Geology and meteorology of Saharan dust

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GEOLOGY AND METEOROLOGY OF SAHARAN DUST.

Mobilisation and Transport

Saharan dust is picked up by wind action from central regions of the Sahara desert, for example the alluvial plain of Bilma in Niger and the Tibesti massif of Chad. Dust which affects the Canary Islands probably originates on the Ahaggar massif of Libya (Fig 5.1).

It is brought to heights of up to about 6000m by the strong convection currents driven by daytime heating of the ground. The heavier particles are sedimented out of the dust by gravity during the convective mobilisation.

At heights of about 1 to about 6 km transportation of the dust by horizontal high speed winds starts (Fig 5.15). The dust shears into a Saharan Dust Layer. The dust is homogeneous: it has been fractionated to small sizes by the settling process and it is from a single geographical origin (although it may be of heterogeneous primary geological origin e.g. an alluvial soil). (p. 111).

A small degree of further settling by gravity of the larger particles (> 10 micron**)** takes place during horizontal transportation, and there may be erosion of the underside of the Saharan Dust Layer by convection from the ground.

The dust is transported by the prevailing wind from its origin. This wind depends on the season. Dust from the eastern Sahara is transported NW towards Israel and the Indian Ocean. Dust from N Africa is transported into S Europe especially Spain and France but occasionally to Scandinavia. Dust from Central Africa is transported to Nigeria. Dust from both Central Africa and NW Africa is transported over the Atlantic and reaches the Caribbean. Surface wind patterns are in Fig 1.1 and 1.2.

Approximately 200 x 10^6 tonnes/yr. is removed from the Sahara, which is about half the total global supply of dust into the troposphere and about 1/5 the supply of sea spray. The quantities involved have significant effects on ocean plankton nutrition (Fig 3.5) sedimentary deposits on the ocean floors (p. 201), deposits of loess in the Mediterranean, Atlantic and the West Indies, as well as local climate and erosion at the point of origin of the dust.

The Canaries lie on the N boundary of the summer trajectory of Saharan dust produced in W Africa. In winter the prevailing wind from the points of origin at the height of a few km runs west at a latitude of 10^o N well south of the islands. In summer, the jet stream is centred at a height of 10 km above latitude 18^o and controls the westerly flow of air at altitudes above 4 km in the zone from 10^o to 30^o (Fig 1.4). The boundary of this flow reaches the Canaries during June, July and August. After a major pick up, a typical dust plume develops across the whole of the Atlantic to the Caribbean in 5 days (Fig 5.13). Whilst the scale size of the plume is thousands of km, the edges are relatively sharp with a scale size of 100s of km.

Composition, including size distribution

The chemical composition of the dust clearly depends on the point of origin. Since the point of origin and prevailing wind define the place to which the dust is transported, dust transported to different places is different. The size distribution depends also on the point of origin, and also on the details of the method by which the dust is picked up. The size distribution is modified greatly in the initial stages of transport, but after injection into the upper atmosphere the size distribution is not much changed. Dust consists in part of organic material (so-called ‘’coaly rods´´) which appear to be burnt grass stems, but is mostly inorganic sandy material from regions of high turbulent winds.

Observations of composition and size distribution include (p 197)

a) Dust transported to Ευροπε (1901)
   quartz, rounded, 2 to 7 μ in diameter

b) Arosa, Switzerland (1936)
   calcite, < I to 100 μ, median size 5 μ

c) Luxembourg (1947)
   quartz, < 8 μ

d) West of Canary Islands (1962)
   aggregate particles, of quartz mica and clay, calcite and rounded quartz. 80% were between 5 & 30 μ.

[P.M. Game (1964) “A dust fall in the E Atlantic Feb. 1962” J Sediment. Petrol. 34 355-359.1

In summary, the principal component is quartz, part of which is rounded (by rolling in the wind) and frosted. Next in abundance is calcite or mica, then plant remains. There is no volcanic glass. Off the W African coast sizes are generally < 15 μ but in Europe smaller.

[O E Radczewski (1939) “Eolian deposits in marine sediments”.

in Transk P D (ed) <u>Recent marine sediments</u> Am. Assoc. Petrol. Geol. 496 - 505.1

The size distribution for aeolian marine sediments off the N African coast (near the Balearic Islands) is in Fig 10.4 and may be representative of the Canary Island Saharan dust.

> The mean refractive index (p 91) of Saharan dust collected from Atlantic stations (Canaries, Cape Verde islands and West Indies) is 1.56 - 0.006 i.

“Turbidity”

This is the term used by meteorologists for the opacity of the atmosphere. It can be crudely measured by the visibility of landscape features at various distances. Fig 7.16 is a distribution map for oceanic visibility. Visibility (V. km) is correlated with the concentration of dust (M gm m^{-3}) via an empirical relation

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

where C (given in m^{-3} km) is a parameter dependent on the particle size distribution and other detailed properties. The value of the exponent γ is experimentally found to be close to 1.

	C(g m^{-3} km for $\gamma = 1$)
Urban environment	1.8×10^{-3}
Deserts with no local erosion, but in the general region where dust storms occur	2×10^{-2}
Thousands of km from the dust source	1.4×10^{-3}

For the Canary Islands when Madeira can be seen from Roque V > 500km, M <= 3×10^{-6} gm m^{-3} (i.e. 3 µg m^{-3}). On the other hand in a typical dusty period in the Canaries when Teide cannot be seen from Roque, V <= 150 km, M >= 1×10^{-5} gm m^{-3} (10 µg m-3). Compare Fig 2.1

Note: density of quartz is 2.6 gm cm-3

Atmospheric turbidity is defined as the extinction of solar radiation by suspended particles (aerosols) with radii in the range 0.1 to 10 µ (p 172). The Volz turbidity is B in the expression for the intensity of the Sun:

$$S\ I_{\lambda} = I_{0\lambda} \times 10^{-(\tau_{R\lambda} + T_{o\lambda} + B_{\lambda})MP/P_{0}}$$

where

I_λ = irradiance at wavelength λ at the observing point
$I_{0\lambda}$ = extraterrestrial irradiance at mean Earth-Sun distance
S = correction for eccentricity of Earth orbit
$\tau_{R\lambda}$ = Raleigh scattering coefficient of air
$T_{o\lambda}$ = absorption coefficient for ozone
B_λ = extinction coefficient for the aerosols
M = air mass
P = station pressure
P_0 = standard pressure at sea level

Angstrom's wavelength exponent α for scattering is given by σ approx. $\lambda^{-\alpha}$ and can be calculated by $\alpha = (\ln(B_{\lambda 1}/B_{\lambda 2}))/(\ln(\lambda_2/\lambda_l))$. Typical continental aerosols (sea level) have $\alpha = 1$. Arid region aerosols have α approx. 0.

Turbidity for Atlantic stations in the summer 1974 ranged up to B_{500} approx. 0.7 (Isla da Sal). B_{500} is close to the astronomer's Visual extinction. α values during dust haze at Miami were typically 0.5. (Fig 8.4 - 8.6): mean values at Sal and a marine station at 15^o N 32^o W were 0.36 and 0.26 respectively.

Saharan dust over La Palma

In 1984 the Saharan dust period over La Palma lasted from June 10 to August 30; in 1985 from June 5 to September 5. In 1984 45% of the nights in June, July and August were affected, in 1985 35%. Visual extinction in the zenith ranged up to 0.8 mag. Jones (1984; La Palma Technical Note No. 10) found that Saharan dust in June 1984 had an Angstrom wavelength exponent of $\pm\alpha < 0.02$ for the range $350 \text{ nm} < \lambda < 550 \text{ nm}$.

A sample of dust which had precipitated onto La Palma in 1983 was characterised by orange, yellow and transparent quartz-like round particles in the size range 20 to 60 microns. (Murdin 1985; Vistas in Astronomy 28, 449 section 3.5).

Reference: Saharan Dust: Mobilisation, Transport and Deposition SCOPE 14 Workshop 1977. Ed. C. Morales J. Wiley 1979. [Page and Fig references are to this book].

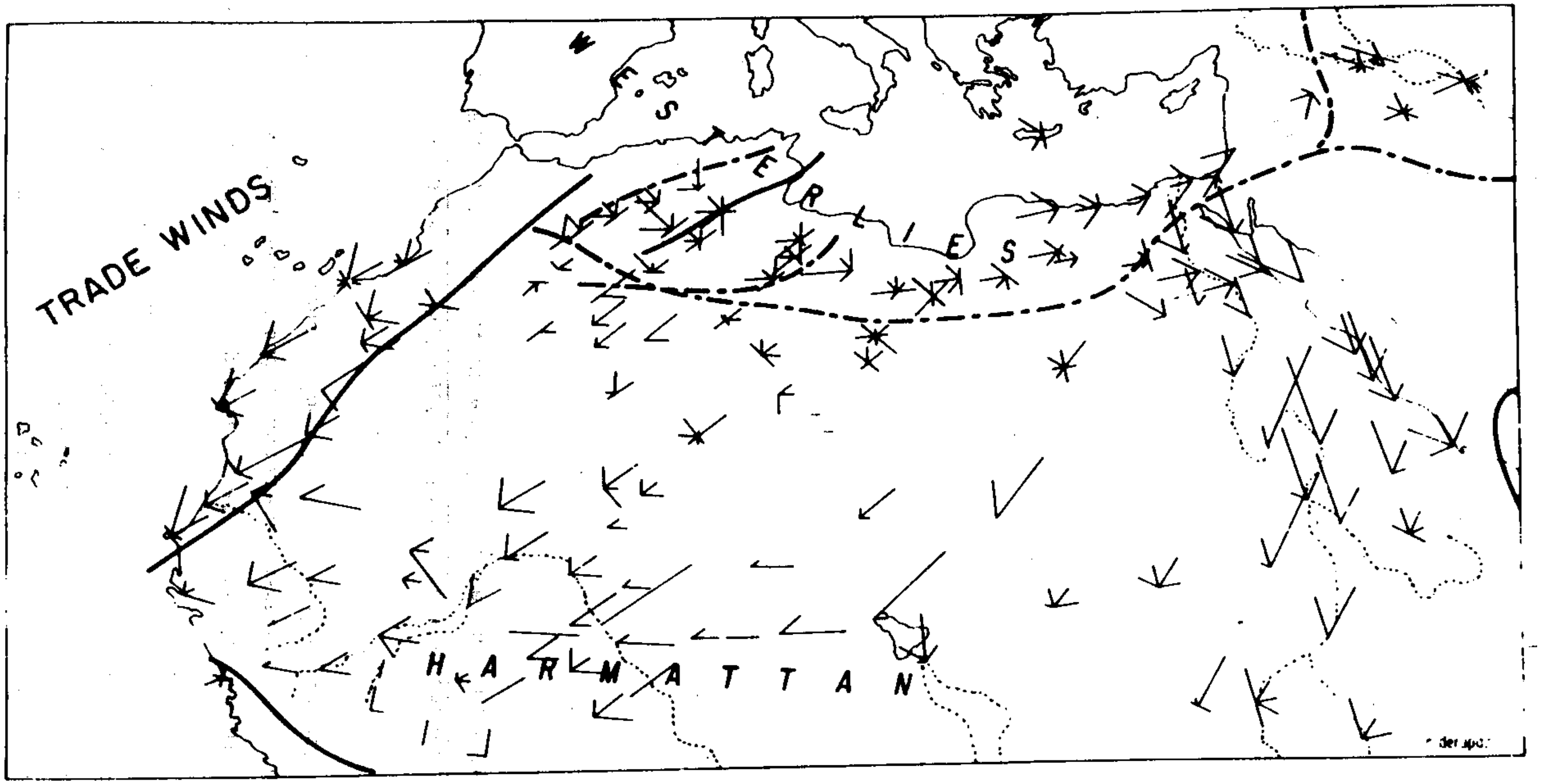

Figure 1.1 Winds in January

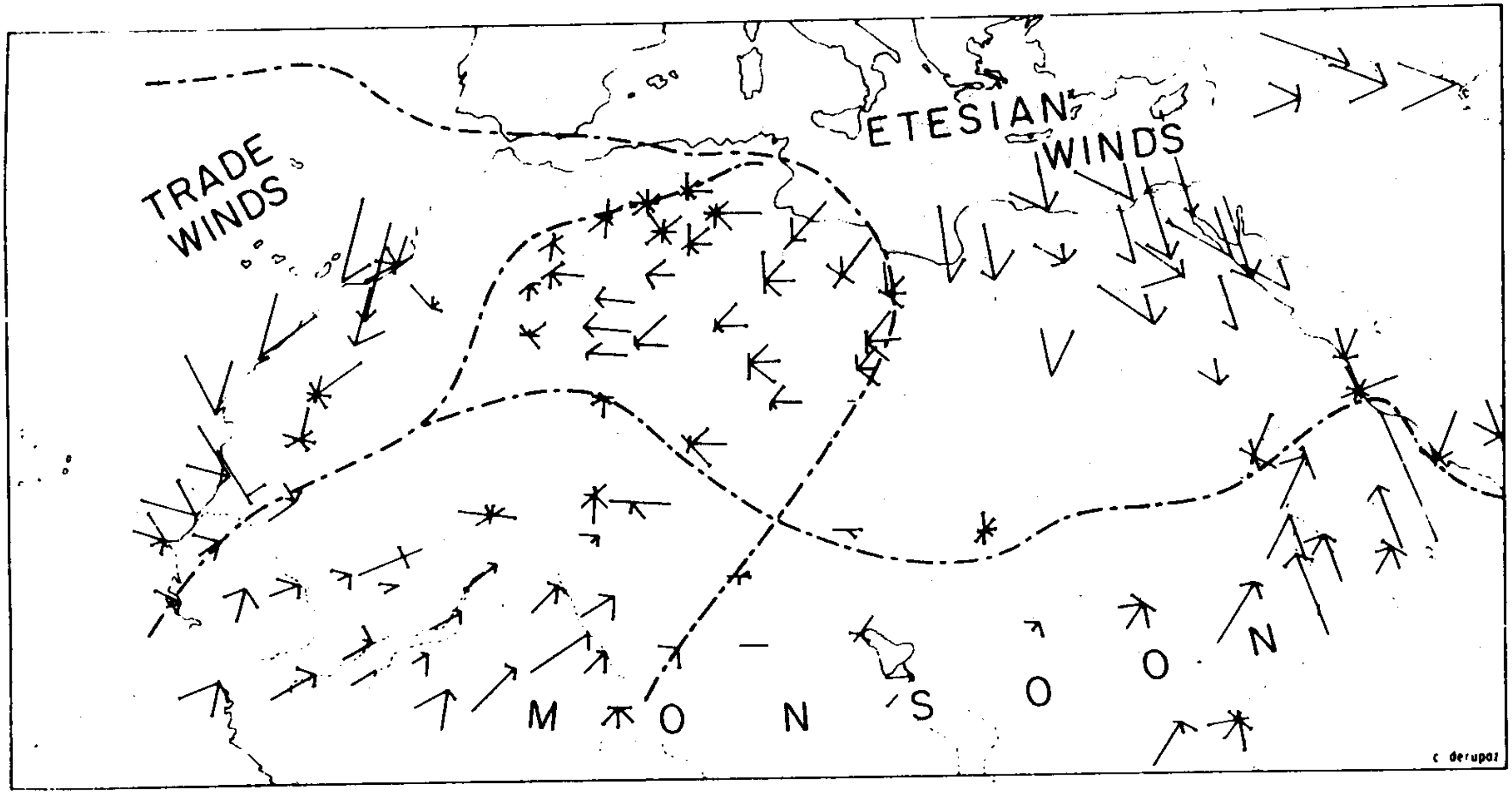

Figure 1.2 Winds in August

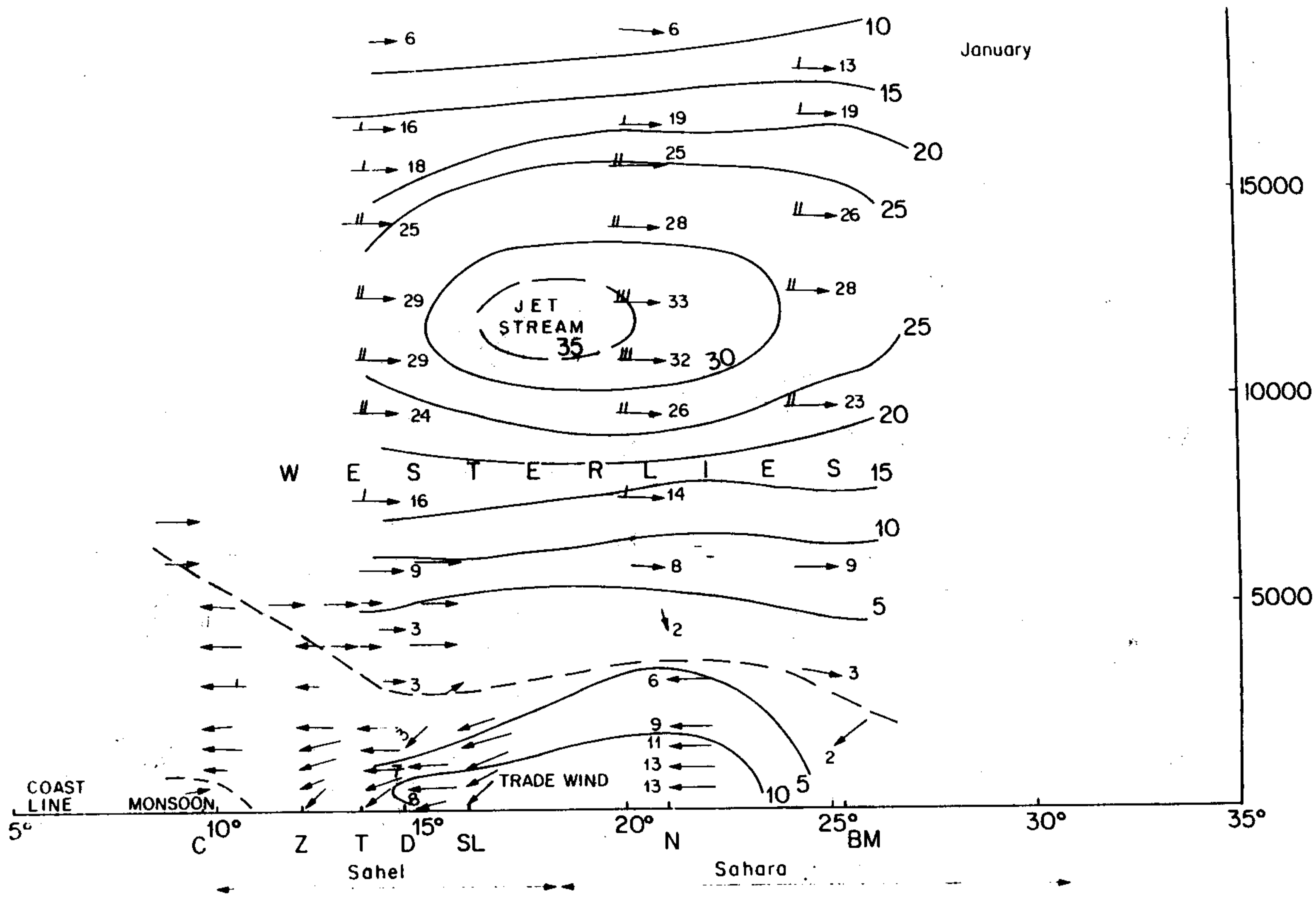

Figure 1.4 Predominant wind directions and mean wind speeds at upper levels at the African Atlantic coast in January

C = Conakry
Z = Ziguinchor
T = Tabacounda
D = Dakar
SL = Saint Louis
N = Nouadhibou
BM = Bir Moghreim

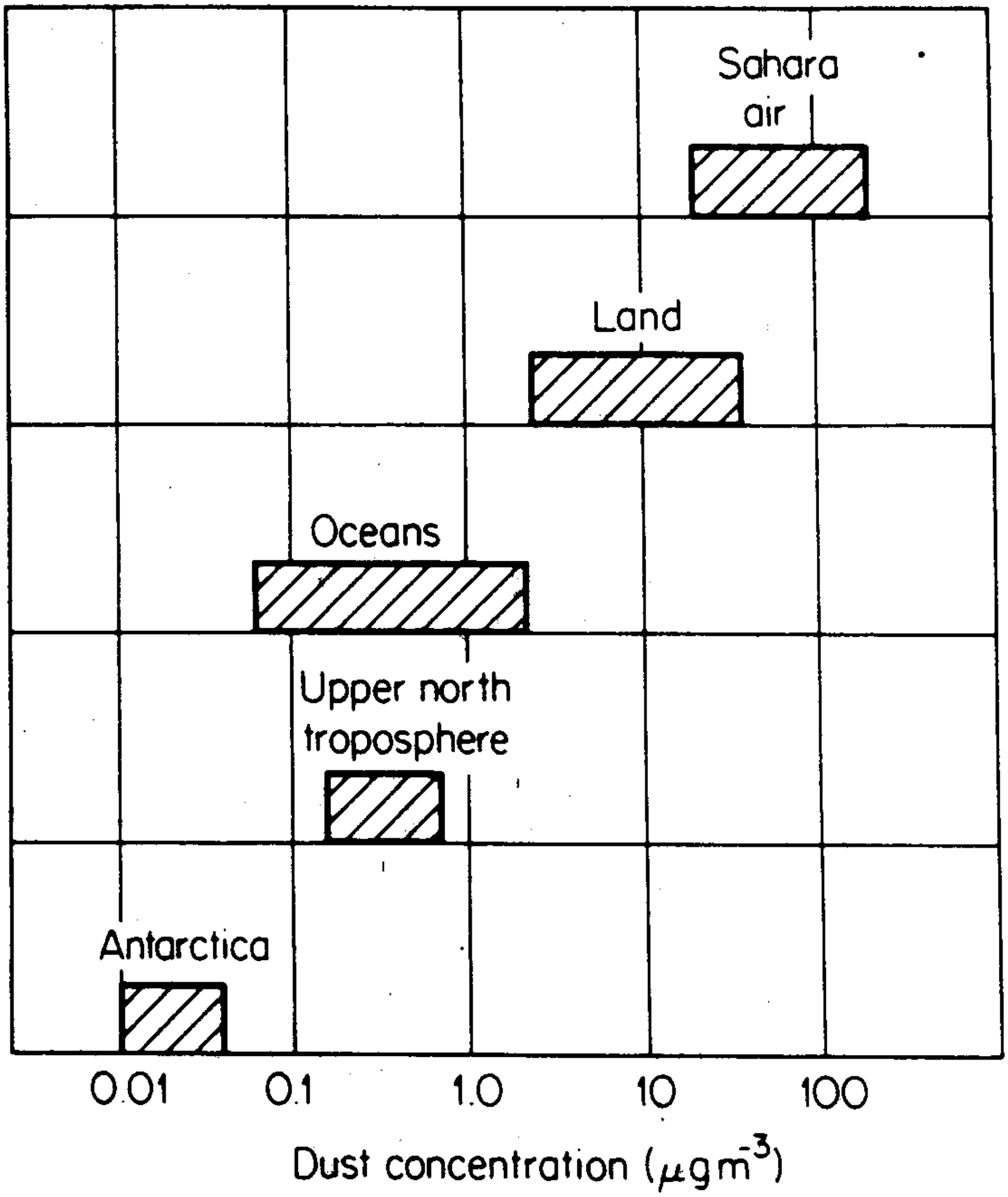

Figure 2.1 Survey of the concentration ranges of mineral dust in the troposphere based on numerous studies by – among others – the following authors: Blifford, Chesselet, Duce, Ferguson, Hoffman, Gillette, Goldberg, Griffin, Jaenicke, Prospero, Rahn, Schütz, Winchester, Zoller

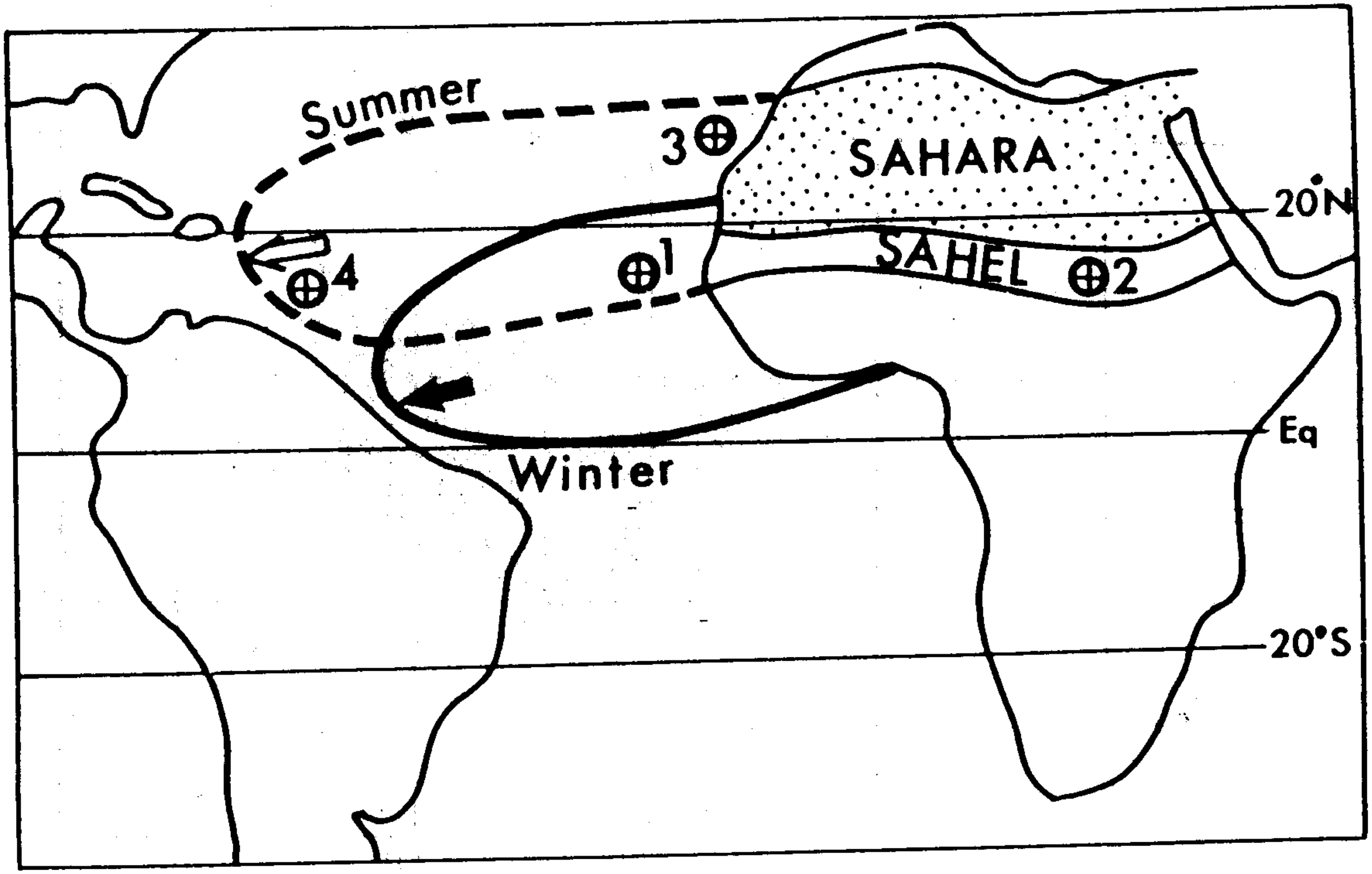

Figure 3.3 Dust transport over the Atlantic (after Rapp, 1974)

3 ~~1~~ = Tenerife
2 = Jebel Marra Mountains, Sudan
1 ~~3~~ = Cape Verde Islands
4 = Barbados

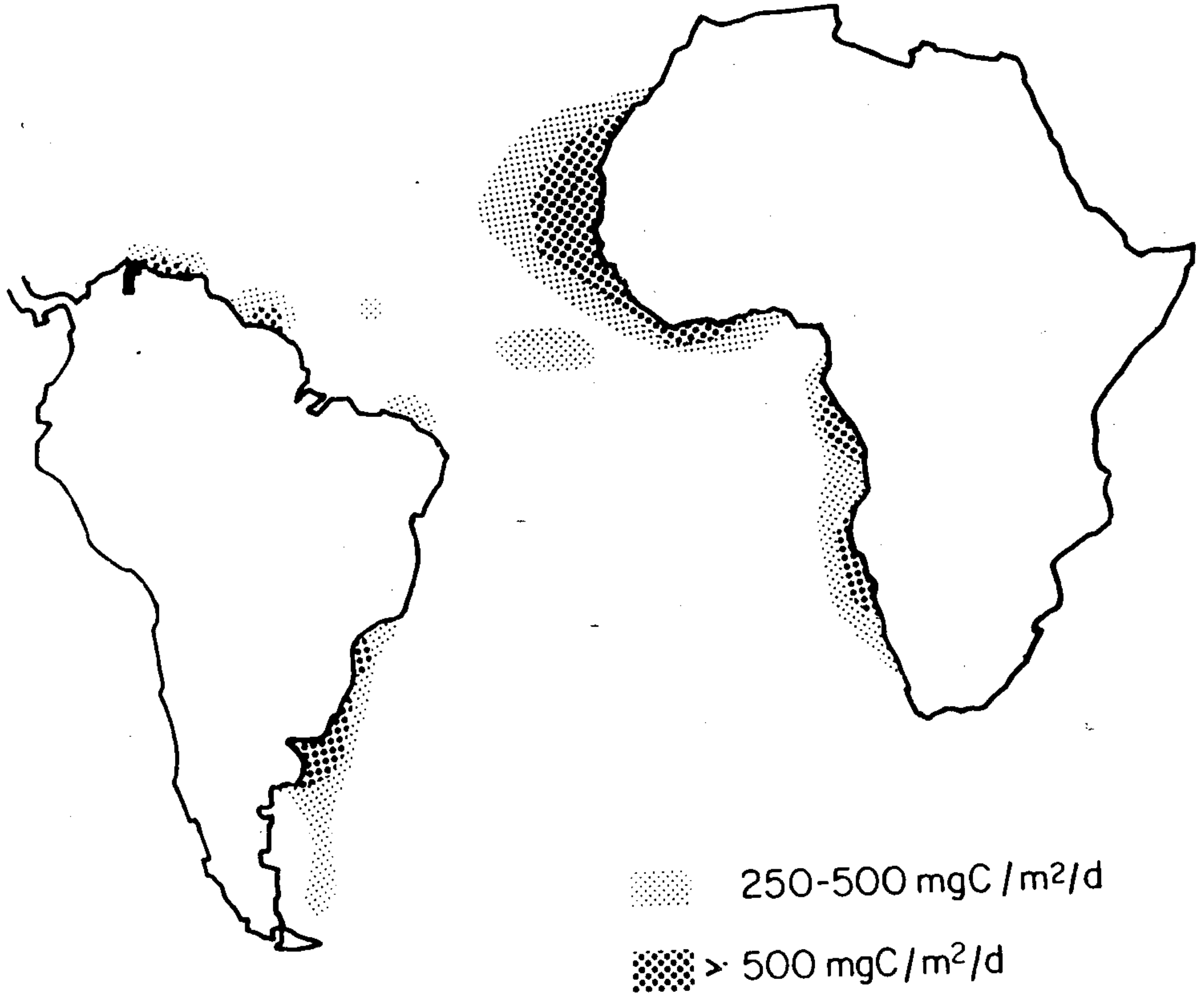

Figure 3.5 Phytoplankton production based on information from *Atlas of the Living Resources of the Sea* (FAO, Rome, 1972)

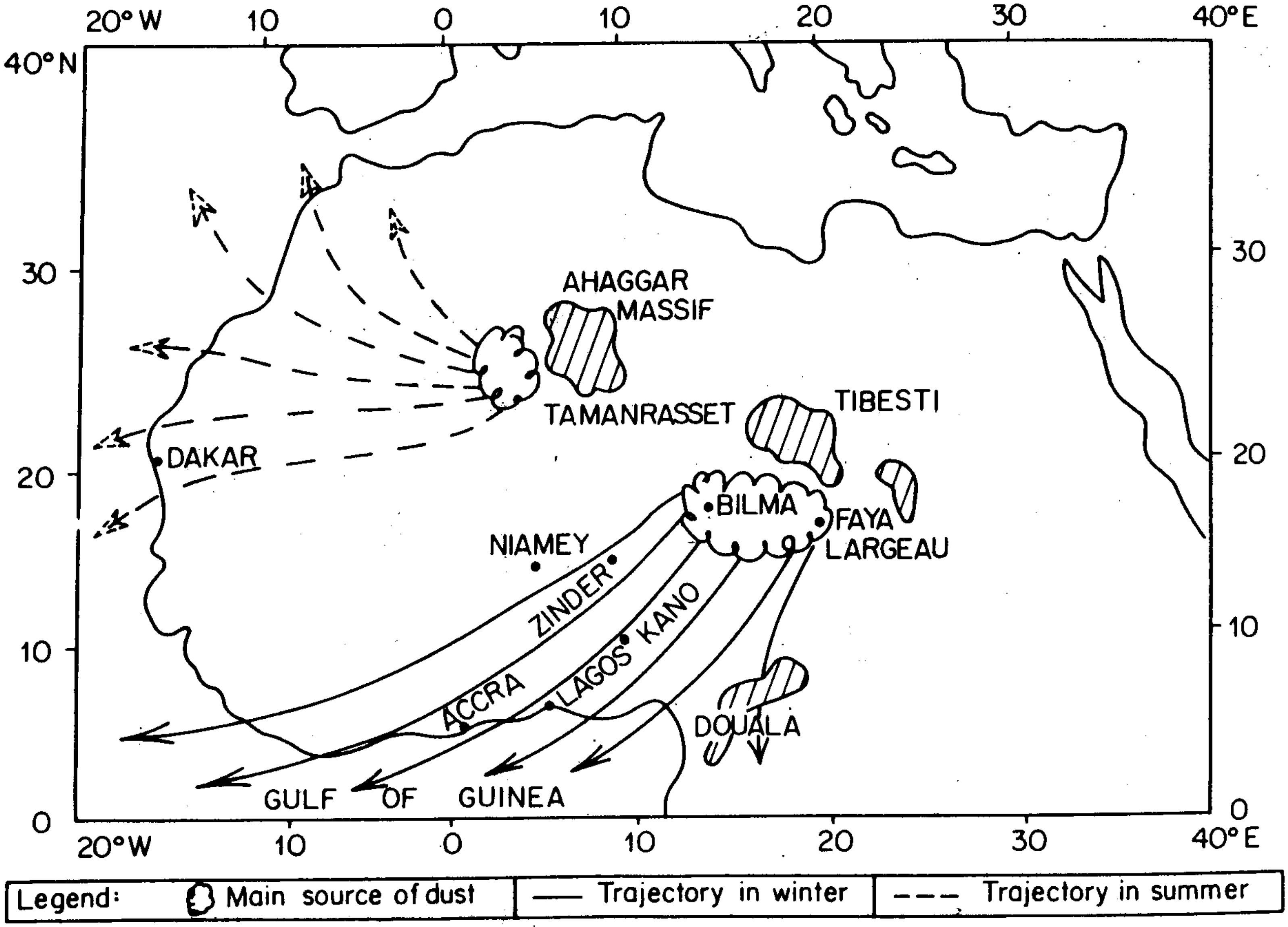

Figure 5.1 Saharan dust trajectories across West Africa

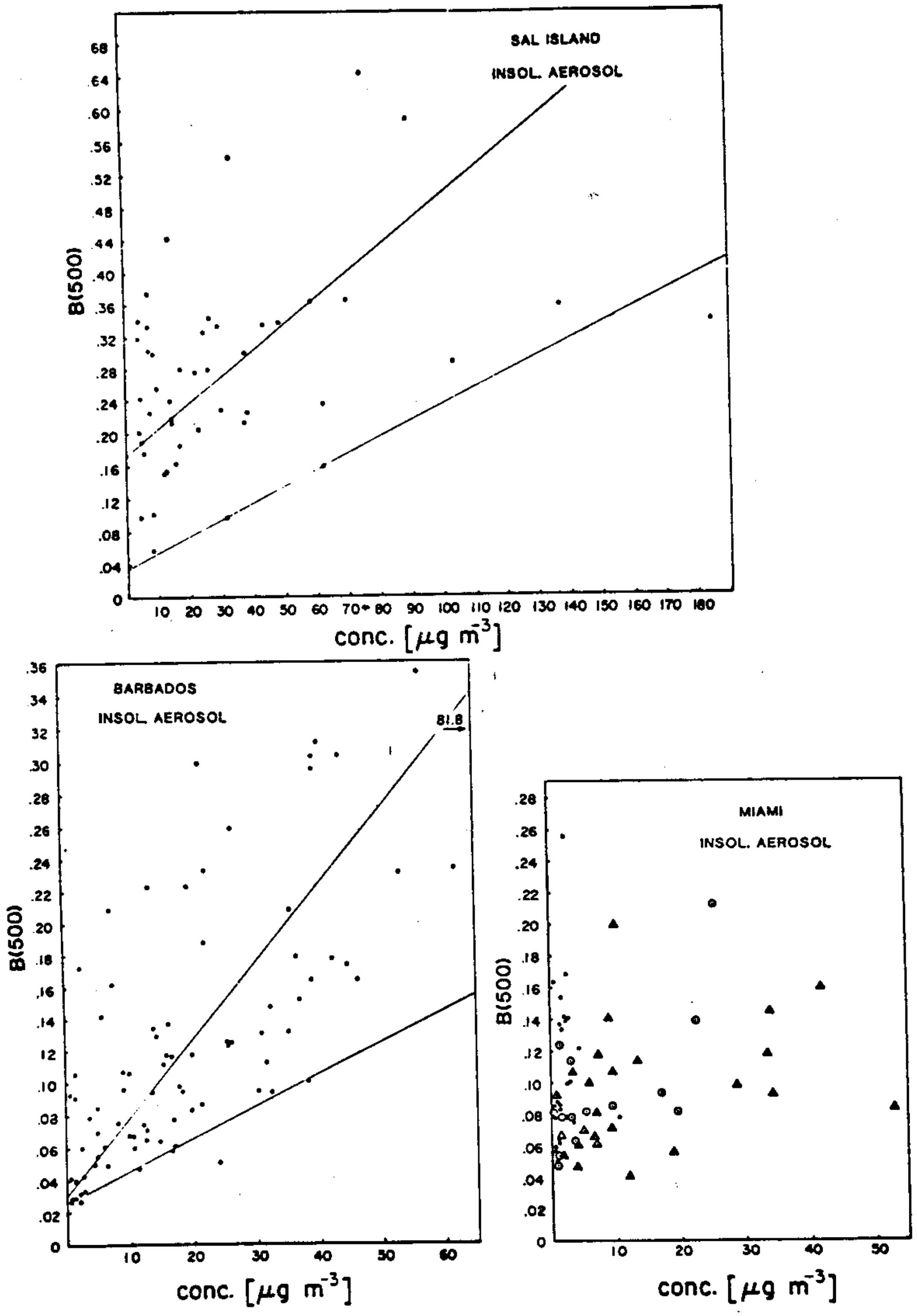

Figure 8.4 Volz turbidity *vs.* mineral aerosol concentration (10^{-6} g m^{-3}). The upper line in the Figures Sal Island and Barbados represents a least squares fit to the data points. The lower line is drawn by eye to show the trend of the lower limit of B *vs* aerosol concentration

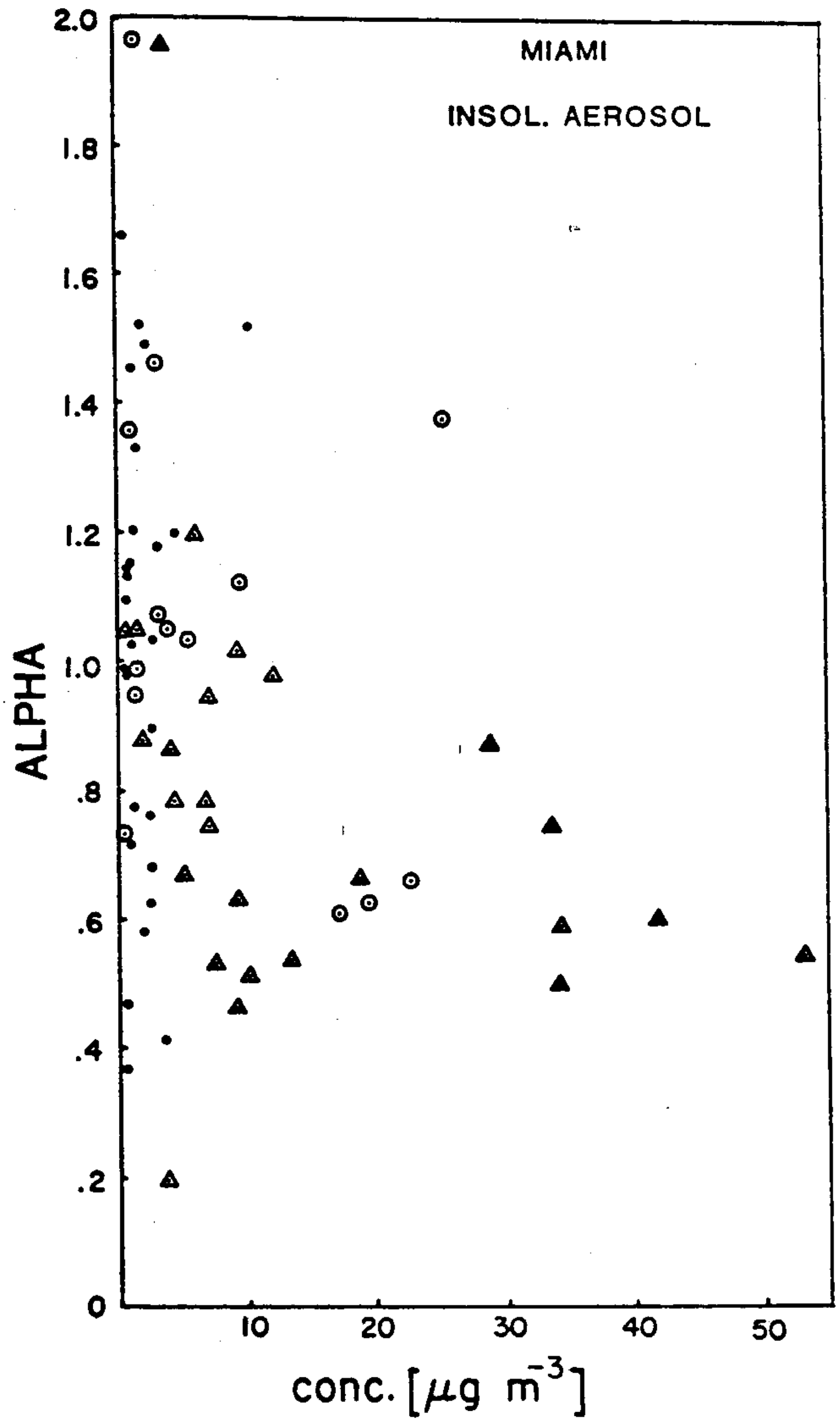

Figure 8.6 Alpha *vs.* mineral aerosol concentration (10^{-6} g m^{-3}), Miami

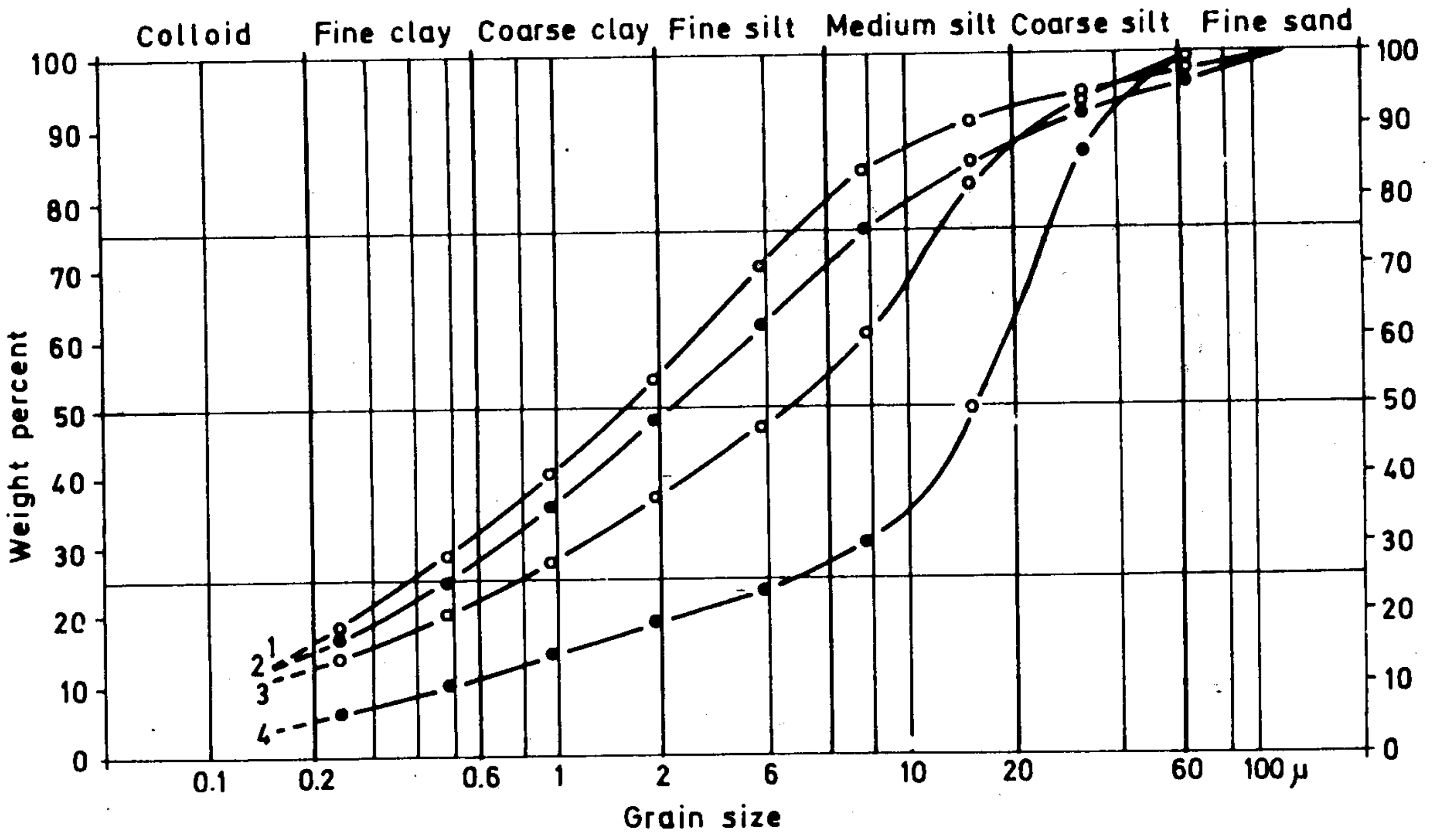

Figure 10.4 Cumulative curves of sediments from Core No. 210

(1) Ordinary, homogeneous sediment; mean of 77 pipette analyses.

(2) Ordinary, heterogeneous sediment; mean of 22 pipette analyses

(3) Microstratified wind-borne sediment; mean of 8 pipette analyses from bed 312.0–284.0 cm.

(4) Microstratified wind-borne sediment; sample 309.5 cm

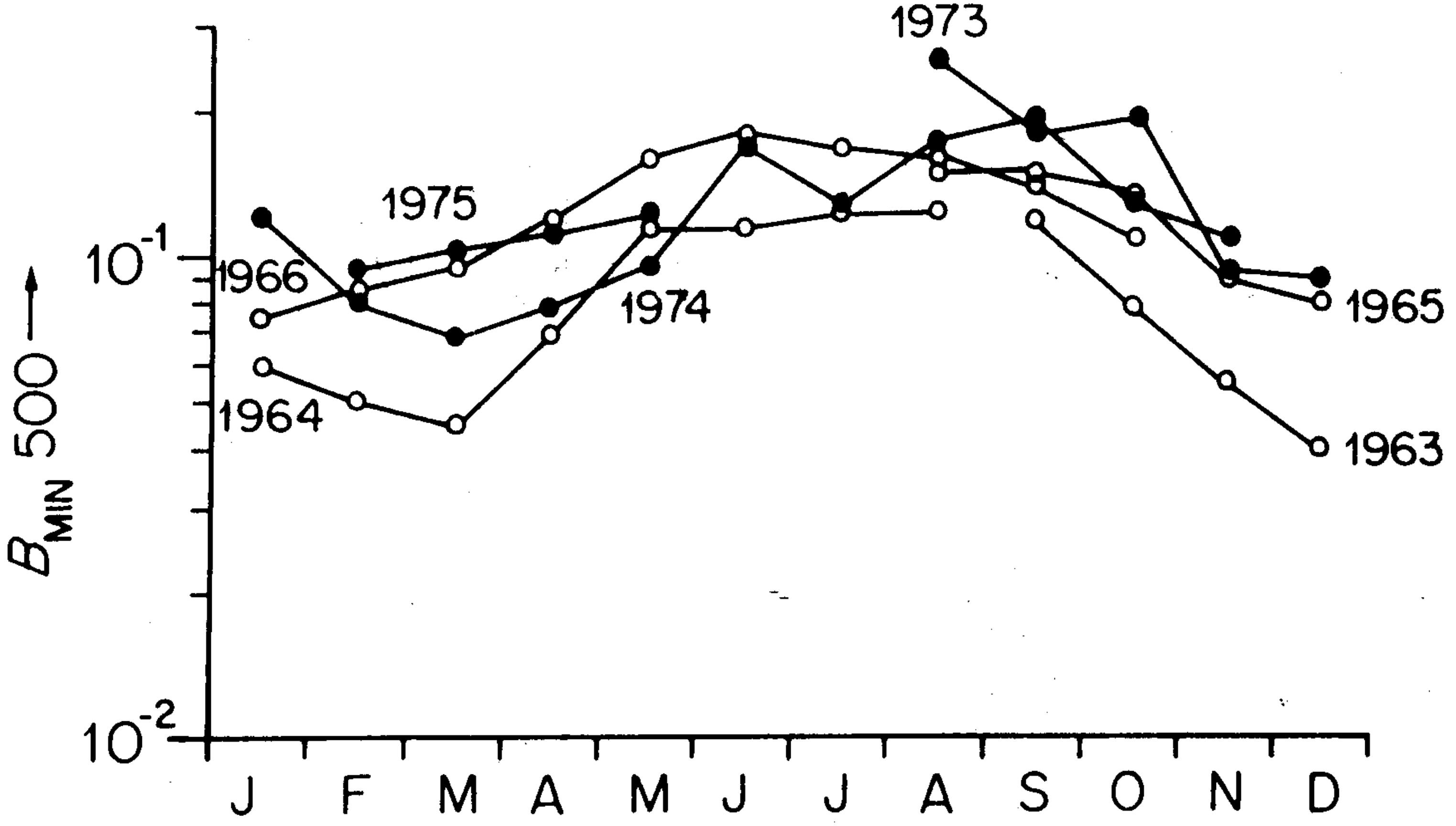

Figure 12.4 Turbidity data from the Cape Verde Islands during 1963–1966 (Volz, 1959) and 1973–1975 (Jaenicke and Kasten, 1977) $B_{\mathrm{min}\ 500}$ was calculated according to Volz