diminish the albedo and increase the amount of solar radiation absorbed by the Earth.

The discussion of loess formation clearly indicates that apparently considerable changes in atmospheric dust content have occurred during and after the ice age. There are also clear indications that man has increased the erosion rates of soil and modified the plant cover over large parts of the world due to agricultural activities over the last 5000 years or so, and is apparently still continuing to do so. This may be accompanied by large scale natural changes of the extent of the arid zones with corresponding changes of dust production. Turbidity data from meteorological observatories in the USSR, some of which are located in remote areas, apparently not affected up to the present time by industry and air pollution show, for instance, trends towards higher values over the last 30 years. It is possible that this trend is due to a large scale increase of dust production over Eurasia as a result of man's activity (Machta, 1972). It is for these reasons that mineral dust has in recent years attracted the interest of meteorologists and climatologists.

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