EOLIAN SEDIMENT TRANSPORT,

SLIMS RIVER VALLEY, YUKON TERRITORY

BY

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A Thesis Submitted to the School of

Graduate Studies in Partial Fulfilment

of the Requirements for the Degree of

Doctor of Philosophy

Department of Geography and Regional Planning

University of Ottawa

Ottawa, Canada

1975

to my grandfather

ABSTRACT

The amount of sediment transported by the wind in surface creep, saltation and suspension was measured in the Slims River Valley, Yukon Territory. Results of this investigation showed that the quantity of sediment transported in saltation and creep varied approximately with the cube of shear velocity. This power relationship is similar to the theoretical and empirical models presented by other investigators.

Although the suspended sediment flow rate during dust storms was also significantly related to shear velocity, the suspension flow appears to be more directly controlled by the amount of air turbulence.

A large number of factors other than shear velocity have been shown to affect the amount of sediment transported by the wind. In the Slims River Valley two of these appear to be of particular importance: (1) surface moisture content, and (2) the presence of soluble salts at the surface. Both of these factors tend to stabilize the surface by holding individual particles in place. This effectively reduces the number of grains which can be entrained into the air stream which causes a corresponding decrease in the sediment flow rate.

Although eolian sediment transport is common in the Slims River Valley high saltation creep flow rates and major dust storms appear to be associated with a distinct set of atmospheric and surface conditions. These conditions appear to be best developed on warm clear days following periods of heavy or extended rainfall.

Grain size analysis of the transported sediment indicated that the suspended sediment was significantly finer and better sorted than the sediment transported in either surface creep or saltation. The grain size data also showed that during the dust storms the mean particle size of the suspended sediment decreased as a power function of height.

The grain size distributions of the sediment transported in suspension and surface creep were usually characterized by weak positive skewness. This results from the selective removal of surface particles small enough to be transported by the wind at a given velocity.

In contrast the grain size distribution of the saltation samples were usually positively skewed. This positive skewness may result from the selective removal of the finer particles into suspension from those particles initially lifted from the surface into the air stream. Thus the proportion of sediment initially entrained into the air stream which returns to the surface in saltation will lack the fine grains removed in suspension and will therefore tend to be less positively or negatively skewed.

RESUME

Des mesures de la quantité de sédiments transportés par le vent par reptation, saltation et suspension ont été effectuées dans la vallée de la Rivière Slims au Yukon (Canada). Les résultats de cette étude montrent que le volume de sédiments transportés par reptation et saltation est proportionnel à la vitesse de cisaillement. Cette relation est en accord avec les modèles théoriques et empiriques existants.

Pendant les tempêtes de poussière le déplacement des sédiments en suspension est relié nettement à la vitesse de cisaillement mais semble plus directement controlé par la turbulence de l'air.

Un grand nombre de facteurs, en plus de la vitesse de cisaillement, influence le volume de sédiments transportés par le vent.

Dans le cadre de la rivière Slims l'humidité des sols et la présence de sels solubles en surface jouent un rôle prépondérant. Ces deux facteurs tendent à stabiliser la surface en retenant les particules à leur place, réduisant de ce fait le nombre de particules mobilisables par le vent et le taux de déplacement du matériel.

Bien que le transport par le vent de sédiments soit chose courante dans la vallée de la Rivière Slims, les tempêtes de poussière et les déplacements importants de sédiments par reptation et saltation semblent liés à des conditions du sol et de l'atmosphère particulières.

Ces conditions favorables semblent réunies par les journées claires et chaudes qui suivent une période de pluies intenses ou prolongées.

Les analyses granulométriques mettent en évidence la texture plus fine et un meilleur classement des sédiments transportés en suspension par rapport à ceux transportés en reptation et saltation.

Ces analyses montrent aussi que pendant les tempêtes de poussière la taille moyenne des sédiments en suspension décroit proportionnellement à une fonction puissance de la hauteur, La distribution granulométrique des matériaux transportés par reptation et suspension est caractérisée habituellement par une légère assymétrie positive, résultant de l'abandon des particules suffisamment fines pour être prises en charge par le vent à une vitesse donnée.

Par contre la distribution granulométrique des sédiments transportés par saltation est habituellement marquée par une assymétrie positive, qui pourrait s'expliquer par une prise en suspension des particules les plus fines. Les particules retournant au sol par saltation auront perdu ces particules fines et les distributions auront des assymétries moins positives ou négatives.

ACKNOWLEDGEMENTS

It is impossible to give full credit to the many people who assisted me during the preparation of this dissertation. In particular, I would like to thank my advisor, Dr. P.G. Johnson for his continued guidance, encouragement and support in the field as well as in the writing of the text. I am also indebted to Drs. D. Bennett, H. French, J.P. Johnson, Jr., J.K. Torrance and T.P. Wilkinson for the many hours of helpful discussions which enabled me to clarify my thoughts. I would also like to give special thanks to; Mr. J.P. Windish of the Industrial Hygiene Unit, Health and Welfare Canada for his suggestions and advice regarding the sampling of suspended sediment; Dr. W. Jolly and Dr. P. Peach of the Department of Geological Sciences, Brock University for their assistance in analysing many of the collected sediment samples and Mr. A. Pendlington of the Department of Geography, Carleton University for helping to design and fabricate many of the instruments used in this investigation. I would like to express my appreciation to; the Arctic Institute of North America, Icefield Ranges Research Project for their unfailing financial and logistical support; Messrs. A. Inksetter, G. Catleugh and T. Kent for their assistance during various parts of the field research; Miss B. Gray and Miss L. Cowie for their help in the laboratory; Mrs. L. Payne for her patience in typing the manuscript; Mr. D. Bell for the drawing of the maps and diagrams; my colleagues in the Department of Geography at Brock University for their support and last but in essence first, my wife Nancy, whose continued assistance, encouragement and understanding has made this work possible.

September, 1975 W.G. Nickling

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CHAPTER I

INTRODUCTION

The Problem and Previous Work

The windblown silts of the Yukon and Alaska were first described by Spurr (1898) who applied the name Yukon Silts to deposits observed in the valleys of the Yukon River and its tributaries. Although a considerable amount of work has been carried out on the eolian deposits of Alaska by such authors as Capps (1915), Eakin (1916), Tuck (1938), Taber (1943), Black (1951) and Péwé (1955), only a few investigations of similar deposits have been made in the Yukon Territory by Bostock (1948, 1952), Denton and Stuiver (1966), Borns and Goldthwait (1966), and Nickling (1972).

A large volume of literature has arisen from investigations of these and other loess deposits in North America and Europe with the main emphasis on a descriptive approach so that these sediments could be placed into their respective stratigraphic sequences.

The movement of soil particles by the wind has been studied by a number of investigators such as O'Brien and Rindlaub (1936), Chepil and Milne (1939), Bagnold (1941), Chepil (1945 a and b, 1957), Zingg (1952), Belly (1964) and Williams (1964). The majority of this work, however, has been primarily concerned with the sand sizes with little attention being given to the eolian transport of silt sized particles.

Although the manner in which sand is transported by the wind was presented in a general way by Udden (1894) it was not until relatively recently that a comprehensive study with a theoretical base was available (Bagnold, 1941). Bagnold suggested that when dune sand is transported by the wind three different types of movement can take place:

(1) very fine particles can be carried in true suspension

(2) relatively larger particles can move downwind by a series of bounces (saltation)

(3) surface creep whereby particles are pushed or rolled along the surface by the impact of saltating grains.

Bagnold (1941) found that the greatest proportion of sand movement was by saltation with sediment transport in creep usually being less than 25 per cent of the total sand transport. He also concluded from theoretical calculations that for average dune sand, the suspension flow, even under relatively strong winds, would not exceed twenty per cent of the flow in saltation and creep.

Chepil (1945a) was the first investigator who attempted to assess quantitatively the proportion of sediment carried in true suspension by the wind. Unlike Bagnold's earlier work which was primarily concerned with sand transport in desert environments, Chepil's investigations were focused on the relative amounts of soil carried by wind over cultivated fields in the prairies (southern Saskatchewan).

In these investigations surface creep, saltation and true suspension were measured over open cultivated fields. Saltation and surface creep were measured in traps similar to those used by Bagnold (1941) and the suspended fines in modified impinger tubes first developed by Greenburg and Smith (1922). Suspended sediment was collected over several different soils at four heights (2.5, 15, 30, 61 cm). The limited sample number and the relatively low height at which the samples were collected are closely related to the problems associated with sampling suspended particulate load.

Some of the problems associated with both sampling techniques and sampling instruments in the measurement of suspended sediment have been discussed by Chepil and Milne (1939), Chepil (1945a), and more recently by Watson (1954), Glauberman (1962), and Sehmel (1970).

Despite the limitations of Chepil's sampling technique, his observations showed that the mechanism of soil transport by wind is similar to that of dune sand as described by Bagnold (1941). Chepil noted, however, that in addition to the movement in surface creep and saltation there was in many cases a greater proportion of soil carried in true suspension than suggested by Bagnold's theoretical calculations.

The work of O'Brien and Rindlaub (1936), Chepil and Milne (1939), Chepil (1945 a and b), and Williams (1964) has demonstrated empirically the mechanics of eolian transport of sand and silt near the surface. However, the more theoretical predictive models developed by such authors as Bagnold (1941), Kawamura (1951), and Zingg (1952) are primarily based on wind tunnel investigations of the transport of sand size particles below a height of 0.5 metres. Little consideration has been given to either the suspended sediment concentrations above one metre during eolian transport or the complicating effects of boundary layer conditions.

Chepil and Woodruff (1957) have attempted to assess suspended sediment concentrations above one metre. This work, however, was primarily based on a few selected samples below 3 metres with estimates of sediment concentrations at higher elevations obtained by extrapolation of best fit curves.

Sediment is most susceptible to wind erosion when the surface particles are found in a dry and loose state. These conditions are rarely found, except in sandy deserts. Particles are more often aggregated in various ways to produce erosion resistant structural units (Cooke and Warren, 1973). These structural units commonly referred to as clods must be broken down if wind erosion is to proceed beyond the removal of existing loose particles. The susceptibility of these clods to wind erosion has been found to vary inversely with their mechanical stability, which is a function of interparticle cohesion (Smalley, 1970). Interparticle cohesion is related to the structure of the soil, the organic matter content, soluble salt content[1] and the soil moisture (Chepil, 1945).

Chepil and Woodruff (1963, p.262) have observed that the

> "relative effectiveness of silt and clay as bonding agents depends somewhat on their relative proportions to each other and to the sand fraction. The first five per cent of silt or clay mixed with sand is about equally effective in creating cloddiness, but the quality of the clods is different. Those formed with clay and sand are harder and less subject to abrasion by windborne sand than those formed from silt and sand. For proportions greater than five per cent and up to one hundred per cent, the silt fraction creates more clods, but these are softer and more readily abraded

[1]Defined as a compound formed when the acid hydrogen of an acid is wholly or partly replaced by a metal or metal-like radical.

> than those formed from clay and sand. The greatest proportion of non-erodible clods exhibiting a high degree of mechanical stability and low abradibility is obtained in soils having 20 to 30 per cent of clay, 40 to 50 per cent silt and 20 to 40 per cent sand."

Soluble salts can also affect the interparticle cohesion and susceptibility of soils to wind erosion. The relationship between salt concentration and susceptibility of a sediment surface to wind erosion is dependent on the types and amounts of salt present. In general, relatively low salt concentrations have only marginal effects on interparticle cohesion. However, large quantities of soluble salts, such as calcium carbonate, tend to decrease the cloddiness and mechanical stability of soil and increase erodibility (Cooke and Warren, 1973).

The susceptibility of soil to wind erosion is also strongly dependent on the surface moisture content. Water held within the pores promotes particle cohesion and restricts erodibility (Cooke and Warren, 1973). Belly (1964) in his wind tunnel investigations of sand transport, defined two sources of surface soil moisture, these being:
(1) moisture derived from air humidity, and (2) moisture derived from other sources such as precipitation, flooding, rising of underground water and wave or tidal action.

When a steady stream of moist air blows over a sand surface with a lower moisture content, moisture is transferred from the air to the sand until an equilibrium is established. Belly (1964) has found a high correlation between air humidity and surface moisture content. However, he has shown that the maximum amount of water that can be imparted to the sand by atmospheric humidity is less than 0.25 per cent dry weight.

Although this variation is measurable, it remains rather small for the usual range of air humidities and consequently does not seriously alter the wind speed necessary to initiate sand movement. In contrast, when the surface is moistened by other sources and the soil moisture attains values of 2 - 3 per cent dry weight, the wind strength necessary to initiate movement increases considerably. Field investigations by Woodruff and Siddoway (1965) have shown that the rate of soil movement by the wind varies approximately inversely with the square of the surface moisture.

Of the numerous variables in the wind erosion system the most important and complex is the nature and strength of the wind. Over level ground and away from any obstructions, the neutral wind profile below 20 metres follows a simple logarithmic relationship, first defined by Prandtl (1932). This relationship can be expressed by

$$U_z = A \log z + B \qquad \text{..............1.1}$$

where U_z = velocity (cm/s) at height z (cm)

Further studies (e.g. Von Kàrmàn, 1934) have shown that this can be rewritten as

$$U_z = \frac{2.3}{k} \sqrt{U_*} \log \frac{z}{z_o} \qquad \text{..............1.2}$$

where U_z = wind speed (cm/s) at height z (cm)

U_* = shear velocity (cm/s) defined as $\left(\frac{\tau}{\rho} \right)$ where
τ = shear stress (dynes/cm^2) and
ρ = air density (gm/cm^3)

k = Von Kàrmàn's constant (≈ 0.4)

z_o = roughness length (cm)

Bagnold (1941) in extending Von Karman's (1934) initial work assumed a value of 0.4 for k. More recent work by Zingg (1952), Belly (1964) and Williams (1964) has shown that the relationship is improved if one accepts a value of 0.375 for the Von Kàrmàn constant, k.

The roughness length, z_o , is the actual height above the surface at which the wind velocity becomes zero and is directly related to the maximum height of surface protuberances. The z_o value remains constant for all wind velocities over a given surface, providing that the surface configuration (i.e. roughness) and the grain size distribution of the surface sediments remain unchanged. When the wind blows across a surface, it is slowed down by any surface protuberance (i.e. vegetation, ripples in sand surfaces or individual grains of the surface sediments). The protuberances cause the wind speed to be reduced to zero above the actual surface. The roughness length can be calculated by extrapolation of the curve on the velocity against log height plot (Fig. 1.1).

From empirical investigations, Bagnold (1941) has suggested that the roughness length over a smooth sand surface can be estimated by

$$z_o = \frac{d}{30} \qquad \text{.............1.3}$$

where d = the mean grain diameter in millimetres.

Zingg (1952), however, on the basis of more detailed investigations, proposes the equation

$$z_o = 0.081 \log \frac{d}{0.18} \qquad \text{..............1.4}$$

which holds true over a wider size range.

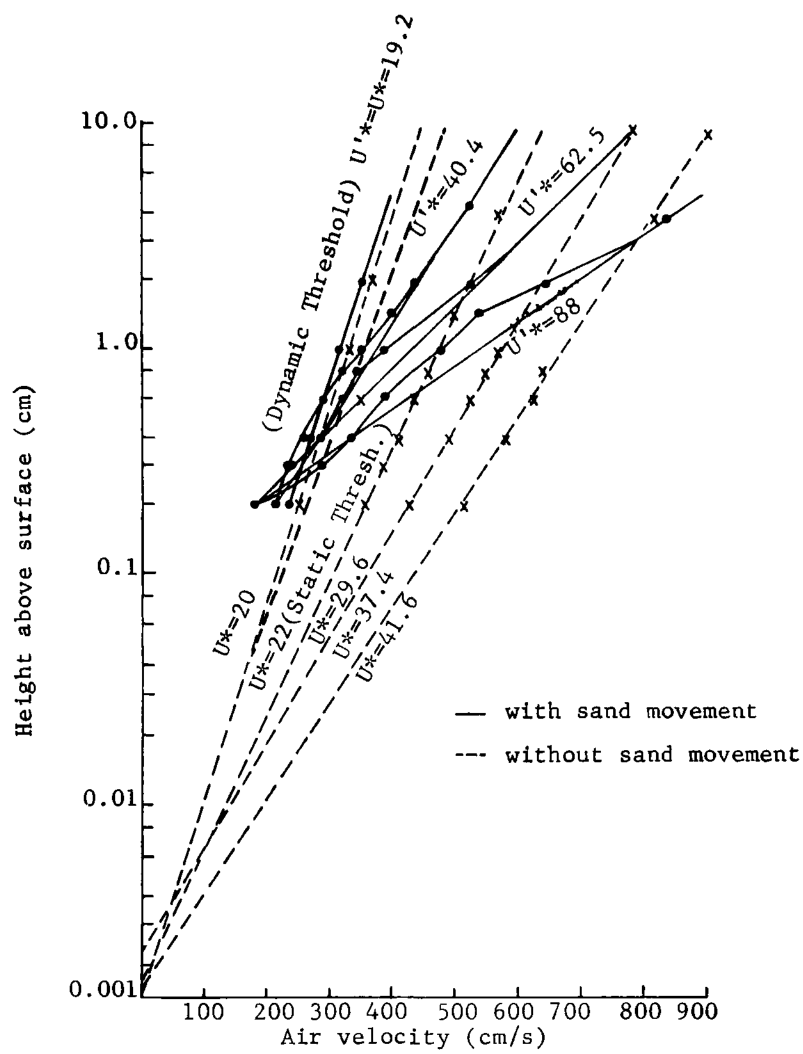

FIGURE 1.1 WIND VELOCITY DISTRIBUTION OVER A STATIONARY AND MOVING SAND SURFACE. (after Bagnold, 1941)

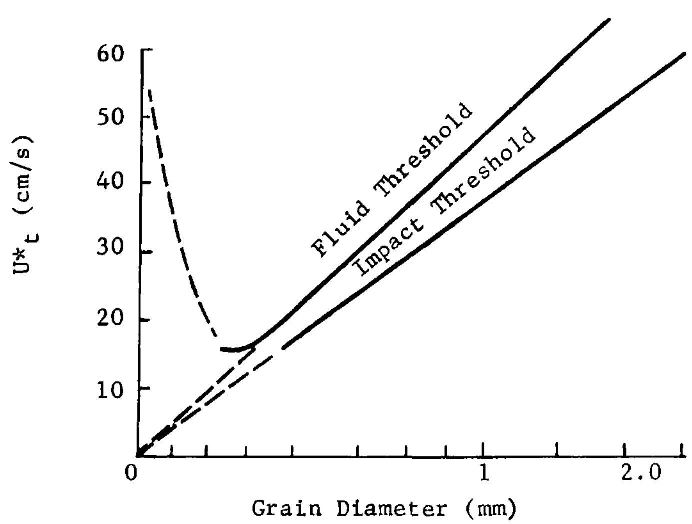

FIGURE 1.2 - VARIATION OF THE THRESHOLD VELOCITY WITH GRAIN SIZE. (after Bagnold, 1941)

Previous investigations have shown that the entrainment and transport of sediment by the wind is primarily a function of the rate of increase of velocity with the logarithm of height (U_*) rather than a function of the velocity at a single height. The value of the shear velocity (U_*)[1] can be determined graphically from the relationship

$$U_* = \frac{\Delta U}{6.13\ \Delta \log z} \qquad \text{..............1.5}$$

where: ΔU is the difference in wind velocity (cm/s) between any two points on the velocity-log height curve and $\Delta \log z$ is the difference between the log height values of the two points.

(after Williams, 1964).

As the shear velocity over a stationary surface is increased, a critical point is reached when the surface grains begin to move by the direct pressure of the wind. This critical velocity has been termed the fluid threshold velocity by Bagnold (1941) and can be expressed as

$$U_{*_t} = A\frac{(\sigma - \rho)}{\rho} . g . D \qquad \text{..............1.6}$$

where U_{*_t} = fluid threshold velocity (cm/s)

σ = density of sand grains (gm/cm^3)

g = acceleration due to gravity (cm/s^2)

D = grain diameter (cm)

A = constant for sand grains larger than 0.01 cm.

[1]The shear velocity (U_*) is also referred to as velocity gradient, friction velocity, shear velocity, shearing stress velocity or drag velocity.

The fluid threshold velocity for a given sediment surface is primarily dependent on two factors: (1) the grain size distribution of the surface and (2) air turbulence near the boundary layer. For grains larger than approximately 0.1 mm, the threshold velocity increases because of the increasing mass of the particles (Fig. 1.2). However, for grains less than 0.1 mm, the threshold velocity also increases because the surface is aerodynamically smooth. In this case, individual grains do not cause small eddies and a non turbulent layer exists around each grain. The drag, instead of being carried by a few more exposed larger grains, is distributed more evenly over the whole surface (Bagnold, 1941, p.89). Consequently, a relatively greater drag (i.e. fluid velocity) is required to initiate movement.

Although the fluid threshold can be precisely defined for a uniform sediment size, it can not be so defined for most natural sediments. Natural sediments, no matter how well sorted, usually contain a range of grain sizes. In such sediments, the fluid threshold is partially dependent on the spatial distribution and protuberance of the larger grains which cause turbulent eddies to form in the air stream. The faster moving air of the turbulent eddies may initiate movement immediately downwind from the grains causing the turbulence.

If the grains which were initially moved attain sufficient speed, they may begin to bounce or saltate downwind. Once entrained into the air stream, the velocity of these grains is increased before they fall back to the surface. Bagnold (1941) has shown that the saltating grains usually strike the surface at a flat angle of between 10° and 16° depending on the size of the grain, its height of rise and the speed of the wind. Bagnold also suggests that two things may occur when these grains strike the surface: (1) they may ricochet off other grains and become re-entrained in the air stream, or (2) they may become embedded in the surface. In both cases, energy is dissipated in disturbing a large number of surface grains. If the particles disturbed by the saltating grains are relatively small, they may also be moved into saltation. Alternatively, if the struck grains are relatively large, they may only be knocked forward by the saltating grains. The force of a saltating grain can move particles with weights far too great to be moved by the drag of the wind alone. A grain moving in saltation under normal conditions can, by impact, move a surface grain six times its diameter or more than two hundred times its own weight (Bagnold, 1941).

Once the wind begins to move surface particles, the wind profile above the surface is altered by the sediment movement. If several wind profiles with different shear velocities (U_*')[1] are taken above the moving sand surface and plotted on semi-log paper all curves will meet at a particular point termed the focus (Fig. 1.1). Thus, no matter how great the shear velocity is once sediment transport begins, the wind velocity (U') at this particular height (z') remains constant. Bagnold (1941) and Zingg (1952) have shown that the wind velocity (U') associated with the focus defines the impact threshold of that particular sediment surface (Fig. 1.2).

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[1]It is conventional to express the shear velocity as U_*' once sediment transport begins.

The height of the focus (z') is associated with the height of the ripples which form on the surface and is somewhat analogous to the relationship between the roughness length (z_o) and the height of surface proturbances. Bagnold (1941, p. 59) suggests that the focus height varies from approximately 0.3 cm for uniform sand in which ripple amplitude is very low, to about 0.8 cm for mixed sand which produces ripples with considerable amplitude.

Bagnold (1941), from field and laboratory experiments, has also suggested that the entrainment of sand grains is only attributable to the drag of the wind (τ) on the grains. Other authors such as Chepil (1945), Von Karman (1953) and Chepil and Woodruff (1963) have argued that particles especially in the fine sand and silt sizes are primarily lifted from the surface by the vertical movements of air associated with air turbulence. Malina (1941) has concluded that entrainment is more likely the result of the interaction of several forces including those which tend to lift the particles from the surface (i.e. drag and turbulence) and those which tend to hold the particles in place (i.e. electrostatic forces, and the presence or absence of moisture). He suggests that, in general, surface drag is probably most important in the case of larger particles with air turbulence becoming of increasing importance as grain size decreases.

The role of air turbulence in the entrainment and transport of sediments, although of possible importance, is as yet not well understood. The difficulties involved in the measurement and mathematical modelling of air turbulence have hindered investigations into the role of air turbulence in eolian sediment transport.

Richardson (1920) derived a non dimensional parameter which specifies the critical point when the atmospheric conditions change from those in which turbulent eddies tend to increase to those in which eddies tend to die out (Sellers, 1965). This parameter is given by

$$R_i = \frac{g[(\partial T/\partial z)+\Gamma]}{T(\partial \bar{u}/\partial z)^2} \qquad \text{..............1.7}$$

g = acceleration due to gravity

T = temperature in degrees Kelvin

z = height

Γ = numerical value of the dry adiabatic lapse rate.

The Richardson number is negative in superadiabatic conditions, positive in stable conditions (i.e. inversion) and zero in dry adiabatic conditions. In the first case air which is displaced upwards will be warmer (i.e. less dense) than the surrounding air and will therefore tend to return to its initial position.

A simplified form of the Richardson number derived by Deacon (1949) has been used by Yeo and Thompson (1954) in their investigation of the dispersion of aerosols released from aircraft. The stability parameter is very similar to the Richardson number and can be defined as

$$F_s = \frac{(T_2 - T_1) + \Gamma(z_2 - z_1)}{\bar{u}^2} \qquad \text{..............1.8}$$

where T_1 and T_2 = the temperatures at heights z_1 and z_2 respectively

Γ = numerical value of the dry adiabatic lapse rate

$\bar{u}$ = the mean wind velocity at an intermediate height between z_1 and z_2.

These authors suggest that the F_s parameter may be more advantageous than R_i for practical applications because it is probably less sensitive to changes in surface conditions than $(\frac{\partial \bar{u}}{\partial z})$. From this it would follow that F_s would also be less sensitive to changes in surface roughness and therefore could be used as a comparative measure for atmospheric turbulence in situations where the surface conditions change over time (Yeo and Thompson, 1954).

Barad and Haugen (1959) and Seale and Couchman (1961) have used a modified form of the above equation in their investigations concerned with the diffusion of effluents emitted from smoke stacks. It can be seen from equation 1.8 that the lapse-height term (Γ $(z_2 - z_1)$) is a constant for a given set of observations and therefore, can be eliminated. Thus, a more frequently used parameter termed the stability ratio can be defined as

$$SR = \frac{T_1 - T_2}{\bar{u}^2} \qquad \ldots\ldots\ldots\ldots\ldots 1.9$$

where T_1 and T_2 = temperatures at heights z_1 and z_2

$\bar{u}$ = mean wind speed at an intermediate height between z_1 and z_2.

The stability ratio is zero under isothermal conditions, positive under inversion conditions and negative in the presence of lapse conditions.

Previous investigators have clearly demonstrated the mechanics by which material in the sand sizes is transported by the wind. Little attention however, has been given to the mechanics or rates of transport of material finer than sand. Also, the majority of these investigations have been carried out in wind tunnels and, for the most part, have ignored the complicating effects of field surface conditions. In this study, the rates of silt and sand transport by the wind are investigated in field conditions with respect to surface and wind conditions.

In this investigation it is hypothesized that

(a) the amount of silt transported by the wind in suspension is directly related to the

    (1) wind strength and air turbulence,

    (2) surface soil moisture,

    (3) surface salt concentration, and

    (4) amount of sand moving in saltation and creep across the surface

(b) previous empirical and theoretical models derived to describe sand transport in saltation and creep near the surface can be expanded to include

    (1) the complicating effects of surface conditions

    (2) material carried in suspension at higher elevations

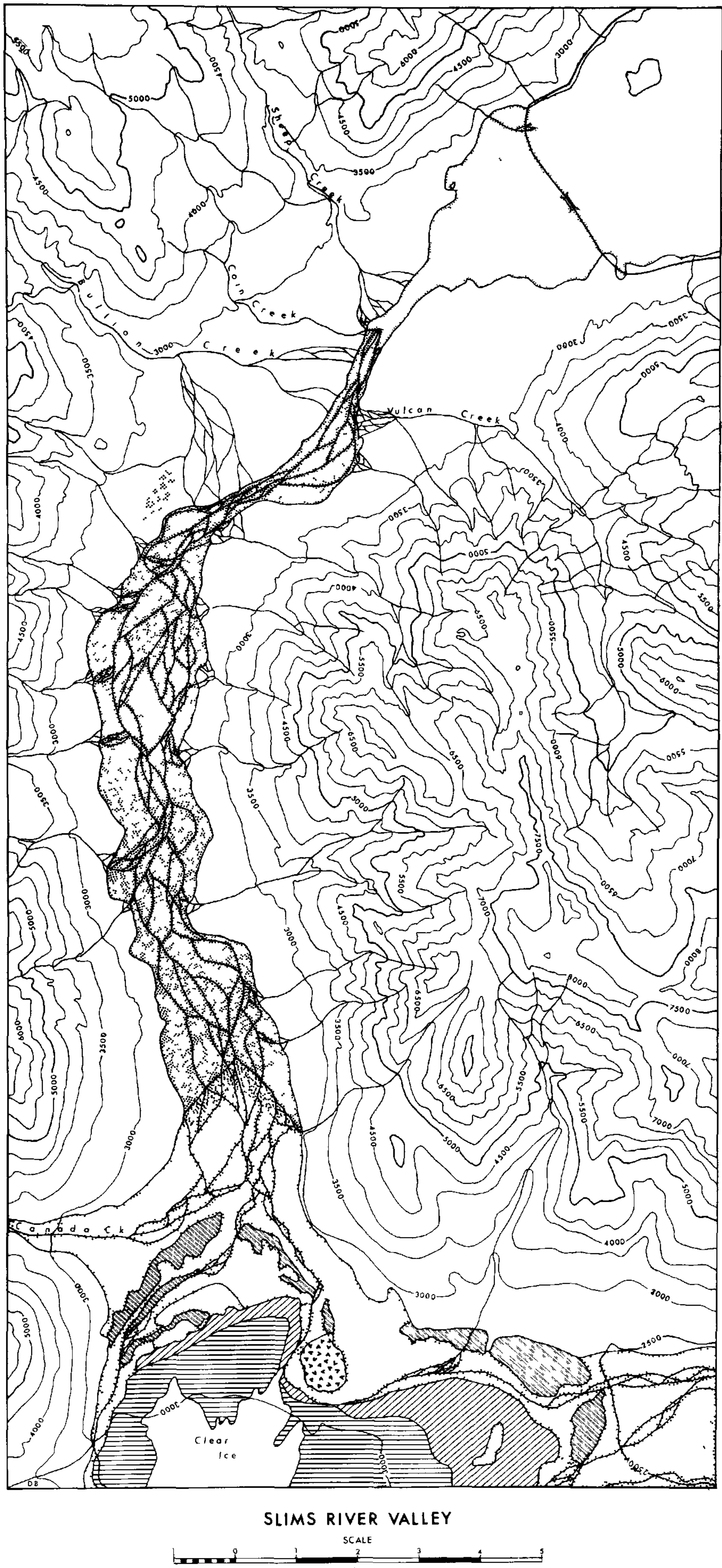
Study Area

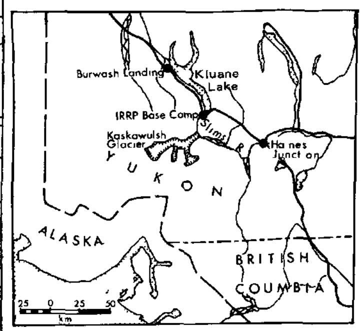
Physiography

Field investigations for this study were carried out in the Slims River Valley, Yukon Territory (Fig. 1.3). The physiography of the general area has been described and various subdivisions named by Bostock (1948, 1952) and Muller (1967). The area lies astride two major physiographic subdivisions, the Yukon Plateau to the northeast and the St. Elias Mountains to the southwest. They are separated by a major structural break, the Shakwak Trench, the deepest part of which is occupied by Kluane Lake (785 m, A.S.L.).

The Kluane Ranges, which are the front range of the St. Elias Mountains, rise from the southwest side of the Shakwak Trench to a maximum elevation of 2590 metres (Muller, 1967). Within this area these ranges are deeply dissected by transverse valleys of the Slims, Duke and Donjek Rivers. These rivers continuously deposit large amounts of silt, sand and gravel on wide floodplains which were formerly occupied by large valley glaciers.

In the Kluane Lake area, relatively uninterrupted loess transport has occurred from the late Neoglacial (maximum 2640 B.P.) to the present (Denton and Stuiver, 1966). Transported sediments found throughout the Kluane Lake area are derived primarily from the active valley trains of the Duke, Donjek and Slims River Valleys. At present, the greatest amount of sediment transport by wind action occurs in the Slims River Valley (Nickling, 1972).





LEGEND

Ablation Drift on Glacier Ice ....

Kaskawalsh Drift Excluding Outwash ... .

End Moraines ....

Dunes ....

Outwash Sediments

Coarse ....

Medium ....

Fine ... .

Ponds .. . .

Bedrock Outcrop . . ..

Map compiled from 1 50,000 and 1 250,000 N.T S. sheets, 1973 air photographs, and base map from Denton and Stuiver, 1966.

Figure 1.3

SLIMS RIVER VALLEY

SCALE

1 0 1 2 3 4 5

km

The Slims River Valley (Fig. 1.3) begins at the terminus of the Kaskawulsh Glacier in the Icefield Ranges. The valley traverses the Kluane Range in a northeast direction, opening into the Shakwak Trench at the southern end of Kluane Lake, approximately 23 km from the glacier (Bostock, 1952, and Muller, 1967). A major bend occurs in the valley approximately 10 km from the glacier. The upper valley axis runs south - north and the lower valley axis southwest - northeast. In the lower valley, the valley walls rise approximately 1000 metres above the delta surface.

The valley is approximately 2000 metres wide except where it is constricted by tributary stream alluvial fans. The Slims River, which drains the meltwater of the Kaskawulsh Glacier, has a braided pattern for most of its course and shows near perfect proximal to distal sorting of the channel sediments. Chaudhuri (1963) has also demonstrated that the geometric mean size of these sediments decreases exponentially from the glacier to the Slims River delta.

The Slims River delta proper is approximately 7 km^2 and is comprised primarily of fine sand and silt (Fig. 1.4). The surface is virtually flat and smooth except for small undulations, surface cracks caused by drying and scour pits resulting from wind erosion. Maximum local relief of the delta sediments is less than one metre, the highest point being on a gentle sloping terrace formed during a high water stage of the Slims River (Fahnestock, 1966).

The Alaska Highway, which crosses the lower delta, has been built up approximately 2 metres above the delta surface and forms a linear barrier to the down valley movement of wind and transported sediments.

AERIAL PHOTOGRAPH OF THE SLIMS RIVER DELTA

FIGURE 1.4

Vegetation is sparse and confined primarily to the sediments near the valley walls. Salt resistant grasses and willows account for the greatest percentage of the vegetation cover.

The lack of vegetation and the fine grained nature of the sediments forming the delta and braid bars in the river make this material extremely susceptible to wind erosion. Consequently, the Slims River Valley has frequent dust storms generated by strong off glacier winds (Muller, 1967). Although dust storms have been observed throughout the year, they are most frequent from May to July. During this period, the Slims River is at a relatively low stage, exposing a larger area of recently deposited silt and fine sand. Temperatures are warm enough in summer to dry the surface of the delta and the surface of braid bars.

Weather Conditions

Weather patterns in the St. Elias Mountain area are primarily due to the position of the mountains as a high coastal barrier blocking westerly air flow in the sub-polar low pressure belt (Benjey, 1969). Summer conditions in this area are characterized by weak pressure gradients with a ridge along the Alaska-Yukon border indentifiable on surface charts (Taylor-Barge, 1969). Circulation associated with the Aleutian low dominates throughout the year in this region. Benjey (1969) has shown that during the summer months of 1965, low pressure systems dominated in this area 82 per cent of the time, while highs were present only 7 per cent of the time.

As a result of the lack of permanent meteorological stations near the study area, only general mean monthly temperature and precipitation data can be given. The closest permanent stations are located along the Alaska Highway at Haines Junction and Burwash Landing. Sporadic meteorological data was also available from observations taken at the Icefield Ranges Research Project (IRRP) base camp (mile 1054) operated by the Arctic Institute of North America. Daily temperature and precipitation data were also collected during the summer months of 1972 and 1973 in the Slims River Valley as part of the present investigation. Temperature and precipitation data from the above sources are given in Tables 1.1 to 1.3.

Coldest mean monthly temperatures usually occur in the Kluane Lake area during January (-30°F, (-34.4°C) to -18°F, (-27.8°C)) with mean monthly maximums in June and July (50°F, (10.0°C) to 56°F, (13.3°C)). Mean annual precipitation in this area is relatively low (11.2", (28.2cm) to 11.38", (28.9cm)) because of the study areas position in the lee of the St. Elias Mountains. Maximum rainfall usually occurs during June and July (1.06", (2.7 cm) to 2.53", (6.4cm)) with maximum snowfalls in November and December (8.7", (22.0cm) to 10.7", (27.2cm)).

TABLE 1.1

TEMPERATURE AVERAGES FOR THE
KLUANE LAKE AREA

BURWASH LANDING, Y.T.
(Based on the period October 1966-October 1972)

	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Year
Mean Daily Temperature (°F)	-18.1	0.2	10.3	26.1	40.7	51.2	53.4	49.9	40.0	25.1	7.2	- 6.5	24.2
Mean Daily Maximum Temperature	- 6.6	13.7	25.1	38.1	52.4	65.3	65.1	61.6	50.4	35.1	17.7	5.0	36.0
Mean Daily Minimum Temperature	-29.5	-13.3	-4.5	14.0	28.9	37.7	41.5	38.3	29.5	15.0	-3.4	-18.0	12.4
Extreme Maximum Temperature	39	50	49	56	70	89	82	79	67	58	51	52	89
No. of Years	6	6	6	6	6	6	6	6	6	7	6	6	
Extreme Minimum Temperature	-66	-67	-56	-25	9	25	26	21	0	-20	-40	-57	-67
No. of Years	6	6	6	6	6	6	6	6	6	7	6	6	
No. of Days with Frost*	31	28	31	29	25	8	4	5	20	28	30	31	270

*Based on the period Jan. 1971- Oct. 1972.

...continued

TABLE 1.1--continued

TEMPERATURE AVERAGES FOR THE
KLUANE LAKE AREA

HAINES JUNCTION, Y.T.

	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Year
Mean Daily Temperature (°F)	- 6.9	3.5	14.2	29.9	41.9	50.8	54.0	50.7	42.8	28.7	9.8	-3.8	26.3
Mean Daily Maximum Temperature	4.4	17.4	29.1	41.6	55.2	65.3	67.5	64.6	55.4	39.7	19.9	6.7	38.9
Mean Daily Minimum Temperature	-18.1	-10.4	-0.8	18.2	28.6	36.3	40.3	36.7	30.1	17.7	-0.5	-14.2	13.7
Extreme Maximum Temperature	54	59	60	69	82	91	88	83	78	66	57	51	91
No. of Years of Record	26	26	26	26	26	26	26	26	26	26	27	27	
Extreme Minimum Temperature	-65	-65	-45	-23	10	20	26	12	1	-23	-53	-65	-65
No. of Years of Record	26	26	26	26	26	26	26	26	26	26	27	27	
No. of Days with Frost	31	28	31	29	24	10	4	10	19	29	30	31	276

TABLE 1.2

PRECIPITATION AVERAGES FOR THE
KLUANE LAKE AREA

BURWASH LANDING, Y.T.
(Based on period October 1966-October 1972)

	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Year
Mean Rainfall (inches)	0.00	T	0.00	T	0.43	1.43	2.53	1.03	0.94	0.04	T	0.00	6.41
Mean Snowfall	8.7	3.3	5.8	8.3	4.6	0.1	0.0	0.7	2.6	6.1	8.7	5.6	49.7
Mean Total Precipitation	0.87	0.33	0.58	0.83	0.89	1.44	2.53	1.10	1.20	0.65	0.87	0.56	11.38
Greatest Precipitation in 24 Hours	0.48	0.21	0.90	0.44	0.69	1.45	1.51	0.51	0.73	0.27	0.59	0.58	1.51
No. of Years	6	6	6	6	6	6	6	6	6	7	6	6	
No. of Days with Measurable Rain	0	0	0	0	4	9	12	9	6	1	0	0	41
No. of Days with Measurable Snow	12	7	7	7	3	*	0	1	3	8	12	10	65
No. of Days with Measurable Precipitation	12	7	7	7	7	9	12	10	9	9	12	10	106

*less than 1/2 day

T trace

...continued

TABLE 1.2--continued

PRECIPITATION AVERAGES FOR THE
KLUANE LAKE AREA

HAINES JUNCTION, Y.T.

	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Year
Mean Rainfall (inches)	0.03	0.02	0.02	0.07	0.48	1.06	1.41	1.04	1.19	0.53	0.23	0.21	6.29
Mean Snowfall	8.4	5.5	4.0	2.6	1.1	0.1	0.0	T	0.6	7.9	9.9	10.7	50.8
Mean Total Precipitation	0.81	0.52	0.39	0.31	0.58	1.07	1.41	1.05	1.25	1.32	1.18	1.23	11.12
Greatest Rainfall in 24 Hours	0.50	0.40	0.15	0.50	0.81	1.30	1.12	0.79	1.28	2.17	2.03	2.30	2.30
No. of Years of Record	26	25	25	26	25	26	26	26	25	26	27	26	
Greatest Snowfall in 24 Hours	13.1	11.5	5.6	5.0	3.2	1.4	0.0	1.2	3.0	26.5	11.8	9.4	26.5
No. of Years of Record	26	24	25	26	26	26	26	26	25	26	26	26	
Greatest Precipitation in 24 Hours	1.31	0.76	0.56	0.50	1.01	1.30	1.12	0.79	1.28	2.65	2.15	2.30	2.65
No. of Years of Record	26	24	25	26	26	26	26	26	25	26	26	26	
No. of Days with Measurable Rain	*	*	*	1	4	6	9	8	8	3	1	*	40
No. of Days with Measurable Snow	8	6	5	3	1	*	0	*	*	5	8	10	46
No. of Days with Measurable Precipitation	8	7	5	4	5	6	9	8	8	8	8	10	86

TABLE 1.3

SUMMER TEMPERATURE AND PRECIPITATION MEANS

		MEAN (°F)						PRECIPITATION (INCHES)				
		MAXIMUM		MINIMUM		TEMPERATURE						
		Slims River Valley	IRRP Base Camp	Slims River Valley	IRRP Base Camp	Slims River Valley	IRRP Base Camp	Slims River Valley		IRRP Base Camp		
								Total	Days	Total	Days	
May	1971		48.6		27.7		38.2			0.05	1	
	1973	53.7*		33.4*		43.5*		0.16*	1			*last 20 days
June	1946		68.5		41.3		54.9					
	1970		58.0		37.6		47.8					
	1971		61.8		37.7		49.8			1.51	9	
	1972	60.7*		41.5*		51.1*		0.53*	4			*last 11 days
	1973	59.0		40.9		50.0		1.51	10			
July	1970		61.4		41.9		51.7					
	1971		68.1		42.8		55.5			0.22	4	
	1972	66.2		43.9		55.1		2.16	14			
	1973	67.3*		44.9*		56.1*						*first 9 days
August	1970		59.8		38.7		49.3					
	1971		62.4		40.7		51.6			2.20	9	
	1972	62.7*		44.7*		53.7*		0.69*	6			*first 7 days

CHAPTER II

FIELD SAMPLING PROCEDURE

Introduction

Although fine grained sediments are transported by the wind throughout the Slims River Valley, most of the material is initially derived from three major source areas (Fig. 1.3): (1) adjacent to the terminus of the Kaskawulsh Glacier, (2) at the major bend in the valley, and (3) in the delta area where the Slims River enters the southern end of Kluane Lake.

Between and within these areas there is a considerable range in the amount of sediment supplied to the air stream. These differences appear to be a function of the amount of material present, the grain size distribution of the sediments and the length of time since their deposition.

At the Kaskawulsh terminus strong off-glacier winds continually winnow fines from sediments deposited by the Slims River and tributary creeks. The majority of the material removed by the wind from this area is in the sand size range. This is directly related to the grain size distribution of sediments deposited in the terminus area by the Slims River.

The Slims River and associated creeks, even during low water stages, have relatively high velocities. Thus, the materials deposited by these streams are predominantly gravels and sands, the finer material being carried downstream in traction and suspension by the Slims River.

Borns and Goldthwait (1966) have shown that a large amount of the sand from the terminus area is carried down valley by the wind and trapped at the major bend in the Slims Valley (Fig. 1.3). This sand has accumulated in the form of dunes and covers an area of approximately 0.65 km^2. These dunes are comprised of medium to fine sand and are actively migrating down valley by saltation and creep.

Sand, silt and clay transported down valley by the Slims River are deposited in great abundance in the form of a delta where the river enters the southern end of Kluane Lake. The delta is comprised primarily of fine sand and silt and is approximately 7 km^2 in area size (Fig. 1.4).

Although the Slims River forms a heavily braided pattern on the delta, the greatest percentage of meltwater is confined in one or two main channels. This is partially the result of meltwater being forced into a single channel in the upper delta where the river passes under a bridge of the Alaska Highway.

The surface of the Slims River delta is by far the most susceptible to wind erosion of all areas in the valley. The regularly occurring down valley winds, abundance of fine grained sediments and the distinct lack of vegetation all contribute to the frequent removal of sediments by the wind. Unlike other active areas in the valley, a large percentage of the sediment is transported by the wind in true suspension. This is evident by the frequent "dust storms" which originate in the delta area. Dust issuing from the Slims River Valley is carried into the Shakwak Trench and across Kluane Lake where it is deposited as far north as the Ruby Range (Denton and Stuiver, 1966 and Nickling, 1972).

Despite the frequency of eolian activity on the Slims River delta, there is a great spatial and temporal variability. It has been found that the most active areas on the delta are associated with the most recent deposits. Although the Slims River changes its course frequently, these changes usually occur within the active floodplain which is approximately 1 km wide. The floodplain is deliniated by a set of terraces, approximately 0.5 metres above the active floodplain, which were cut by high water stages of the Slims River (Nickling, 1974). Sediments within this area are continually being deposited and eroded with changes in the seasonal and diurnal flow of the Slims River. Sediments east and west of this active channel area have remained relatively undisturbed since their deposition, and at present appear less susceptible to wind erosion. The greater susceptibility of the more recent sediments may be directly related to the relative amount of salts present at the surface.

Nickling (1973) has shown that surface salt concentration of the Slims delta increases from the main channel area towards the valley walls. It was also noted that a sharp increase in surface salt concentration occurred above the terrace which marks the normal lateral extent of the Slims River. This area of relatively high salt concentration appears to coincide with the area most susceptible to wind erosion on the Slims River delta.

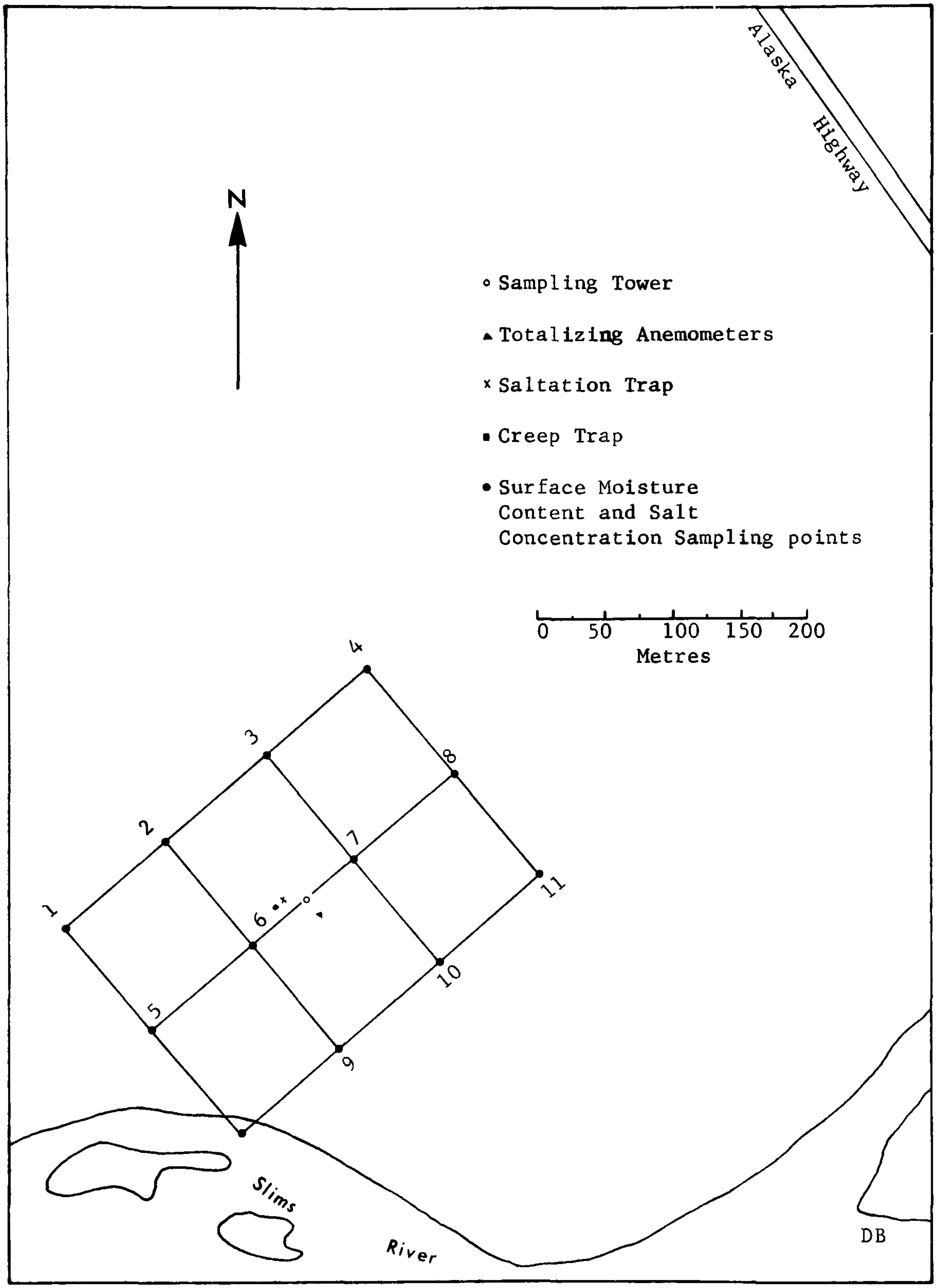
Field Instrumentation

Introduction

An experiment was designed to investigate the hypothesis that the amount of sediment transported by the wind is, at least in part, a function of the wind speed and the surface conditions of the eroding sediment. It was decided that a suitable method to investigate such a hypothesis would be to measure the amount of sediment transport and changes in the surface conditions of a relatively small area where eolian transport was actively taking place. A small study area, or grid was selected and instruments, designed to measure the amount of sediment transport, were installed. Changes in the surface moisture content and surface salt concentration, as well as the amount of sediment transport, were continually monitored during the study period.

The Sampling Grid

Frequency and variability of eolian activity were the main criteria in the selection of the sampling site on the Slims River delta. It was noted from personal field observations that the greatest activity occurred on the west side of the Slims River adjacent to the active channel. Sediments in this area are continually deposited and



SCHEMATIC DIAGRAM OF THE SAMPLING GRID

FIGURE 2.1

reworked when the Slims floods its main channel. Flooding usually occurs in mid to late August when discharge increases as a result of increased ablation of the Kaskawulsh Glacier.

A 200 by 300 metre grid with 100 metre grid intersections was surveyed on this relatively active area of the delta surface (Fig. 2.1). Tagged stakes were then placed at each grid line intersection. The choice of a sampling grid of this size was made to enable the daily monitoring of surface conditions and eolian activity during the study period. Consideration was also given to the amount of time needed to analyse daily samples. Soil moisture, surface salt concentration and grain size distribution were continually monitored throughout the study period at each grid intersection.

In the preliminary stages of investigation, several methods were used in an attempt to measure surface soil moisture. Soil moisture only affects the susceptibility of the soil to wind erosion if the water is held by the soil in close proximity to the earth air interface. Water held below this boundary has little effect on the stability of the surface and only acts as a reservoir for water which is moving towards the surface by capillary action. It is therefore essential that moisture determinations be made only on the soil which is at or in close proximity to the earth air interface.

Preliminary investigations showed that standard ceramic soil moisture blocks and Soil Test soil moisture modules were of little use in the measurement of surface moisture. Sensors of this nature could not be placed close enough to the surface to give a representative surface moisture content. Moisture blocks installed close to the surface were quickly exposed by wind erosion. Also, this type of moisture sensor gives an average moisture content for the material surrounding the block. Thus, because of the block's size, the surface moisture content was heavily biased by the moisture content of the sediment below the true surface. Problems associated with the calibration of this type of sensor and serious disturbance of the soil during installation were also prohibitive factors in their possible use.

A decision was therefore made to take daily surface samples from the grid intersections and determine moisture contents gravimetrically. A small (5 - 10 g) sample was collected each day from the surface by means of a sharpened spatula within a 0.5 metre radius of the grid intersection marker. By using this method, a "thin slice" of surface material could be collected, more closely approximating the true surface moisture. A similar technique has been used to measure soil moisture in wind tunnel experiments by Belly (1964). Reproducible results appear to be possible using this technique. Several samples were collected from selected grid intersection points on various days during early May 1973. These results are shown in Table 2.1.

After moisture contents were determined for each of the daily samples, they were analysed for total salt concentration by electrical conductivity. Weekly samples were also kept for grain size analysis.

TABLE 2.1

COMPARISON OF SURFACE SOIL MOISTURE DETERMINATIONS
FOR SELECTED GRID POINTS

	Surface Moisture Content (per cent dry weight)											
	Grid Point 11				Grid Point 7				Grid Point 1			
May 7	7.84	7.31	7.49	8.02	7.54	7.06	8.11	7.28	6.94	7.20	6.81	7.52
May 8	8.66	7.95	8.81	8.58	8.91	8.59	8.65	8.94	7.57	8.18	8.01	7.44
May 10	4.16	4.00	4.15	4.23	4.02	3.97	3.81	3.84	3.68	3.57	3.26	3.43
May 12	4.91	5.24	4.95	5.07	4.69	4.74	4.68	4.79	3.94	4.02	3.67	3.83

Measurement of Saltation and Creep

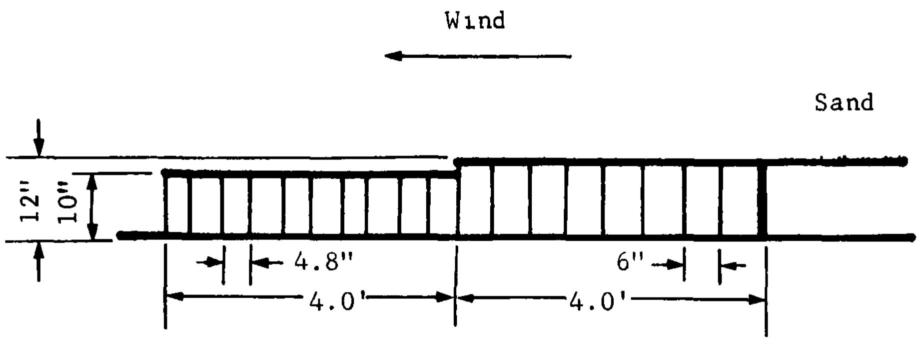
Previous studies (eg. Bagnold, 1941) have indicated that the initial movement of silt grains into suspension is closely related to the movement of sand grains in saltation and creep across the surface. To investigate this hypothesis, saltation and creep traps were installed within the study grid (Fig. 2.1).

Belly (1964) has discussed the efficiency of different trap designs used by various authors. In general, traps fall into two design types: (1) vertical and (2) horizontal.

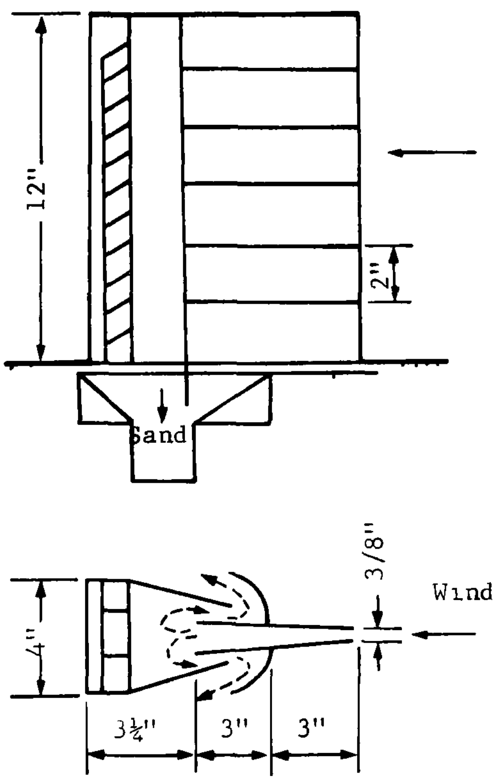
Horizontal traps usually take the form of large pans divided into a number of segments which are open at the top. The pan is buried in the sediment so that the upper lip is level with the sand surface (Fig. 2.2). Sand moving by creep is caught in the first pan segment, while sediment transported in saltation is trapped in segments further downwind. The size of the horizontal trap is dependent on the nature of the problem and the environment in which the trap is to be used. The horizontal trap used by Belly (1964) in his wind tunnel experiments was 0.76m wide and 2.43m in length. Traps of this type are relatively wide (usually 0.3 to 1.0m), normal to the wind direction, and have been used primarily in wind tunnel investigations (O'Brien and Rindlaub, 1936; Kawamura, 1951; Belly, 1964).

Vertical traps which have been used (eg. Bagnold, 1941; Chepil, 1945; Williams, 1964) also vary in size but usually have a relatively narrow intake orifice. Traps of this nature are usually

SCHEMATIC DIAGRAMS OF HORIZONTAL AND VERTICAL TRAPS.



(a) HORIZONTAL SAND TRAP

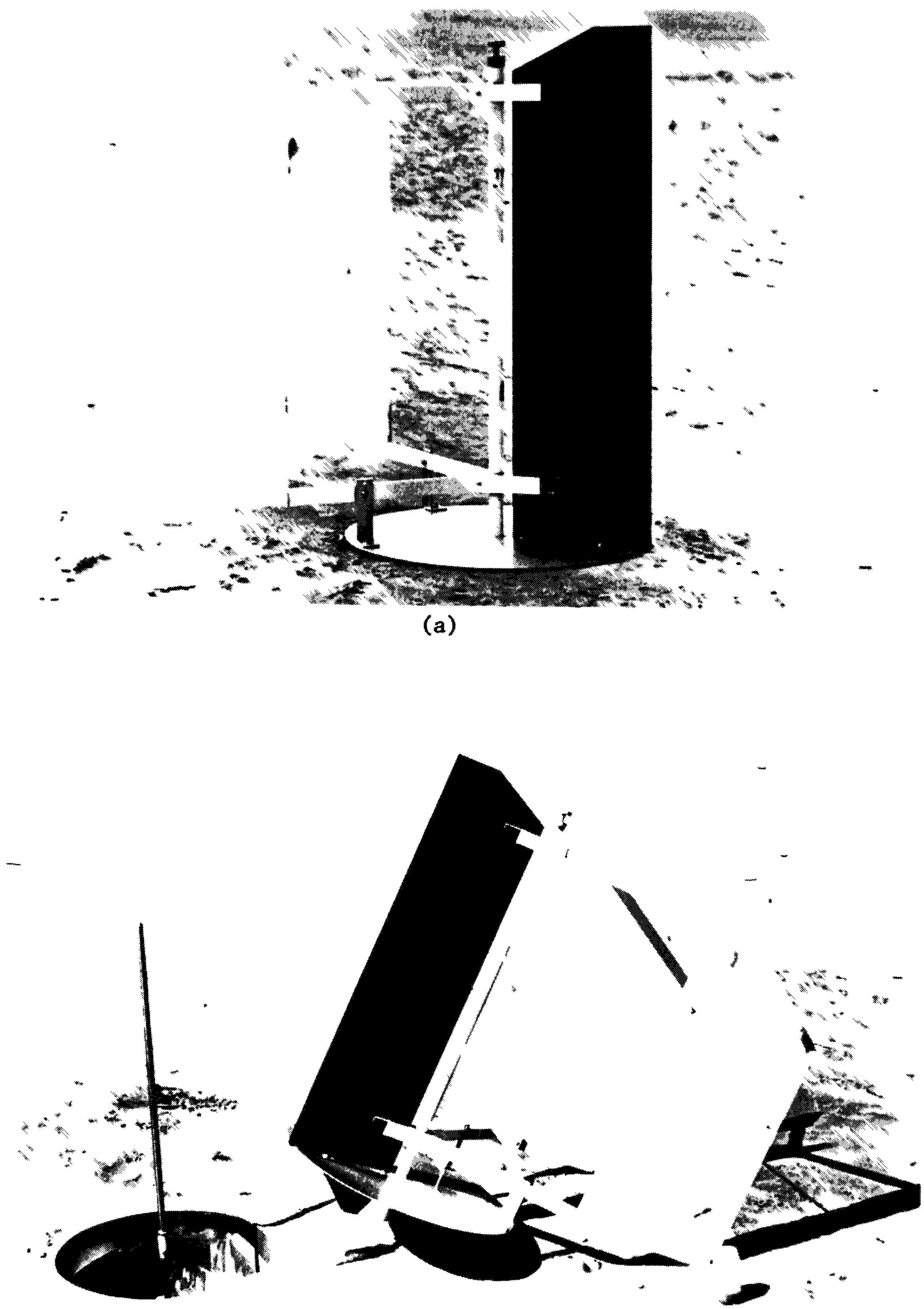


(b) VERTICAL TRAP

FIGURE 2.2 (After Belly, 1964)

tall and thin and are placed in an upright position facing the wind direction (Fig. 2.2).

Belly (1964) has compared the efficiency of the two trap designs in wind tunnel experiments using sand with a mean diameter of 0.5 mm and wind velocities between 7.86 to 10.82 m/s. He argues that the horizontal trap design collects almost all material moving in saltation and creep except under higher wind velocities, when some sand transported in saltation is carried beyond the last compartment. He also demonstrated that the vertical trap design tends to underestimate total sediment transport in saltation and creep, especially at low wind velocities. He suggests that the underestimation is partially the result of scour taking place at the base of the vertical trap. In his experiments, scour usually began immediately at the beginning of a run and eventually undermined the bottom lip of the trap. After this occurred, no material moving in creep could enter the sampling orifice (Fig. 2.2). Underestimation, especially at lower wind velocities, is also caused by the deflection of smaller particles around the sampling orifice. The trap itself impedes the air flow and as a result, air is forced to flow around the trap. A percentage of the smaller particles is carried away from the sampling orifice by the deflected air stream. Larger particles, possessing greater momentum, are not deflected as easily and continue into the sampling orifice (Chepil, 1945). Belly (1964) concluded however, that if wind velocities were relatively high (9.1 to 10.7 m/s) and sampling time was short (< 15 minutes), the vertical trap was sufficiently accurate to be used in such investigations.

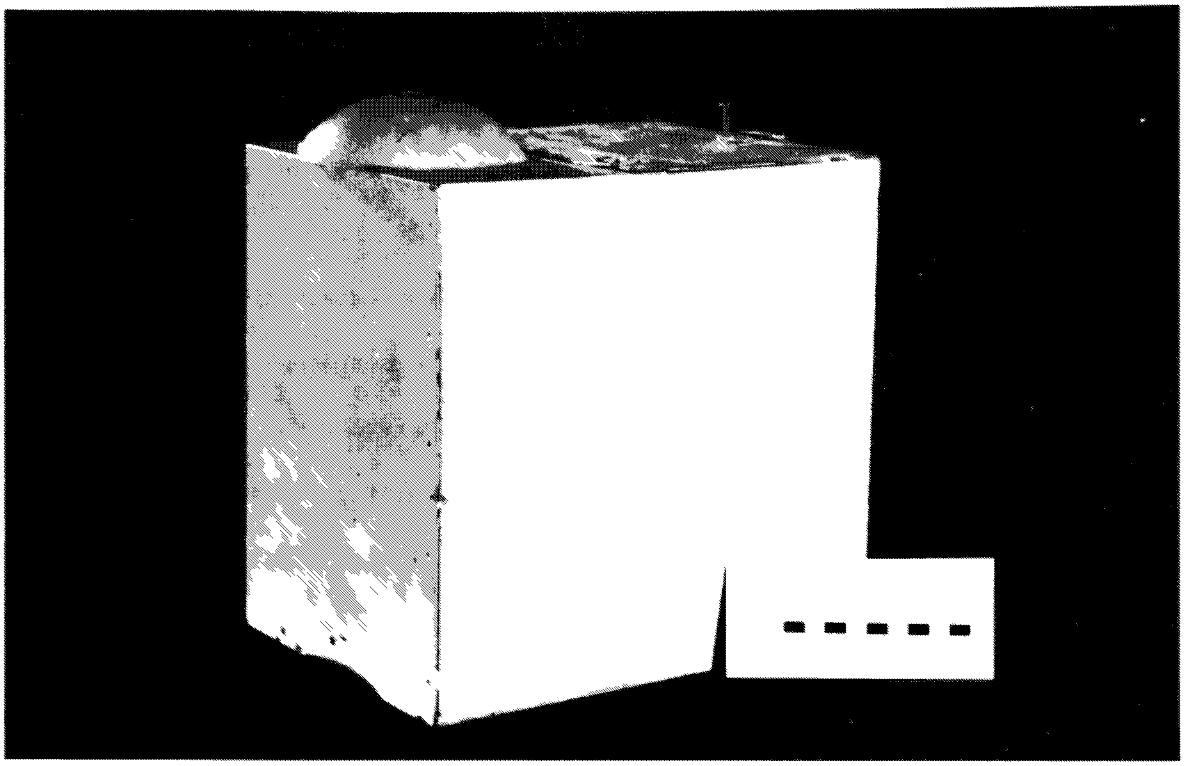


(a)

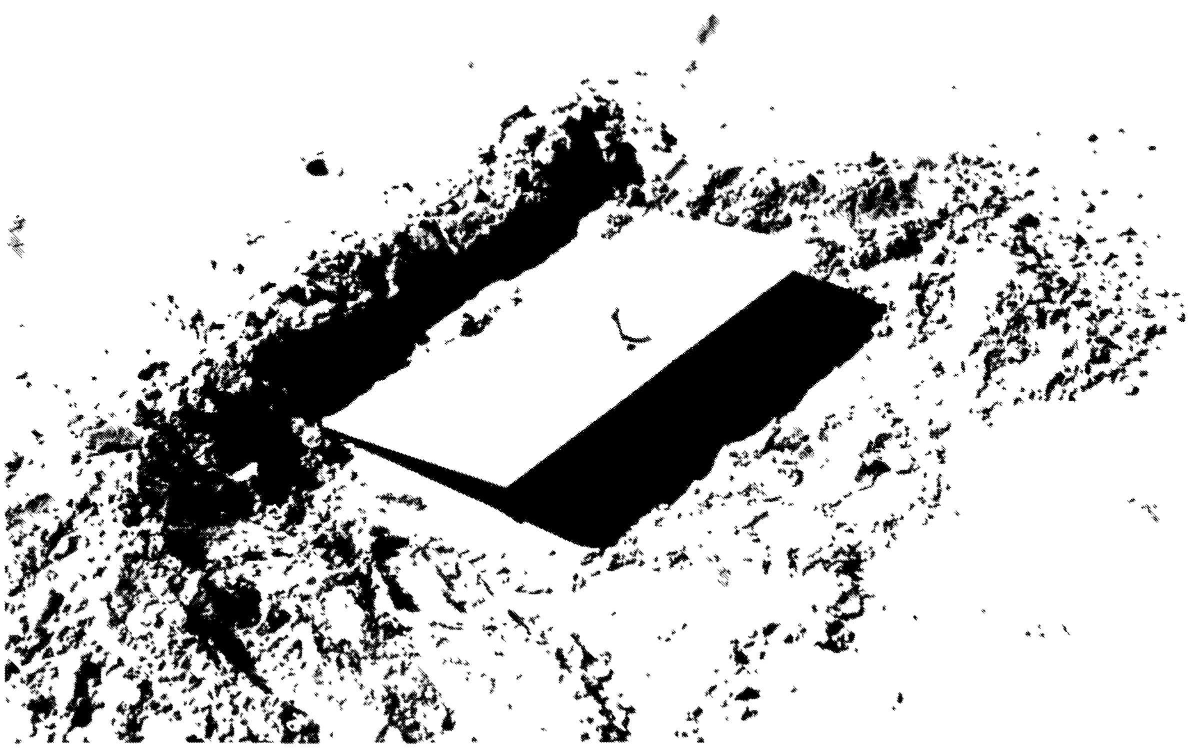
(b)

THE SALTATION TRAP

FIGURE 2.3



(a)



(b)

THE CREEP TRAP

FIGURE 2.4

Both trap designs have major drawbacks if used in field conditions. In wind tunnel studies, wind direction is constant, whereas in field conditions, wind direction can vary during the sampling period.

In order to measure saltation and creep in the Slims River delta, traps were designed to overcome the problems associated with the variations in wind direction and the loss of creep material resulting from scour at the trap base. Traps used in this investigation are similar to those used by Bagnold (1941) and Chepil (1945), but have the advantage of being automatically oriented into the wind (Figs. 2.3 and 2.4).

Problems associated with scour at the base of the vertical traps are almost unavoidable. It was therefore decided to use two trap types - one for saltation and another for creep. This is similar to the sampling method used by Bagnold (1941). The saltation trap was 1 metre in height and had a sampling orifice 1.0 cm in width (Fig. 2.3). The trap was mounted on a shaft with teflon bearings to which two large fins were also connected. The fins continually kept the sampling orifice normal to the wind direction, even with low wind velocities. Sediment entering the sampling orifice was channelled to the back of the trap and fell into a collection bottle at the base, housed in an aluminum pan and buried below the surface (Fig. 2.3). The bottom of the sampling orifice was kept as close as possible to 0.05 cm of the active surface. This was accomplished by means of an adjustment screw on the pivot shaft which allowed the trap to be raised or lowered. This effectively collected all saltation material while excluding sand moving in creep.

A trap similar to that used by Bagnold (1941) was used to collect sand moving in creep. Bagnold's trap consisted of a bottle with a rectangular opening 1.3 x 3.8 cm affixed to the top. The bottle was buried below the surface so that the sampling orifice was level to the sand surface and the long axis of the opening normal to the wind direction. After the sampling period, the bottle was removed and the sediment weighed.

The amount of sediment caught in a trap with a rectangular sampling orifice is partially a function of wind direction. The sample opening is effectively decreased if the wind direction deviates from normal to the long axis of the opening. To overcome this problem, the original trap used in this study had a circular intake orifice with a 1.0 cm diameter. Other difficulties however, were encountered with this trap design.

Shortly after installation, scour began to develop around the sampling orifice. Once scour had occurred, material moving in creep was no longer able to enter the trap. To compensate for this, a dome with a 1.0 cm hole through the centre was soldered to the top of the sampling orifice (Fig. 2.4). The trap was then buried so that the apex of the dome was level with the surface. Although some scour did occur around the sampling orifice, sand moving in creep was able to move up the dome and into the trap.

The creep and saltation traps used by Bagnold were removed after each sampling run. This was not possible in this investigation however, because of the nature of the sediments making up the Slims River delta. These sediments differ from the desert sands of Bagnold's experiments in that they are predominantly fine sand and silts and are only dry near the surface. As a result, these sediments are not completely cohesionless and possess a weak sub-angular blocky structure. If the sediments are disturbed during the installation or removal of the traps, the surface cannot be restored to a condition similar to the undisturbed state. It was therefore essential to disturb the surface as little as possible during the collection of samples.

This condition was met in the designs of both the saltation and creep traps. The sampling orifice and dome of the creep trap were permanently mounted to a rectangular plywood box which housed a removable sample bottle, also affixed to the intake orifice. A small lid was attached to the surface of the box at the end opposite the intake orifice. The box was buried at the beginning of the field season so that the trap door was positioned downwind from the dominant wind direction. After installation, the surface was smoothed as well as possible and the disturbed area sprayed with water until the surface became saturated.

After a sample run, only the sediment covering the trap door was removed. It was then possible to reach into the upwind end of the container, remove the sample bottle, and replace it with a new one. The lid was then closed and the sediment replaced on top of the door and sprayed with water. In this manner, the material surrounding the intake orifice was disturbed as little as possible.

It was necessary, however, after extremely windy days, to repair scour around the intake orifices. This was done by spreading surface material around the orifice and gently moistening the surface.

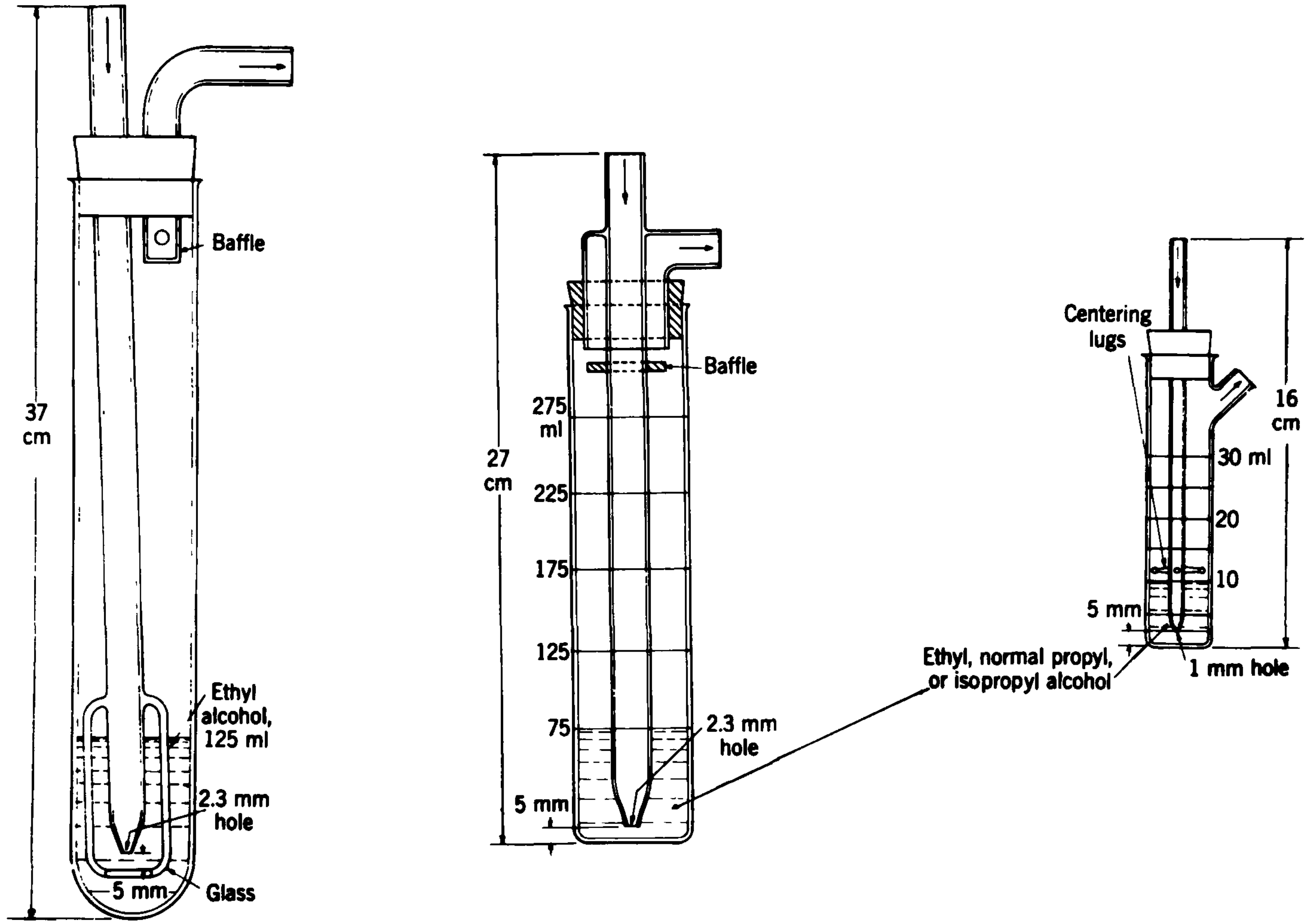
Material from near the trap and as little water as possible were used to avoid serious alteration of the surface salt concentration. During repairs, a cork was placed in the orifice and left until the surface dried to approximately the same moisture content of the surrounding area. This was dependent on the temperature and wind velocity, but usually required less than 45 minutes.

Removal of the sample bottle from the saltation trap did not affect the surface. The trap could be removed from its installation point by simply sliding the pivot shaft off the spindle permanently attached to the base pan. It was necessary, however, to occasionally repair scour around the lip of the buried pan. This was accomplished by the same method used to repair scour around the intake orifice of the creep trap.

Although scour had to be repaired once or twice a week depending on wind and temperature conditions, the problem was not as severe as indicated by other investigators working with sand. This is directly related to the greater cohesion of the Slims River delta sediments as a result of the high silt content.

Measurement of Suspended Sediment

The measurement of the amount of sediment carried in suspension by the wind is more difficult than the measurement of either creep or saltation load. Suspended sediment is seriously affected by instantaneous changes in the velocity and direction of the wind (Chepil, 1945). As well, the sampling rate and intake orifice design can produce serious sampling errors.



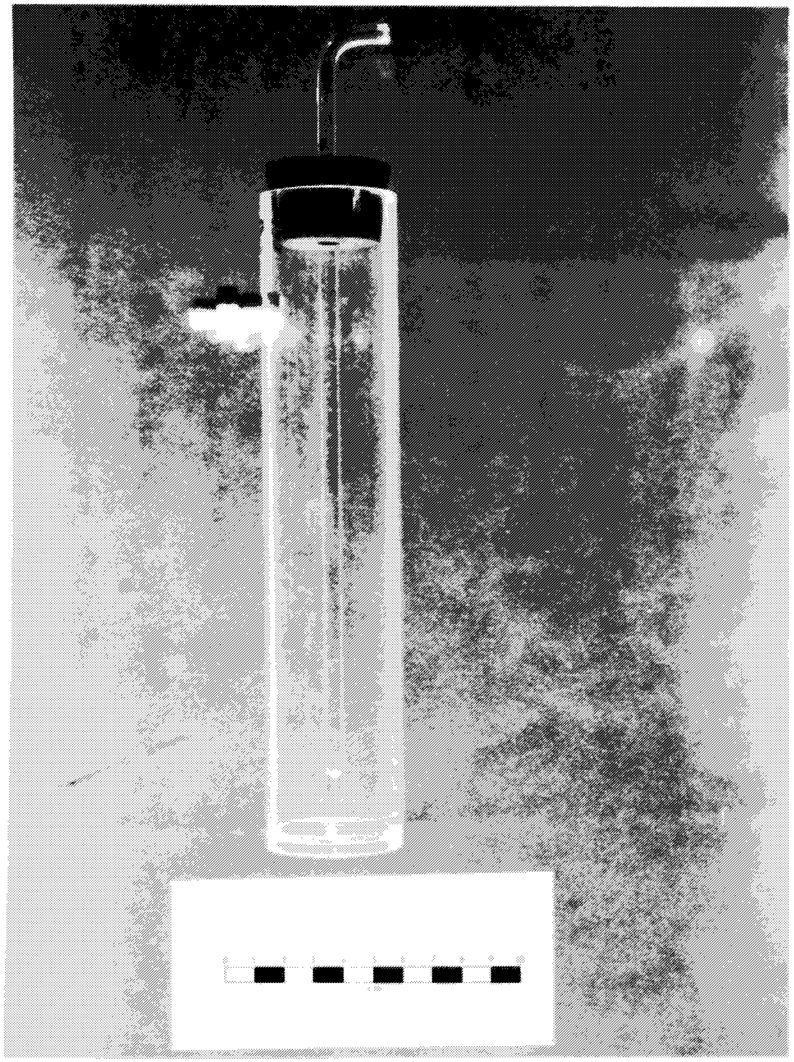
a b c

Figure 2.5 Various Impingers Used by the Bureau of Mines, (after Brown and Shrenck, 1938)

One of the earliest methods used to sample fine particulate matter from the atmosphere was the impinger tube technique developed by Greenburg and Smith (1922) for the United States Bureau of Mines and United States Public Health Service. In this technique the particle-laden air is drawn into the impinger tube and is forced through a small nozzle which is immersed in alcohol or water (Fig. 2.5). After passing through the nozzle, the particles carried in the air stream are impinged either on the bottom of the outside container holding the liquid or on a small plate affixed to the intake tube (Fig. 2.5). In both cases the nozzle is placed 0.5 cm from the impingement surface. Particles are impinged on this surface in order to stop the movement of the particle long enough for them to become wetted by the liquid. As a result of wetting, the particles remain in suspension in the liquid and are therefore not drawn out of the impinger tube through the evacuation line. After sampling, the suspension is removed from the impinger tube and the alcohol evaporated off. The amount of sediment collected can be calculated by weighing the residue or counting individual particles under a microscope.

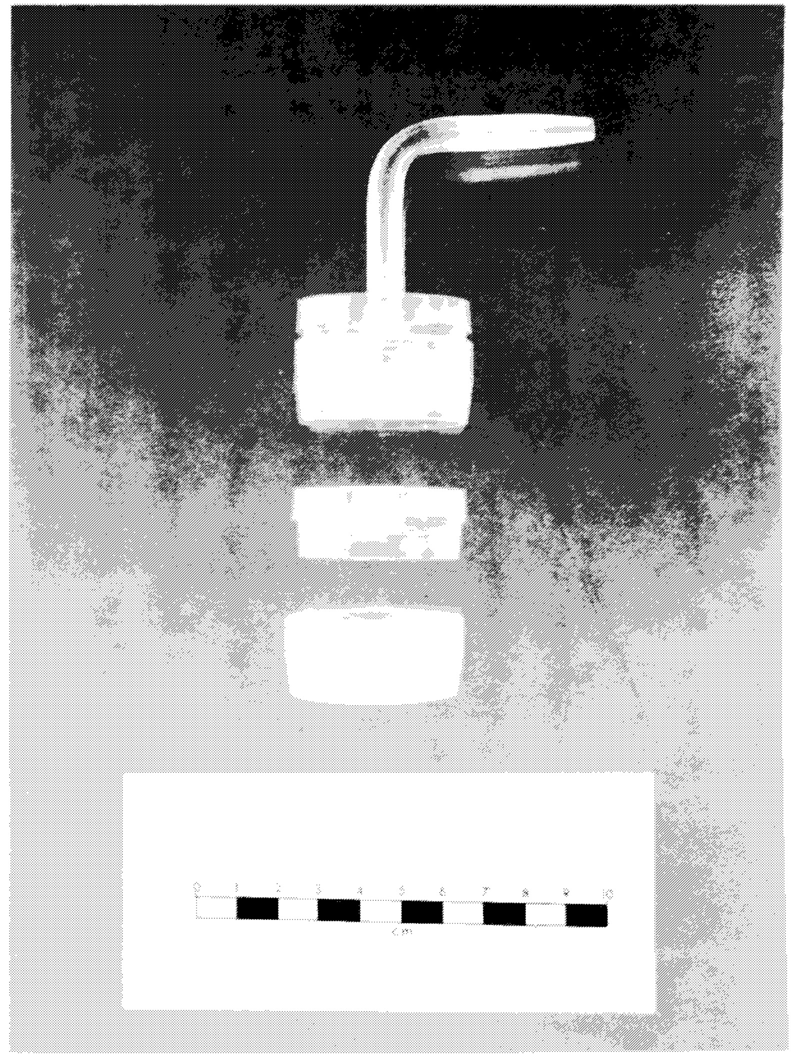
A modified form of the impinger tube was used during the 1972 field season (Fig. 2.6). It was found, however, that the impinger tube was impractical for the measurement of suspended sediment transport during dust storms in the field because of inherent problems associated with the technique.

The transfer and subsequent drying off of the suspension proved to be somewhat laborious in that it was often difficult to thoroughly rinse out all particulate matter from the impinger tube. This problem



MODIFIED IMPINGER TUBE SAMPLER

FIGURE 2.6



FILTER CASSETTE SAMPLERS

FIGURE 2.7

appeared to be critical, especially when the suspended sediment concentrations were very low and the amount of suspended sediment collected small.

A second more serious type of error was thought to have occurred during the sampling itself. The impinger tube had been designed primarily for the measurement of dust concentration in mines and industrial plants. In these types of environments, the movement of air around the impinger tube and especially the sampling nozzle is usually relatively small. However, when suspended sediment concentrations are measured during dust storms a great deal of turbulence is created because of the passage of air around the impinger tube. This problem is in part due to the relatively large size of the impinger tubes which obstruct the air flow.

In order to overcome these problems suspended silt load in this investigation was measured by drawing air through fine cellulose type filters held in small plastic cassettes (Fig. 2.7). A similar method of sampling industrial dust has been adapted by the Canadian and the United States Bureau of Mines (Knight and Lichti, 1970).

Gelman filters (Metricel VM-1) 3.7 cm in diameter with a pore size of 5.0 microns were used. The filter holders were standard 37 mm Millipore Sampling Cassettes modified to include a streamlined sampling nozzle. The sampling nozzle was made of 0.79 cm acrylic tubing with an inside diameter of 0.47 cm. The nozzle was bent into an "L" shape, each limb being 5 cm in length. The bottom limb was inserted into a 0.79 cm hole which had been bored in the upper lid of the sampling cassette. The upper limb was turned on a lathe into a tapered streamline form (Fig. 2.7).

Although the suspended sediment concentrations measured during dust storms in the Slims River Valley can not be checked against an absolute value, the filtering cassette sampling method appears to give more reliable results than the impinger tube technique. This was checked by simultaneously sampling at the same height (2 m) with three impinger tubes and three filtering cassettes spaced horizontally at 0.5 m intervals. The results of this comparison (Table 2.2) showed that the filtering cassettes gave more reproducible results and that the impinger tubes in comparison irregularly underestimate the quantity of sediment collected in the filtering cassettes. It is suspected that this underestimation results from excessive air turbulence around the intake nozzle because of the impinger tube's relatively large size.

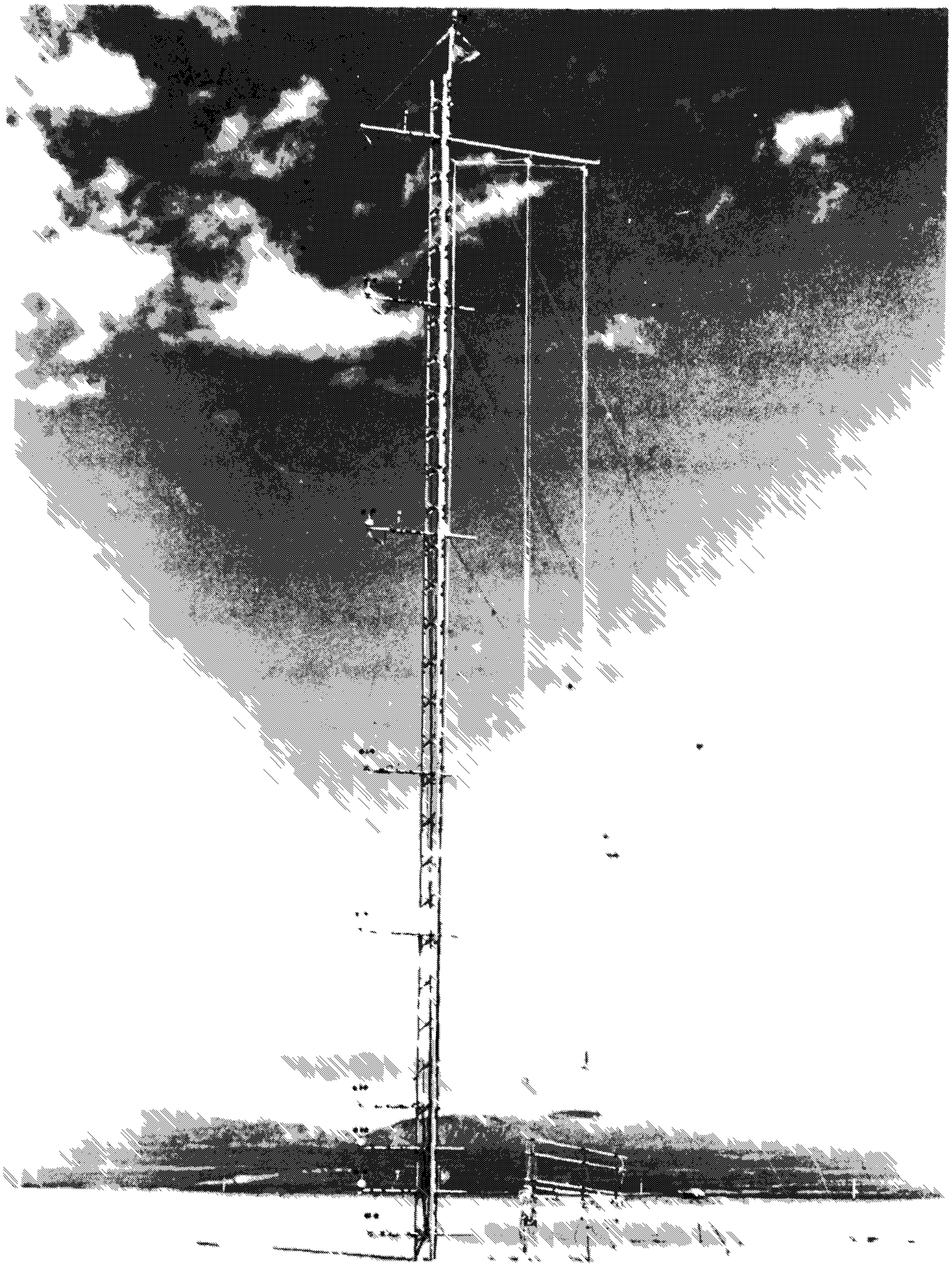
During the sampling periods, eight to ten cassettes were placed on a tower at heights of 0.5, 1, 1.5, 2, 4, 6, 9 and 12 metres (Fig. 2.8). Each cassette was connected by 0.6 cm Tygon tubing to a vacuum pump driven by a 3 horsepower gasoline engine located 15 metres from the tower base.

In order to facilitate rapid installation of the cassettes at the various heights before each sampling run, a mechanical method of positioning the samples was devised. Eight sliding booms were attached to two parallel steel cables (0.6 cm diameter). The cables were stretched taut between a permanent boom at the top of the tower and the ground surface. The sliding booms were joined together by 0.16 cm steel cable so that when extended upward, the booms were positioned at the predetermined intervals. The complete assembly of sliding booms was

TABLE 2.2

COMPARISON OF THE SAMPLING EFFICIENCY
OF THE MILLIPORE FILTER CASSETTES AND THE
MODIFIED IMPINGER TUBES

	Concentration (mg/m^3)			
Millipore Cassettes				
1	39	79	14	27
2	44	72	16	29
3	47	84	11	24
Impinger Tubes				
1	21	58	8	16
2	36	66	15	22
3	27	39	6	9
Wind Velocity at 2.0 metres (cm/s)	690	644	551	586



THE SAMPLING TOWER

FIGURE 2.8

lifted into position by a winch located at the tower base. The sampling cassettes were affixed to the sliding boom by small pinching type clamps and when in position were located 2.25 metres out from the tower. The evacuation tubing for each cassette was extended across the boom and down the inside guy line. This tubing, which then extended 15 metres from the tower base, was buried 0.5 metres below the surface.

By using this method, sampling cassettes could be positioned in less than 15 minutes. The sampling apparatus also produced less disturbance to air flow than if the samplers had been mounted directly on the tower.

In order to use this method to calculate the amount of sediment transported by the wind in suspension, it was essential to know the volume of air in which the sediment was suspended. To do this a constant flow rate of 0.1 l/s was maintained across the filters during the sampling period. This was accomplished by the incorporation of a needle valve and flowmeter (Dwyler, VM-1) in each line. This particular flow rate gave an intake nozzle velocity of 5.7 m/s.

The choice of this particular flow rate and corresponding nozzle velocity was not completely arbitrary because the intake velocity can directly affect the accuracy of measurement. If the intake velocity of the sampling nozzle is below the wind velocity (subisokinetic), the amount of the sediment collected will be greater than the amount that was originally in the air that entered the probe. Some of the air that approaches the intake orifice is diverted around the nozzle because of the cushioning effect of the slower moving air in the nozzle. As a result of their inertia, particles moving the air stream are not as easily deflected and consequently are carried into the sampling orifice (Sehmel, 1970). Thus, more particles are collected than were in the air that entered the probe. The overestimation increases with the weight of the particles and the difference between nozzle and air velocity.

Similarly, if the nozzle velocity exceeds the wind velocity, the number of particles collected will be less than the number that were originally in the air that entered the nozzle. For this flow condition, some of the particles in the air stream will not be drawn into the probe because of particle inertia. Rather, these particles will continue to move in the direction of air flow past the intake orifice. The errors related to nonisokinetic nozzle velocities can in many cases be overshadowed by larger errors resulting from turbulence generated by the sampler itself (Sehmel, 1970).

The selection of a 5.7 m/s intake velocity was made in an attempt to approximate the range of wind velocities encountered during the dust storms. Mean wind velocities during the storms ranged from 3.96 to 12.99 m/s, the modal velocity being approximately 6.0 m/s. Although a higher intake velocity may have been somewhat more desirable, it was impossible because velocities greater than this caused excessive wear on the vacuum pump used during the experimental runs.

Wind speed and temperature were measured at each height during the sampling periods. This was accomplished by means of sensitive anemometers (Casella make) and self-aspirating thermocouples positioned on the tower at each of the eight sampling heights. Two standard totalizing anemometers (Belfort, 12.7cm cup) were placed on a separate mast 10 metres from the tower near the saltation and creep traps.

A standard meteorological station was also maintained throughout the study. This station was positioned on the northwest side of the valley adjacent to the study grid. It was found, however, that the silt laden air quickly clogged the gears and pen nibs of the recording instruments. It was therefore necessary to move the station towards the valley wall where dust conditions were less severe. The location of all meteorological and sampling instruments is shown in figure 2.1.

Sampling Procedures

Two sampling schemes were used in the investigation of eolian transport in the Slims River Valley. In the first scheme, the amount of material moving in saltation and creep was continually measured each day throughout the field season. Sediment was collected from the saltation and creep traps each day during the early afternoon. This sediment was then weighed and selected samples were kept for grain size analysis. Mean daily wind speed was also recorded from the totalizing anemometers. Surface samples used to measure surface soil moisture and salt concentration were also collected each day during the early afternoon. Because of the variability of eolian transport in the Slims River Valley, it was necessary to adopt a second sampling scheme to measure total sediment transport in creep, saltation and true suspension.

Although the eolian transport of sediments occurs regularly on the Slims River delta, daily as well as hourly changes in surface and climatic conditions cause great spatial and temporal variability in the intensity of eolian transport. During the night and morning, relatively low evaporation rates allow the surface to become moistened by water rising from below the surface by capillary action. Even with relatively strong winds, little eolian transport occurs during this period. As evaporation increases because of higher daytime temperatures and increased wind speeds, the surface begins to dry. The drier surface and relatively strong daytime winds initiate the movement of surface material in saltation and creep. The rate of drying of the surface is not uniform throughout the delta area. Variations in the grain size distribution, proximity to the river, height above the river and daily variations in the velocity and turbulence of the wind throughout the valley all contribute to the spatial variability of drying of the delta surface. As a result, the eolian transport of sediments does not begin, nor is it maintained uniformly, in the delta area. This is especially true of material carried in suspension.

Large plumes of dust are indicative of relatively large amounts of sediment being carried in true suspension. The size and frequency of dust plumes in the delta area is extremely variable. Occurrence can range from small isolated plumes to a complete obscuring of visibility throughout the lower Slims River Valley.

The variability in frequency, distribution and intensity of sediment transport in true suspension dictated the sampling of suspended material only during relatively large "dust storms". This was necessary because of the permanent location of the sampling tower and the amount of sediment needed for analysis. That is, in order to sample suspended material during a dust storm, it was necessary that suspended material be transported through the study grid. The intensity of the transport was important in two other respects. One of the principal objectives of the investigation was to establish the distribution of sediment during dust storms to a height of 12 metres. It was therefore necessary that material be carried by the wind to at least this height during sampling.

Preliminary investigation had also shown that the amount of sediment (mg/m^3) carried in suspension, even during relatively intense dust storms, was extremely small, especially at higher elevations. This had also been found by Chepil and Woodruff (1957) in their investigation of the effects of suspended dust concentrations on visibility. It was therefore necessary to obtain a sample of suitable size to offset any errors which might result from the transfer and weighing of sediment samples. It was found that a one hour sampling period with a 0.1 l/s flow rate was necessary to obtain a sufficient sample.

During the one hour sampling period, the amount of sediment transported in saltation and creep was measured. At the beginning of each one hour sampling run, the daily sample bottles in the saltation and creep traps were exchanged for clean bottles. After the sampling period, the daily collection bottle was replaced. The weight of sediment caught in the saltation and creep traps during the one hour period was added to the daily weight record for that day for the respective trap.

Wind speed and temperature were continually recorded at each of the eight sampling heights. Temperatures were recorded from a microvoltmeter at five minute intervals. Mean temperature for each sampling height was then calculated. The anemometers used in this investigation employed digital counters to measure wind speed. Mean wind speed for a given time period (60 minutes) could then be obtained from the calibration curve for each anemometer.

During each dust storm, surface sediment samples were also collected from the grid intersections. The samples were then analysed in the laboratory for moisture content and total soluble salt concentration.

Thus, during the one hour sampling period of each dust storm the following variables were measured

(1) sediment transport in saltation and creep,

(2) sediment transport in suspension at eight heights (0.5, 1.0, 2.0, 4.0, 6.0, 9.0, 12.0 m),

(3) mean velocity and temperature at each sampling height,

(4) surface salt concentration, and

(5) surface soil moisture.

In total, fifteen dust storms were sampled during the study period. From the fifteen storms, 136 suspended sediment samples were collected and analysed for weight and grain size distribution.

CHAPTER III

LABORATORY PROCEDURES

Introduction

The samples collected during the investigation were subjected to laboratory analysis either in the field (Icefield Ranges Research Project Base Camp Laboratories, Y.T.) or in the Geography Department laboratories at the University of Ottawa and Carleton University. The samples can be divided into two groups: (1) those collected from the surface both daily and during dust storms, and (2) those collected from creep, saltation and suspended sediment traps.

Surface samples were analysed daily for moisture content and salt concentration. Weekly samples from each grid point were also analysed for grain size distribution. Selected samples from the trap were kept for grain size analysis. Suspended material caught during dust storms from each of the sampling heights was weighed to 0.1 mg. Grain size analysis was carried out on all suspended sediment samples.

Surface Moisture Content and Salt Concentration

The moisture content and salt concentration of the surface samples were determined each day by the following procedure. A 5 - 10 gram portion of each of the daily surface samples collected from the grid intersections was weighed and then dried for 24 hours at 110°C.

After drying, the samples were reweighed and moisture content, expressed as per cent dry weight, was determined. The mean moisture content of the eleven samples[1] was then taken as a representative value of the average surface moisture conditions during the transport period. Moisture contents were determined only once a day unless a dust storm occurred 45 minutes before or after sample collection. This procedure was followed for two reasons. Eolian transport in the Slims River Valley usually occurs from mid to late afternoon. It was thought, therefore, that samples collected during the peak transport period would be sufficient as an indicator of average surface moisture conditions. The second and more fundamental of the reasons is directly related to the time necessary for laboratory analysis. It was found that with the time, equipment and assistants available in the field, only 10 to 15 samples could be analysed each day. It was therefore necessary to reduce the number of samples and frequency of collection so that laboratory analysis could keep pace with sample collection. Mean daily surface moisture contents are given in Appendix A.

After soil moisture contents were determined, the samples were analysed for total soluble salt concentration. Five grams of each oven dried sample were placed in a 100 ml nalgene bottle with 25 ml of distilled water. After shaking in a mechanical shaker for twelve hours, the samplers were allowed to stand for eight hours.

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[1]Only eleven samples were collected each day because the southeast corner of the grid was flooded by the Slims River during early May, 1972.

Conductivity measurements were then taken on the supernates with a Fisher conductivity bridge. After each day's conductivity measurements, a calibration curve was calculated from conductivity measurements taken on standard NaCl solutions of a known concentration. Total salt concentration for each sample, expressed as milliequivalent of NaCl, was determined from the daily standard curve. The mean daily total soluble salt concentrations for the delta surface are given in Appendix A.

Selected samples were also analysed for various ions present in the sediments. As can be seen from Table 3.1 the soluble surface salts are dominated by calcium with lesser amounts of potassium and sodium respectively.

Grain Size Analysis

The Grain Size Distribution of Unconsolidated Sediments

Most unconsolidated sediments have a grain size distribution which approaches log normality; that is, they tend to plot as a straight line on logarithmic probability paper. Various statistical parameters have been developed to characterize the grain size distribution of sediments and to portray their deviation from log normality.

Folk (1966) and Friedman (1967) have reviewed the development, refinement and efficacy of graphic measures which can be calculated directly from the cumulative frequency curve plots. Of the variety of measures which have been introduced, four have proven to be especially

TABLE 3.1

RELATIVE ION CONCENTRATION
OF THE SURFACE SALTS

Sample	Calcium (ppm)	Potassium (ppm)	Sodium (ppm)
1	135	60	13
2	105	62	17
3	160	110	28
4	120	78	22
5	125	95	32
6	145	92	21

useful in describing sediments in relation to the nature of the transporting medium and the depositional environment. These are mean diameter (representing the average size of the sample), the sorting or dispersion about the mean, the skewness or asymmetry of the distribution and the kurtosis or peakedness of the curve.

A second method of obtaining statistical parameters from the grain size distribution is the method of moments. This is a computational method in which every grain in the sediments affects the measure. Griffiths (1967) argues that the computation of size parameters by the method of moments has fulfilled the objectives and assumptions of the development of statistics from frequency distributions. Folk (1966) also suggests that the method of moments probably gives a truer picture than graphic methods, which rely on a few selected percentiles; but despite its aesthetic satisfaction it does have some serious drawbacks for natural sediments which may not make it much superior to the graphic method.

Erroneously sized sieve openings can produce errors, especially in the sensitive third and fourth moments. Also, it is assumed that particles within a given class interval have a centre of gravity at the mid-point of that class. Folk (1966) agrees with Friedman (1962,b) that the method of moments, while measuring a slightly different property than the graphic measures, is equally valid for a comparison of a suite of samples because sample to sample variation is usually so large that it outweighs the small statistical differences between the two methods.

Friedman (1961, 1962b, 1967) has been a strong advocate of the method of moments and has successfully used this method in his investigations of the variations in the grain size distributions of beach, dune and river sediments.

In the present study, the first three moments of the grain size distribution have been used to characterize the surface and eolian sediments collected in the Slims River Valley. The formulae used to calculate the moments are identical to those given by Friedman (1967).

$$\text{Mean } (\bar{x}) = 1/100 \; (\Sigma f M\phi)$$

$$\text{Standard Deviation } (\sigma) = (\Sigma f (M\phi - \bar{x}_\phi)^2/100)^{1/2}$$

$$\text{Skewness } (\sigma_3) = (1/100)\sigma^{-3}\Sigma f (M\phi - \bar{x}_\phi)^3$$

Where $M\phi$ = class mid point in phi (ϕ) units

f = frequency

$\bar{x}_\phi$ = mean size in phi (ϕ) units.

Various authors, notable Koldijk (1968), Friedman (1967) and Folk and Ward (1957) have demonstrated the environmental significance of these statistical parameters and have established a correlation between the size frequency distribution of a sediment and its depositional environment. The ability of these parameters to do this lies in the fact that the frequency distribution is dependent upon sediment supply, the mode of transport and the competence of the transporting medium (Friedman, 1961). This dependence on the mode and competence of the transporting medium can be demonstrated in the case of the mean size of the sediment as influenced by sediment supply. For example, Mason and Folk (1958) have shown a distinct decrease in mean size of sediment as one moves back from a beach environment to the eolian environment behind it (i.e. from a high energy environment to a low energy environment).

Sorting, represented by graphic standard deviation and the second moment is a measure of the spread about the mean. As with the mean size, sorting is dependent on the competence and energy level of the transporting medium as well as the sediment supply. Quantitative scales for the sorting values have been formulated to describe the sorting of sediments (Folk and Ward, 1957; Friedman, 1962b). Folk (1966) points out that these scales must not be considered absolutes because sorting is a rather closely controlled sinusoidal function of mean size. That is, in natural sediments, sorting tends to improve with a decrease in size of material, and then becomes poorer as size decreases further. The generalization that sorting increases with transport in many cases is simply due to the fact that the mean size of the sediment changes with transport, and the improvement in sorting is dependent only on the decreasing mean size, not the distance of transport (Folk and Ward, 1957).

Size frequency measures which describe the shape of a distribution or variation in its tails seem to be the most environmentally sensitive. The third moment, skewness, is one such measure. Skewness is a measure of the asymmetry of the distribution where a symmetrical distribution has a value of zero. Although the mathematical limits of graphic skewness are +1.00 and -1.00 few natural sediments have skewness values outside ±0.80 (Folk, 1966). Distributions having a tail of fine material have positive skewness values, whereas those having a coarse tail have negative values.

The investigations of Koldijk (1968), Friedman (1967) and Folk and Ward (1957) have shown that the statistical parameters derived from the grain size distribution can be used in conjunction with bivariate plots of these parameters to distinguish between sedimentary environments without prior knowledge of the mode of deposition. The ability of the bivariate plots to distinguish between sedimentary environments results from difference in the energy regimes of the transporting media. That is, the grain size distribution of a given sediment which can be characterized by the first four moments is a function of the competence and energy regime of the mode of transport.

In these investigations the purpose was to identify the environment in which the sediment was deposited without regard to the source of the material. Nickling (1972) has also demonstrated that the same techniques can be used to identify source areas of sediments in a unimodal transport system. In this investigation, statistics derived from the grain size distribution were used to identify possible source areas of known earlier deposits. It was shown that variations in the grain size distributions of different loess deposits could be related to spatial and temporal variations in the wind regime and grain size distributions of the source materials. It was also suggested that the grain size distribution of the loessial deposits will be similar to the grain size distribution of the sediment immediately after it moves into suspension from the source surface. It is argued that the grain size distribution will not change during transport but will only be modified as a result of transport distance and changes in the wind regime.

In order to investigate the relationships between grain size distribution and wind regime, grain size analysis was carried out on sediment moving in creep, saltation and suspension in the Slims River Valley. The size distribution of sediments actively being transported by the wind in the Slims Valley are compared to eolian deposits known to have originated in this valley.

Grain Size Analysis of Creep, Saltation and Surface Samples

Grain size analysis was performed on the surface samples collected weekly from the grid and on all saltation and creep samples which were of sufficient size.

Krumbein and Pettijohn (1938) have outlined several techniques for deriving the distribution of particles over a given number of size classes. Sieving of samples has become the most successful and widely used technique in the sand size range.

A variety of techniques has been employed in the analysis of silt and clays, the most common and accurate of which are the pipette methods. These methods are based on the sedimentation of silt and clay particles in water, which is assumed to follow Stoke's Law (Folk, 1969).

A relatively new pipette technique, the falling drop, developed by Moum (1965), facilitates the rapid simultaneous analysis of the silt and clay fractions of numerous samples. This technique has demonstrated the same good reproducibility of standard pipette methods and has proven to be much less time consuming (Nickling, 1972). For this reason, the falling drop technique was used in this investigation.

Nickling (1973) has shown that standard grain size techniques are not completely satisfactory for the analysis of sediments collected from the surface of the Slims River delta. Unusually high salt concentrations, a strong unimodal character and the fact that the modal diameter usually occurs near or at the crossover between sieving and pipetting, necessitated the adoption of a technique suitable to these samples. A method was derived which is similar to that presented by Folk (1969) for samples with a predominance of fines.

All samples were air dried for at least 24 hours. If the samples were of a sufficient size, they were split into units of approximately 10.00 grams, placed in a tared vessel and weighed to 0.001 grams. It was necessary to keep the samples at less than 10.00 grams because of the tendency for larger samples to clog those sieves near the modal diameter of the sediment. In many cases, however, the collected sample was considerably less than 10.0 grams and in these situations the total collected sample was used.

All samples were weighed, then washed and centrifuged three times to remove soluble salts. On completion of salt removal, 10 ml of 40 per cent calgon solution was added to each sample. Following this, the sample was dispersed with a sonic dismembrator. The suspension was then wet-sieved through a series of three five centimetre sieves (3.75ϕ, 0.07 mm; 4.25ϕ, 0.053 mm; 4.75ϕ, 0.037 mm).

The suspension passing through the 4.75ϕ sieve was collected in either a 250 or 500 ml volumetric flask, depending on the original sample size. The sediment trapped on each sieve was dried and then weighed to 0.001 g.

The dried material collected on the 3.75ϕ (0.074 mm) sieve was then dry-sieved at either 0.25 or 0.50ϕ class intervals, the finest sieve always being 3.75ϕ. The material caught on each sieve was weighed and the material passing through the 3.75ϕ (0.074 mm) sieve added to the volumetric flask. The suspended material in the volumetric flask was then analysed for grain size distribution by the falling drop technique.

As in the standard pipette method, samples were drawn from a sedimentation tube at predetermined depths and times. However, in this technique, only a drop of suspension is removed from the sedimentation tube by means of a micro-pipette. The suspension held in the micropipette is ejected into a column of organic fluid, anisole, which has a density slightly less than that of water. After the drop reaches its terminal velocity in the anisole column, it falls with a constant velocity.

The time the droplet takes to fall a certain distance is a measure of the density of the drop and the particle concentration in the sedimentation tube from where the drop was taken (Moum, 1965, p. 343). The density of the drop is obtained from a calibration curve, constructed by timing the fall of droplets of a known concentration through the known distance.

The density of anisole, and hence the speed at which a drop of a given concentration will fall, is highly temperature dependent. Although the cylinder containing the anisole in this study was surrounded by a temperature controlled water bath, it was necessary to calibrate for each sample run. This resulted from the inability of temperature control devices to maintain temperatures within 0.1°C over a long period.

A detailed procedure for the analysis of silt and clay by the falling drop technique is given in Appendix B.

Size classes were determined prior to the analysis and sampling times established to correspond to these predetermined size classes. From the falling drop analysis the weight of material in each of the above size classes was calculated. These weights were grouped with the weights of the material caught on each of the wet and dry sieves. By the three methods outlined, a maximum of eighteen size classes ranging from greater than 1.5ϕ to less than 9.00ϕ were established for each sample. The weight of sediment in each size class for each of the analysed samples is given in Appendices C and D.

Size Analysis of Suspended Sediments

At the conclusion of each one hour sample period during the dust storm, the suspended sediment sampling cassettes were sealed and returned to the laboratory for analysis. The pre-weighed filter paper and collected sediment were transferred from the cassette to a tared weighing crucible. Material adhering to the interior of the cassette by electrostatic charges was rinsed into the weighing crucible with isopropyl alcohol. Samples were oven dried at 80°C for eight hours and then weighed on a sensitive balance to 0.1 mg.

The quantity of suspended material collected during dust storms (0.8 - 233.0 mg) makes the use of standard sedimentation techniques, such as the falling drop, virtually impossible. However, several different techniques have been developed for the measurement of small quantities of material with a relatively mean size.

Friedman (1962b) has carried out grain size analysis by the direct measurement of grains through a microscope using a measuring graticule. Although there are inherent inaccuracies in this technique, some of the problems can be overcome by taking photographs of the sample slide through the microscope. Measurements can then be made from enlarged prints.

Even using this procedure, grain size analysis on a large number of samples is time consuming, laborious and relatively inaccurate. Friedman (1962b) suggests that at least 150 grains from each slide should be measured in order to obtain a representative sample.

Size of the individual particle, expressed as an equivalent volume, can be calculated several ways. A mean volume can be estimated by taking the average of the length and width of the particle. With a suitable correction factor, depending on particle shape, the volume can be estimated. Volume can also be estimated by fitting a best fit circle to the individual grain. In both cases, however, there is an assumption that the particles are almost equidimensional.

In the initial laboratory investigations, grain size analysis was attempted by the above microscope technique and best fit circles. It was found, however, that the technique was unsatisfactory because of problems associated with the resolution of the photographs, the relatively inaccurate method of measurement and the amount of time required to analyse a single sample.

More rapid and accurate techniques using a Coulter Counter have been developed for the grain size analysis of fine grained sediments (Sheldon and Parson, 1967, and Kranck, 1973). In this technique a quantity of sediment is placed in solution and then drawn through capillary orifices of increasing known diameters. Each time a single grain passes through an orifice, it is counted. Grain size distribution can then be stated in terms of equivalent spherical diameter of the particle volume relative to concentration in parts per million (Kranck, 1973). This technique, although rapid and relatively accurate, requires a sample of at least 0.5 - 1.0 g, which is greater than the size of the suspended sediment samples collected during this investigation.

Grain size analysis of small quantities of fine grained sediments has also been carried out using image analysing computers, initially designed for crystalographic investigations. Peach and Perrie (1975) have used a Quantimet 720 to analyse sediment collected from very closely spaced units in finely laminated sediments. This technique has permitted the resolution of variations within different phases of single units of varved clay.

In this technique, a small portion of the sediment is dispersed on a standard microscope slide and placed under the Quantimet 720's optical microscope. The observed field is then viewed with an optical line scanner which converts the image into the signals required by the Quantimet's detectors. The detector used in this case was a sensitive densitometer which can distinguish 62 gray levels. Thus, by setting a lower threshold, all particles can be distinguished from the white (i.e. less gray) background field. Although the image is viewed by a line scanner the entire field is divided into a large number of discrete units of known area termed picture points (500,000 picture points in a single live frame). By setting class limits, the Quantimet 720 can be programmed to measure the total particle area larger or smaller than any of the given limits. In practice, accuracies of better than one per cent can be obtained when the measured area covers more than five per cent of the viewing field (Quantimet 720, Operating Manual, 1971). In using this technique for grain size analysis the size limits used are based on the diameter of a circle with an equivalent area.

Although grain size analysis of extremely small samples can be carried out by the Quantimet 720, two sample mounting techniques are used depending on sample size. For relatively large samples ($\geq$0.01 gm) the grains are rapidly sedimented out of an aqueous dispersion onto the surface of a gelatin coated slide. As a result of the hygroscopic properties of gelatin, the water is quickly absorbed, leaving the individual grains dispersed and cemented to the slide (Perrie and Peach, 1973).

This mounting technique was used on all suspended samples which were of a sufficient size. It was also necessary to analyse the fifteen creep samples collected during the dust storms because they were of insufficient size to be analysed by the falling drop method. Results of this analysis are given in Appendix D.

Extremely small samples collected during dust storms were usually associated with a relatively small mean diameter and were most often collected at the 6.0, 9.0, or 12.0 metre heights. In these cases, the grains usually adhered to or became imbedded in the acetate filter paper in the sampling cassette. As a result, this sediment could be analysed without being transferred.

The filter paper was mounted directly on a suitably sized glass slide and wetted with xylene. Wetting with xylene effectively turns the cellulose filter paper translucent and allows enough light transmission through the filter paper so that the darker sediment grains can be analysed by the Quantimet 720. Using the two above mounting techniques, all suspended samples collected during dust storms were analysed for grain size distribution. Eight size classes were defined: less than 2.96ϕ, 2.96ϕ to 3.96ϕ, 3.96ϕ to 4.96ϕ, 4.96ϕ to 5.96ϕ, 5.96ϕ to 6.96ϕ, 6.96ϕ to 7.96ϕ, 7.96ϕ to 8.96ϕ, and greater than 8.96ϕ. The results of the grain size analysis are given in Appendix E and are expressed in terms of per cent volume in each size class.

Computation of the First Three Moments

The first three moments of the grain size distribution were calculated by means of a computer program outlined in Appendix F. For the saltation, creep and surface samples, the moments were calculated on the basis of percentage of total weight in each size class. Moments for suspended sediments and creep samples collected during the fifteen dust storms are computed from per cent of total volume in each size class. The first three moments of the grain size distribution of all samples are given in Tables 4.1, 5.7, 5.8, and 5.9.

CHAPTER IV

MEAN DAILY SEDIMENT TRANSPORT IN SALTATION AND CREEP

Sample Size

Saltation and creep were measured continuously in the Slims River Valley for a total of 59 days. Sediment was collected in traps and wind speed was measured by pre-calibrated totalizing anemometers (Belfort). Sediment in each trap was weighed daily. The amount of sediment transported both in saltation and creep is expressed as a mean daily flow rate defined as weight per unit width in unit time (mg/cm.s) (Appendix A).

Although sediment transport was measured for 59 days, not all days have been used in the analysis. Twice during the experimental run, high wind velocities damaged the saltation trap, making it inoperable for at least part of the day. Since the sediment caught in the trap did not represent transport for the total 24 hour period and since only an estimate of when the damage occurred could be made, these data points were discarded.

The sediment transport itself also caused damage to the 0.5m level anemometer on the first three consecutive days of the experiment.

Fine sand and silt penetrated the bearing of the anemometer making the recorded wind velocities unreliable. The problem appeared to be at least in part related to the fact that the anemometer bearings had been lubricated with fine machine oil which held the fine sand and silt particles. To overcome this problem, a second anemometer was lubricated with graphite and installed. To ensure the problem did not occur again, the anemometers were cleaned and lubricated regularly with graphite.

On some days, although wind velocities were relatively high, little or no sediment transport occurred. These days were associated with periods of extended rainfall which caused the surface to be continually wet for at least 24 hours. Conversely, on several occasions a relatively large amount of sediment was found in the sediment traps despite the fact that the surface was wet and wind velocities were moderate. The sediment collected in the traps during these periods was usually wet and was attributed to splash caused by the impact of falling raindrops. This problem was most severe in the case of the creep trap. The principal problem was the fact that it was impossible to know in many cases, how much of the collected material could be attributed to rain splash.

This situation points to a rather obvious conclusion. That is, eolian transport of sediment does not occur when the eroding surface is saturated or wet. As previously mentioned, Belly (1964) has shown in wind tunnel experiments with fine sand that the wind velocity necessary to initiate movement increases sharply once the surface moisture content reaches 3 - 4 per cent dry weight (Fig. 4.1). As can be seen from this figure, once the sand's surface moisture content reaches approximately 8 - 10 per cent, the wind velocities

VARIATION OF THE THRESHOLD SHEAR VELOCITY
WITH MOISTURE CONTENT

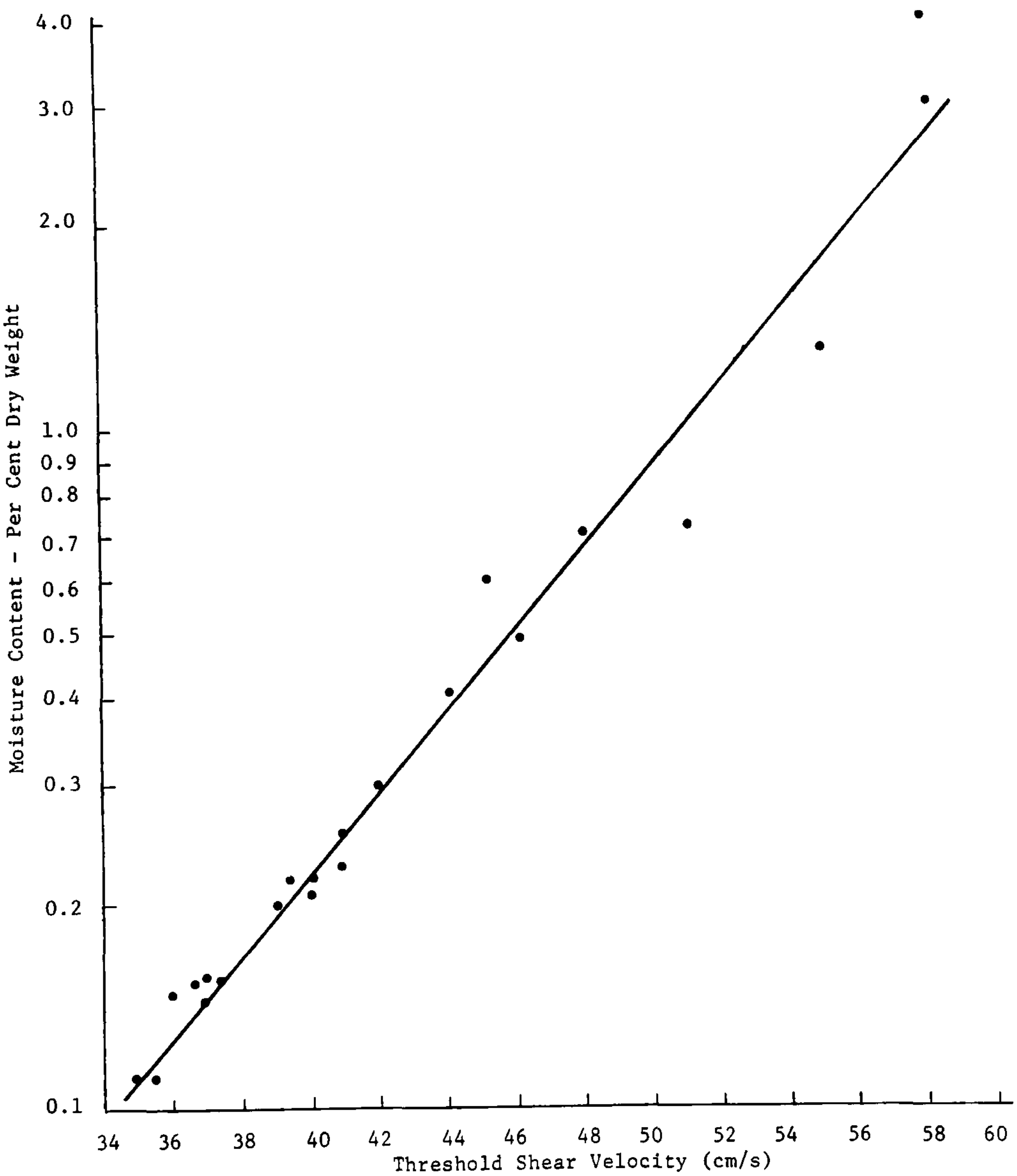


FIGURE 4.1

(after Belly, 1964)

necessary to initiate movement are outside the normal range of naturally occurring wind velocities.

An attempt was made to try to deliniate a threshold moisture content above which eolian transport did not occur in the Slims River Valley. However, the lack of data points in the critical moisture content and wind velocity ranges made this impossible.

Under these circumstances it was concluded that the rejection of sample points should not be made on the basis of surface moisture content but on the basis of precipitation. For this reason all days which had greater than 0.01 inches (0.25 mm) of rainfall were eliminated from the analysis. Those points which have not been included in the analysis are included on the scatter plots.

The data was checked for normality by the Kolmorgov-Smirnov one sample test. As might be expected the data was not normally distributed and had to be transformed. In all cases transformation to Log_{10} gave acceptable approximations to the normal curve.

Saltation-Creep Relationships

A relatively strong linear relationship exists on the Slims River between the amount of sediment transported in surface creep and saltation ($r = 0.78$). It was also found that on average the amount of sediment transported in creep represented only about 5 to 7 per cent of the total sediment load in saltation and creep. This percentage is considerably lower than values reported by other investigators. Bagnold (1941) in his investigations of dune sand found that creep was usually about 25 per cent of the total load. Similarly, Chepil (1945) has reported a value of 15.7 per cent for sand between 0.15 and 0.25 mm and 24.9 per cent for sand from 0.25 to 0.83 mm. Horikawa and Shen (1960) also found values of about 20 per cent and concluded that the proportion of creep load to total load is independent of wind velocity.

The relatively low percentage of creep load may be related to the nature of the delta sediment. Ishihara and Iwagaki (quoted in Horikawa and Shen, 1960) found in their field investigations that the percentage of transport in creep ranged from 6.5 to 16.6 per cent and appeared to be related to the grain size of the surface sediments. Sharp (1964) has also suggested, on the basis of his field observations that the percentage of sediment transported in creep increases as the proportion of coarser grains in the surface sediment increases. Thus, the relatively low percentage of sediment transport in surface creep on the Slims River delta may be related to the fact that the delta is comprised of fine sand and silt which are considerably finer than the sediments considered by other investigators (Fig. 4.2).

Sharp's (1964) argument that the proportion of sediment transported in creep increases as the proportion of coarse grains increases does not necessarily hold true for all situations. In some cases the proportion of sediment transported in creep may increase as the sediment size decreases. The reason for this is that individual grains are not always carried as discrete units. Rather, many fine grains may join together and move as aggregates. This bonding can be

TYPICAL GRAIN SIZE FREQUENCY CURVES
FOR THE SURFACE SEDIMENTS

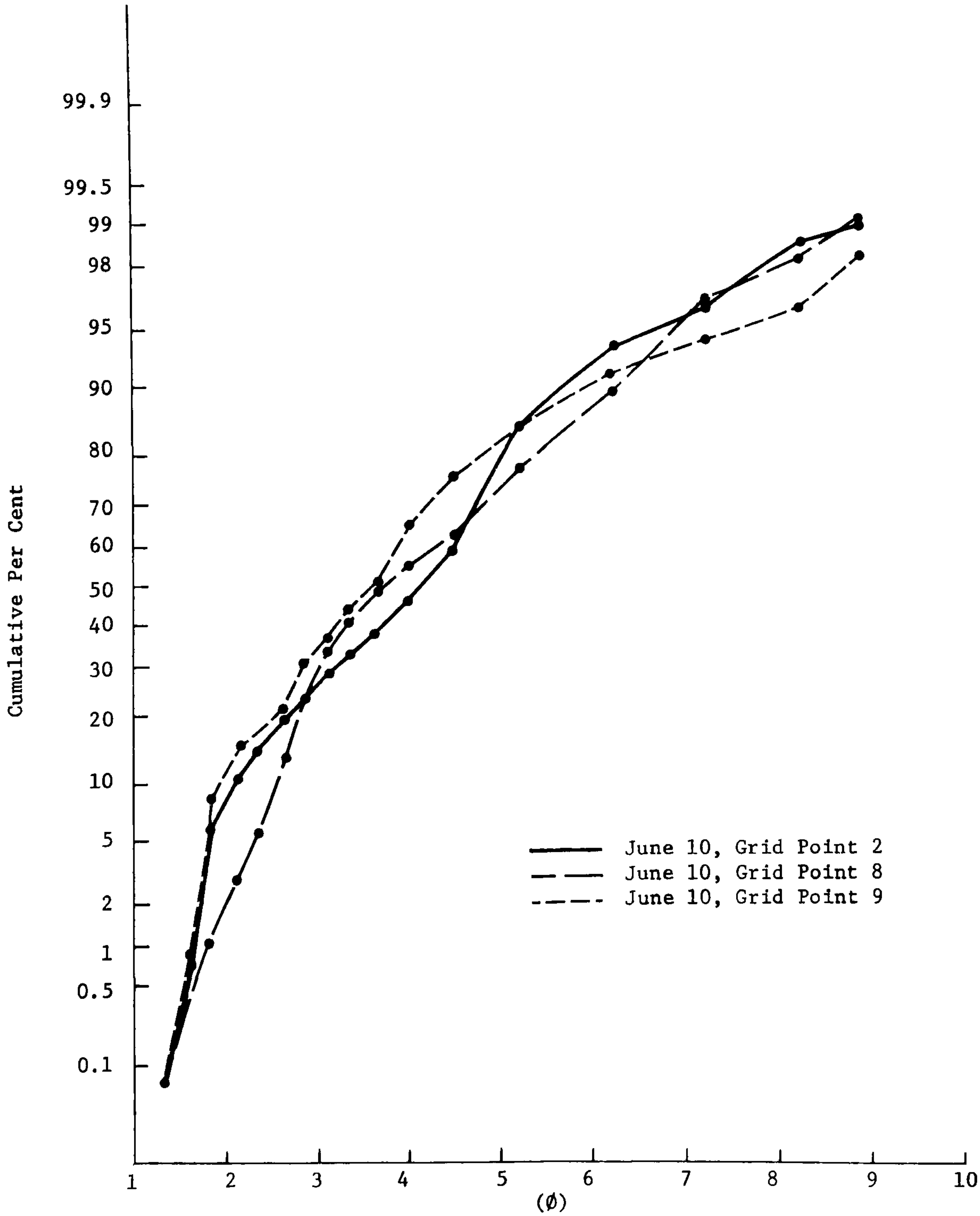


FIGURE 4.2

TYPICAL GRAIN SIZE FREQUENCY CURVES
FOR THE SURFACE SEDIMENTS

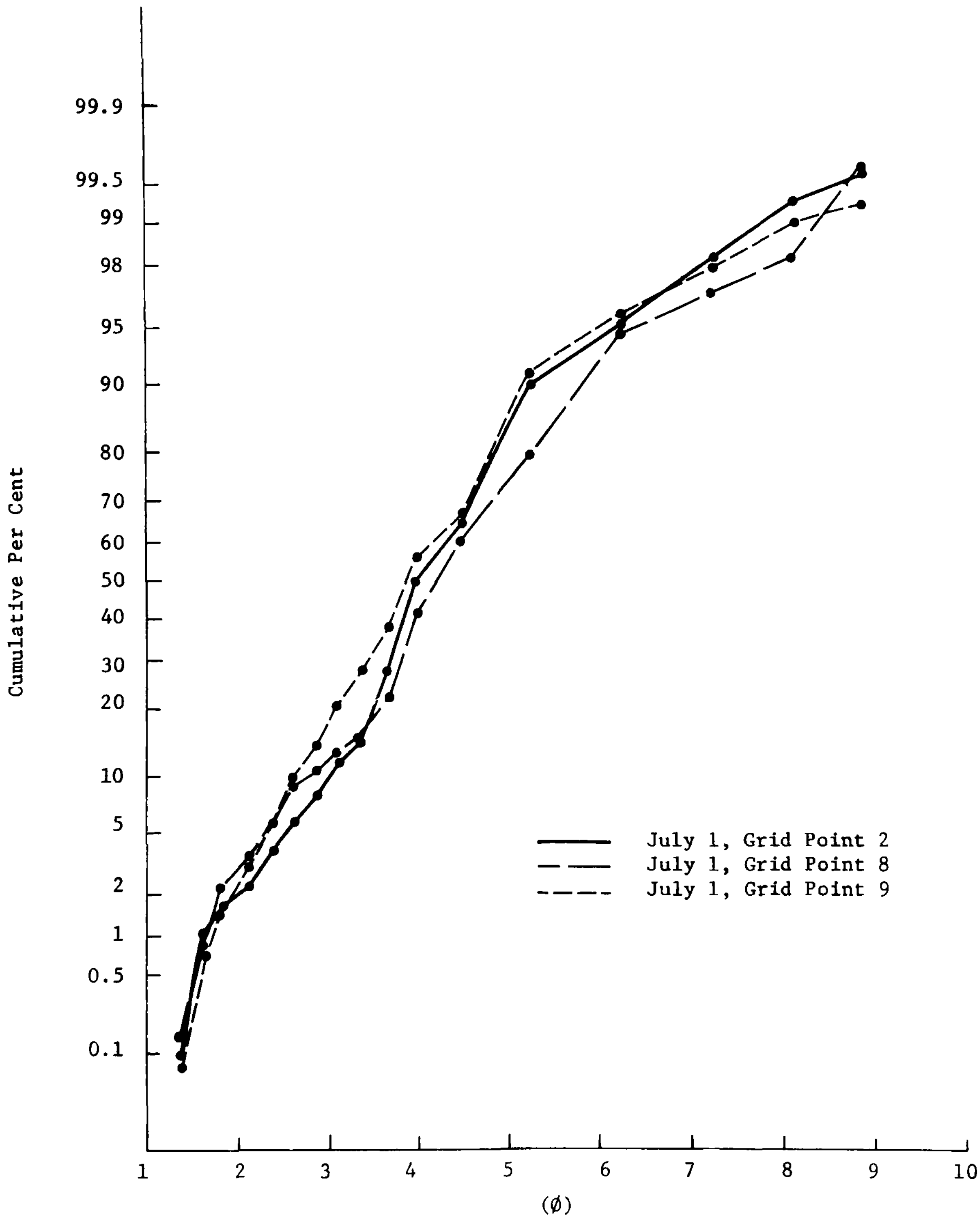


FIGURE 4.2 (continued)

caused by many agents such as organic residues, salts and clay particles, the most common being clay (Chepil, 1945).

Chepil (1945) has shown that in the case of four different textured soils, the lowest percentage of sediment transported in creep was associated with a pulverized loam and the highest with a coarsely granulated heavy clay. On analysis, Chepil found that the normally finer textured clay had a large number of aggregates bonded together by clay. Thus, in effect, the heavy clay was coarser than the loam and consequently had a higher percentage of transport in creep.

Sediment collected in the creep and saltation traps was frequently analysed by hand lens to check on the degree of aggregation. Although the number of large aggregates was usually relatively small, the majority were almost always found in the creep trap. In an analysis of 15 creep samples the total weight of the aggregates never exceeded two per cent of the total weight caught in the trap. The small number of aggregates found is probably due to the fact that the delta sediments have a relatively low clay content (Fig. 4.2). From this analysis it was concluded that this form of aggregation did not seriously affect the proportion of sediment transported in creep.

In addition to the large aggregates comprised of many fine grains bonded together, a second form of aggregation was also noted. In all samples, extremely fine grains in varying number adhered to the larger grains in the sample. This attraction of the fine grains to the larger ones appeared to be caused by electrostatic charges. Although this type of aggregation does not appreciably alter the size of the larger grains, it does alter the nature of the grain size distribution of the total sample.

Grain size analysis was carried out on saltation and creep samples from 28 of the 40 observation days.[1] The first three moments of the grain size distribution were calculated for each sample using the equations on page 62.

The first three moments for the 50 creep and saltation samples are given in Table 4.1 and per cent weight in each size class in Appendix C.

It can be seen from the grain size data that the mean size for both the creep and saltation samples usually falls within the very fine sand range (3.00ϕ - 4.00ϕ). Although these sediments have a finer mean size than one might expect of sediments transported in creep and saltation, they reflect the grain size distribution and fine mean size of the delta sediments as well as the nature of the transport. Since the surface sediments are fine and relatively well sorted, it is not surprising that sediment removed from the surface by a lower energy erosion system is similar in grain size distribution.

The mean size of the creep and saltation samples is also affected to some extent by the two forms of aggregation, especially the adhering of small grains to larger ones by electrostatic charges. In the grain size analysis these grains are represented as discrete particles, but in fact were often transported as large aggregates or as "dust" adhering to larger grains. In this respect there are

[1]Only those samples greater than 5 - 7 grams could be analysed by the combined sieving and falling drop techniques.

TABLE 4.1

GRAIN SIZE ANALYSIS OF CREEP
AND SALTATION SAMPLES COLLECTED
DURING DAILY OBSERVATIONS

Sample Number	Mean (ϕ)	Standard Deviation (ϕ)	Skewness
	Creep		
C07D*	3.72	1.06	0.11
C08D	3.73	0.71	0.33
C10D	3.83	0.77	-0.09
C15D	3.87	0.82	0.18
C28D	3.42	0.76	0.54
C30D	3.87	0.80	0.14
C35D	3.90	0.78	-0.20
C56D	3.80	0.78	0.05
C02D	3.65	0.61	0.83
C03D	3.69	0.64	0.66
C04D	2.83	0.96	0.07
C09D	3.77	1.00	0.22
C11D	3.59	0.91	-0.14
C12D	3.01	1.14	0.18
C14D	3.63	0.54	0.26
C19D	3.28	0.58	0.50
C20D	3.17	1.08	-0.03
C24D	3.37	0.59	0.39
C25D	3.07	0.99	0.26
C26D	3.02	0.67	0.35
C27D	3.07	0.76	0.15
C29D	3.69	0.61	0.19
C31D	3.57	0.89	0.38
C46D	3.53	0.52	0.70
C47D	3.75	1.00	0.70
	Saltation		
S02D	4.02	1.00	-0.62
S03D	4.02	1.02	-0.60
S04D	3.73	0.99	-0.06
S07D	4.06	0.98	-0.58
S08D	4.07	0.72	-0.14
S09D	4.12	0.65	-0.26
S10D	4.19	0.65	-0.46
S11D	3.97	1.06	-0.55

* surface creep 7th observation day
see appendix I

...continued

TABLE 4.1--continued

Sample Number	Mean (ϕ)	Standard Deviation (ϕ)	Skewness
S12D	3.81	1.08	-0.24
S14D	3.98	1.04	-0.59
S15D	4.52	0.81	-1.09
S19D	3.90	0.77	-0.12
S20D	3.86	0.76	-0.11
S24D	3.90	0.76	-0.16
S25D	3.85	0.97	-0.14
S26D	3.83	0.80	0.18
S27D	3.85	1.02	-0.18
S28D	3.90	0.78	0.24
S29D	4.02	1.07	-0.61
S30D	4.29	0.67	-0.10
S31D	3.93	1.00	-0.63
S35D	4.79	0.60	-1.55
S46D	3.91	0.74	-0.18
S47D	4.10	0.97	-0.67
S56D	4.18	0.73	-0.45

more fine particles accounted for in the distribution than should be, which effectively makes the mean size of the sample finer.

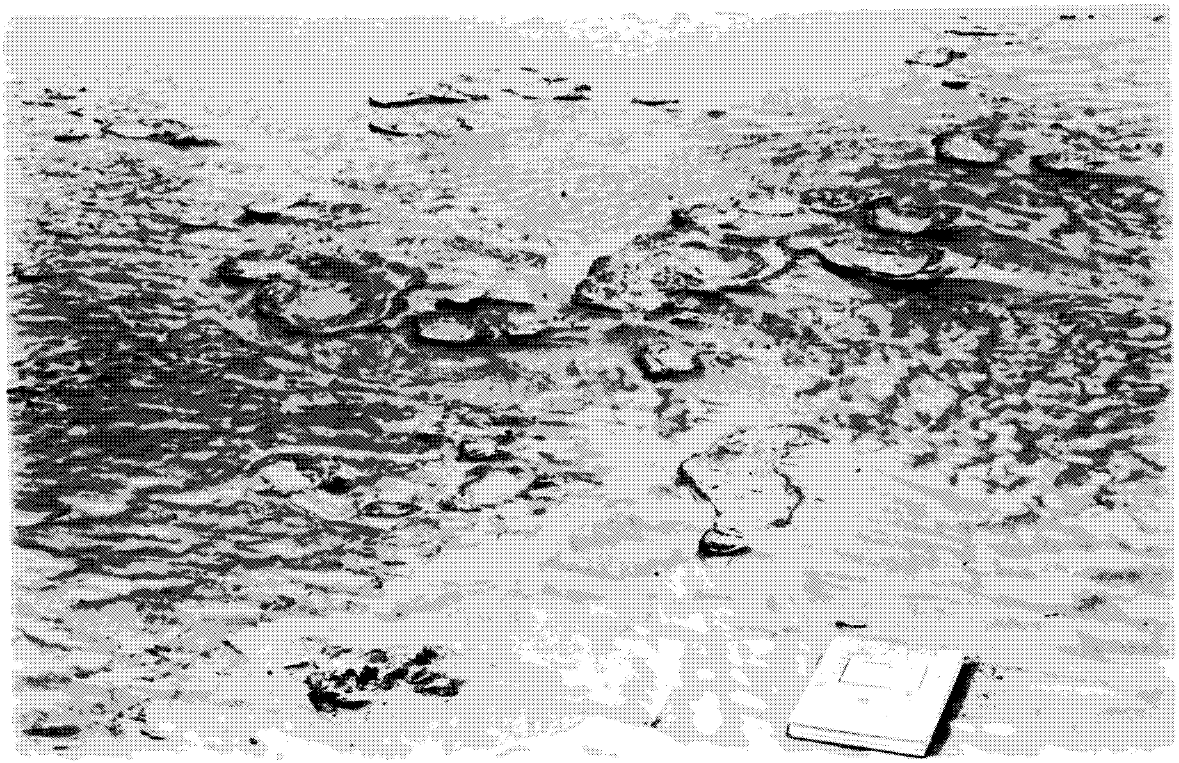
Although some aggregation was found, it does not seriously alter the relative proportion of sediment carried in saltation and creep. The reason for this is that particles do not usually bond together as large transportable aggregates because of the lack of clay. Rather, the Slims River delta sediments are transported for the most part as discrete grains, except for the adhesion of a few extremely fine grains to the larger ones. Thus, because of the fine size and lack of aggregation, the majority of sediment is able to be transported in saltation, rather than in creep.

Chepil (1945) also suggests that the nature of the surface can affect the amount of sediment transported in creep. Irregularities in the surface, such as ridges or depressions, tend to trap material moving in creep without seriously altering the amount transported in saltation.

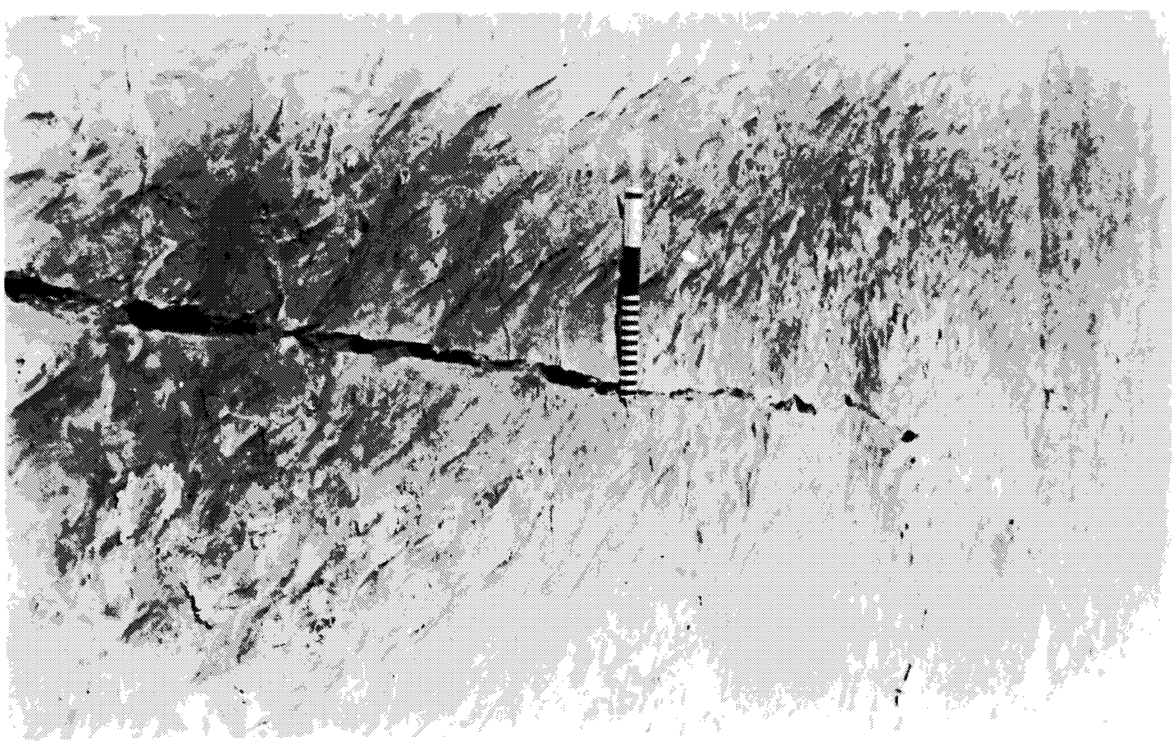
The surface of the Slims River delta, although relatively smooth, is covered with surface irregularities caused by wind erosion (Fig. 4.3). These surface features, which range from a few millimetres to several centimetres in height, effectively trap or slow down sediment moving in creep.

Thus, a low percentage of transport in creep would be expected, given the fine mean diameter of the delta sediments, the lack of aggregation and the presence of surface irregularities.

Variations in the first three moments of the frequency distributions for both the saltation and creep samples also appear to be



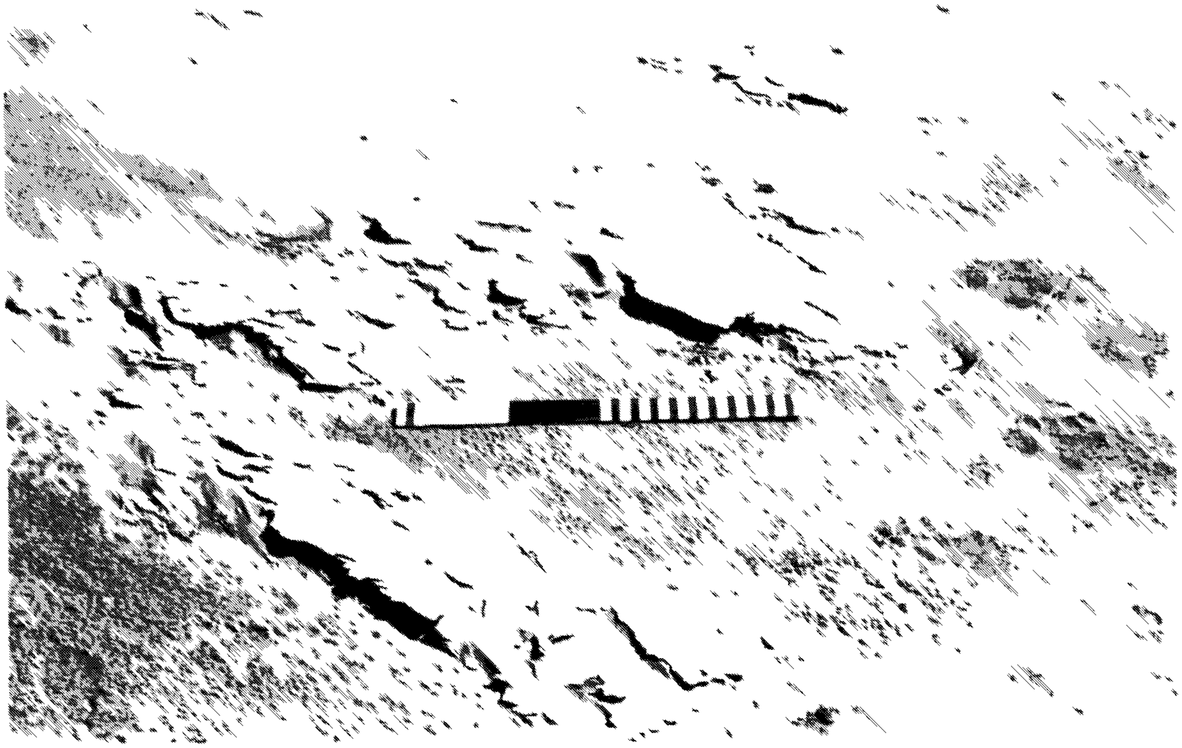
(a)



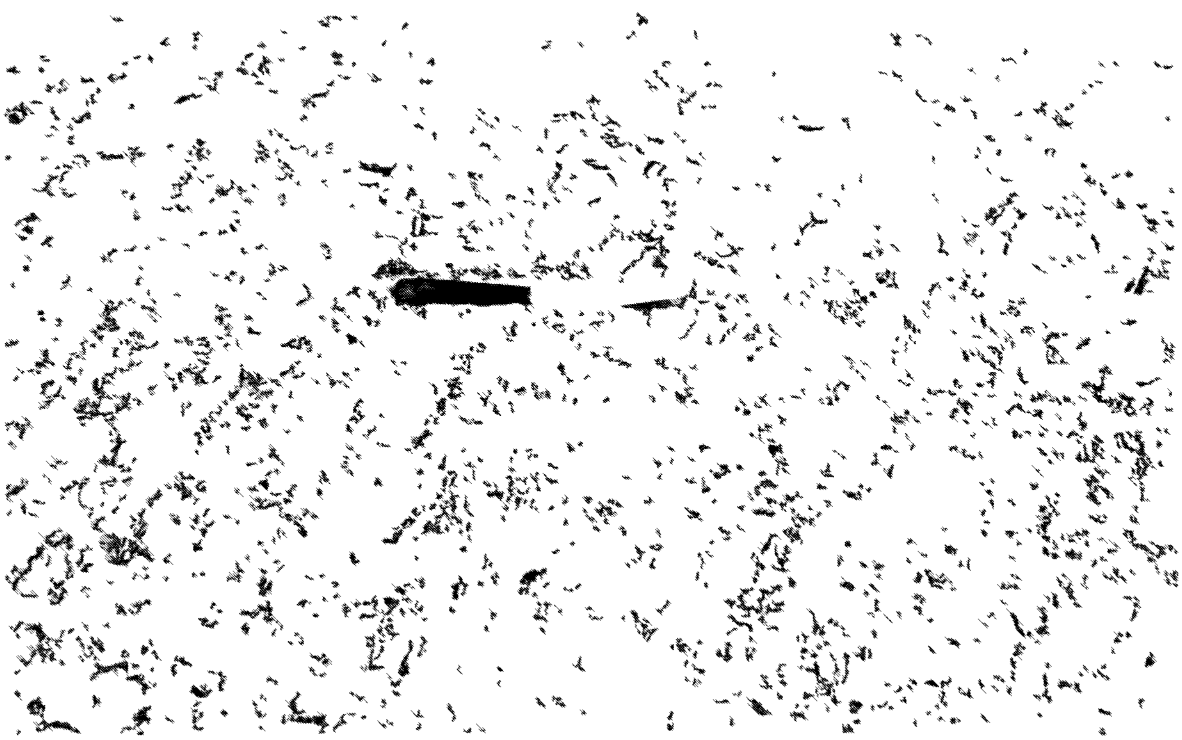
(b)

SURFACE FEATURES DEVELOPED ON THE
SLIMS RIVER DELTA BY WIND EROSION

FIGURE 4.3



(c)



(d)

FIGURE 4.3 (continued)

related to the nature of the transport and the size distribution of the delta sediments.

The standard deviations (i.e. second moment) of the creep samples were compared to those of the saltation samples by means of a Student's t test to check if the sorting in the saltation samples was significantly better. Results of the t test demonstrate that the sorting in saltation samples did not differ significantly from that in the creep samples ($t = 1.36 < t$ for 0.01). These results differ somewhat from those obtained in wind tunnel experiments by Williams (1964), who found that sediment moving in saltation was usually better sorted than material moving in creep. These differences, however, were usually very small and were not tested statistically. Williams, (1964) has also found that the creep samples were always coarser than the corresponding saltation samples. Similar results were obtained in the analysis of the mean size of the creep and saltation samples collected on the Slims River delta ($t = 6.08 > t$ for 0.01) which showed that the creep samples were significantly coarser than the saltation samples.

The most interesting difference between the creep and saltation samples is shown by the third moment, skewness. In all but four cases, the creep samples have positive skewness and the saltation samples negative skewness. Thus, the creep samples have a tail of fines and the saltation samples a tail of coarser grains.

It is thought that the positive skewness (i.e. tail of fine grains) in the creep samples may be related to the adhering of small particles to larger ones by electrostatic charges. The creep samples are usually comprised of relatively coarse grains which do not get moved directly into the air stream because of their size. As previously mentioned, the grains often have many small particles adhering to them which would normally be transported either in saltation or suspension. Thus, within the creep samples there is a percentage of fine material present which may not have been present except for the adherence of these fine grains to the larger ones.

The tail of coarse material in the saltation samples may be directly related to the size distribution of the creep samples. Negative skewness in samples can also be caused by the removal or truncation of fine sediment which would normally be present (Martins, 1965; Folk, 1966). Since some of the fine material which should be transported in saltation or suspension is being effectively held in creep, at least part of the fine tail of sediment will be missing from the saltation samples. This situation would result in the saltation samples having an apparent coarse tail or negative skewness.

Mean Daily Flow Rate

The amount of sediment transported through a unit width in unit time has been termed the sediment flow rate by Bagnold (1941). Mean daily flow rates were calculated for the sediment transported in saltation and creep on the Slims River delta. Since the intake orifices of both the saltation and creep traps were 1.0 cm this could be done by dividing the total daily catch in each trap by the length of the sampling period. An attempt was made to keep the sampling period as close to 24 hours (86,400 seconds) as possible. The mean daily flow rates (mg/cm.s) in saltation and creep for each observation day are given in Appendix A. Mean daily shear velocities (U_*') were also calculated using equation 1.5 from daily observations of the two totalizing anemometers located near the sediment traps. The mean daily shear velocities (cm/s) are given in Appendix A.

On the Slims River delta the logarithms of saltation and creep flow are related to the logarithm of the mean daily shear velocity. It can be seen from Table 4.2, however, that a stronger linear relationship exists between U_*' and saltation flow ($r = 0.70$) than between U_*' and creep flow.

The surface of the Slims River delta is covered with small irregularities which tend to obstruct the movement of sediment in creep, but not in saltation. It is suggested that the interference to creep flow by the surface irregularities is at least partly responsible for the relatively weak relationship between creep flow and U_*'.

In investigations of wind erosion, the usual concern is with the total amount of sediment transported rather than the amount transported in a single mode. Thus, further discussion of the data will be centered on total sediment flow by creep and saltation (q) so that results of this study may be compared with related work.

On the Slims River delta a significant relationship ($r = 0.71$) exists between the sediment flow (q) in saltation and creep and the shear velocity (U_*') (Fig. 4.4).

TABLE 4.2

CORRELATION MATRIX FOR MEAN DAILY OBSERVATIONS

	Saltation Flow Rate	Saltation-creep Flow Rate (q)	Mean Daily Shear Velocity	Mean Daily Surface Moisture Content	Mean Daily Surface Salt Concentration
Creep flow rate	0.78	0.97	0.59	-0.69	-0.51
Saltation flow rate		0.99	0.70	-0.64	-0.60
Saltation-creep flow rate (q)			0.71	-0.67	-0.59
Mean daily shear velocity				-0.42	-0.39
Mean daily surface moisture content					0.48

VARIATION OF THE MEAN DAILY
SALTATION-CREEP FLOW RATE
WITH SHEAR VELOCITY

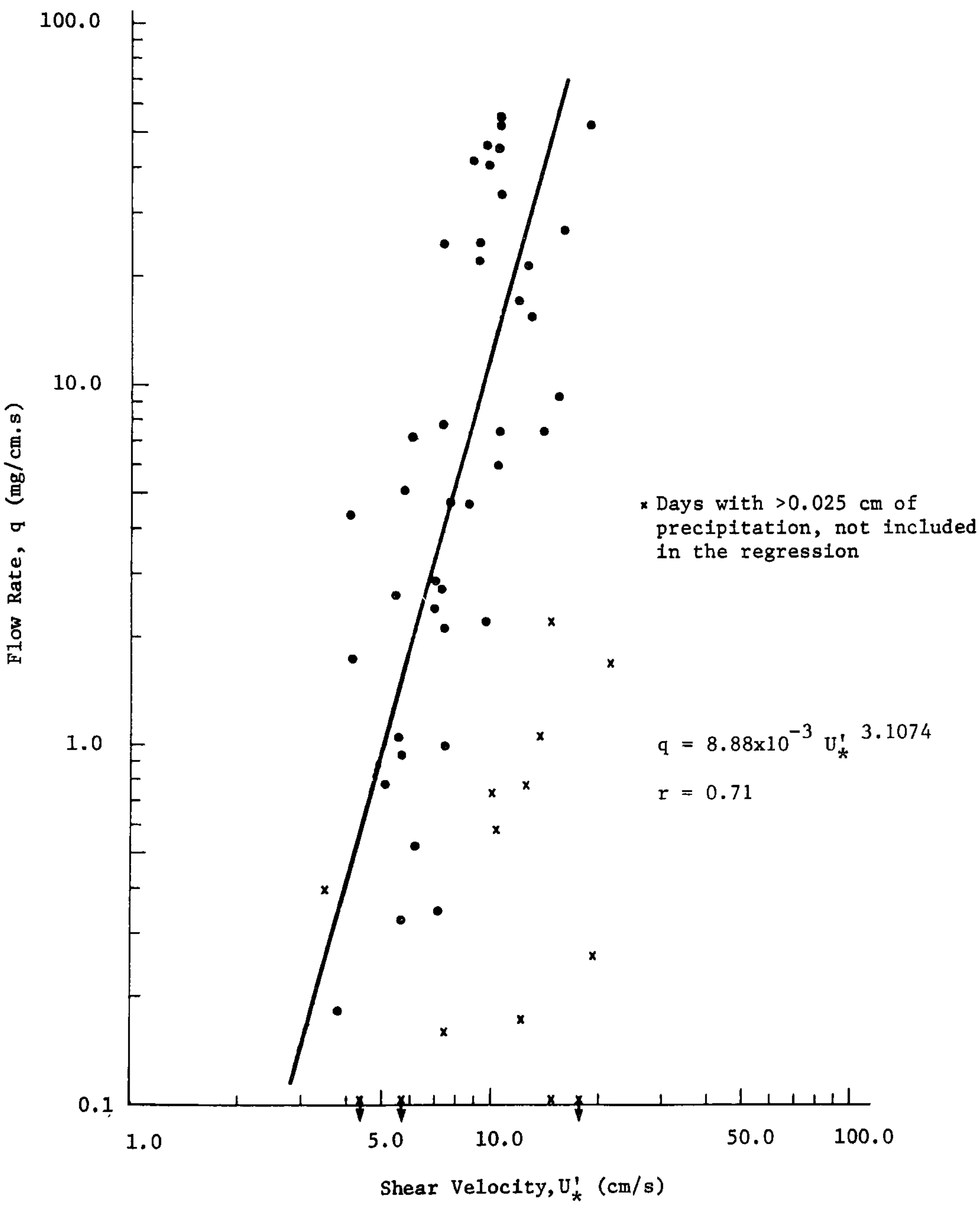


FIGURE 4.4

This relationship can be expressed as a power function (log-log) of the form

$$q = aU'^{\,b}_{*} \quad \text{(mg/cm.s)} \qquad \ldots\ldots\ldots\ldots\ldots 4.1$$

where $a = 8.8 \times 10^{-4}$

$b = 3.11$

Similar power relationships have also been obtained from theoretical and wind tunnel investigations. For example, Bagnold (1949), from his theoretical investigations has proposed

$$q = c\left(\frac{d}{D}\right)^{0.5} \frac{\rho}{g} U'^{\,3}_{*} \quad \text{(g/cm.s)} \qquad \ldots\ldots\ldots\ldots\ldots 4.2$$

where d = average grain diameter (mm)

D = 0.25 mm (diameter of a standard sand)

ρ = specific weight of air

g = gravitational acceleration

U'_{*} = shear velocity

c = empirical constant
whose value increases with a decrease in the sorting of the original surface
1.5 for nearly uniform sand
1.8 for naturally graded sand
2.8 for sand with a wide range of grain sizes

From later wind tunnel experiments, Zingg (1952) modified the Bagnold equation to

$$q = c\left(\frac{d}{D}\right)^{0.75} \frac{\rho}{g} U'^{\,3}_{*} \quad \text{(g/cm.s)} \qquad \ldots\ldots\ldots\ldots\ldots 4.3$$

Kawamura (1951) proposed a similar formula based on theoretical work:

$$q = k \frac{\rho}{g} (U_* - U_{*_t}) (U_* + U_{*_t})^2 \quad (\mathrm{g/cm.s}) \quad \ldots\ldots\ldots\ldots\ldots 4.4$$

where U_* = friction velocity

U_{*_t} = threshold friction velocity

ρ = specific weight of air

k = empirical constant

The k value varies with the sand size and sorting of the initial sediment and must be determined by wind tunnel experiments. Kawamura obtained k = 2.78 for moderately well sorted sand with a mean diameter of 2.5 mm.

A more recent wind tunnel investigation by Williams (1964) has yielded a relationship similar to that found on the Slims River delta:

$$q = a\, U_*^b \quad (\mathrm{g/cm.min}) \quad \ldots\ldots\ldots\ldots\ldots 4.5$$

where a and b are empirical constants which vary with particle shape

Shape Type	a	b
spheres	5.5×10^{-7}	4.1068
sand	1.27×10^{-5}	3.4277
crushed quartzite	1.853×10^{-4}	2.7619

Although the above theoretical and empirical wind tunnel formulae do differ somewhat in form they give approximately the same results (i.e. flow rates) if suitable constants are chosen. For example, Belly (1964) has found that both the Bagnold and Kawamura formulae fit his wind tunnel data quite closely if $c = 2.5$ in Bagnold formula and $k = 3.2$ in Kawamura formula (Fig. 4.5).

Belly (1964) suggests, however, that because of the wide scatter found in most field data, the theoretical formulae are still of use in the description of a particular situation when suitable constants are chosen (Fig. 4.6).

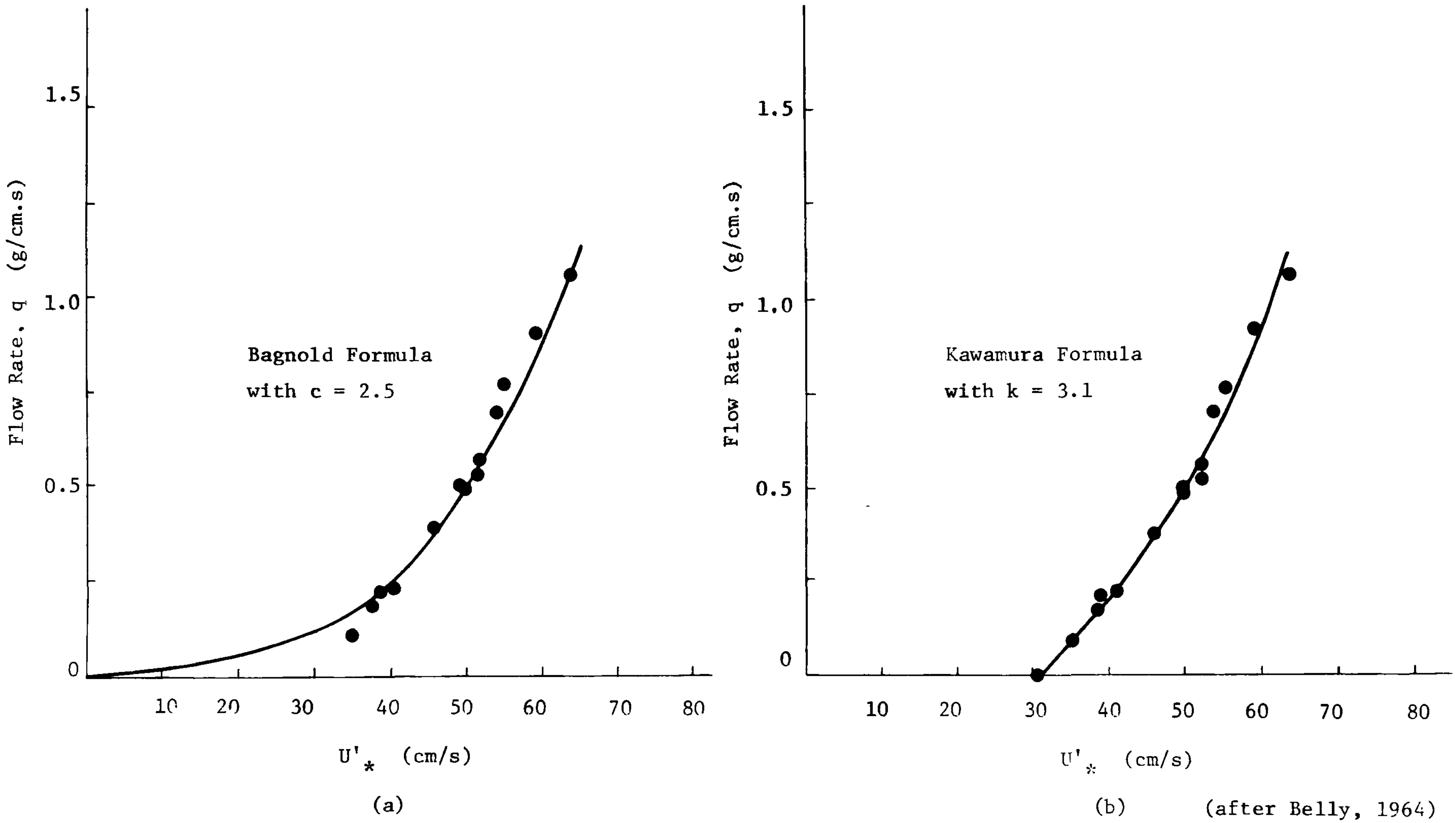
It can be seen from Fig. 4.6 that the slopes of the theoretical and observed curves are very similar. It is also evident that the Slims River Valley curve in this particular shear velocity range falls between the upper and lower limits defined by Williams' (1964) empirical observations and Bagnold's (1941) theoretical calculations.

Some parameters such as grain size, degree of sorting and particle shape are taken into account by the constants in the theoretical formulae by such authors as Bagnold (1941) and Zingg (1952). Changes in these constants result in a shifting of the curves as shown in Fig. 4.6. The range of constants partially explains the close agreement between the theoretical formulae and wind tunnel results obtained by Belly (1964) and Williams (1964).

In wind tunnel investigations other variables in the wind erosion system are closely controlled and eliminated, and dry cohesionless sediments with relatively narrow size ranges are usually used. This, however, is not always the case in field situations where a wide

COMPARISON BETWEEN BELLY'S EXPERIMENTAL RESULTS AND THE BAGNOLD FORMULA

COMPARISON BETWEEN BELLY'S EXPERIMENTAL RESULTS AND THE KAWAMURA FORMULA



(after Belly, 1964)

FIGURE 4.5

COMPARISON OF THE MEAN DAILY FLOW RATE CURVE
WITH THEORETICAL AND EMPIRICAL CURVES
PRESENTED BY OTHER AUTHORS

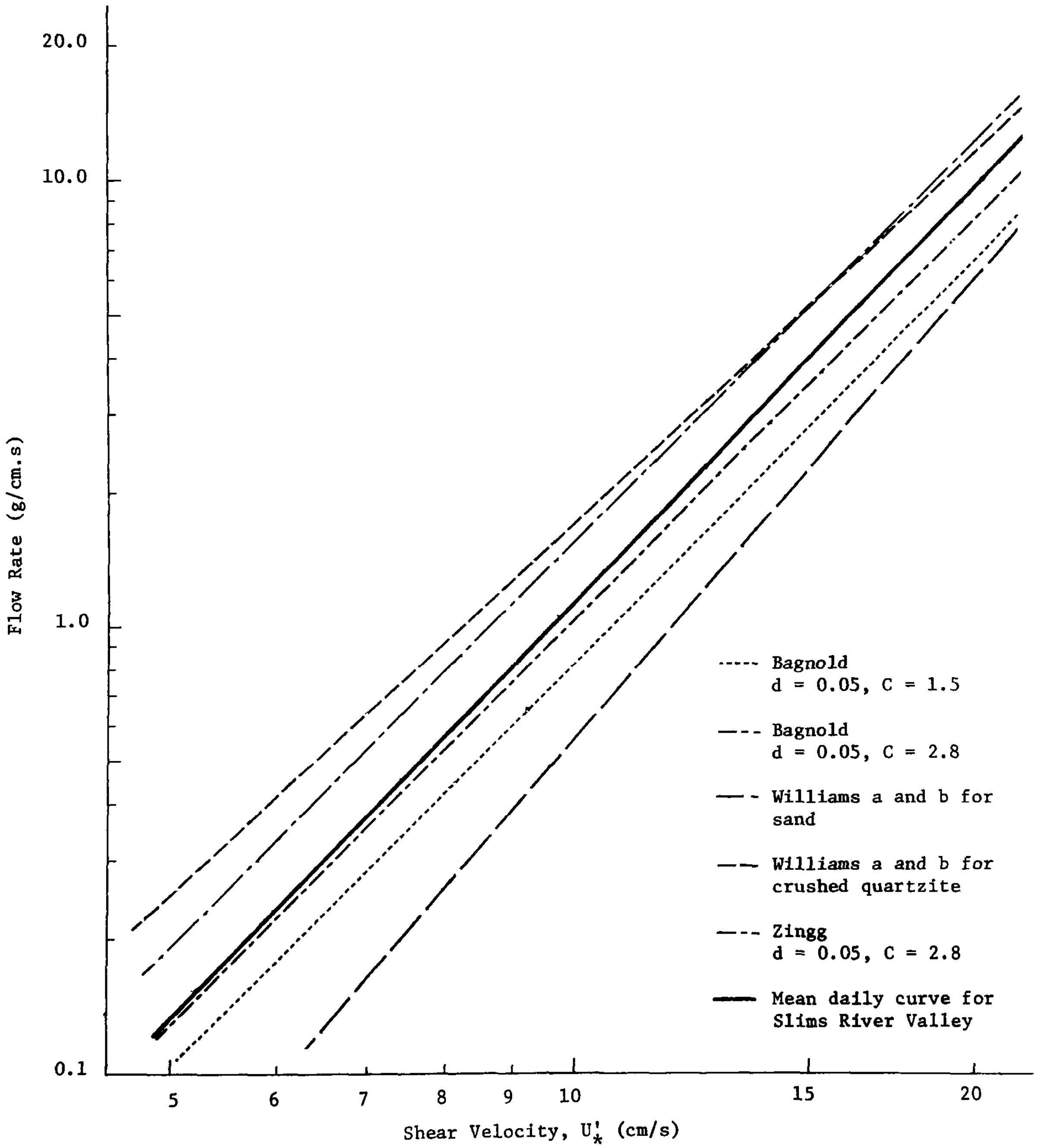


FIGURE 4.6

variety of factors, which fluctuate daily and hourly, affect the ability of the surface to supply sediment to the air stream. It is suggested, therefore, that the scatter of the data on the flow rate against shear velocity plot (Fig. 4.4) results from the complex interaction of variables which are not accounted for in the above bivariate relationships.

Although a large number of factors can affect the rate of supply to the air stream, two appear to be of particular importance on the Slims River delta. These are surface moisture and the presence of surface salts. Both surface moisture and surface salt concentration tend to stabilize the surface by holding individual particles in place.

The relationship between sediment flow (q) and soil moisture is shown in Fig. 4.7. Although the relationship is not particularly strong ($r = 0.67$), it is significant at the 99 per cent confidence level and indicates that sediment transport decreases rapidly with increasing moisture content.

Chepil (1956) suggested that a decrease in erodibility with increasing water content is a function of the cohesive force caused by the water which surrounds the sediment particles. In a granular mass, most of the water is held in the pore spaces, which is analogous to the situation in a capillary tube. If a capillary tube is placed in a liquid such as water, the liquid will rise up in the tube as a result of a pressure difference between the air and the liquid. This pressure difference is caused by the surface tension between the liquid and the wall of the capillary tube. In a glass capillary tube, the pressure of the water is always less than the pressure of the air. The pressure

VARIATION OF MEAN DAILY SALTATION-CREEP
FLOW RATE WITH SURFACE MOISTURE CONTENT

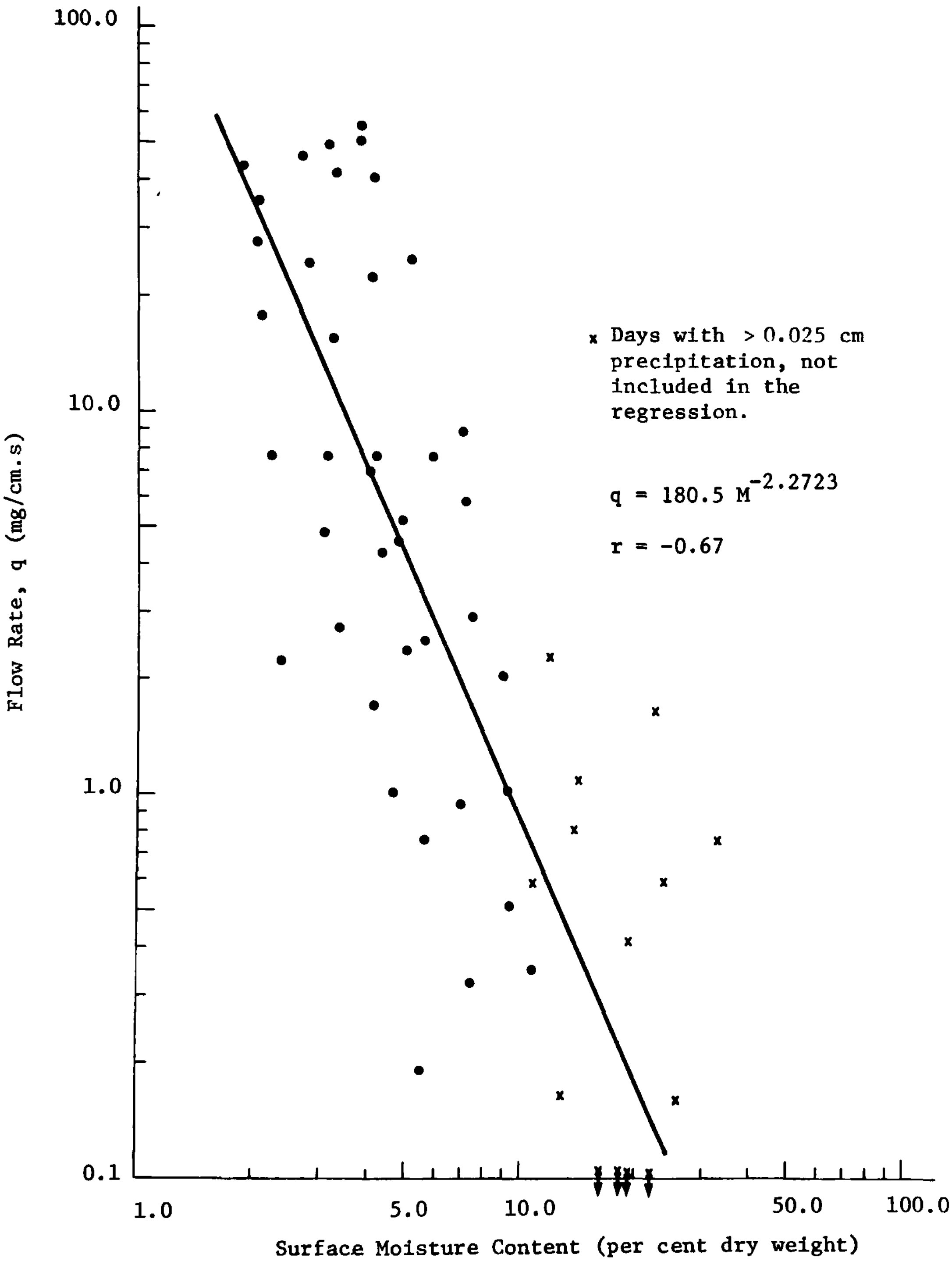


FIGURE 4.7

difference can be shown to be:

$$Pa - Pw = \frac{2\omega}{r} \cdot \cos\alpha \qquad \ldots\ldots\ldots\ldots\ldots 4.6$$

where Pa = pressure of the air

Pw = pressure of the water

ω = surface tension

r = radius of the interface

α = contact angle between the liquid and the capillary tube wall.

The situation in granular materials is analogous to the capillary tube. The pore system within granular masses can be viewed as a series of various sized interconnected capillary tubes which obey the above relationship. In the case of water and granular materials, the contact angle, α, is generally assumed to be zero degrees and the radius of curvature of the meniscus, r, equivalent to the radius of the pore in which the air water interface lies (Williams, 1966 b). The pressure difference, or suction, developed in the pores of the granular masses can therefore be defined as:

$$Pa - Pw = \frac{2\omega}{r} \qquad \ldots\ldots\ldots\ldots\ldots 4.7$$

From the above relationship, it is evident that the pressure difference is negative as long as the pores are not completely filled with water. This negative pressure or suction can reach very high values with decreasing moisture content and effectively pulls the individual grains together. Thus, it would appear to be contrary to argue that increasing water content decreases erodibility.

However, when a soil dries, the remaining water is held in progressively finer pores. As water recedes into the smaller pores, the contact area between the water and mineral grains is significantly reduced with a resultant increase in the suction per unit area of contact. There would, however, be a decrease in the total suction force between grains (i.e. the product of contact area and force would decrease). Chepil (1956) suggests that the total area of contact increases more rapidly than the curvature of the menisci (i.e. 1/pore radius) decrease. Consequently, the total force of attraction or suction between grains increases as more water is added.

The velocity at which the wind will first begin to move the uppermost grains of a dry cohesionless sediment is a function of the particle's mass and the air density. The relationship has been investigated theoretically by Bagnold (1941) and Zingg (1952), who have shown that:

$$U_{*_t} = A\left(\frac{\sigma - \rho}{\rho} \; . \; g \; . \; d\right) \qquad \ldots\ldots\ldots\ldots 4.8$$

where U_{*_t} = threshold shear velocity

d = grain diameter

g = acceleration of gravity

σ = specific weight of the grain

ρ = specific weight of air

A = constant (0.1) for all sands greater than 0.25 mm.

This relationship, however, does not account for any external forees such as moisture, which tend to hold the grains in place. Belly (1964), from wind tunnel investigations has found that the Bagnold formula for threshold velocity can be modified to account for moisture content by:

$$U_{*_t} = A\left(\frac{\sigma - \rho}{\rho} .g.d\right)^{0.5} (1.8 + 0.6 \log M) \quad \ldots\ldots\ldots\ldots 4.9$$

where M = the moisture content expressed as per cent dry weight.

Thus, only the force of the wind in excess of that required to overcome the force of gravity acting on the uppermost particles and the force of cohesion between these and other particles will contribute to the movement. That is, for a given sediment surface and wind velocity, sediment transport will be lower over the surface with the higher moisture content.

Moisture in the delta sediments is derived from three sources: atmospheric humidity, precipitation, and by capillarity. Belly (1964) has suggested that the maximum amount of moisture that can be imported to a sand surface from atmospheric humidity is approximately 0.25 per cent. Although this can be of significance in wind tunnel experiments, it is relatively unimportant in field situations where other sources of moisture completely overshadow the effects of atmospheric humidity.

The amount of precipitation falling in the delta area during the summer months is relatively small (Table 1.4). It has been found that surface moisture derived from precipitation only alters the surface's susceptibility to wind erosion for a maximum of a few days after the rainfall.

By far the most important source of surface moisture is that derived from ground water by capillary flow. It can be seen that although the immediate surface (<1.0 cm) is usually relatively dry, the sediments at depths greater than this are continually damp. This permanent wet condition resulting from capillarity is directly related to the grain size distribution of the delta sediments.

The rate and height to which water moves in a granular mass is a function of the pore size distribution and the permeability. Clays have extremely fine pores and consequently can draw water to great heights. However, because of the small pores, water can not migrate through the clay very quickly. Conversely, sands can only hold a relatively small column of water, but because of the large pores, the water can migrate very quickly. The optimum relationship between suction and permeability is found in silts where relatively high suctions can be developed without greatly reducing the permeability. In the case of the delta, water evaporating from the surface can be quickly replaced by capillary flow. Therefore, the wind must almost continually overcome the cohesive force of water before sediment transport is initiated. Thus, the quantity of sediment transported at any given wind velocity will be reduced. (Fig. 4.7)

The presence of moisture at the surface does not appear to be the only factor affecting the rate of sediment transport on the Slims River delta. As can be seen from Fig. 4.8 the presence of surface salts is also significantly related to the sediment flow rate. The relationship demonstrates that transport decreases rapidly with increasing surface salt concentration. Although the scatter is relatively wide, the relationship is significant at the 95 per cent confidence level ($r = -0.5949$). Several reasons for the degree of scatter can be presented. Perhaps the most significant is that the surface salt concentration values represent the daily mean of the twelve grid point samples which were collected in the early afternoon each day. Thus, variations could result from fluctuations in salt concentration during the day at any of the sample points. A second problem may arise from the rise of a single mean value. The problem is that in some cases this value may not adequately represent the surface conditions of the area from which the majority of the sediment was derived on a given day. To overcome these possible inadequacies, it would have been necessary to have an extremely dense sampling grid from which samples could be taken at frequent intervals. This type of sampling was beyond the manpower available for this study. It is felt however, that the relationship based on the gross average is representative of the situation in the Slims River delta. Mean daily surface salt concentrations for the grid during the study period varied from 14.0 to 59.9 m.e./ 100 grams of soil.

Salts at or near the surface of the delta result from four primary factors:

(1) the location of the delta in a low precipitation environment;

(2) excessive evaporation of moisture from the delta surface due to almost constant winds in the valley;

(3) presence of a permanently high water table, and

(4) rapid upward movement of capillary water.

Most of the water in the delta sediments is derived from the Slims River which crosses the delta in a braided pattern. During the summer months, the Slims River water contains a relatively high quantity of dissolved load (Bryan, 1972; Nickling, 1974). Water migrating laterally through the delta transports this dissolved material with it.

The dissolved concentration of this water is then augmented by the continuous contact with the sediment through which it is migrating.

When the capillary water evaporates, it leaves behind the previously dissolved material at or near the surface of the delta. The term "salt" has been used to describe this residual material in that it includes any soluble compound which has been formed by the partial or total replacement of hydrogen by a metal or metal-like radical. On warm dry days, the deposited salts form a thin white film on the delta surface (Fig. 1.4). Although the quantity of salt present at the surface is a function of the amount of evaporation, it is also related to the amount and intensity of precipitation just preceding the measurement. When precipitation falls on the surface, some of the soluble salts go into solution. This salt is then carried downward with the rainwater as it percolates into the delta. If only a trace of rain falls, this is almost immediately evaporated. However, if a relatively large amount of precipitation falls in a short period of time, a considerable proportion of the surface salts can be leached below the surface. This is evident by the sharp drop in salt concentration which occurred after heavy rainfalls (Appendix A). However, once evaporation begins to occur, capillary water transports the soluble salts to the surface where they are redeposited. This recurrent build-up of salts appears to be associated with a decrease in the susceptibility of the surface to erosion by the wind.

It is suggested that the susceptibility of the surface to erosion is reduced because of bonds formed by the adhesion of the sedimentary particles to precipitated salt crystals. As a soil initially begins to dry from a very wet condition, all but the largest pores contain water. In the case of the delta, this water has a relatively high dissolved salt concentration. In this very moist state, the salt would have little effect on the erodibility of the soil. Thus, at these higher moisture contents, the resistance of surface grains to movement by the wind is caused primarily by the cohesive force of the water. However, when the water held at the surface evaporates, the salt remains as small crystals which form bonds between individual grains. It should be noted that even when the immediate surface is almost dry, the sediments just below the surface (<0.5 cm) often remain moist. This high moisture gradient is able to develop because of the extremely high evaporation rate caused by the winds blowing across the surface such that during the strong wind storms, the surface of the delta becomes noticeably drier. If the wind dies down for as little as an hour, the surface begins to darken. This indicates that the surface is becoming increasingly wetter. This water moving towards the surface also transports salt in solution which is subsequently deposited when the water evaporates.

The stabilizing influence of surface salts can be very noticeable. Nickling (1973) has shown that the average surface salt concentration in the delta area increases as one moves away from the river. This has been related at least in part to the age of the delta sediments and flood characteristics of the Slims River. In the summer months increased meltwater supply from the Kaskawulsh Glacier produces flooding in the delta area which extends as much as 400 - 500 metres west of the main channel. The soluble surface salts in this area are therefore seasonally removed or reduced. Sediments outside the normal area of seasonal flooding are not usually affected by inundation and therefore only lose soluble salts by leaching caused by the percolation of snow melt and rain water. During August of 1973, the salt concentration of the surface sediment 200 metres from the northwest valley wall was approximately 75 - 80 m.e./100 grams of soil (Nickling, 1973). In this area, the surface sediments have a fluffy character and visually, appear to have a very loose structure (Fig. 4.3d). The fluffy appearance, however, is caused by the presence of salts which surround and separate the individual grains. Although the salt does cause the sediment to have a loose packing arrangement, it also effectively bonds the individual grains together, making them less susceptible to wind erosion. Field observation has shown that in this particular area, the sediments are not transported by the wind, even with wind velocities in excess of 15 m/s.

Although the bonding together of surface grains by soluble salts is not as pronounced in the study grid, it does seem to have a significant effect on the sediment transport rate (Fig. 4.8).

From the above discussion it would appear that the presence of both moisture and surface salts can significantly decrease the erodibility of the surface. A comparison of Fig. 4.7 and 4.8 also suggests that salts can decrease the erodibility of the surface even if it is not completely dry. This would appear to negate the argument that the salts are deposited as a result of the evaporation and subsequent drying of the surface. Two mechanisms can be given to explain this apparent discrepancy.

RELATIONSHIP BETWEEN MEAN DAILY FLOW RATE
IN SALTATION AND CREEP AND
SURFACE SALT CONCENTRATION

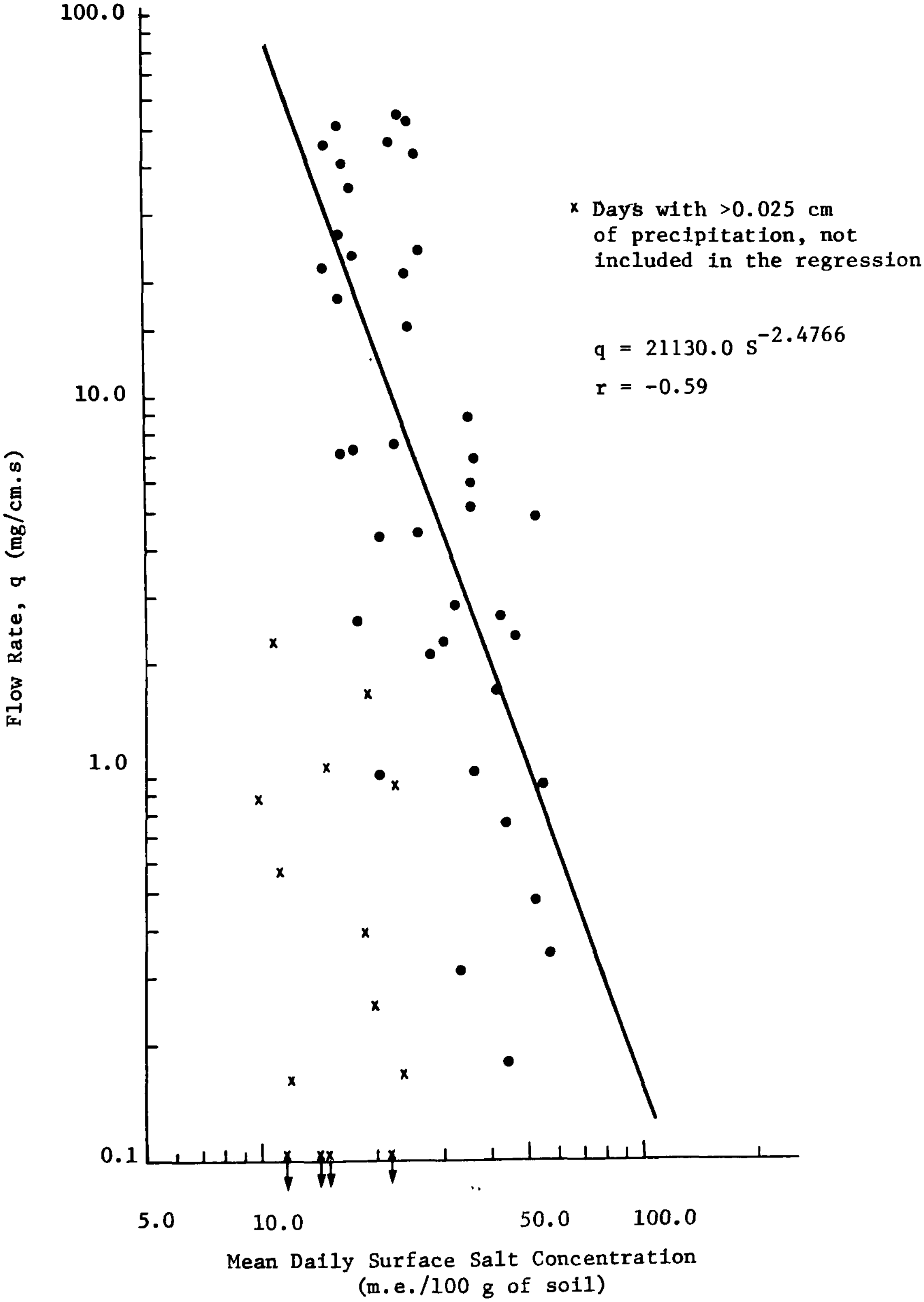


FIGURE 4.8

As a soil initially begins to dry from a saturated state, all but the largest pores contain water, with most of the soluble salts being in solution. In this case, the cohesive force of the water is almost totally responsible for the stability of the surface. A decrease in the surface moisture content by evaporation would cause an increase in the salt concentration of the remaining water. If the concentration exceeded the saturation point, some of the salt may be crystallized out in the larger pores. It is thought that the crystals will adhere to at least some of the grains which encompass the pore space, thereby forming a cement-like bond between these grains. Further decreases in surface moisture content would continue to cause increase in the concentration of the held water. Since water retreats into the smaller pores as the moisture content decreases, new bonds would be formed between grains which enclose the smaller pore spaces. Salt which crystallizes out, however, will grow preferentially on the original crystals in the larger pores.

A second reason why the salts could act as a stabilizing agent, even though the surface was not completely dry, may be related to spatial variation in drying of the delta surface.

The rate at which a smooth sediment surface will dry in still air is a function of the sediment's grain size distribution, or rather the pore size distribution. In order for water to be evaporated from the surface, the vapour pressure of the overlying air must be less than the vapour pressure of the evaporating surface (Rose, 1966). The vapour pressure of the water held within a granular mass is a function of the suction developed in the pores in which the water is being held. If the pore size is large, the vapour pressure gradient will be such that water vapour will be transferred to the atmosphere. Conversely, if the pore size is very small (i.e. a high suction) the vapour pressure of the water may be equal to or less than that of the air and no evaporation will occur. Thus, within any soil, at a constant temperature and air vapour pressure, evaporation will continue until the vapour pressure gradient equals zero. This particular vapour pressure will correspond to a critical suction or pore size, below which no evaporation will occur. Although the Slims River delta sediments are fairly uniform, there is spatial variation in the grain size and consequently, pore size distribution of the sediments. Thus, for any given air vapour pressure, there may be certain areas on the delta where the pore size distribution is such that almost all the pores are larger than the critical diameter. These sediments would therefore give up most of the water held within the pore system while other areas with a greater number of fine pores would retain moisture. In this manner, some areas of the delta could dry almost completely and in doing so, crystallize out the stabilizing salt even though other areas of the delta were still moist. This situation could also be enhanced by variation in the degree of near surface air turbulence.

As shown in Fig. 4.3 the surface of the delta is covered with small surface irregularities which cause the air flow near the surface to be turbulent. In these cases, if the wind has been blowing over the surface for a period of time, a distinct difference in the moisture content between the crests and depressions of the irregularities is evident (Fig. 4.3). It is thought that this results from the greater amount of turbulence (i.e. greater air flow) around the crests of the obstructions, causing evaporation to be greater around the peaks. Moisture and salt concentration measurements from the crests and depressions of fairly large irregularities (>1.0 cm) for two relatively windy days are shown in Table 4.3. As can be seen from this data, the crests of the irregularities are almost completely dry, while the troughs are still somewhat moist. Also, the crests have a significantly higher salt concentration which would indicate that they have a somewhat higher evaporation rate.

The difference in salt concentration between the peaks and crests of the large irregularities is visually apparent on almost any dry day. In such cases, the crests, or protruding edges, have a distinct white film of salt covering them, while no salt is visually apparent in the depressions.

Thus, because of variations and the mechanisms of surface drying, it appears that surface salts and surface moisture can simultaneously affect the susceptibility and transport rates of surface sediments in the Slims River delta.

From the above discussion, it is evident that in field situations, sediment transport is not necessarily a function of the grain size and wind velocity alone. Rather, complicating surface conditions can also affect the amount of sediment transported at any given wind velocity. The cohesive forces of surface moisture and surface salts may explain a great percentage of the variance in the simple bivariate relationship between flow rate and shear velocity (Fig. 4.4). Klovan (1966) argues that most graphs and diagrams of this type fail to describe field

TABLE 4.3

MOISTURE CONTENT AND SALT CONCENTRATIONS FOR THE CRESTS AND DEPRESSIONS OF SURFACE IRREGULARITIES FOR TWO WINDY DAYS

	Crests		Depression	
	Moisture Content (%)	Salt Concentration (m.e./100g of soil)	Moisture Content (%)	Salt Concentration (m.e./100g of soil)
May 21	2.46	38.2	5.7	24.4
	1.71	27.4	5.3	19.6
	2.57	41.7	6.6	27.4
	1.88	29.0	4.8	20.1
July 7	2.04	59.6	4.16	35.3
	1.95	60.1	5.04	41.5
	1.47	47.8	5.84	38.4
	2.13	55.7	4.27	39.6

situations adequately because they attempt to express a complex multi-dimensional situation in only two dimensions. Thus, the flow rate at any given time can only be predicted with a high degree of confidence if all known variables are accounted for.

Shear velocity, surface moisture content and surface salt concentration, when log transformed, are all linearly related to the log of the flow rate. This would indicate that all three variables influence the flow rate simultaneously and that the combined effect is algebraically additive. A convenient technique for investigating such a relationship is Multiple Linear Regression by the method of least squares.

Multiple Linear Regression

In the case of two linearly related variables, the relationship can be expressed by a best fit line in two dimensional space, which has the equation:

$$y = a + bx + e \qquad \text{..............4.10}$$

where y = the dependent variable

x = the regression coefficient or weight, which is a measure of the slope

a = the intercept on the y axis

e = random error term

Similarly, when a single dependent variable is influenced by two or more independent variables, the relationship can be represented by a general equation:

$$y = a + b_1x_1 + b_2x_2 \ldots\ldots b_nx_n + e \ldots\ldots\ldots\ldots\ldots 4.11$$

which locates a best fit hyperplane in n dimensional space. The b coefficients in the above equation are similar to the b's in a simple linear relationship and are called partial regression coefficients. The coefficients indicate the average increase in y resulting from a unit increase in that particular independent holding all other independent variables constant. The partial regression coefficients however, take into account not only the variation in the dependent variable (y), which is directly due to that particular independent, but also the variation which is due to the other dependents correlated with the particular dependent (Kerlinger and Pedhazur, 1973). In general, these regression coefficients differ from the total regression coefficients which are the simple regressions of each individual x on y.

The size of the partial regression coefficient is also a function of the units of the associated independent variable. In order to directly compare the partial regression coefficients, the dependent and independent variables can be converted into units of standard deviations.

The new standardized partial regression coefficients are termed beta coefficients (β) and can be used to help rank the importance of the individual independent variables.

The simple linear correlation coefficient, r, is a measure of the strength of a linear relationship between two variables. The statistic indicates the goodness of the fit of a line in two dimensional space fitted by the method of least squares. In the case of more than two independent variables, a similar statistic, the coefficient of multiple correlation, R, can be used as a measure of the degree of relationship between the combined effect of the independent variables and simple dependent. In this case, the coefficient measures the closeness of fit of a hyperplane in N dimensional space to the observed variables.

The square of the multiple correlation coefficient is termed the coefficient of multiple determination and shows that proportion of the variance in the dependent variable that can be explained by the concommitant variation in the values of the independent variables.

It should be noted that the coefficient of multiple determination calculated by least squares is positively biased and therefore tends to overestimate the true R^2 value. This results from the fact that the method of least squares does not account for the variation resulting from the inclusion of an increasing number of independent variables. The coefficient must therefore be adjusted for the change in the number of the degrees of freedom caused by the addition of more than one independent variable (Yamane, 1964). The adjusted coefficient of determination can be calculated from

$$\bar{R}^2_{y.x_1x_2....x_n} = 1 - \frac{S_E}{S_{yy}} \cdot \frac{n - 1}{n - k - 1} \qquad ....4.12$$

where $\bar{R}$ = adjusted coefficient of determination

n = number of observations

k = number of independent variables

S_E = unexplained sum of squares

S_{yy} = total sum of squares

The adjusted coefficients of multiple determination and regression are a measure of the combined effect of the total number of independent variables. It is often desirable to know the relative importance of each independent taken separately while simultaneously allowing for the effects of the other independents.

This can be done by the use of partial correlation coefficients. The partial correlations measure the importance of each independent variable in the relationship by determining how much the particular independent reduces the variation in the dependent after all other variables are taken into account (i.e. held constant) (Ezekiel and Fox, 1959).

The classical multiple regression model by least squares is based on the assumption that all independent variables are represented by known fixed values, measured without error. Also, the dependent variable must be a random normal variable. In the case of this study, the log transformed dependent variable (flow rate) and the independents (shear velocity, surface moisture content and surface salt concentration) satisfy the above basic assumptions.

In order to make valid tests of the significance of the estimated parameters, the error term (i.e. residuals) must also conform to several strong assumptions. If the error term does not fulfill the assumptions, the estimates may be biased and valid statements about their significance can not be made (Davis, 1973).

It is assumed that the error term (residual) is a normally distributed random variable, with a mean of zero and constant variance. Also, the error term should not be autocorrelated and must be independent of the independent variables (Davis, 1973). This is necessary because of the problem of multicollinearity, which can be present. Multicollinearity occurs when the independent variables are not mutually statistically independent (Yamane, 1964). If an independent variable is not independent of the other variables, it can affect the value and reliability of the coefficients of the other independents.

The log transformed values of the mean daily flow rates (q) were run in a multiple linear regression against the log transformed values of the mean daily shear velocities (U_*'), surface moisture contents and surface salt concentration (Table 4.4). In standardized linear form the calculated regression model can be written as

$$\log q = 0.4366U_*' - 0.3876 \log M - 0.2301 \log S \qquad 4.13$$

where q = mean daily sediment flow rate

U_*' = mean daily shear velocity

M = mean daily surface moisture content

S = mean daily surface salt concentration

TABLE 4.4

MULTIPLE LINEAR REGRESSION RESULTS
FOR THE MEAN DAILY OBSERVATIONS

Dependent Variable: Mean Daily Saltation-Creep Flow Rate (q)

Variable	Coefficient	Beta Coefficient	Calculated F Value
Constant	1.2307		
Mean Daily Shear Velocity	1.9227	0.4366	17.01*
Mean Daily Surface Moisture Content	-1.3127	-0.3876	13.82*
Mean Daily Surface Salt Concentration	-0.9440	-0.2301	4.62*

Standard Error of the Estimate	0.3932
Coefficient of Determination	0.7069
Multiple Correlation Coefficient	0.8408
Adjusted Coefficient of Determination	0.6825
Adjusted Multiple Correlation Coefficient	0.8261

Analysis of Variance for the Regression

Source	Degrees of Freedom	Sum of Squares	Mean Squares	F
Regression	3	13.4227	4.4742	28.94*
Error	36	5.5654	0.1546	
Total	39	18.9881		

Durbin-Watson d Statistic 1.76**

**Significant at the 0.05 confidence level
*0.01 confidence level

or in power form

$$q = U_*'^{0.4366} . M^{-0.3876} . S^{-0.2301} \qquad \ldots\ldots\ldots\ldots\ldots 4.14$$

It is evident from Table 4.4 that the above regression model is significant at the one per cent confidence level (F = 28.94 >F for one per cent) and explains approximately 68 per cent of the observed variance in the sediment flow rate ($\bar{R}^2 = 0.6825$). It can also be seen from the F values for each of the independent variables that shear velocity, surface moisture content and surface salt concentration all contribute significantly to the explanation of the dependent variable (flow rate).

The absolute values of the beta coefficients can be used to rank the relative importance of each of the independent variables in the regression model. As might be expected, shear velocity ($\beta = 0.4366$) appears to be the most important of the three measured variables followed by surface moisture content ($\beta = -0.3876$) and surface salt concentration ($\beta = -0.2301$) respectively.

Although the above model is significant, its validity is dependent on the fulfillment of the assumptions of multiple linear regression. The residuals of the regression were checked for homoscedasticity (i.e. constant variance) by running a simple linear regression between the residuals and each of the independent variables. In all three cases the slopes of the regressions were not significant at the 95 per cent confidence level, indicating that the residuals do have constant variance (Davis, 1973).

The possibility that the residuals were autocorrelated was also checked by using the Durbin-Watson d statistic (Yamane, 1964).

Autocorrelation in the context of multiple regression means that the residuals tend to occur in "groups" of adjacent deviations around the regression plane and may indicate that the regression model is inappropriate (Davis, 1973). Since the calculated d statistic for the flow rate regression is greater than the critical upper limit ($d = 1.7639 > d_u$ at one per cent) it can be assumed that the residuals of the regression are not autocorrelated.

It can be seen from the simple correlation matrix (Table 4.2) that a certain degree of multicollinearity does exist between the independent variables. This is most pronounced in the relationship between surface moisture and surface salt concentration. This is not unexpected because the salt is transported to the surface by capillary water. Thus, it can be seen that the salt concentration at any given time may well be at least in part a function of the moisture content at that moment.

The degree of multicollinearity is important in the interpretation and use of the regression model because it can seriously affect the precision of the partial regression coefficients of the related independents. The degree of multicollinearity which is acceptable in a given regression has been discussed by various authors (Huang and Bolch, 1974; Yamane, 1964; and Klein, et al, 1967). Klein, et al (1967) suggests that the multicollinearity is tolerable if the simple regression coefficient of the related independents is considerably less than the multiple regression coefficient.

However, Huang and Bolch (1974) also suggests that one must consider the strength of R. They argue that the effects of multicollinearity increase the closer R is to zero, even if r_{ij} R. No critical limits for R or r_{ij} can be set and multicollinearity should be avoided if possible. They conclude, however, that the effect of multicollinearity is rather small when R is large and r_{ij} small.

It can be seen from Table 4.2 that the relationships between the independent variables are just significant at the 0.05 confidence level ($r_{critical\ 0.05} = 0.3980$). Since the correlation coefficients are small, it is suggested that the effect of multicollinearity in the multiple regression model is also small and therefore does not seriously affect the partial regression coefficients of the independent variables.

Conclusions

The multiple regression model (equation 4.14) does appear to satisfy the assumptions of multiple linear regression and explain a relatively high proportion of the total observed variance ($\bar{R}^2 = 0.68$). It should be noted, however, that 31 per cent of the total observed variance is unexplained by the above model. It is suggested that a large proportion of this variance is related to other independent variables not included in the model.

One of the most important of these may be the effect of surface irregularities which can alter the amount of sediment transported in creep. Variations resulting from surface irregularities would be extremely difficult to account for in field situations like those encountered in the Slims River delta. Two reasons for this can be put forward. First, the surface irregularities on the delta surface change almost continuously. Thus, within a few minutes under strong wind conditions, irregularities which were hindering flow may themselves be removed by wind erosions, while others may be formed. Secondly, the surface irregularities may not be uniformly distributed over a given surface. Thus, the degree to which the flow rate is affected by the surface irregularities will in part be a function of the direction from which the sediment is being transported. If the direction of the wind fluctuates, material being transported at any given time may well be transported over surfaces offering varying degrees of interference to the flow. Problems of this nature could only be overcome if the variability of the wind and the exact nature and change in surface irregularities was known for the sampling. Complex problems of this nature would be extremely difficult to assess in field situations.

Further field investigations coupled with wind tunnel experiments may help to overcome the unexplained variance related to the variability in surface conditions.

CHAPTER V

SUSPENDED SEDIMENT TRANSPORT

Introduction

The transport of suspended sediment by the wind was measured in the Slims River Valley during the summer months of 1972 and 1973. During the 1972 field season very few dust storms were generated in the valley as a direct result of the high frequency of rainfall (28 of the 49 observation days had a measureable amount of precipitation, Table 1.4). Consequently, only five dust storms of sufficient size and duration could be sampled during the 1972 field season.

Suspended sediment was collected at eight heights between 0.5 and 12.0 metres by modified impinger tube samplers (see page 46). The data showed that this type of sampler was inadequate for field investigations of this nature because the sampler's size and shape creates a great deal of air turbulence around the intake nozzle. The sampler design results in an irregular underestimation of the suspended sediment concentration because less material enters the nozzle than is in the air stream. As a result of the suspected underestimation of the suspended sediment concentrations the data collected during the 1972 field season has not been included in the following analysis.

During the 1973 field season modified millipore sampling cassettes (Fig. 2.7) were used to sample suspended sediment during the dust storms. This type of sampler proved to be much more satisfactory than the impinger tube technique.

The frequency of dust storms was much higher during the 1973 field season than the previous summer. In total, fifteen relatively large dust storms occurred and were sampled between May 12 and July 9, 1973. The greater number of storms would again appear to be at least in part related to the frequency and amount of rainfall (Table 5.1). Nine of the 15 storms occurred between May 18 and June 10. In this 24-day period only 3 days had a measurable amount of precipitation. The remaining six dust storms occurred in the dry intervals between periods of extended rainfall (Table 5.1).

The occurrence of dust storms during the two field seasons demonstrates the close relationship between the frequency and intensity of rainfall and the frequency of dust storms. From the 1972 data it would appear that very small amounts of rainfall at frequent intervals accompanied by cool daytime temperatures effectively decreases the number of dust storms in the Slims River Valley. As a result of the delta surface being almost continually moist the wind is unable to effectively erode the surface sediments. The close dependence on even small changes in surface moisture content is shown in the daily pattern of sediment transport. Very little transport occurs in saltation creep or suspension in the early morning, although winds may be relatively high. It is thought that this lack of sediment transport results from an increase in surface moisture content during the night because of lower night time evaporation rates and migration of water towards the surface by capillarity. During the mid-morning when air temperatures

TABLE 5.1

TEMPERATURE AND PRECIPITATION DATA
FOR THE SLIMS RIVER VALLEY

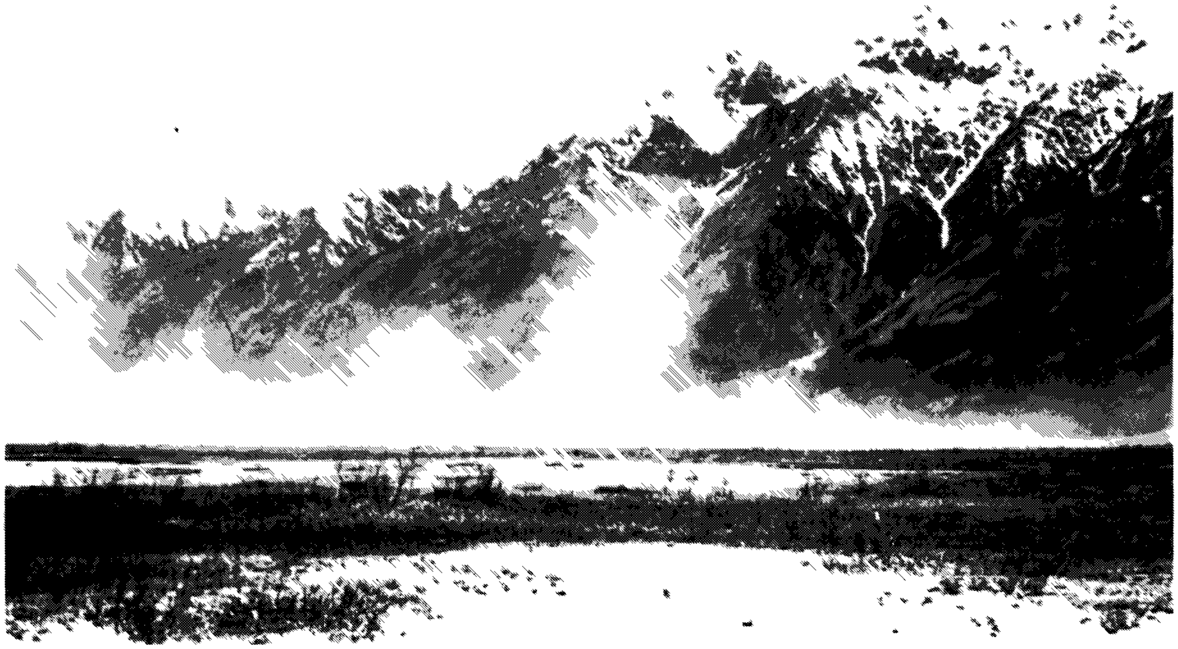
Date (1973)	Max (°C)	Min (°C)	Precip (mm)	Dust Storm No.	Date (1973)	Max (°C)	Min (°C)	Precip (mm)	Dust Storm No.
May 12	10.0	-2.2			June 11	10.0	5.6		
13	11.2	5.7			12	16.8	7.8	4.1	
14	20.0	9.4			13	19.5	6.1	1.8	
15	14.8	1.1			14	16.1	4.4	0.7	
16	13.3	3.2	2.0		15	18.5	7.7		
17	12.3	1.5	3.8		16	14.6	1.1		
18	11.0	1.2		1	17	16.2	5.6	4.8	
19	12.7	0.6		2	18	12.2	4.0	1.3	
20	8.8	-0.5			19	10.7	2.2	2.8	
21	9.9	-1.1			20	16.1	5.4	9.7	
22	11.2	0.0			21	16.0	4.8	0.5	
23	8.7	2.2			22	20.2	7.9		10
24	10.1	0.1	2.3		23	16.6	8.4		11
25	10.9	1.1	4.1		24	17.8	6.7	T	
26	12.6	0.2		3	25	16.8	4.4		12
27	10.0	0.1			26	16.7	5.4		
28	10.8	0.5		4	27	16.3	4.9		
29	12.0	1.2		5	28	12.1	6.7	11.8	
30	17.0	3.5			29	19.2	7.6		
31	14.6	3.4		6	30	18.8	5.4		13
June 1	10.7	1.6		7	July 1	17.8	5.6		14
2	13.3	3.0			2	19.0	7.1		15
3	12.2	3.2	0.9		3	18.0	3.8		
4	11.2	1.2			4	18.4	7.7		
5	14.4	3.7	T		5	20.2	7.9	T	
6	12.3	2.6			6	20.1	7.8		
7	12.2	4.1			7	19.4	7.7		
8	13.4	3.3			8	22.6	8.8		
9	13.3	2.2		8	9	21.2	8.3		
10	16.8	5.4		9					

and evaporation rates are increasing, the surface of the delta begins to dry, resulting in an increase in the amount of sediment transported in saltation and creep. By mid-afternoon some areas of the delta may become sufficiently dry that fine material is picked up into suspension to form small plumes or "dust devils" (Fig. 5.1). The cyclic moistening and drying of the delta surface can be seen from Fig. 5.2, showing moisture contents determined approximately every four hours from samples collected at two of the grid intersections for 3 continuous days. Although the changes in moisture content are somewhat irregular they do indicate a cyclic trend.

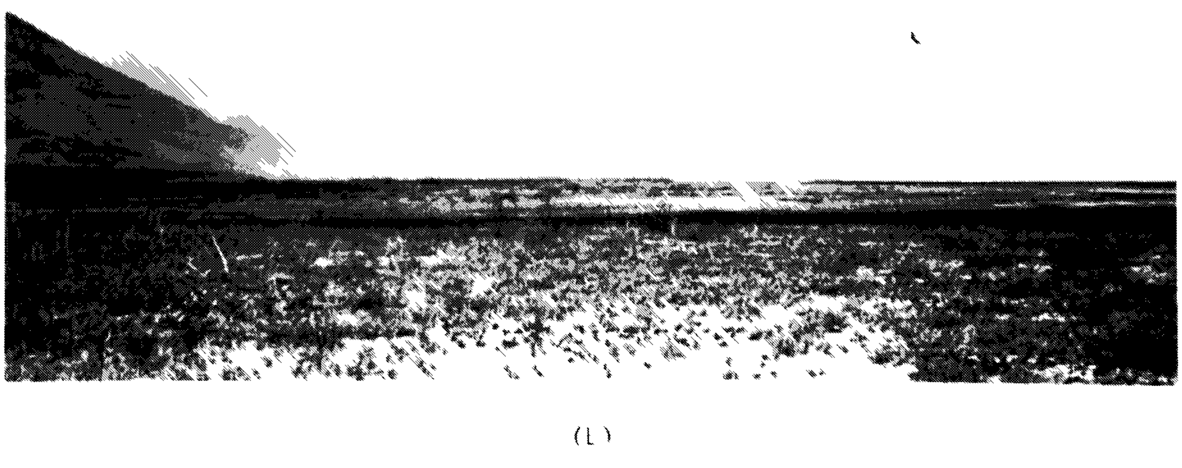
Even with extremely dry surface conditions large quantities of suspended sediment may not be transported in the Slims River Valley because of other controlling factors such as wind velocity, air turbulence, and the presence of surface salts.

Suspended Sediment Flow Rate

Mean suspended sediment concentration at each sampling height for the fifteen dust storms are shown in Fig. 5.3. The suspended sediment concentration for each height was calculated by dividing the total amount of sediment caught during the sampling period by the volume of air evacuated through the filtering cassette. These figures show that a relatively strong linear relationship exists between the logarithm of concentration and the logarithm of height. Similar results have been reported by Chepil and Woodruff (1957) from measurements at four heights below 5 metres during dust storms in Colorado and Kansas. Chepil and Woodruff were primarily concerned with the effects



(a)



(b)

DUST DEVILS RISING FROM THE SURFACE
OF THE SLIMS RIVER DELTA

FIGURE 5.1

CHANGE OF SURFACE MOISTURE CONTENT OVER TIME

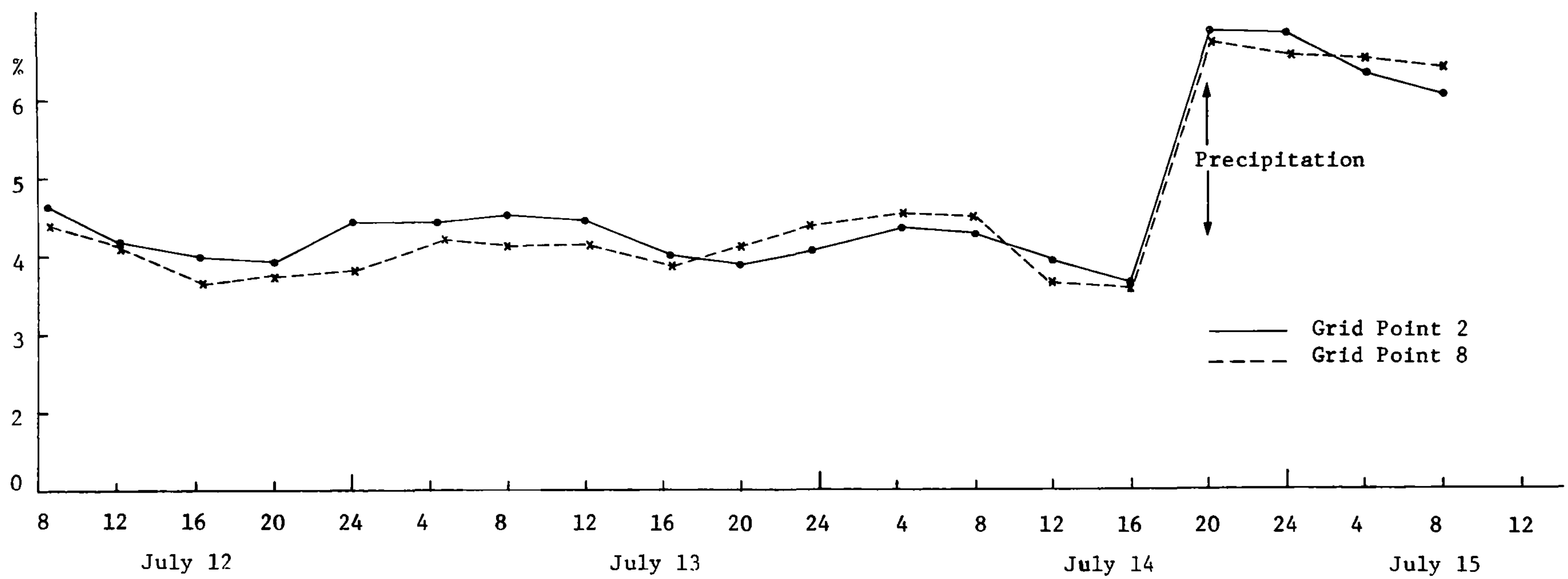


FIGURE 5.2

SUSPENDED SEDIMENT CONCENTRATION

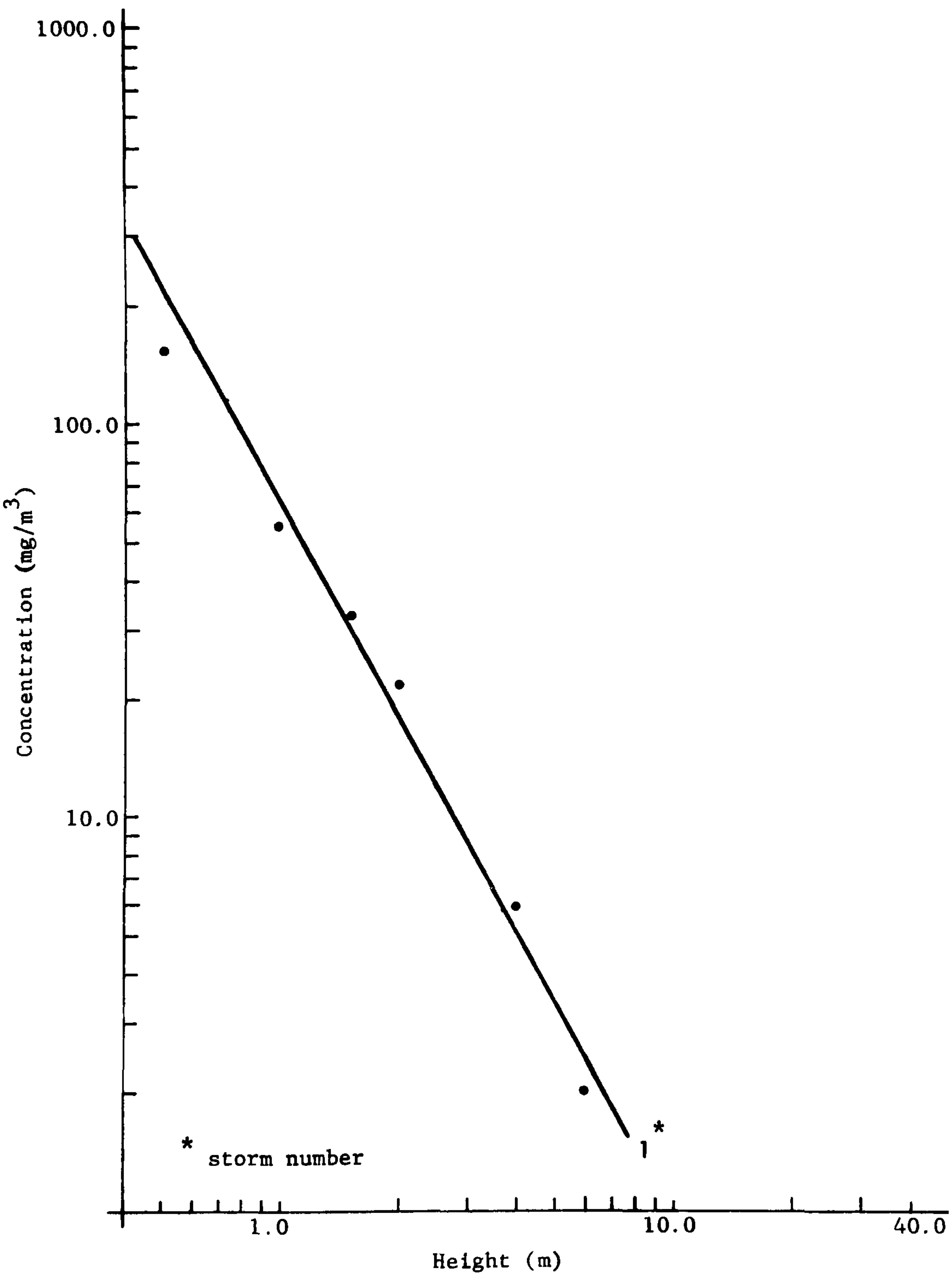


FIGURE 5.3

SUSPENDED SEDIMENT CONCENTRATION

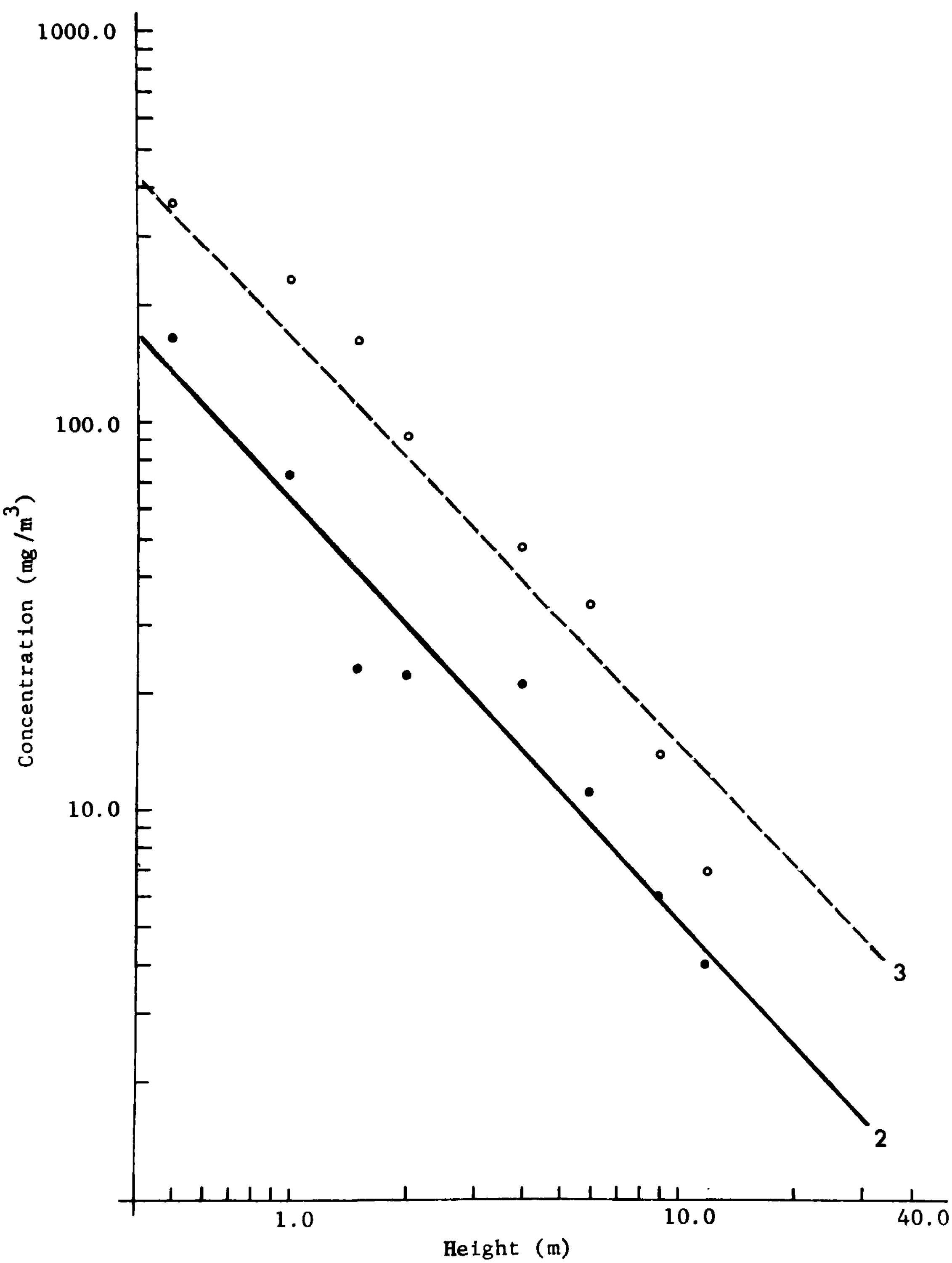


FIGURE 5.3 (continued)

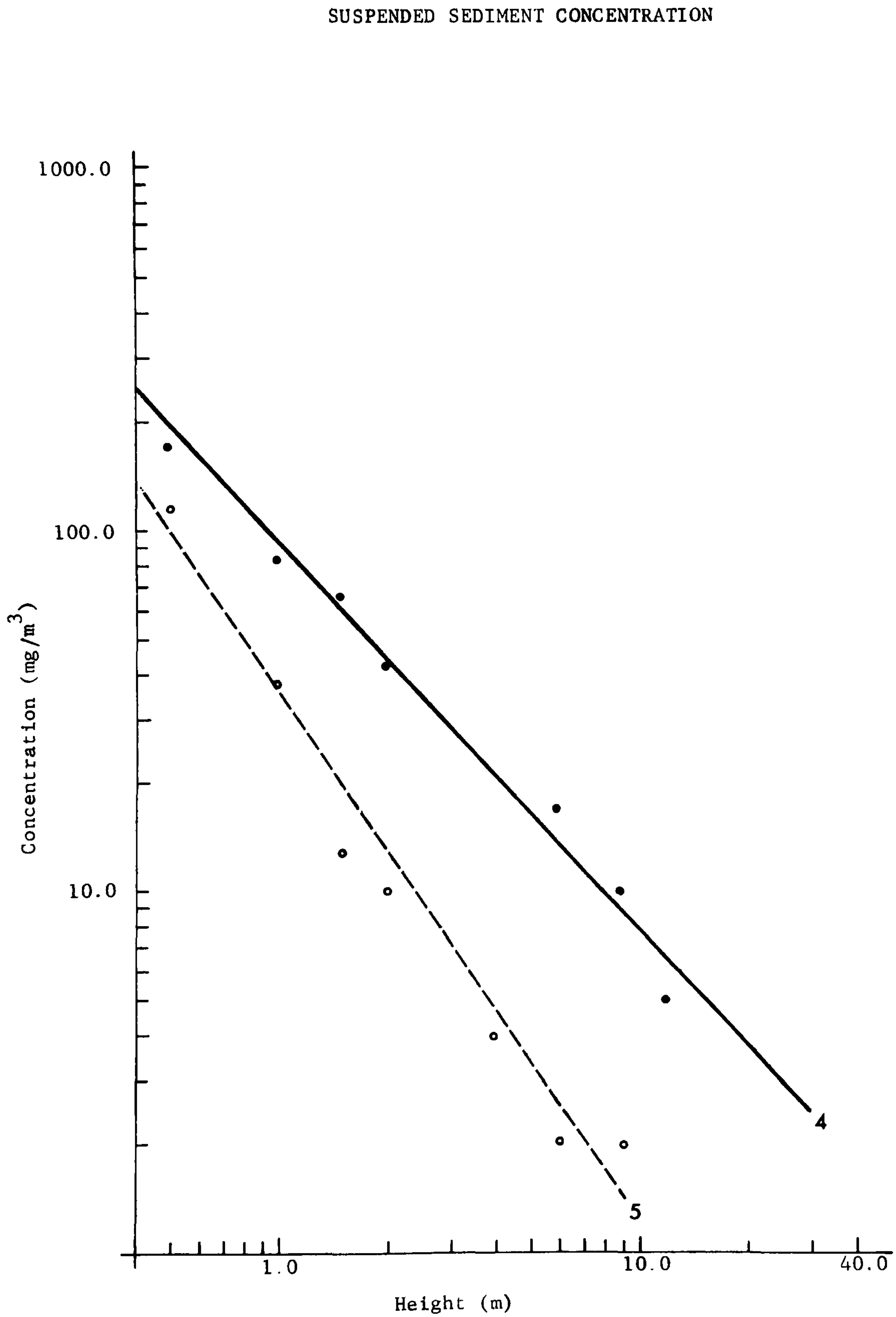


FIGURE 5.3 (continued)

SUSPENDED SEDIMENT CONCENTRATION

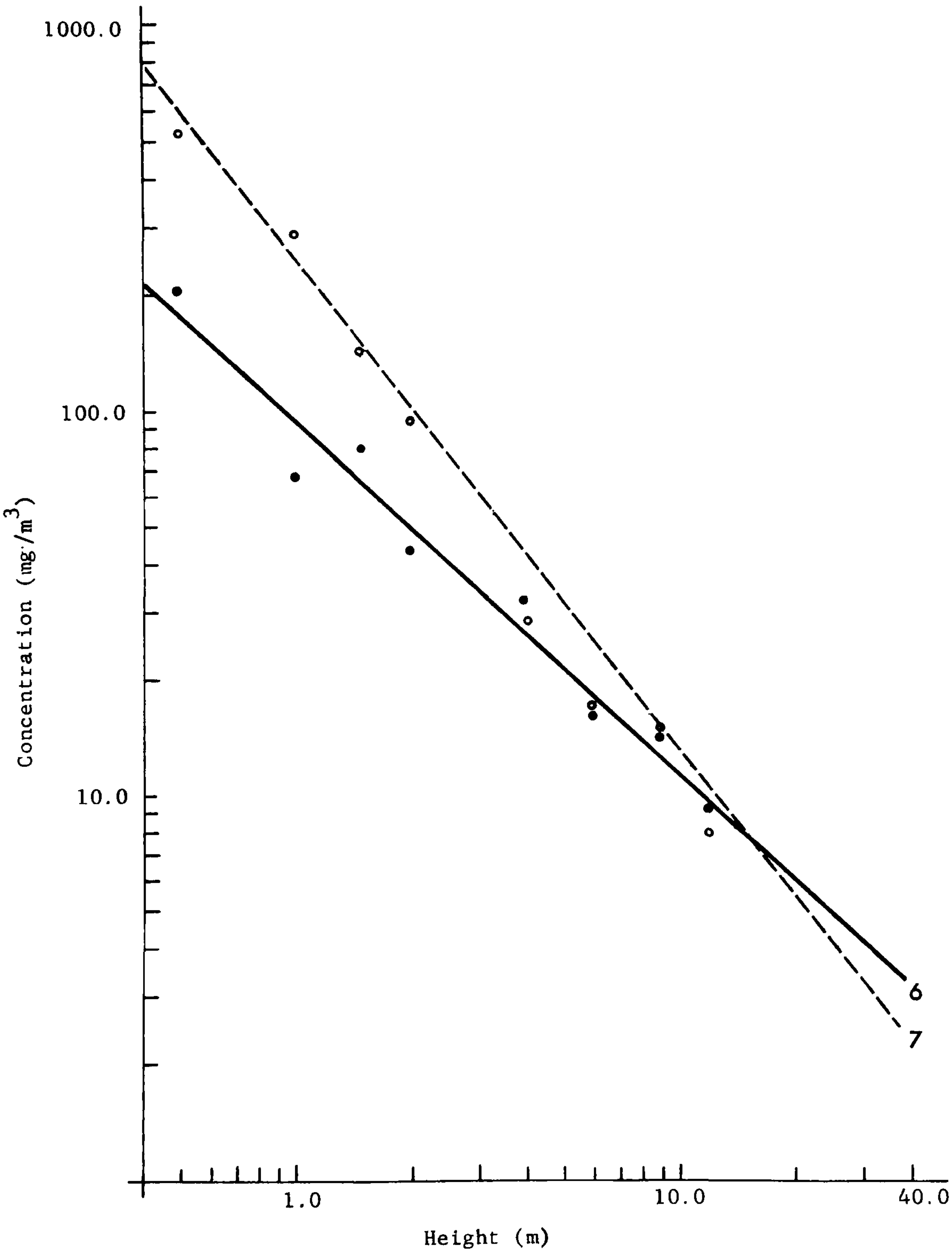


FIGURE 5.3 (continued)

SUSPENDED SEDIMENT CONCENTRATION

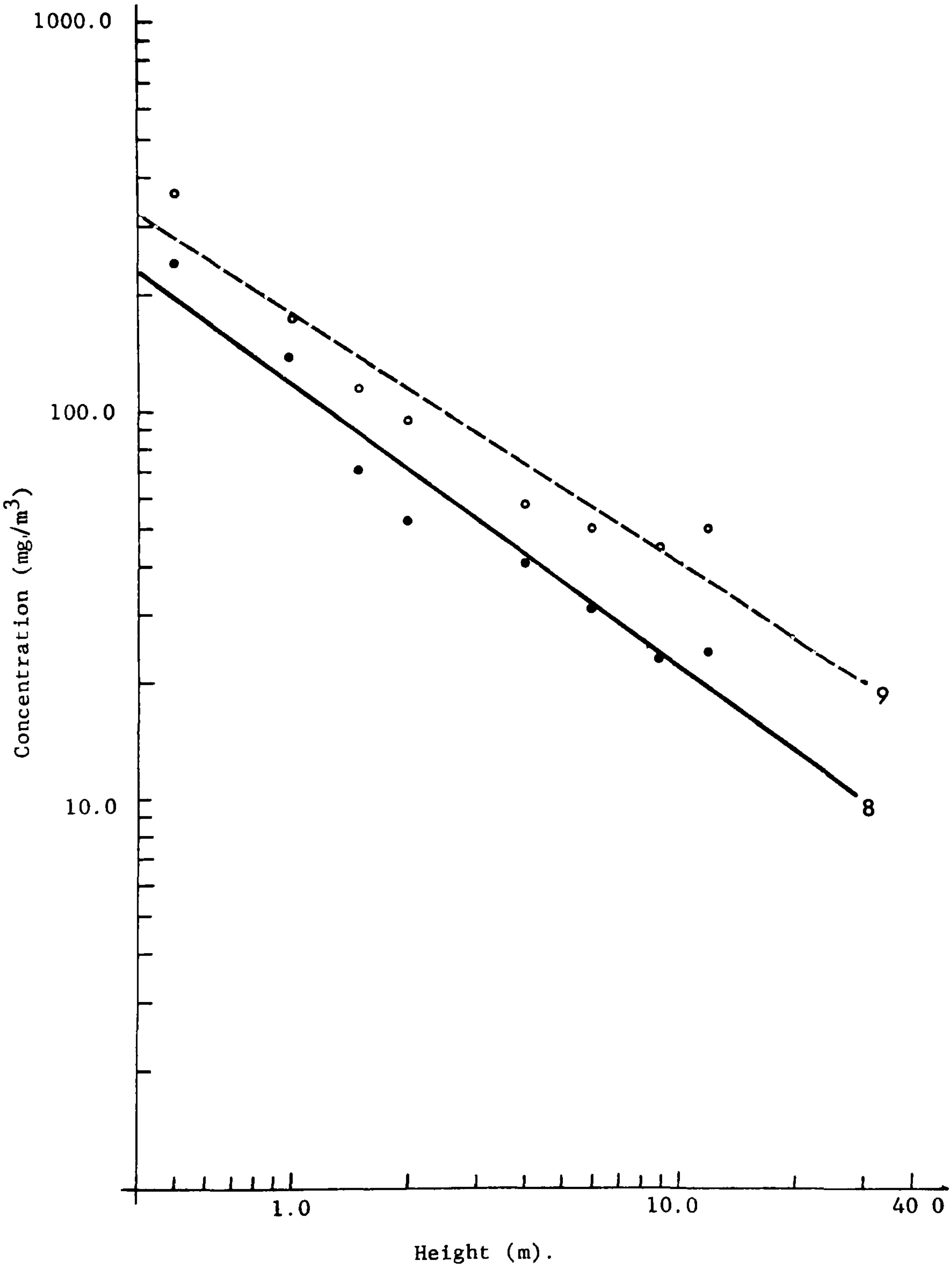


FIGURE 5.3 (continued)

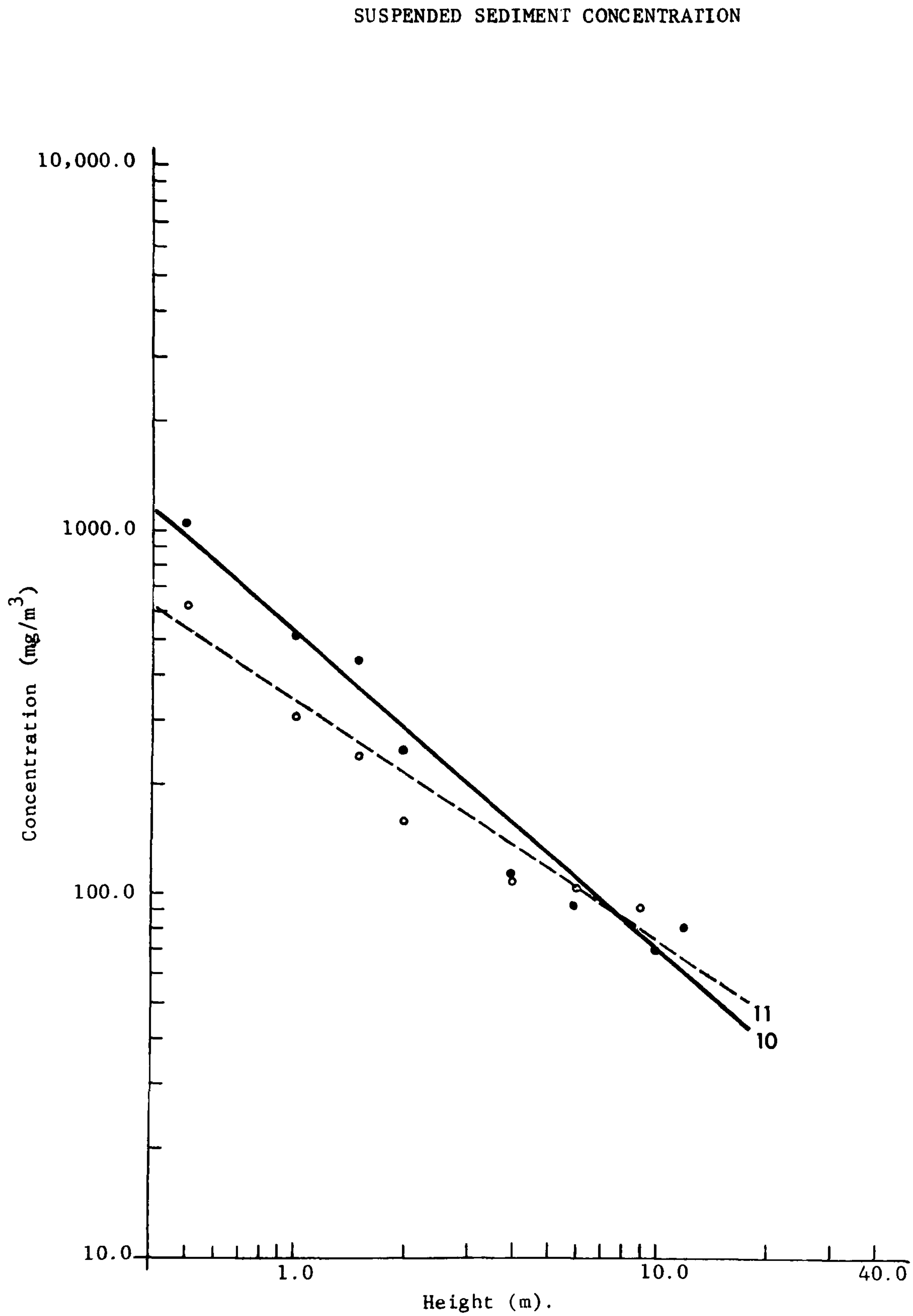


FIGURE 5.3 (continued)

SUSPENDED SEDIMENT CONCENTRATION

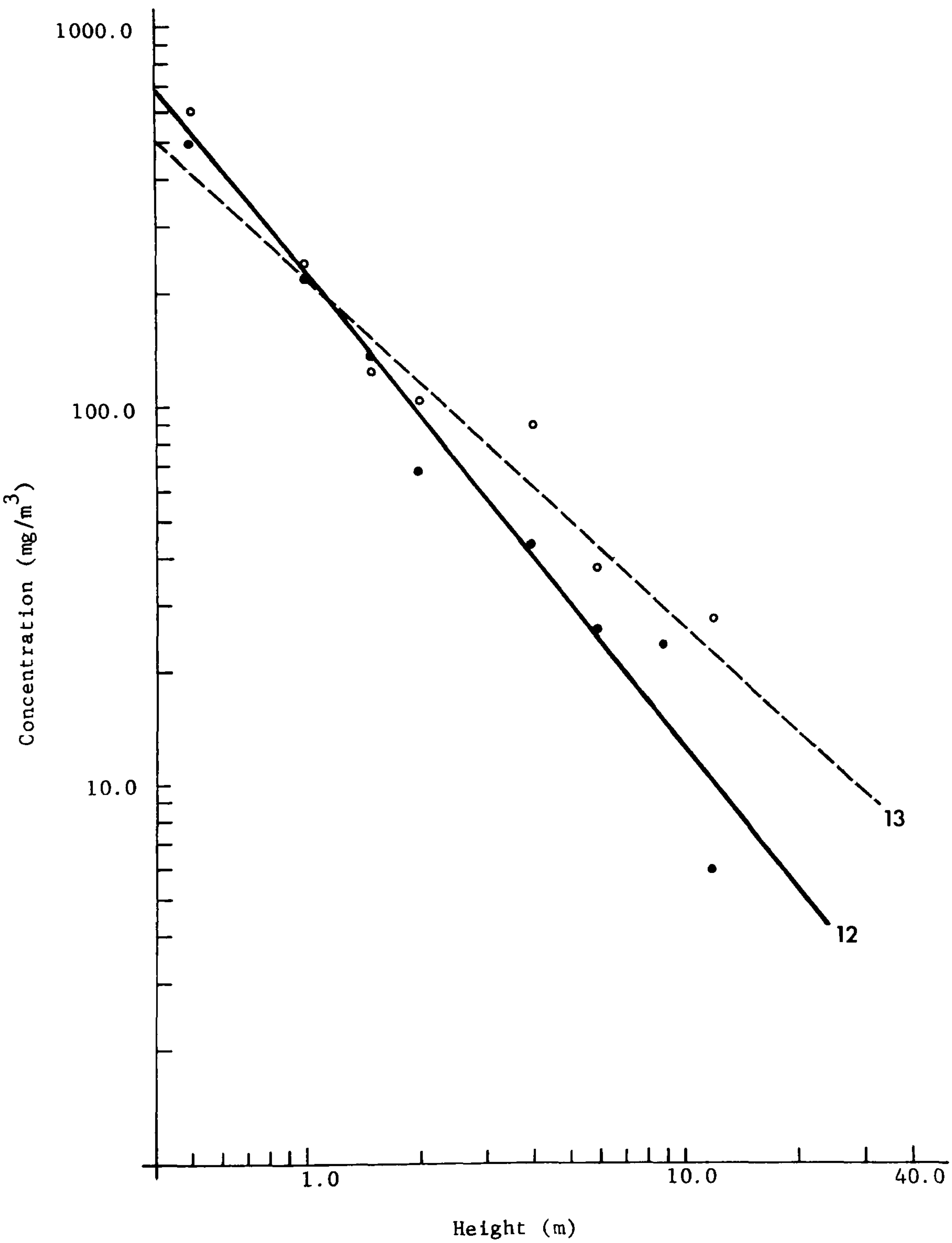


FIGURE 5.3 (continued)

SUSPENDED SEDIMENT CONCENTRATION

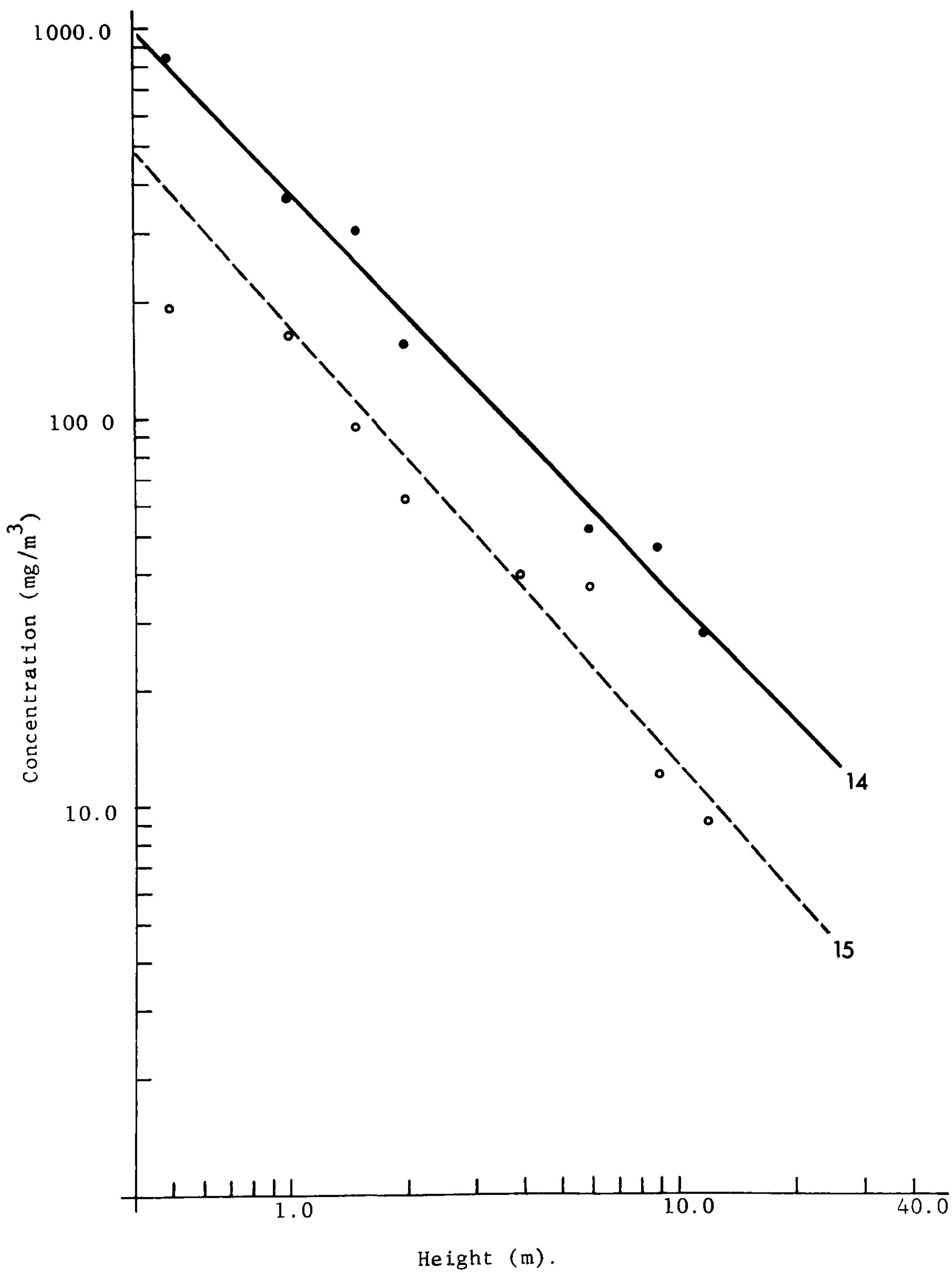


FIGURE 5.3 (continued)

of suspended sediment concentrations on visibility, and they present a mean dust concentration against height curve for the sampled storms. This curve is compared with a similar mean concentration curve from the fifteen storms in the Slims River Valley (Fig. 5.4). On average, suspended sediment concentrations below 0.5 metres were considerably higher and decreased more rapidly with height in the Slims River Valley storms compared with Chepil and Woodruff's results. Two possible explanations for the dissimilarities can be given.

Suspended sediment concentrations near the surface in part reflect the ability of the surface to supply sediment to the air streams. At any given wind velocity the near surface sediment concentrations will be a function of the grain size distribution of the surface sediments and the presence or absence of bonding agents which tend to stabilize the surface. As a result of spatial and temporal variations in surface conditions it is not surprising that samples collected in two different geographical locations have different near surface suspended sediment concentrations. This is demonstrated even further by the storm to storm variation in the Slims River Valley.

Near surface concentrations are also strongly dependent on wind velocity near the surface. It can be seen from the mean wind profiles (Fig. 5.4b) that average wind velocities over the surface (above 0.25 metres) were greater in the Kansas and Colorado dust storms, despite the fact that the mean suspended sediment concentrations were lower close to the surface (below 0.5 metres). This indicates that the surface of the Slims River delta possibly supplies more sediment to the air stream. The surface of the Slims River delta is comprised

MEAN SUSPENDED SEDIMENT AND WIND VELOCITY PROFILES
FOR THE SAMPLED DUST STORMS

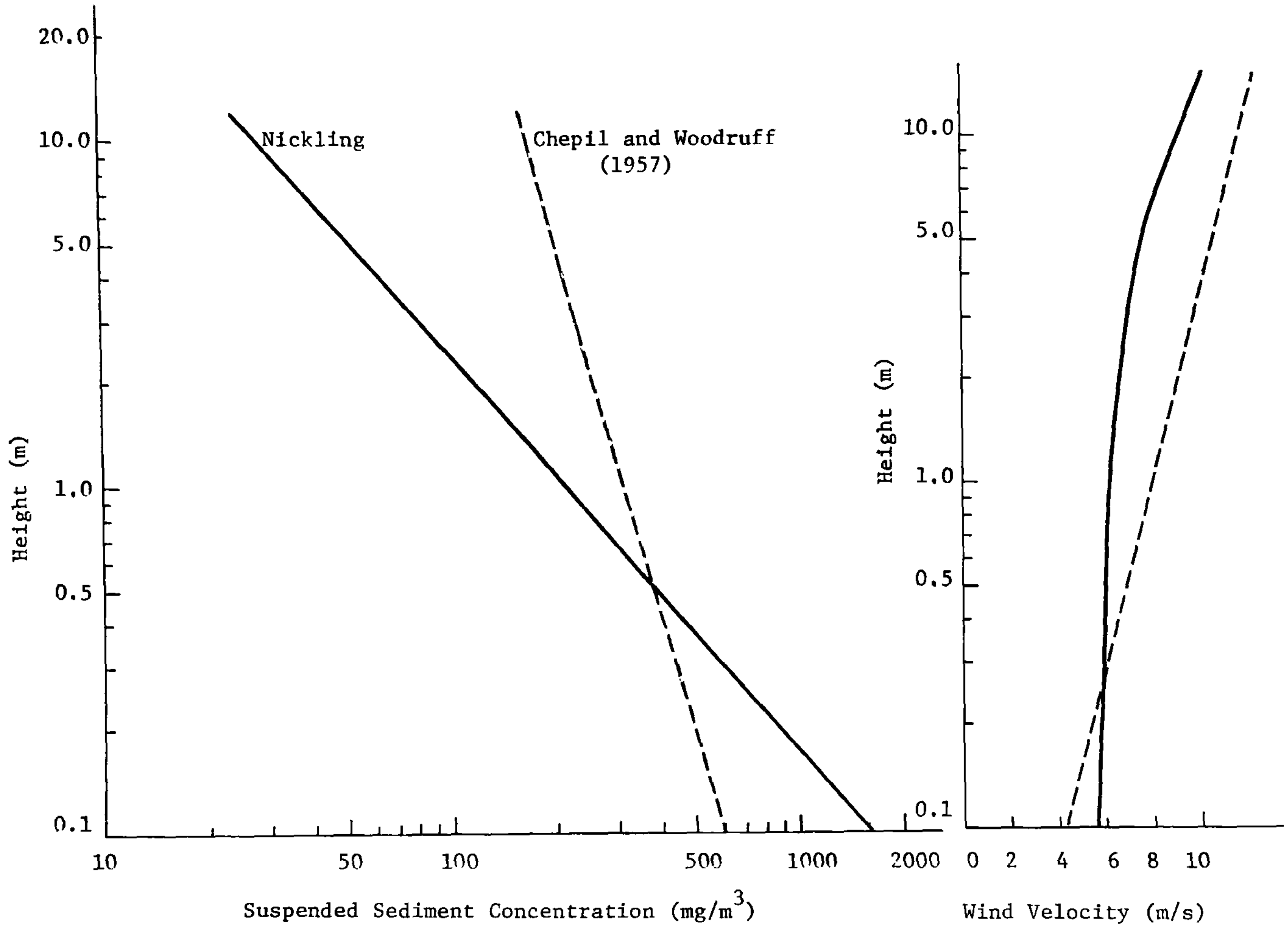


FIGURE 5.4

of dominantly fine to medium silts with a very small percentage of clay (Fig. 4.2). Thus, these sediments, on the basis of texture, lack cohesion and would tend to be more susceptible to wind erosion when dry. In comparison, the suspended material in the Kansas and Colorado dust storms was usually eroded from cultivated fields comprised of loams and clay loams. The fact that these sediments had loamy textures indicated that there was a much higher clay content which would increase the cohesion and cloddiness of the surface sediments. This in turn would reduce the amount of material which is readily available to the air stream.

Although the near surface concentration appears to be closely related to the nature of the eroding surface, the rate of decrease of concentration with height is probably more closely related to wind velocity, or more specifically, air turbulence. Fine particles once entrained into the air stream are kept aloft by turbulent eddies. These eddies tend to lift the fine particles from near the surface vertically as well as distributing them laterally (Bagnold, 1941). Thus, as turbulence increases there would be a tendency for the concentration to become more uniform with height. On the basis of this argument, it could be suggested that, on average, the air was more turbulent during the Kansas and Colorado dust storms than in the case of the Slims River Valley storms. Some evidence to support this hypothesis can be derived from a comparison of the two mean wind velocity profiles (Fig. 5.4). Mean shear velocity as indicated by the slope of the wind profile associated with the Kansas dust storms is somewhat greater than that for the Slims delta storms. Although air turbulence is not a direct function of the shear velocity, the higher shear velocity associated with the Kansas and Colorado storms may indicate a greater tendency for the air to be more turbulent in this case. On the basis of photographs from the Kansas and Colorado study area (Chepil and Woodruff, 1957), it would also appear that the surface was aerodynamically rougher than the surface of the Slims River delta. The Kansas and Colorado samples were usually taken in agricultural areas over tilled fields which would result in greater ground-air friction than in the Slims River delta situation. This higher ground-air friction would also result in greater frictional turbulence (Rose, 1966).

Instability in the atmosphere is strongly dependent on the temperature gradient. The atmosphere becomes unstable when the environmental lapse rate is greater than the dry adiabatic lapse rate (1°C/100m). Unstable superadiabatic conditions are often caused by intensive daytime heating of the lower atmosphere by the earth's surface. This intensive heating causes an increase in free convection or air turbulence (Sellers, 1965).

Although somewhat speculative in the absence of temperature data from the Kansas and Colorado study, it could be suggested that the more uniform suspended sediment concentrations in this case are a function of a greater degree of air turbulence, resulting from higher superadiabatic lapse conditions. This is reasonable if one considers the geographical location and the daily temperature regimes of the two study areas. One would suspect that on average, stronger superadiabatic conditions would be developed in Kansas and Colorado as a result of greater daytime surface heating.

Chepil and Woodruff (1957) suggest that in some cases the measured suspended sediment may not have been derived in the immediate area but rather had been transported many miles. This in itself may also be responsible for the relatively uniform distribution of suspended sediment with height. Fine particulate matter which is carried in a turbulent atmosphere over a distance would tend to become more diffused and attain more uniform concentrations as one moved downwind from the source. An analogous situation would be the emission of effluents from a smoke stack when pronounced superadiabatic temperature conditions exists. Under such conditions the effluents diffuse rapidly and erratically and become more uniformly distributed as they move downwind because of the unstability and convective mixing (Sellers, 1965).

In order to investigate the effects of surface conditions on the amount of sediment transported in suspension, the mean suspended sediment flow rate was calaulated for each height during each dust storm. (Flow rate is the product of unit suspended sediment concentration and mean wind velocity at a given height). This assumes that the velocity of the particles carried in suspension is equal to the wind velocity. Chepil (1945) has shown from wind tunnel experiments that for most practical purposes this assumption is correct.

Log-log plots of the suspended sediment flow rate against height are given in Fig. 5.5. The approximated straight line relationships can be expressed by a power function of the general form

$$F = \frac{a}{Z^b} \qquad \text{..............5.1}$$

where F = the suspended sediment flow rate (mg/cm.s) at height Z(m).

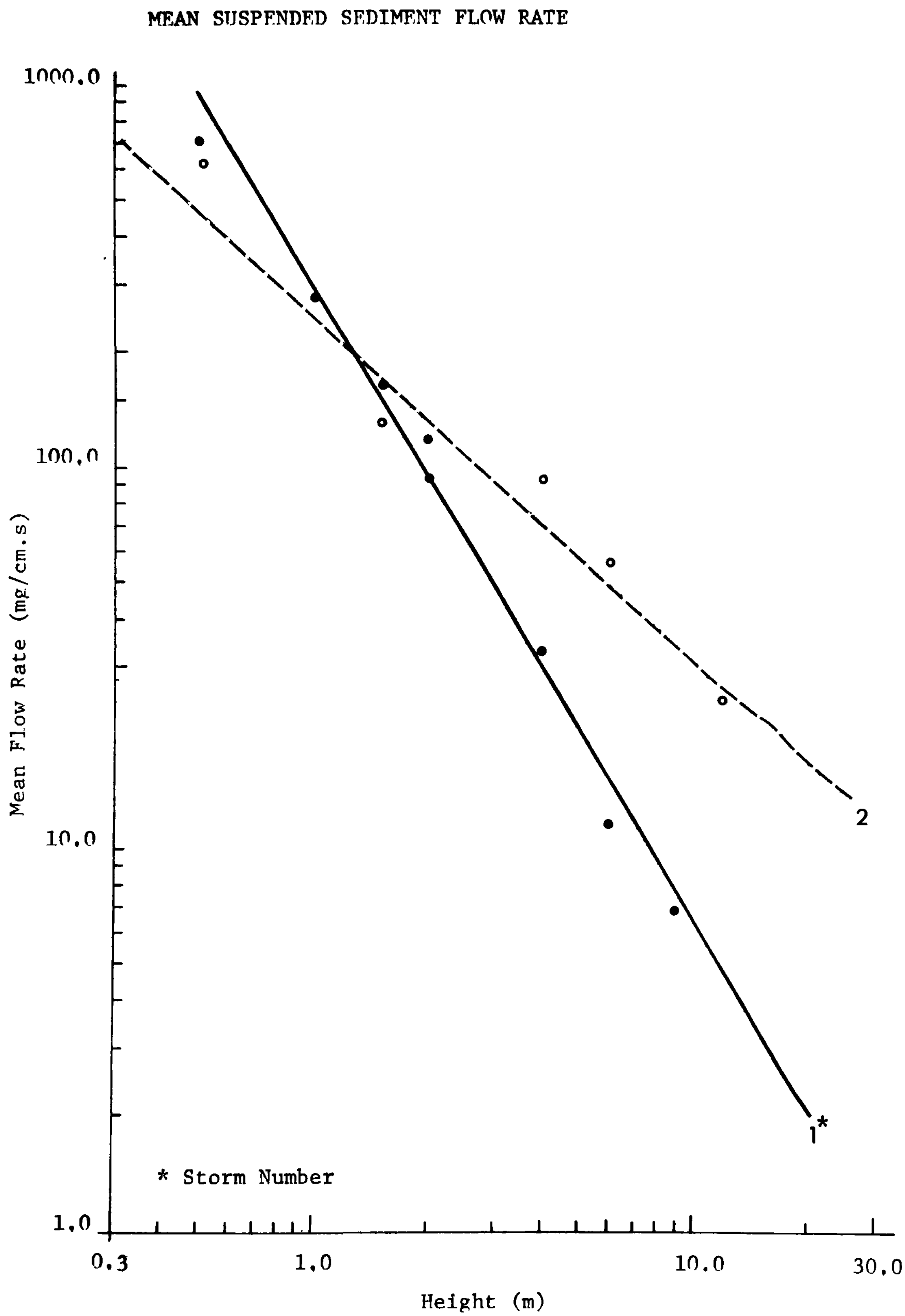

MEAN SUSPENDED SEDIMENT FLOW RATE
1000.0
100.0
10.0
1.0
Mean Flow Rate (mg/cm.s)
0.3
1.0
10.0
30.0
Height (m)
2
1*
* Storm Number

FIGURE 5.5

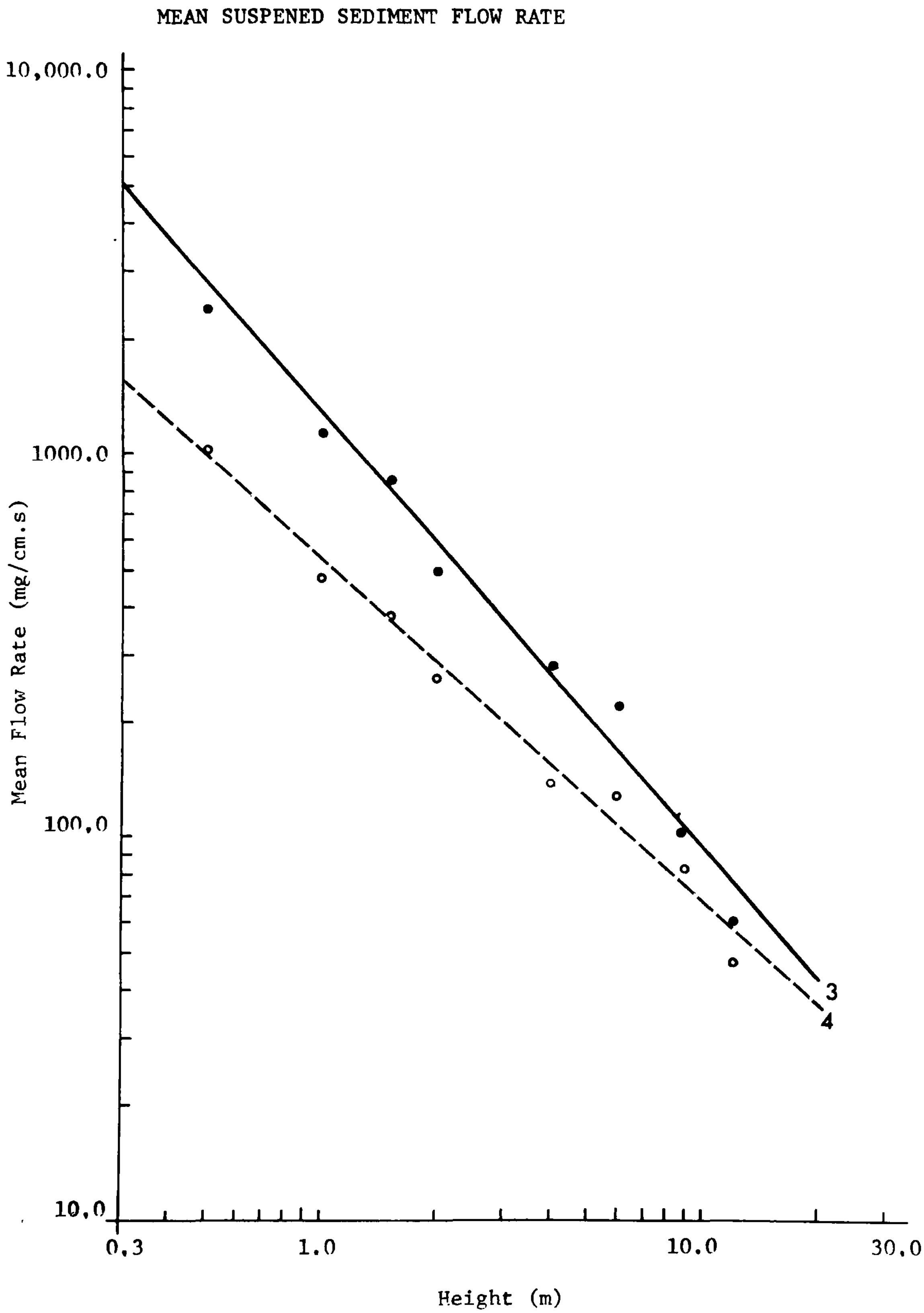


FIGURE 5.5 (continued)

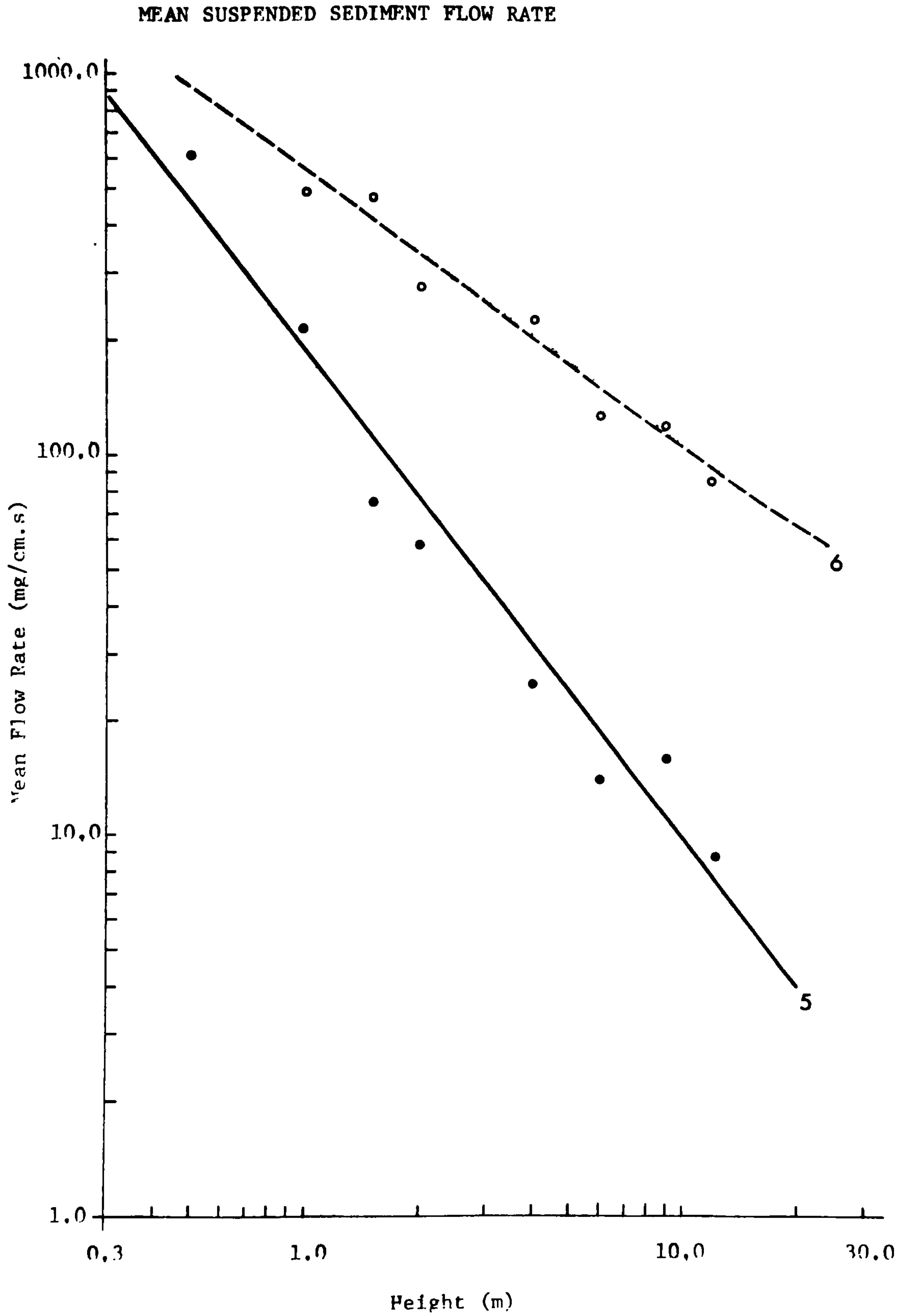


FIGURE 5.5 (continued)

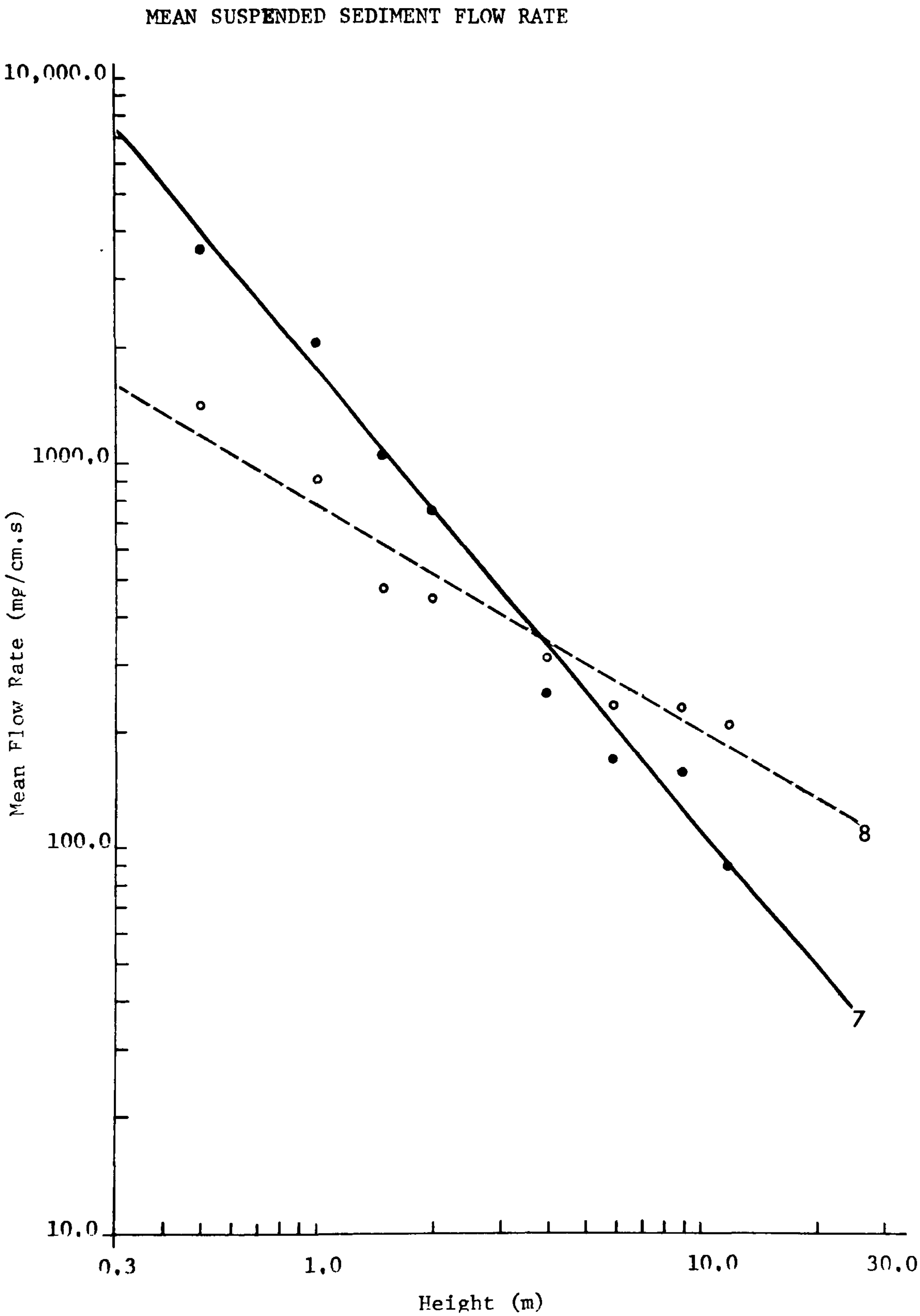


FIGURE 5.5 (continued)

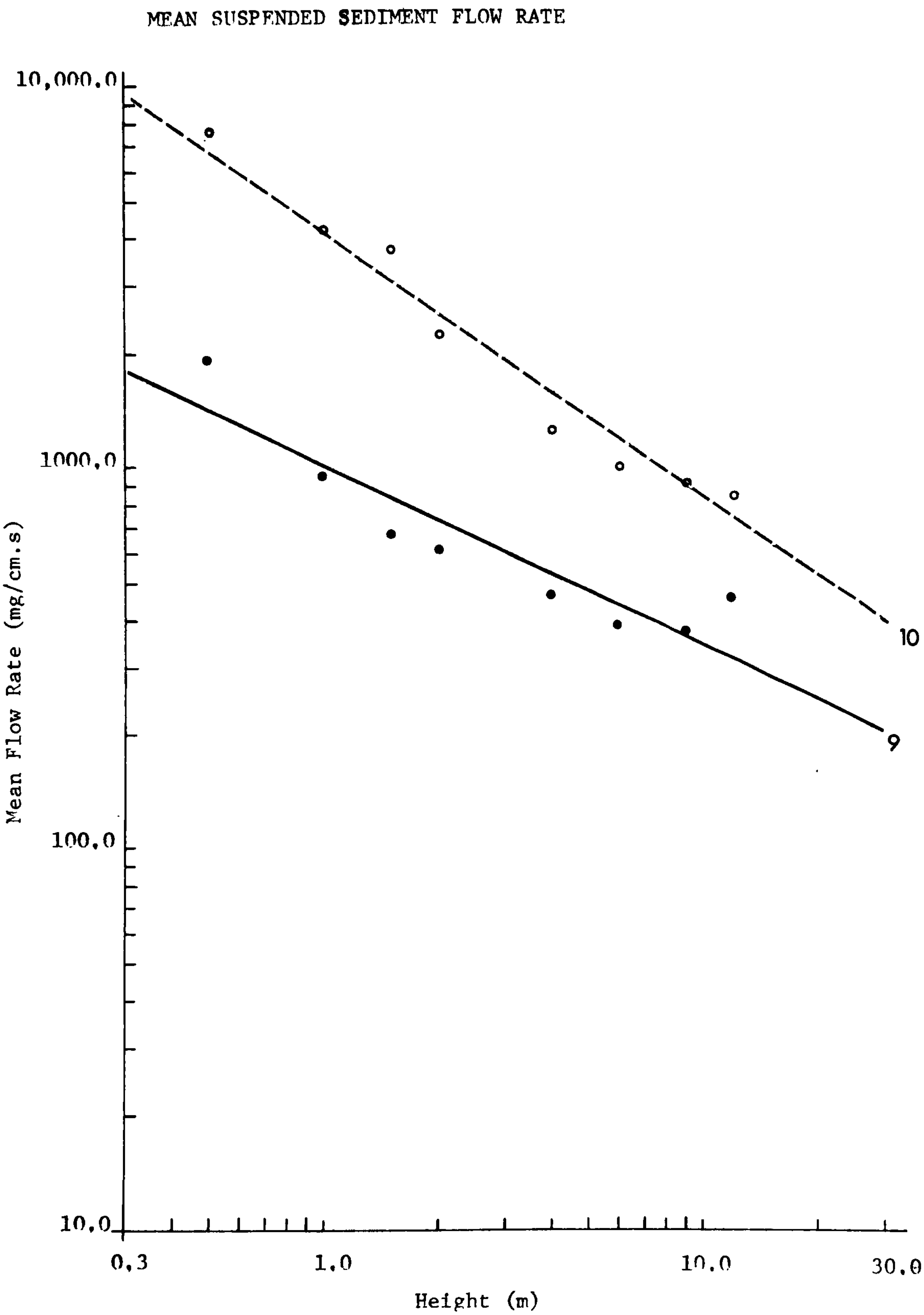


FIGURE 5.5 (continued)

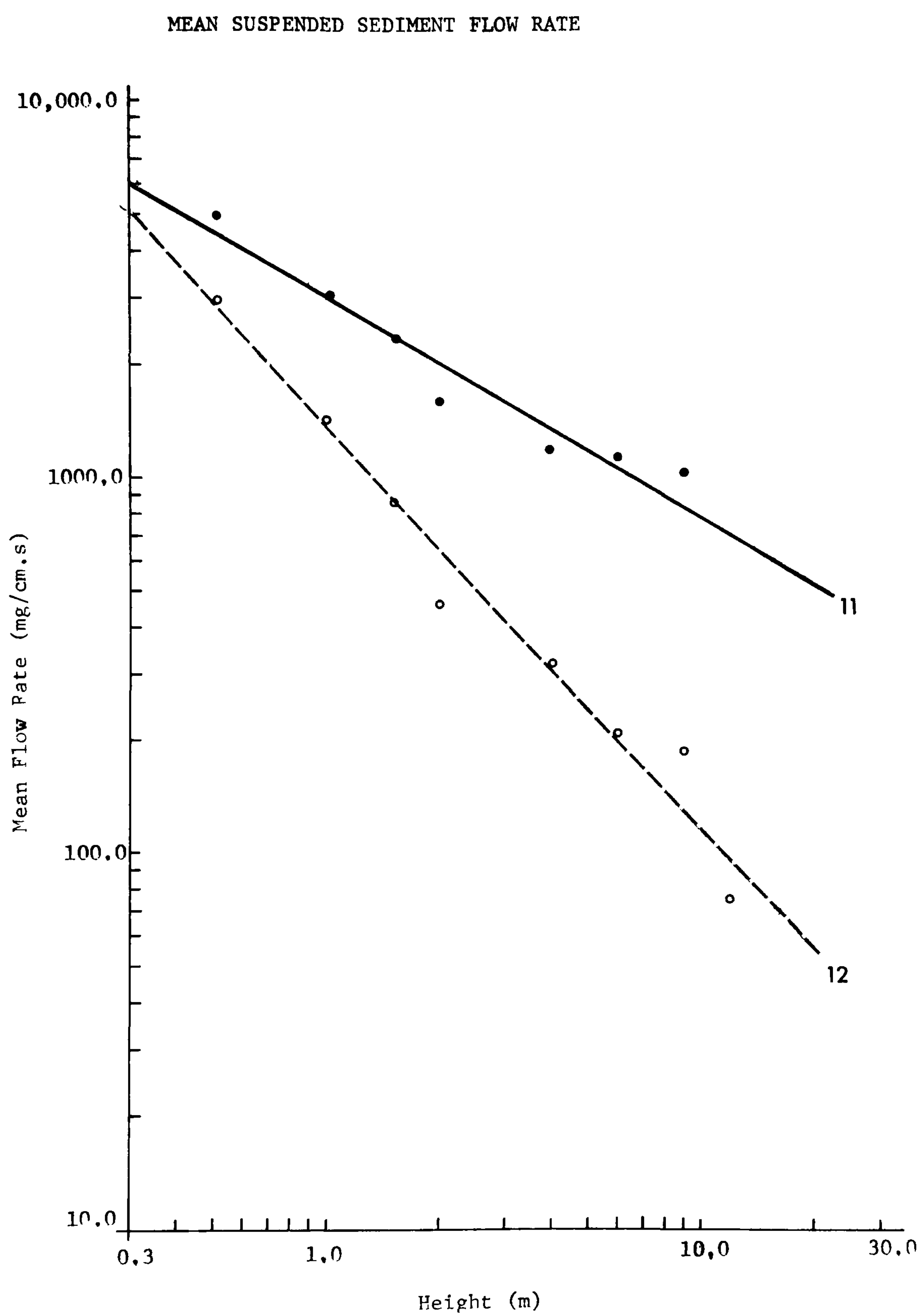


FIGURE 5.5 (continued)

MEAN SUSPENDED SEDIMENT FLOW RATE

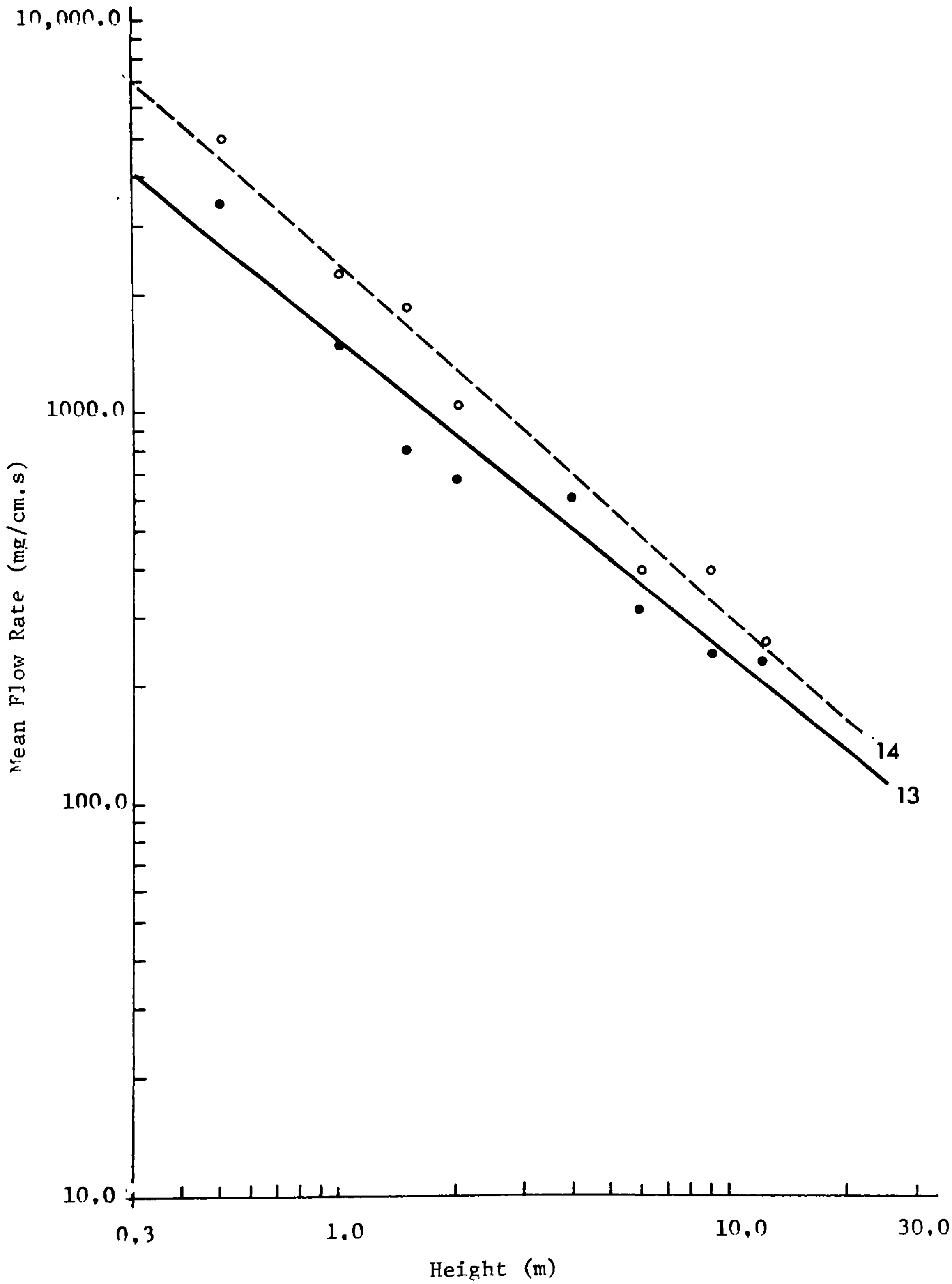


FIGURE 5.5 (continued)

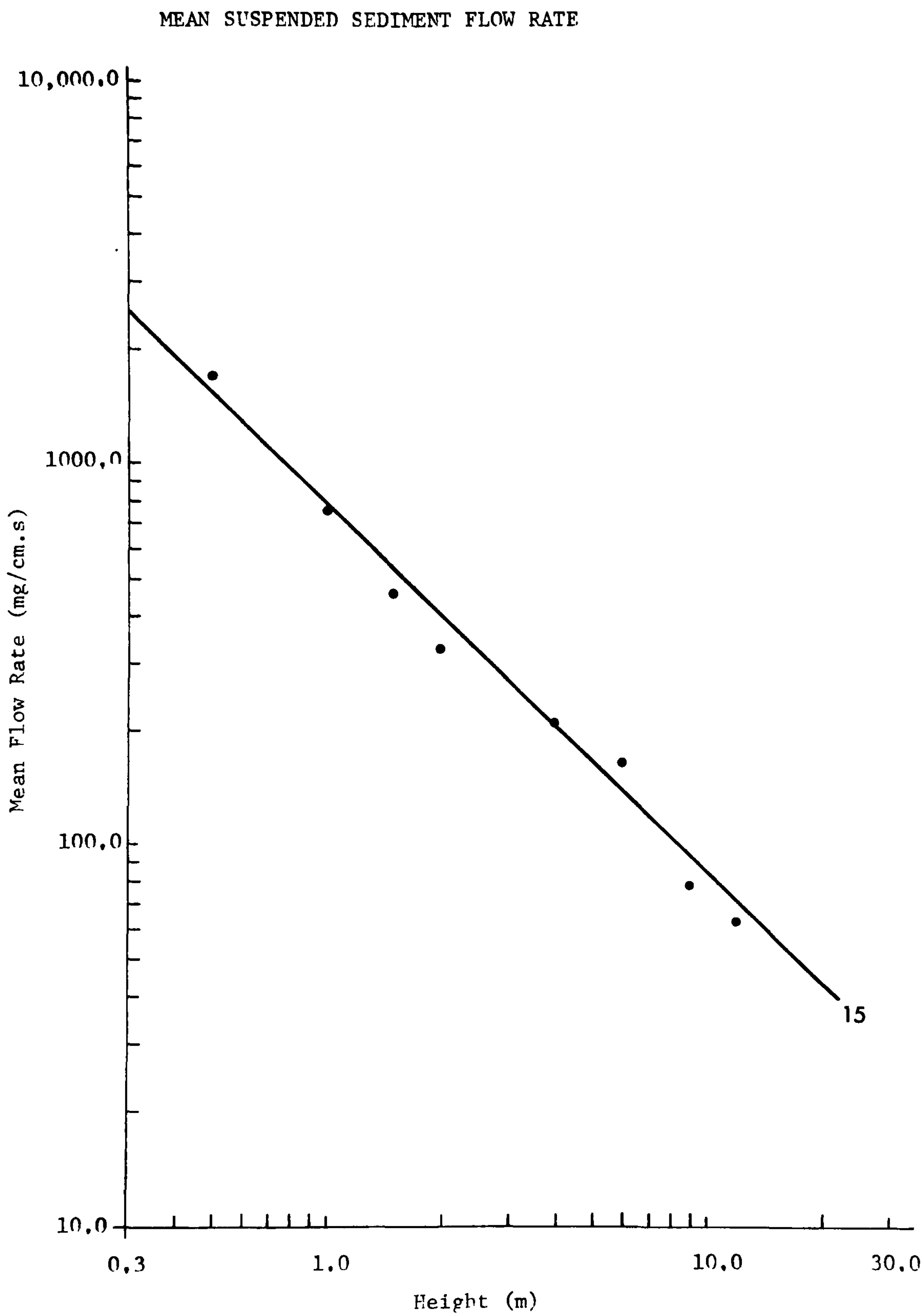


FIGURE 5.5 (continued)

The total amount of sediment transported through a unit width in unit time between any two heights can be calculated by

$$q_{F\,Z_1 - Z_2} = \int_{Z_1}^{Z_2} \frac{a}{Z^b}\,dz = \frac{a(Z_1^{\,1-b} - Z_2^{\,1-b})}{1-b} \qquad \text{............5.2}$$

It is evident that equation 5.2 is a limiting function and that suspended sediment transport at Z = 0 is infinite. Therefore, as an approximation of the total amount of sediment transported through a unit width between the surface and 12.0 metres (Z_2), the lower limit was selected as 0.001 m(Z). The selection of this height was not completely arbitrary. As previously mentioned (p. 7) the wind velocity is reduced to zero at some point above the true surface. This critical height (i.e. roughness length) has been related to the height of surface irregularities and the mean size of the surface sediments. In the Slims River delta area roughness lengths vary from approximately 0.0026 to 0.001 metres. Thus, the selected lower limit (Z_1 = 0.001) approximates the mean roughness length for the delta which corresponds to the height at which the wind velocity is effectively reduced to zero.

The suspended sediment flow rates calculated for the height interval 0.001 to 12.0 metres ($q_{F0.001}$) range from 0.022 to 0.289 g/cm.s (Table 5.2). In comparison the amount of suspended sediment transported between 0.5 and 12.0 metres ($q_{F0.5}$) ranged from 0.008 to 0.186 g/cm.s (Table 5.2). Thus, approximately 20 to 60 per cent of the suspended sediments transported during the fifteen dust storms was carried between 0.001 and 0.5 metres.

In both cases the flow rates show a relatively good linear

TABLE 5.2

ATMOSPHERIC AND SURFACE DATA COLLECTED
DURING THE SLIMS RIVER VALLEY DUST STORMS

Storm No.	Mean Shear Velocity (cm/s)	Mean Stability Ratio (°C/cm.s)	Mean Surface Moisture Content (%)	Mean Surface Salt Concentration (m.e./100 g of soil)	Suspended Sediment Flow Rate (0.001-12.0 m) (mg/cm.s)	Suspended Sediment Flow Rate (0.5-12.0 m) (mg/cm.s)	Saltation/Creep Flow Rate (q) (mg/cm.s)
1	35.69	-0.1772	3.57	29.8	22.5	19.7	28.0
2	26.12	-0.2720	4.49	28.1	22.9	8.7	22.2
3	39.36	-0.1788	4.28	25.3	104.2	43.3	46.2
4	43.14	-0.1040	3.50	22.4	47.0	18.8	114.5
5	39.00	-0.2349	4.62	26.8	12.4	7.6	86.2
6	50.35	-0.1658	2.47	23.7	38.5	22.8	137.8
7	62.25	-0.1323	2.61	18.5	130.0	66.7	154.9
8	54.72	-0.1709	3.91	16.7	53.5	31.8	133.5
9	46.81	-0.1815	3.83	24.9	70.8	57.9	38.3
10	56.26	-0.0936	2.58	14.9	289.9	185.8	231.3
11	56.62	-0.0862	2.33	21.2	198.3	149.1	149.7
12	43.73	-0.1290	2.01	25.8	124.9	47.5	249.9
13	42.90	-0.1369	2.50	18.2	113.3	58.2	95.6
14	40.89	-0.1373	3.10	17.9	219.2	85.2	115.4
15	32.86	-0.2528	3.06	17.2	80.5	25.7	50.0

relationship with shear velocity, surface moisture content, and surface salt concentration when plotted on log-log paper. From Table 5.3 it can be seen that surface moisture content and surface salt concentration are slightly better correlated with $q_{F0.001}$ than with $q_{F0.5}$. This indicates that surface conditions may become of less importance to the amount of sediment transported in suspension as one moves up from the surface. This may also suggest that surface conditions only affect the flow rate at height in that they directly control the initial entrainment of grains into the air stream.

In accepting this latter hypothesis one might expect that the rate of sediment transport to be a strong function of wind velocity. However, it can be seen from Table 5.3 that in both cases the flow rates are not highly correlated with shear velocity. The relatively low correlations may result in part from the fact the shear velocity does not adequately describe the total wind profile in each case.

The mean wind profile for each of the dust storms is given in Fig. 5.6. It should be noted that the profiles do not strictly follow the logarithmic law. In almost every case the wind velocity above 3 to 4 metres increases more rapidly with height than the logarithmic law would predict. Wind profiles of this form are not unusual and are indicative of unstable atmospheric conditions caused by superadiabatic temperature conditions. The mean temperature profiles (Fig. 5.7) from each storm show that the measured lapse rates are considerably greater than the dry adiabatic lapse rate especially below 3.0 metres. Although the temperature profiles continue to be superadiabatic above 3.0 metres they do begin to approach the dry adiabatic lapse rate. This decrease

TABLE 5.3

CORRELATION MATRIX FOR THE DUST STORM OBSERVATIONS

	Suspended Sediment Flow Rate (0.5-12.0m)	Saltation-Creep Flow Rate (q)	Shear Velocity	Stability Ratio	Surface Moisture Content	Surface Salt Concentration
Suspended Sediment Flow Rate (0.001-12.0m)	0.942	0.554	0.524	0.678	-0.640	-0.668
Suspended Sediment Flow Rate (0.5-12.0m)		0.538	0.667	0.754	-0.628	-0.607
Saltation-Creep Flow Rate (q)			0.759	0.710	-0.705	-0.541
Shear Velocity				0.755	-0.518	-0.501
Stability Ratio					-0.651	-0.384
Surface Moisture Content						0.375

MEAN WIND PROFILES

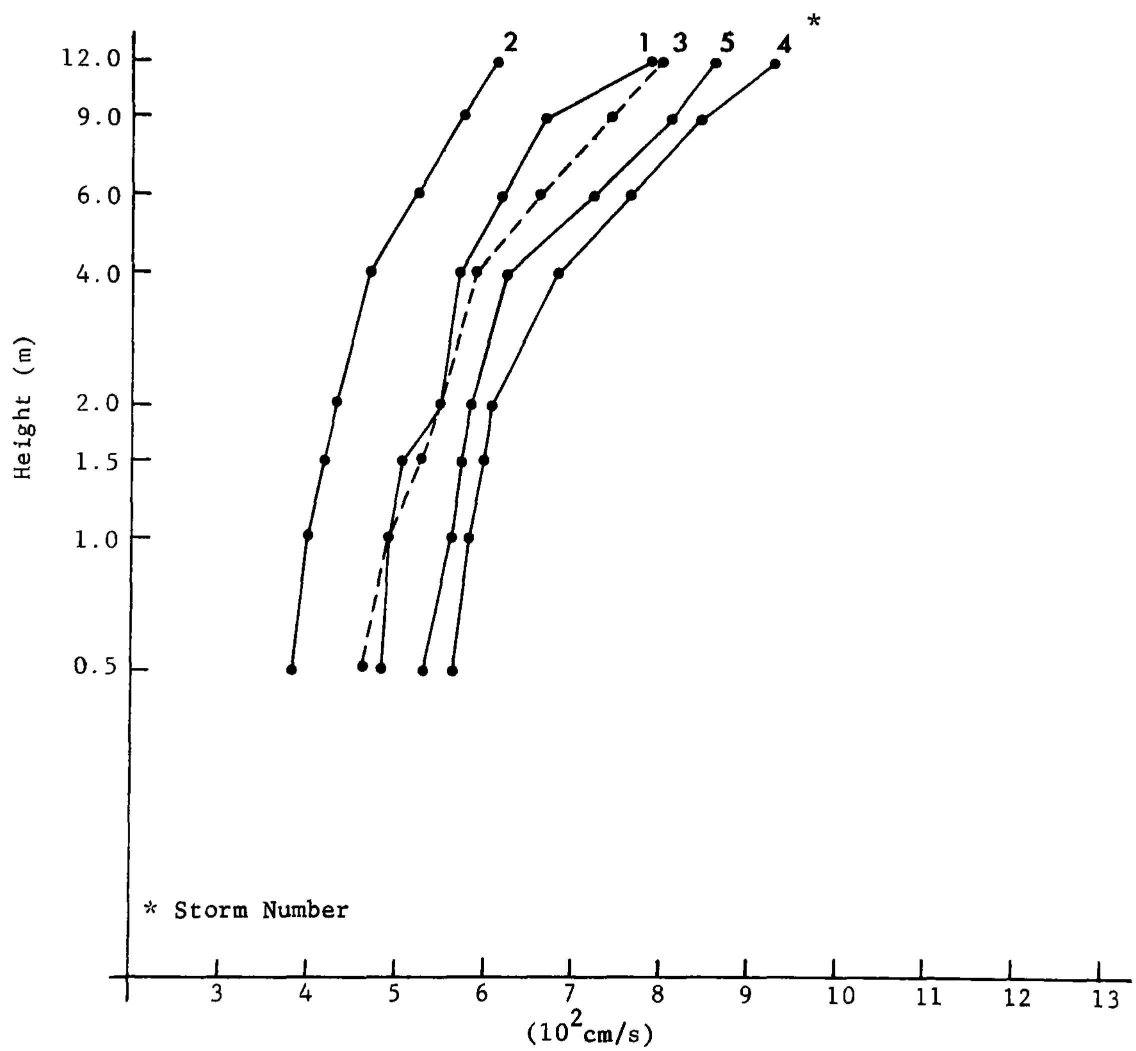


FIGURE 5.6

MEAN WIND PROFILES

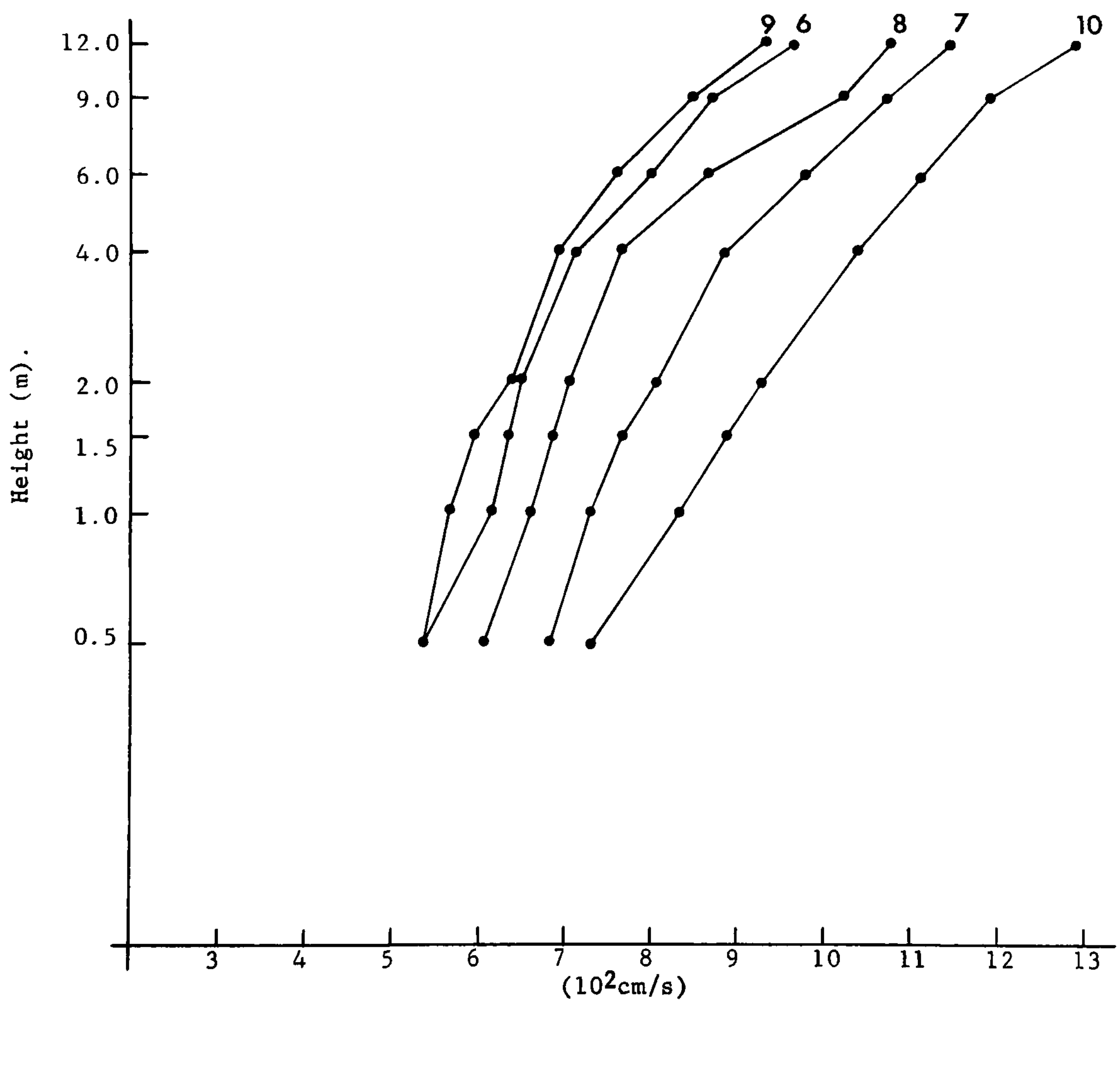


FIGURE 5.6 (continued)

MEAN WIND PROFILES

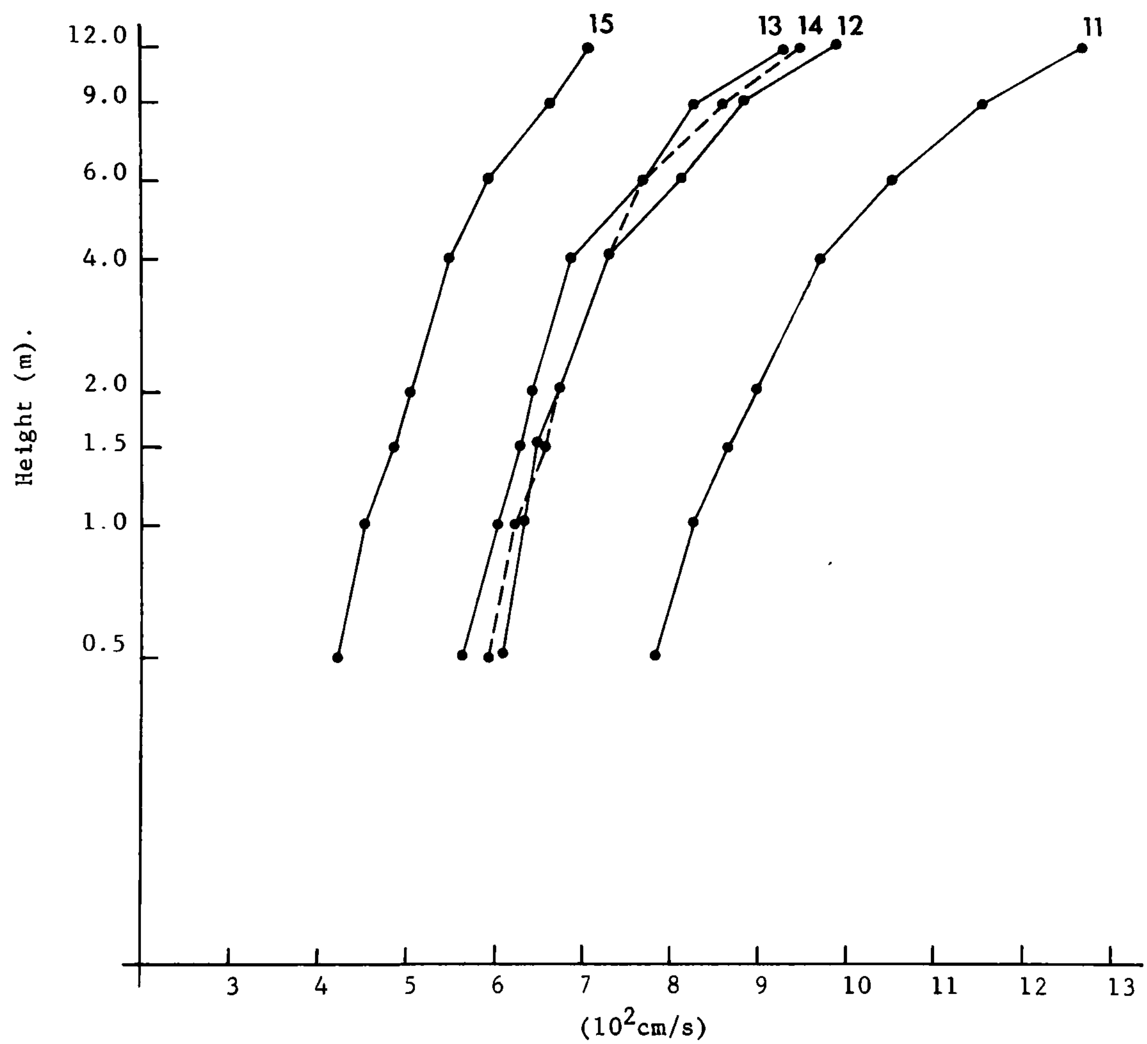


FIGURE 5.6 (continued)

MEAN TEMPERATURE PROFILES DURING THE DUST STORMS

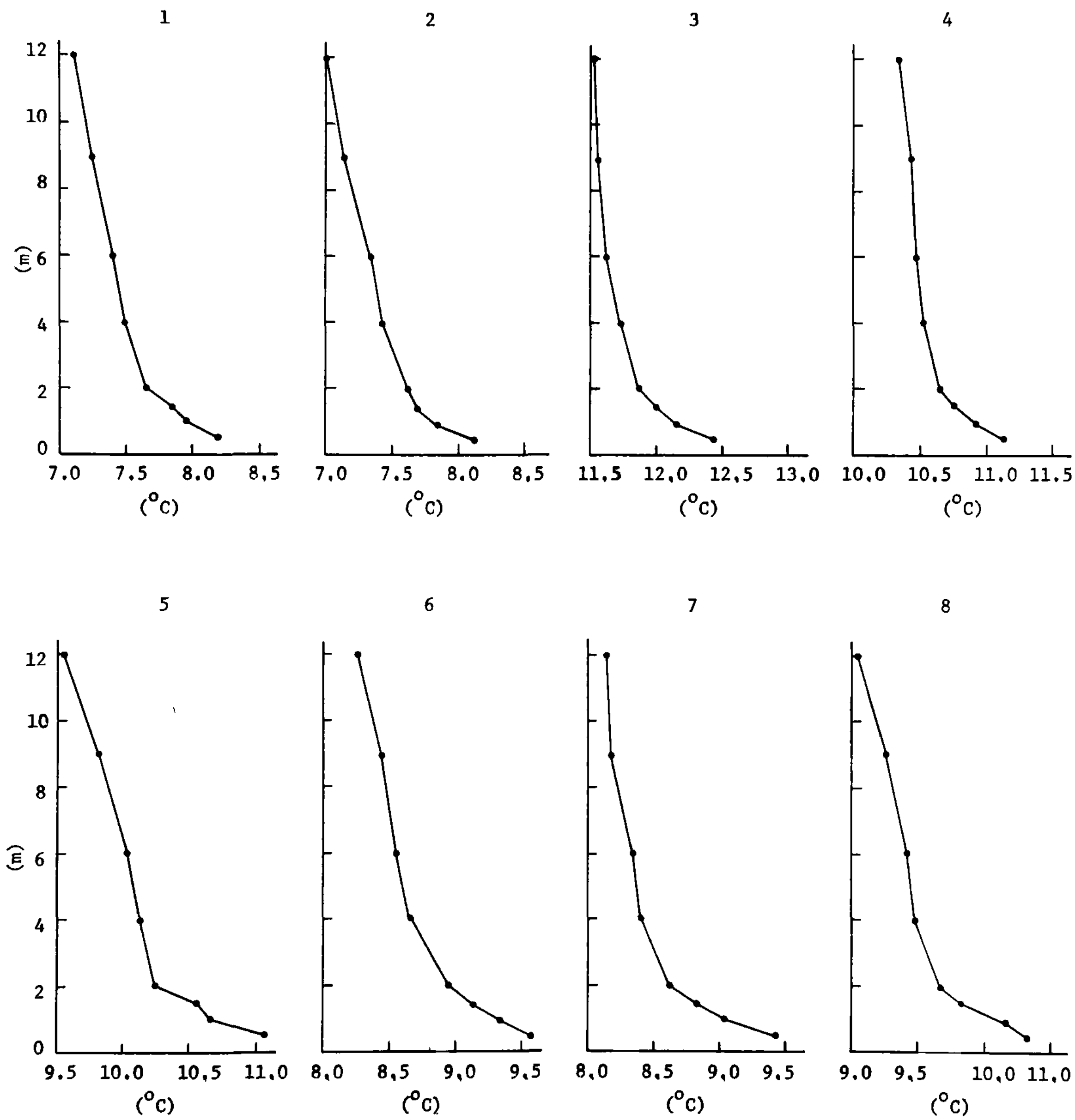


FIGURE 5.7

MEAN TEMPERATURE PROFILES DURING THE DUST STORMS

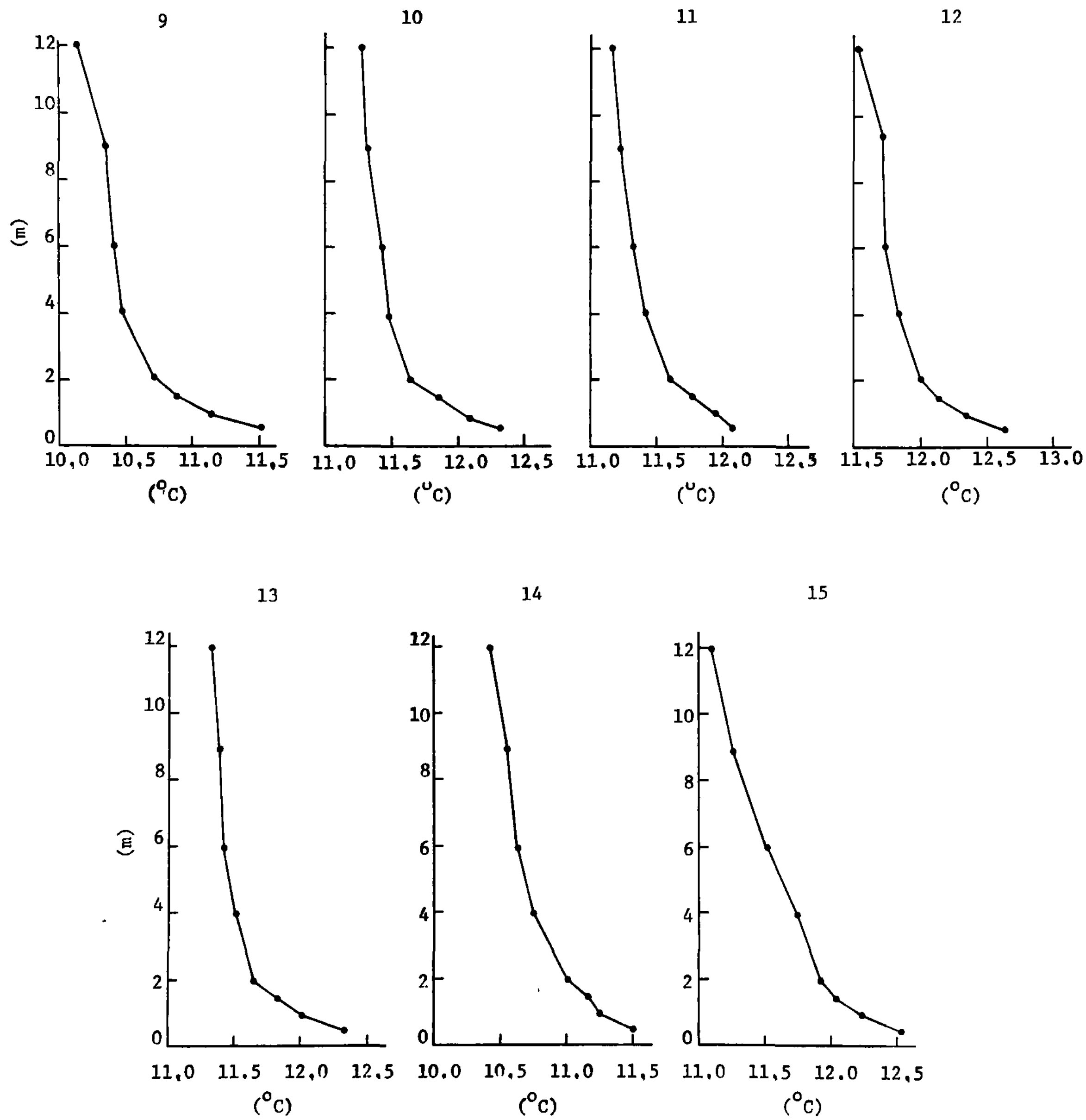


FIGURE 5.7 (continued)

in the lapse rate above 3.0 metres may be the result of relatively high wind velocities which were measured during the same time period. Super-adiabatic conditions are usually established and maintained by relatively clear skies and low wind velocities. Strong superadiabatic conditions are broken down or reduced as wind velocities are increased as a result of the rapid turnover or turbulence of the lower atmosphere. It is common in such situations that the measured lapse rate will begin to approach the dry adiabatic lapse (Sellers, 1965). Such a situation seemed to exist during the Slims dust storms. Although not all the dust storms occurred under clear sky conditions they were usually preceded by relatively low wind speeds during the late morning and early afternoon. This situation would enhance the development of a superadiabatic condition during the early and mid-afternoon which is subsequently broken down during the later afternoon with the increase in wind speed. It should also be noted that the increase in wind speed during the mid-afternoon is a function of the superadiabatic lapse conditions.

The stability ratio for each of the fifteen dust storms was calculated from the measured temperatures at 0.5 and 12.0 metres and the mean wind velocity at 6.0 metres (equation 1.9). In all cases the value of the stability ratio is strongly negative (Table 5.2). The stability ratios were regressed against the flow rate calculated for the two height intervals ($q_{F0.5}$ and $q_{F0.001}$). It is evident from Table 5.3 that in both cases a stronger relationship exists between the stability ratio and the flow rates than between the shear velocity (U_*') and flow rates. This would tend to support the hypothesis that the suspended sediment transport rate is more closely related to the degree of air turbulence than it is to wind velocity alone.

It is also of interest to note that there is a stronger relationship between the stability ratio and $q_{F0.5}$ than between the stability ratio and $q_{F0.001}$. This supports the argument that as one moves closer to the surface, the surface conditions, which closely affect the entrainment of grains, become increasingly more important. This does not necessarily mean that the wind velocity or turbulence is unimportant near the surface. Rather, the surface conditions must also be considered especially when discussing suspended sediment transport near the surface.

In order to investigate this hypothesis the stability ratio and the surface moisture and salt concentration were run in a multiple linear regression against the two calculated flow rates ($q_{F0.5}$ and $q_{F0.001}$). Results of the regressions (Table 5.4) show that the coefficients of multiple regression are significant and very similar ($q_{F.5} = 0.78$; $q_{F0.001} = 0.77$) in both cases. However, the contribution and significance of each independent variable to the regression is quite different. It is evident from the F ratios and beta coefficients of the independent variables that surface moisture content does not appear to significantly add to the explanation of either $q_{F0.5}$ or $q_{F0.001}$. This may appear somewhat unusual in that surface moisture content has been shown to significantly affect the mean daily sediment transport in saltation and creep (p. 103).

Surface moisture content might be considered significant if looked at from another point of view. A certain percentage of fine sediment is almost always carried in suspension whenever coarser

TABLE 5.4

MULTIPLE LINEAR REGRESSION RESULTS
FOR THE DUST STORM OBSERVATIONS

Dependent Variable: Suspended Sediment Flow Rate (0.001-12m)

Variable	Coefficient	Beta Coefficient	Calculated F Value
Constant	5.1890		
Stability Ratio	2.4520	0.3459	3.60*
Surface Moisture Content	-0.8681	-0.2495	1.23
Surface Salt Concentration	-1.8671	-0.4419	5.71*

Standard Error of the Estimate	0.2487
Coefficient of Determination	0.6896
Multiple Regression Coefficient	0.8304
Adjusted Coefficient of Determination	0.6050
Adjusted Multiple Regression Coefficient	0.7778

Analysis of Variance for the Regression

Source	Degrees of Freedom	Sum of Squares	Mean Squares	F
Regression	3	1.5112	0.5037	8.15*
Error	11	0.6802	0.0618	
Total	14	2.1914		

Durbin-Watson d Statistic 1.22**

Significant at the 0.05 confidence level**
0.01 confidence level*

...continued

TABLE 5.4--continued

MULTIPLE LINEAR REGRESSION RESULTS
FOR THE DUST STORM OBSERVATIONS

Dependent Variable: Suspended Sediment Flow Rate (0.5-12m)

Variable	Coefficient	Beta Coefficient	Calculated F Value
Constant	4.4618		
Stability Ratio	3.6966	0.5150	5.42**
Surface Moisture Content	-0.5699	-0.1617	0.54
Surface Salt Concentration	-1.4928	-0.3490	3.71**

Standard Error of the Estimate	0.2468
Coefficient of Determination	0.7019
Multiple Regression Coefficient	0.8378
Adjusted Coefficient of Determination	0.6206
Adjusted Multiple Regression Coefficient	0.7878

Analysis of Variance for the Regression

Source	Degrees of Freedom	Sum of Squares	Mean Squares	F
Regression	3	1.5772	0.5257	8.63*
Error	11	0.6699	0.0609	
Total	14	2.2471		

Durbin-Watson d Statistic 1.49**

Significant at the 0.05 confidence level**
0.01 confidence level*

material is transported in creep and saltation. This is indicated qualitatively by small plumes of dust which rise irregularly from the surface during these periods. As previously shown (p. 99), a measurable amount of sediment was transported in saltation and creep on 40 of the observation days. On these days the mean daily surface moisture contents varied from 1.81 to 10.96 per cent dry weight. Although the transport of sediment in saltation and creep is common on the Slims River delta, the transport of large quantities of suspended sediment, in the form of dust storms, occurs infrequently. It can also be seen that these dust storms are associated with a relatively narrow range of surface moisture contents (2.01 to 4.62 per cent dry weight) which form the lower end of the moisture content range for the total observation period. This might suggest that there is a surface moisture content threshold above which large dust storms do not occur.

In order to investigate this hypothesis the values for the mean daily sediment transport in saltation and creep were again plotted against surface moisture contents, this time indicating the days on which a dust storm occurred. From Fig. 5.8 it can be seen that 14 of the 15 days with dust storms appear to be separated from the mean daily flow rates of the days without dust storms. If one considers the best fit regression line and the scatter of the data, a division between the two groups could be set at approximately 5.0 to 6.0 per cent dry weight. It is suggested that this moisture content range could be viewed as a threshold above which large dust storms do not occur in the Slims River delta. It is obvious that this threshold can not be given an absolute value because moisture content is only one

RELATIONSHIP BETWEEN MEAN DAILY
SALTATION-CREEP FLOW RATE
AND SURFACE MOISTURE CONTENT

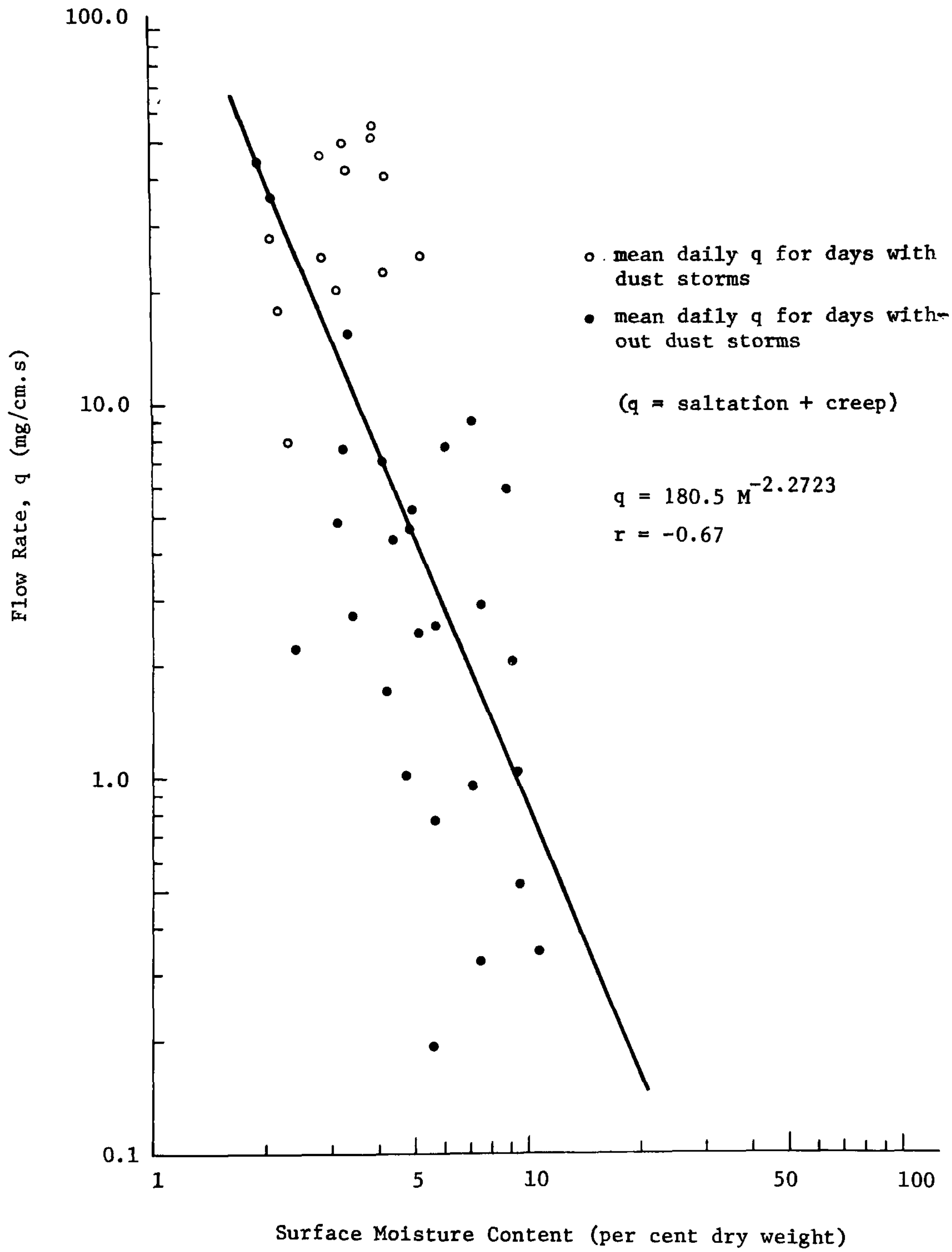


FIGURE 5.8

of many interrelated variables which affect the transport of sediment in suspension. Also, as a result of the relatively wide scatter of the data, the designation of an absolute value would be meaningless. Therefore it can only be suggested, that on average dust storms usually occur when the surface moisture content of the delta is below 5.0 to 6.0 per cent dry weight. However, below this threshold, the surface moisture content does not appear to significantly affect the amount of sediment transported in suspension.

The reason for this is not immediately apparent but it is suggested that it may at least in part be related to the pore size distribution of the delta sediments and overriding effects of other variables when surface moisture contents fall below 5.0 to 6.0 per cent dry weight.

A possible explanation might be given for the two field grouping in Fig. 5.8 if one assumes that the Slims River delta sediments have a large percentage of pores within a narrow size range (i.e. the pore size distribution is strongly unimodal). This in turn may also help to explain why dust storms occur only on days with relatively low moisture contents and high mean daily saltation/creep flow rates.

It is highly possible that the delta sediments would have a strongly unimodal pore size distribution. This is suggested because the pore size distribution of a sediment is closely related to the grain size distribution (Terzaghi and Peck, 1967). As previously mentioned the fine grained delta sediments are relatively well sorted, unimodal and positively skewed (Table 5.8 and Fig. 4.2). This type of grain size distribution would favour the development of a pore size distribution with a dominant mode.

As the surface of the delta dries the water menisci will retreat into the progressively finer pores. The negative capillary pressures which are developed in the pores effectively hold together the grains which are in contact with the menisci (equation 4.7). However, if as a result of continued drying the menisci retreat into smaller pores, the grains which formed the relatively larger pores will no longer be held by capillary suction and will therefore be susceptible to entrainment into the air stream.

A point will be reached during the drying of the sediment when the menisci will be located in the modal pore size. If the drying continues and the menisci retreat into the still finer pores a large number of grains which form the modal pore sizes will no longer be held by surface tension effects. Consequently, when the menisci retreat into the pores which are finer than the modal pore diameter there would be a sharp increase in the number of grains available for transport in creep, saltation and suspension. It is suggested that in the Slims River delta sediments the menisci lie in the modal pore sizes when the moisture content is somewhat above 5.0 to 6.0 per cent dry weight.

Support for the above argument can be found in Fig. 5.9 which shows the suction-moisture curves for three samples collected from the surface of the Slims River delta. These relationships give for any moisture content the negative pressure[1] of the water held within the pore spaces. As already shown (equation 4.7) this negative pressure is

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[1]It is conventional to express this negative pressure as the common logarithm of the height of water (cm) which the suction would support (pF).

SUCTION-MOISTURE AND GRAIN SIZE FREQUENCY CURVES FOR THF SURFACE SEDIMENTS

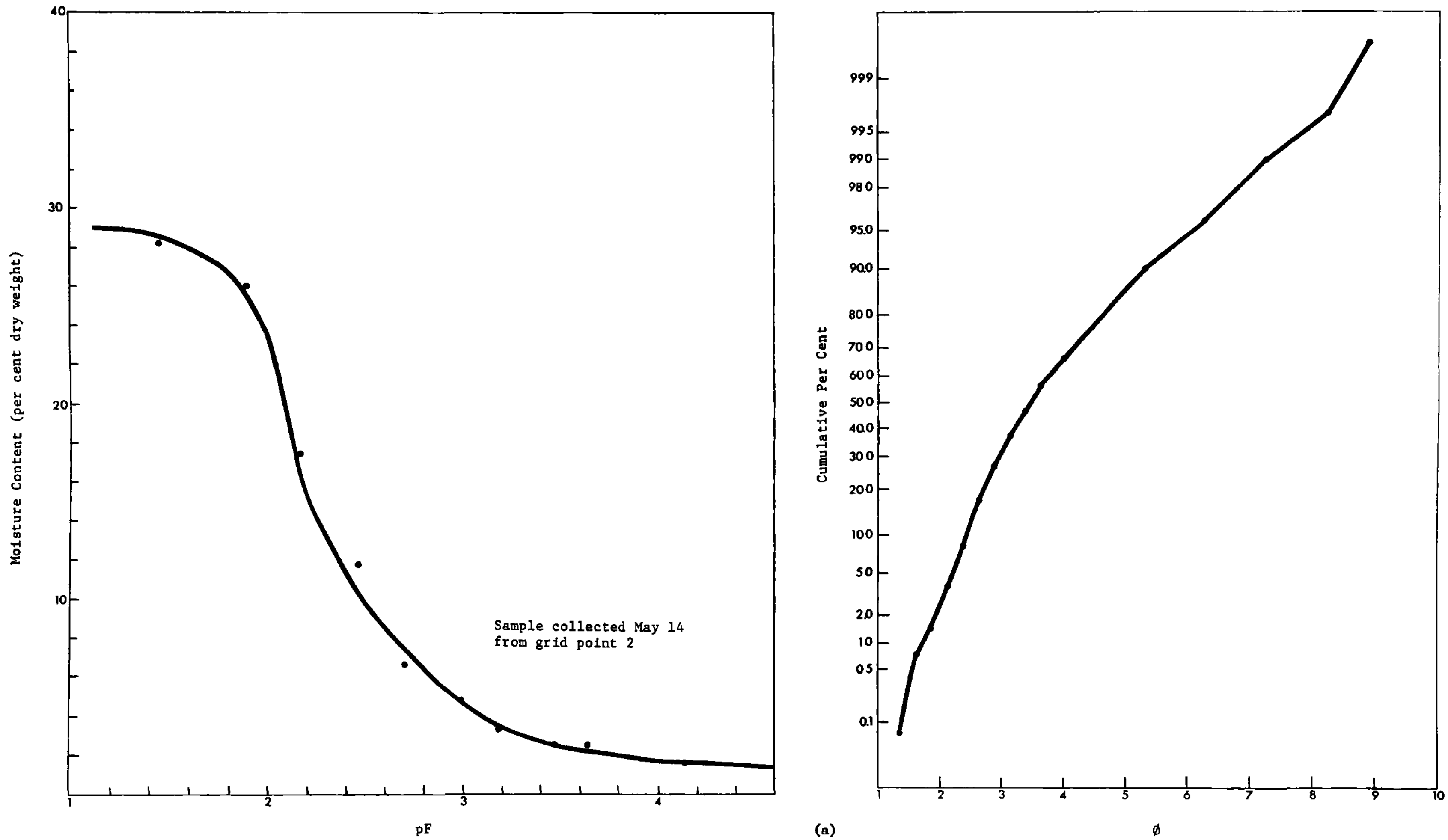


FIGURE 5.9

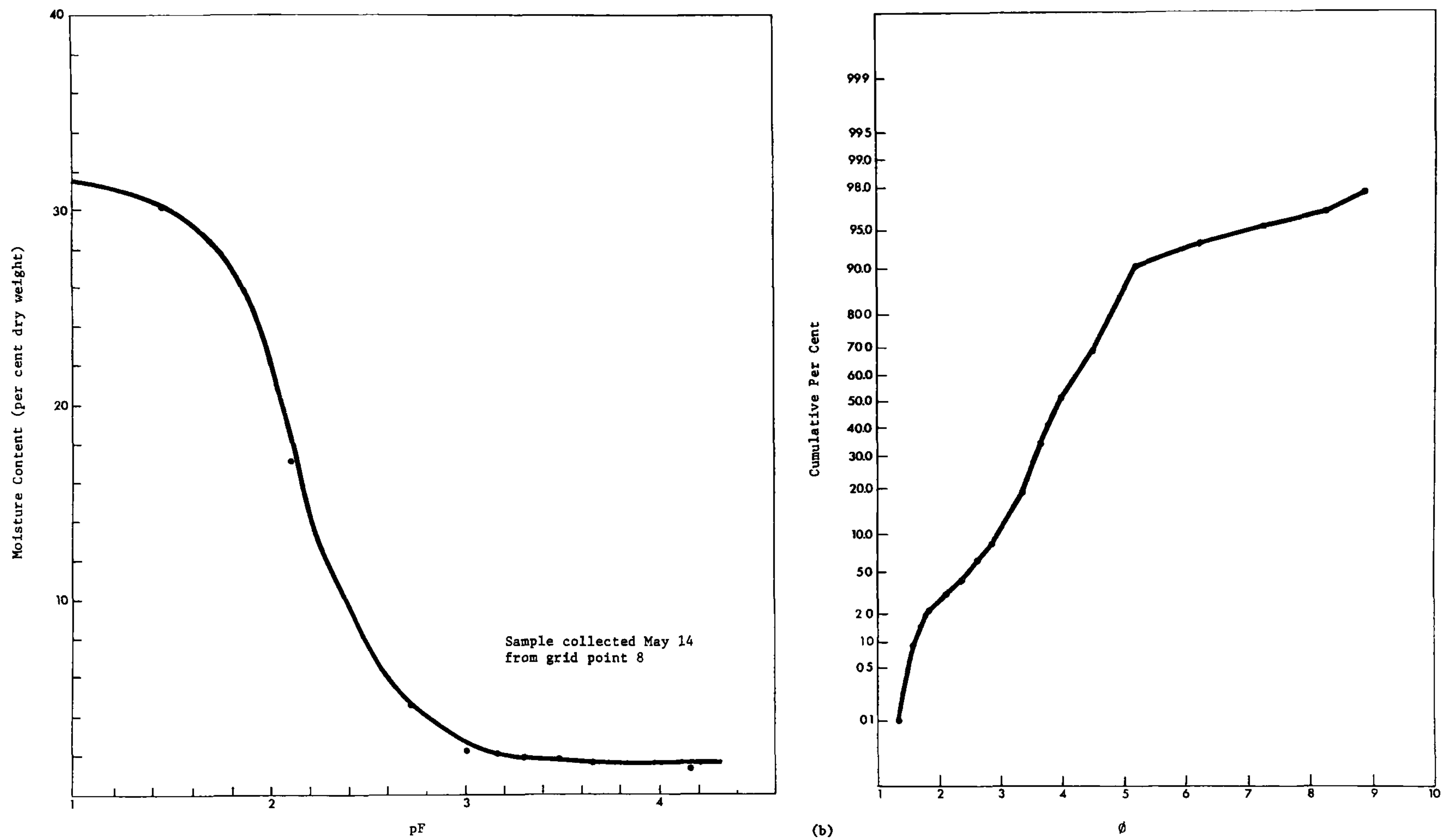
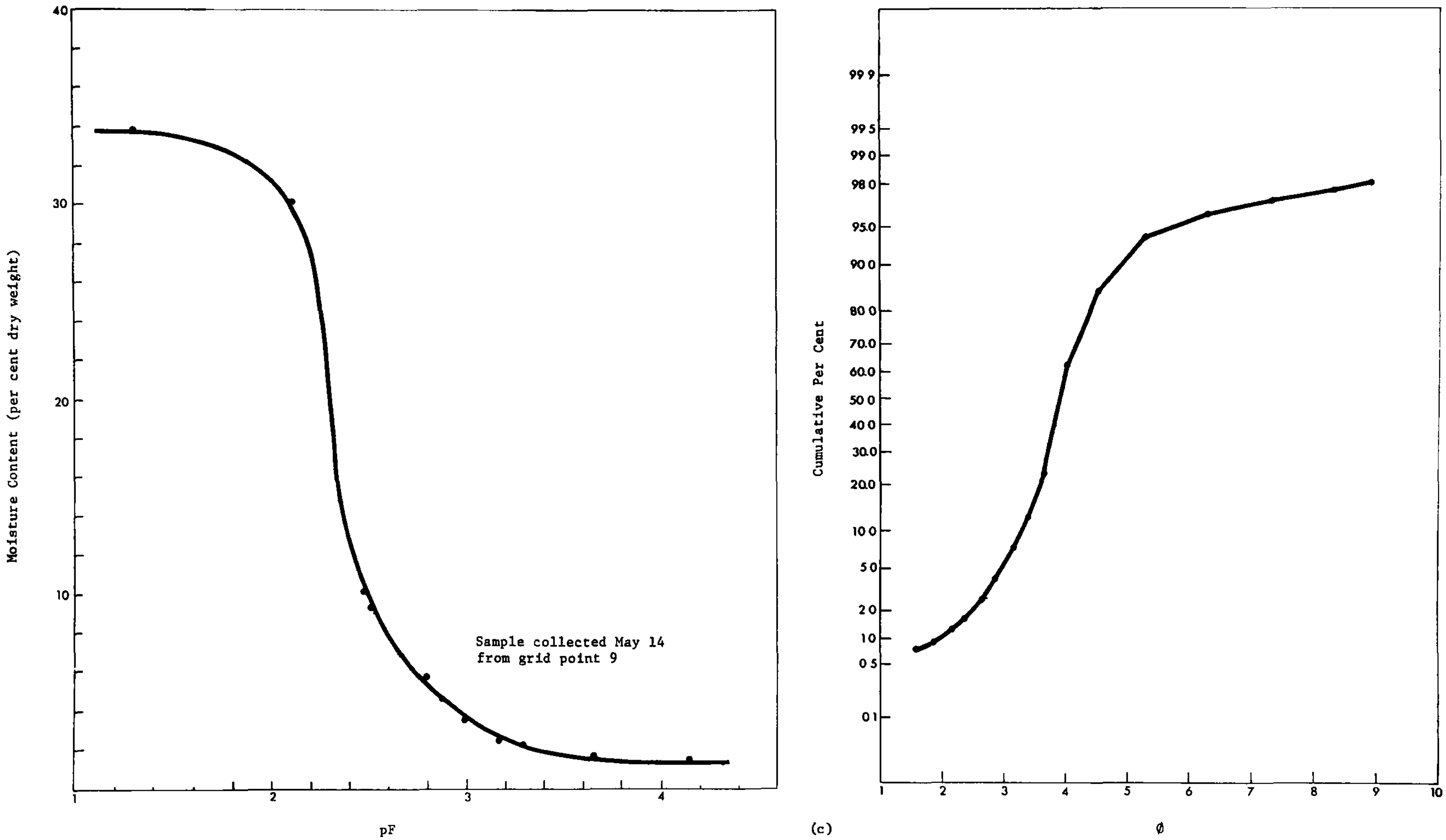


FIGURE 5.9 (continued)

SUCTION-MOISTURE AND GRAIN SIZE FREQUENCY CURVES FOR THE SURFACE SEDIMENTS



(c)

FIGURE 5.9 (continued)

directly related to the radius of the pore in which the water is held.

It is evident from Fig. 5.9 that the moisture content of the surface samples decreases sharply when the negative pressure of the held water rises from approximately 2.0 to 2.5 pF. The sharp decrease in moisture content with a relatively small increase in suction indicates that there are a large number of pores with radii corresponding to this pF range. Thus, the surface samples appear to have a somewhat unimodal pore size distribution with the modal pore sizes (approximately 10-30 microns) corresponding to a negative capillary pressure of approximately 2.0 to 2.5 pF.

It is also apparent that the curves begin to flatten out (i.e. suction increases sharply) when the moisture content falls below 2 to 4 per cent. Below this moisture content the menisci retreat into the pores which are smaller than the modal pore diameter thereby removing one of the forces holding a large number of surface grains in situ. It is also suggested that the pores equal to or smaller than the modal pore diameter will be formed by a large percentage of grains which are small enough to be carried in suspension. Thus, once the menisci retreat into the pores finer than the modal diameter a large number of these fine grains will be susceptible to entrainment into the air stream.

However, from the scatter in Fig. 5.8 it is apparent that large amounts of fine grained material are not always entrained into the wind stream when surface moisture contents fall below 5.0 to 6.0 per cent even if the corresponding wind velocities are relatively high. On these occasions a thin layer of surface material could be picked up in one's hand and would feel dry and powdery. The material would also crumble very easily, and if tossed into the air the majority of the material would be carried away in suspension. This would suggest that another perhaps more critical variable affects the entrainment of fine grained sediment as the surface dries.

It would appear from both the simple and multiple linear regressions that surface salt concentration is a significant variable affecting the entrainment and transport of sediment in suspension. That is, as the surface salt concentration increases the amount of sediment transported in suspension decreases.

When the surface of the delta is wet the majority of the soluble salts will be in solution with the pore water. As the surface of the delta dries and the menisci retreat into the smaller pores the salt concentration of the pore water will increase. If the pore water continues to evaporate a point may be reached when the pore water will become saturated and some salt will begin to precipitate out into the pores. The precipitated salt will effectively bond together any particles in contact with the salt crystals.

A thin crust of white salt can be seen on the surface of some areas of the delta even though the sediment immediately below the encrusted sediment is damp (Fig. 1.4). This may imply that in many cases, because of the high salt concentration in the pore water, relatively little evaporation occurs before salt begins to precipitate out. At these relatively high moisture contents the water menisci will be located in the larger pores which will be formed by a large percentage of the coarser grains. As already suggested (p. 106) the salt will initially precipitate out in these larger pores and will therefore bond together the coarser grains. Thus, the amount of coarser sediment which would be susceptible to entrainment and transport in saltation and creep will be reduced. If the surface continued to dry, salt would be precipitated out in the progressively finer pores, thereby increasing the cohesion between a large percentage of the finer grains which form these pores. Thus, in many cases, even at very low moisture contents, a larger amount of fine grained material will not be entrained and transported in suspension because of the cohesive forces imparted by the precipitated salt. It should be noted that at any instant in time pore water and precipitated salt can both be present at or near the surface. It therefore follows that as the surface dries and the cohesive force caused by the presence of moisture decreases, the cohesive force attributable to the presence of salt crystals in many cases increases.

From the above discussion and Tables 5.2 and 5.3 it would appear that the dust storms usually occur when the wind velocities are relatively high, surface salt concentrations low and moisture contents less than 3.5 to 3.8 per cent. An examination of the storm dates and the precipitation data for the observation period, would indicate that such atmospheric and surface conditions occur on the days following periods of relatively heavy rainfall (Table 5.1).

The explanation for this is probably in part related to the fact that during the rainstorms the soluble salts are leached from the surface as the precipitation percolates into the delta. Several examples of the decrease in surface salt concentration during and following periods of extended rainfall have been given in Table 4.3. Thus, following these periods of precipitation one of the principal cohesive forces opposing the entrainment of fine grained sediment into the wind is substantially reduced. Although the salt concentration of the sediment at points immediately below the surface was not taken during or following the rainstorms it is felt that the salt is not leached more than a few centimetres below the surface. Since sediment is only picked up by the wind from the immediate surface, the salt concentration need only be lowered in the top few millimetres in order to have a significant effect on the amount of sediment which is entrained. Despite the fact that the salt concentration of the surface is substantially reduced by precipitation, the surface sediments will not immediately be susceptible to entrainment because of the increase in surface moisture content. From the Table 5.1 it can be seen that the dust storms usually occur one to three days after periods of relatively heavy precipitation depending on the amount and intensity of the rainfall.

It is also evident from Table 5.1 that the transport of sediment in saltation and creep usually begins very shortly after the precipitation ends (<24 hours). The reason for this is that only a small amount of drying by evaporation needs to occur before water is lost from the larger pores thus removing the cohesive forces holding a large number of the coarser grains. Thus, even though the surface sediment may appear to be relatively moist some larger surface grains will be susceptible to entrainment. As the surface continues to dry a progressively greater amount of sediment is entrained with a consequent increase in the saltation/creep flow rate. If one accepts the hypothesis that water retreats into the finer pores which are formed on average by progressively smaller grains one might expect that the mean grain size of the transported sediment would decrease as the surface moisture content decreased.

The mean grain size of the sediment transported in saltation and creep each day and during the dust storms has been plotted against the respective surface moisture contents (Fig. 5.10 and 5.11). Although the relationships are not particularly strong there is a tendency in each case for the mean size of the transported material to decrease with decreasing surface moisture content.

When the surface moisture content falls below 5.0 to 6.0 per cent a larger number of fine grains will no longer be held by surface tension effects. Since the bonding effect of the soluble salts will also have been decreased by the rainfall these fine grains, which are small enough to be carried in suspension, will be more susceptible to entrainment. However, it is evident in the field that these sediments, even when very dry, are not completely loose (i.e. powdery) but rather maintain a weak structure which causes the surface to remain somewhat resistant to wind erosion. This weak structure probably results from soluble salt and moisture remaining within the fine pores, as well as from the presence of a small amount of clay. Relatively low surface moisture contents and salt concentrations appear to be necessary for entrainment of fine grained sediment, but large dust storms will occur only if the above surface conditions are present and if wind velocities and saltation/creep flow rates are relatively high.

The necessity of relatively high wind velocities is indicated by Table 5.2 which shows that the dust storms occur only when the mean shear velocities are greater than 25 cm/s. The high shear velocities have two basic effects; first, they increase the horizontal shearing force on the surface sediments and second, they increase the amount of

RELATIONSHIP BETWEEN SURFACE MOISTURE CONTENT
AND THE MEAN SIZE OF THE SEDIMENT
TRANSPORTED IN SALTATION

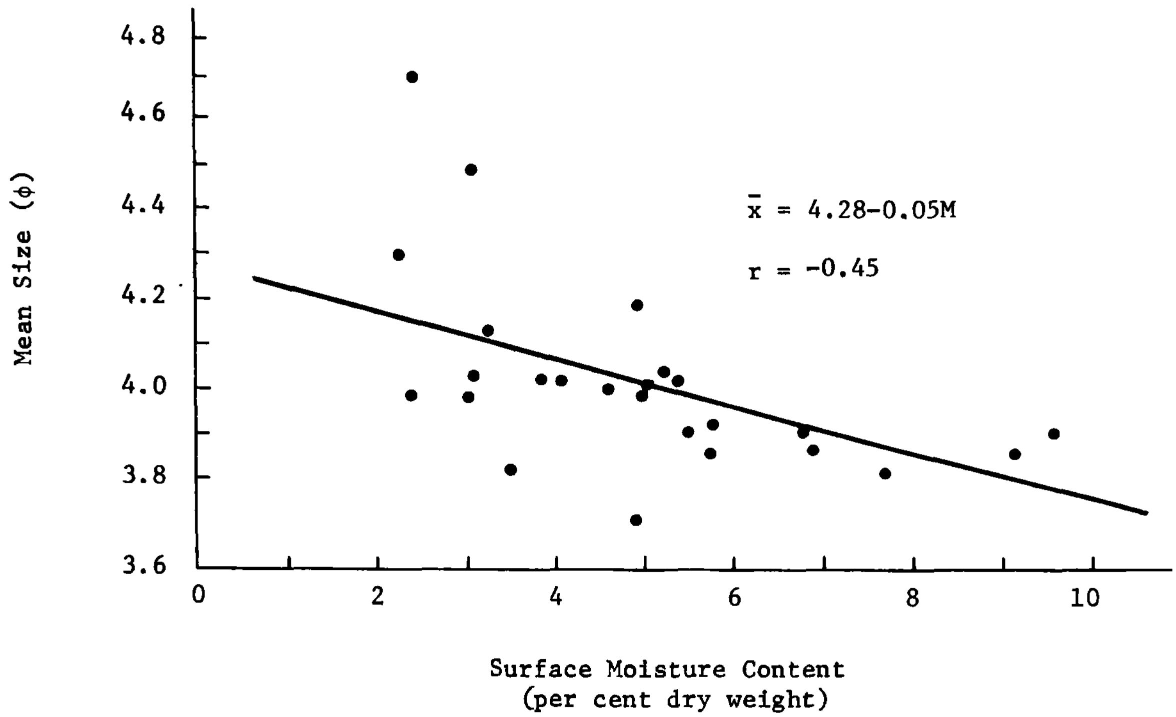


FIGURE 5.10

RELATIONSHIP BETWEEN SURFACE MOISTURE CONTENT
AND THE MEAN SIZE OF THE SEDIMENT
TRANSPORTED IN CREEP

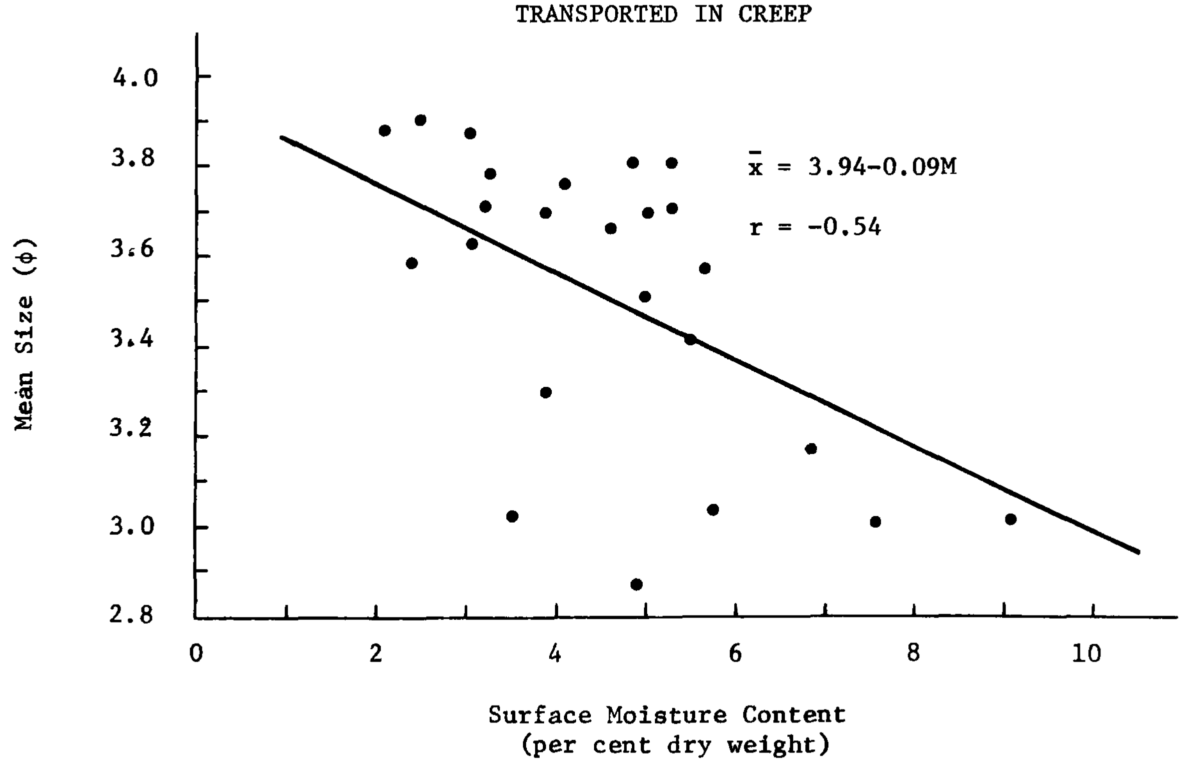


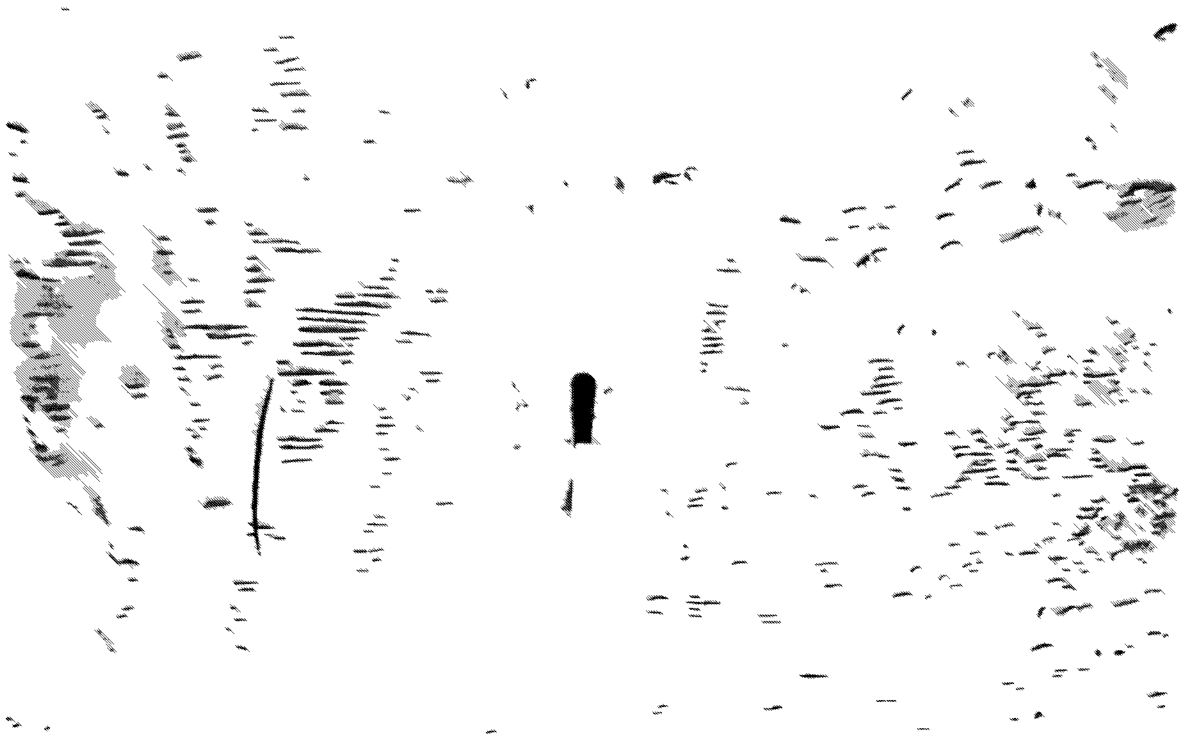
FIGURE 5.11

sediment transported in saltation and creep (Table 5.2).

Bagnold (1941) has shown in wind tunnel experiments that dry, fine-grained sediment is not necessarily entrained into the air stream even if the wind velocity above the sediment is extremely high. This was demonstrated by blowing a steady stream of air over a loosely scattered layer of fine Portland cement powder. In this case no particle movement occurred even when the shear velocity (U'_*) exceeded 100 cm/s. (i.e. shear velocity great enough to move pebbles 4.6 mm in diameter; Bagnold, 1941, page 90). The reason for this is that the surface of the fine grain sediment is aerodynamically smooth and consequently little or no air turbulence is developed immediately above the surface. Since only a negligible amount of turbulence is developed the individual particles at the surface are only acted upon by a horizontal shearing force, rather than by the vertical component of turbulent eddies and thus remain in place. Bagnold (1941) has also suggested that larger particles moving in saltation and creep can cause small particles to become entrained into the air stream. This was demonstrated in a wind tunnel by covering the upwind half of the tunnel floor with sand and remaining half with fine grained Portland cement powder. If the sand was held in place by wetting, no movement of the fine grained sediment occurred even if wind velocities were extremely high. However, if a moderate wind velocity was maintained over the sediment and the sand allowed to dry, a dust cloud immediately appeared when the sand began to move over the fine grained sediment. The entrainment of the fine grained material resulted from the bombardment of the fine grained material by the coarser sand grains. These sand grains affect the surface in two ways. First, the sand grains effectively knock the finer particles up from the surface by impact. Secondly, the presence of the coarser sand grains on the surface of the fine grained material causes the surface to become aerodynamically rougher. As a result the air flow immediately above the fine grained surface becomes more turbulent which increases the vertical forces tending to lift the particles from the surface.

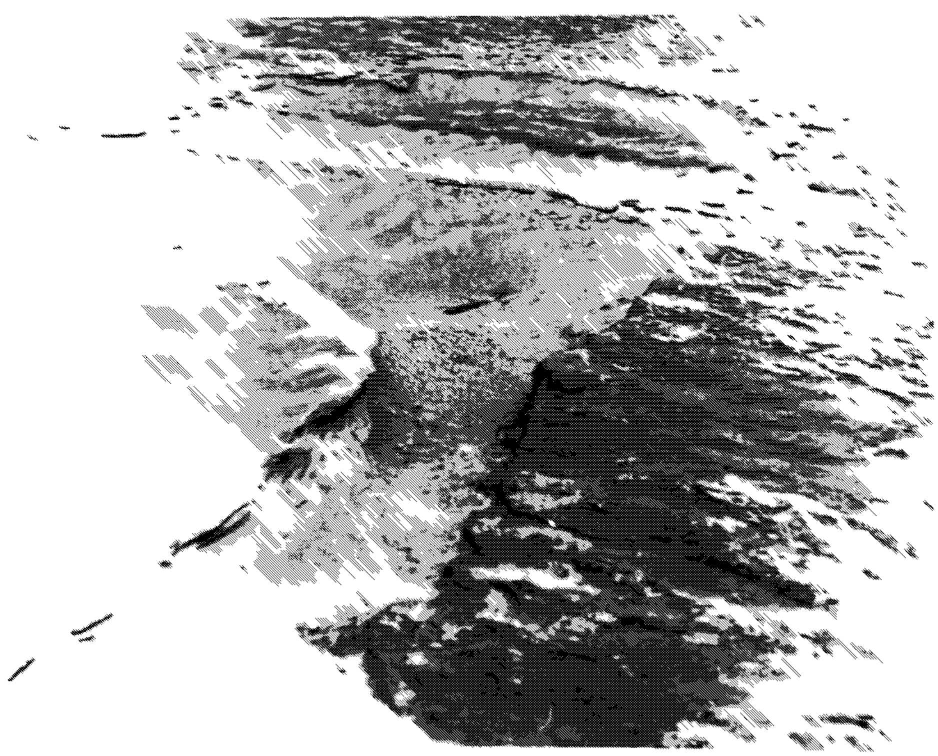
It is felt that a similar mechanism is responsible, at least in part, for moving fine grained sediment into suspension during the dust storms on the Slims River delta. The situation in this particular case is somewhat more complex because of the additional cohesive forces caused by the soluble salts and moisture content.

The effect of coarser material moving in saltation and creep on the entrainment of fine grained sediment is quite visible during the dust storms in the Slims River Valley. One example of this is shown in Fig. 5.12. This photo shows a previously flooded area of the delta covered with very small ripples less than 2-3 mm in height aligned almost parallel to the dominant wind direction. The very fine sediment, in which the ripples were formed had a very firm structure when dry and overlaid a somewhat coarser and more loosely packed sedimentary unit. During eolian activity material being transported in saltation and especially creep was effectively funnelled down the troughs of the ripples. After several days of very strong winds the troughs of the ripples were preferentially abraided by the sediment moving in saltation and creep. As a result the coarser sedimentary unit was exposed below the troughs while the ripple crests remained almost intact. Subsequent



SMALL LONGITUDINAL SCOUR MARKS ERODED
BY SEDIMENT MOVING IN SALTATION AND CREEP

FIGURE 5.12



LARGE DEFLATION PIT IN THE DELTA SURFACE FORMED
BY THE PREFERENTIAL REMOVAL OF FINE SANDS

FIGURE 5.13

to this the material below the troughs also began to be removed by the wind. Since it was relatively coarse a large percentage of this material was transported in saltation and creep and thus increased the rate of abrasion within the ripple troughs. As a result of lateral erosion, the ripples and almost the entire fine upper sedimentary unit were removed.

This phenomena occurred in a similar manner in other areas of the delta. The most dramatic case was found near the south east corner of the study grid. At the beginning of the study period this area of the delta was almost completely flat and smooth except for an arcuate depression approximately 10-15 cm in depth, 3.0 metres long and 0.5 - 1.0 metres wide. This depression appeared to be the remnant of one of the many overflow channels visible on the delta's surface (Fig. 1.4). The surface of the depression was covered with less than a centimetre of relatively fine sediment which overlay a considerably coarser stratigraphic unit[1] (Fig. 5.13). Towards the end of May 1973 the fine cap eroded through near the centre of the depression and exposed the coarser sediment below. Erosion accelerated rapidly and the fine grained cap was quickly removed from a large portion of the depression in line with the dominant wind direction. Since the coarser sediment in the depression was medium to fine sand it was easily entrained and transported by the wind in saltation and creep. As a result the original depression was severely deflated and by the end of the field season was approximately

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[1]This sediment was considerably coarser (3.4ϕ) than any other sediment encountered near the delta surface in the study grid.

30-40 cm in depth. The fine to medium sand which originated in the depression also severely affected the fine grained surface immediately down wind from the depression. In this area, the surface became noticeably more broken and somewhat more loosely packed than the surface in adjacent areas of the study grid. As a result of the daily observations it soon became apparent that small plumes of dust were generated more frequently in this area than in any other part of the study grid. It was also evident that during the fifteen dust storms this particular area appeared to supply more sediment to the air stream than any other section of the study grid.

Comparison of the Mean Daily and Dust Storm Flow Rates

Chepil (1945) has concluded from his field observations that the proportion of sediment transported in creep can vary considerably from one soil type to another, but for any one soil type the proportion is independent of wind speed. This also appears to be the case in the Slims River Valley in that no significant relationship exists between shear velocity and per cent creep transport for either the mean daily or the dust storm observations. There is, however, a significantly smaller proportion of sediment transported in creep during the dust storms which usually occur when shear velocities are comparatively high. (Fig. 4.6 and Table 5.2).

During the fifteen dust storms the proportion of sediment transported in creep accounted for 2.3 to 7.9 per cent (mean, 4.6 per cent) of the total sediment flow in saltation and creep. This percentage of creep transport is significantly lower ($t = 2.95$ t for 0.01) than the proportion found from the mean daily creep flow observations (1.02 to 35.55 per cent, mean 8.9 per cent).

This difference is probably not related to the higher shear velocities during the dust storms, but more likely results from an underestimation of the creep flow rate by the creep trap.

It is thought that underestimation may have resulted from a greater degree of air turbulence around the intake orifice of the creep trap during the dust storms when shear velocities were relatively high. Since the particles transported in creep on the Slims River delta are relatively small (Tables 4.1 and 5.9) the turbulent eddies may have deflected some of the sediment particles away from the intake orifice. Thus, the amount of sediment collected in the trap would underestimate the true creep flow rate. Belly (1964) has also found that this type of trap design underestimates the sediment flow rate and suggests the underestimation may also be caused by the scour of sediment from around the lip of the intake orifice.

The relationship between sediment transport in saltation and creep (q) and shear velocity (U_*') was similar for both the mean daily and dust storm observations. The mean daily relationship can be expressed by

$$q = 8.88 \times 10^{-4} \, U_*'^{\,3.1074} \qquad \text{..............5.3}$$

and the dust storm relationship by

$$q = 8.44 \times 10^{-3} \, U_*'^{\,2.4502} \qquad \text{..............5.4}$$

where q = sediment flow in saltation and creep (mg/cm.s)

U_*' = shear velocity (cm/s).

The above relationships have been plotted in Fig. 5.14 for the range of shear velocities encountered during the observation period. The curves are similar and can be approximated by Bagnold's (1941) theoretical formula. It is also apparent that the slope of the curve for the dust storm observations is significantly less than that for the daily observations.

Bagnold (1941) and Zingg (1952) have suggested on the basis of their theoretical calculations and wind tunnel observations that sediment transport in saltation and creep varies with the cube of shear velocity (equations 4.2 and 4.3). Belly (1964), however, has found from his wind tunnel investigations that in the case of very fine sands, the slope of the q against U_*' relationship tends to decrease with decreasing grain size. He suggests that when the surface sediments are relatively small, a greater proportion of the sediment is transported in suspension and consequently not collected in the creep and saltation traps. A similar argument can be put forward for the lower slope of the q, (saltation and creep) against U_*' (shear velocity) curve for the dust storm observations.

During the dust storms the amount of sediment transported in creep, saltation and suspension was measured. The relative rates of transport in each of the three types of movement are given in Table 5.5. The proportion of sediment transported in suspension was relatively high, ranging from 12.7 to 65.5 per cent (mean, 46.4 per cent) of the total sediment flow. Thus, the loss of this large proportion of sediment in suspension may explain why the slope (b = 2.4502) of the q against U_*' curve (Fig. 5.14) is somewhat lower than Bagnold's

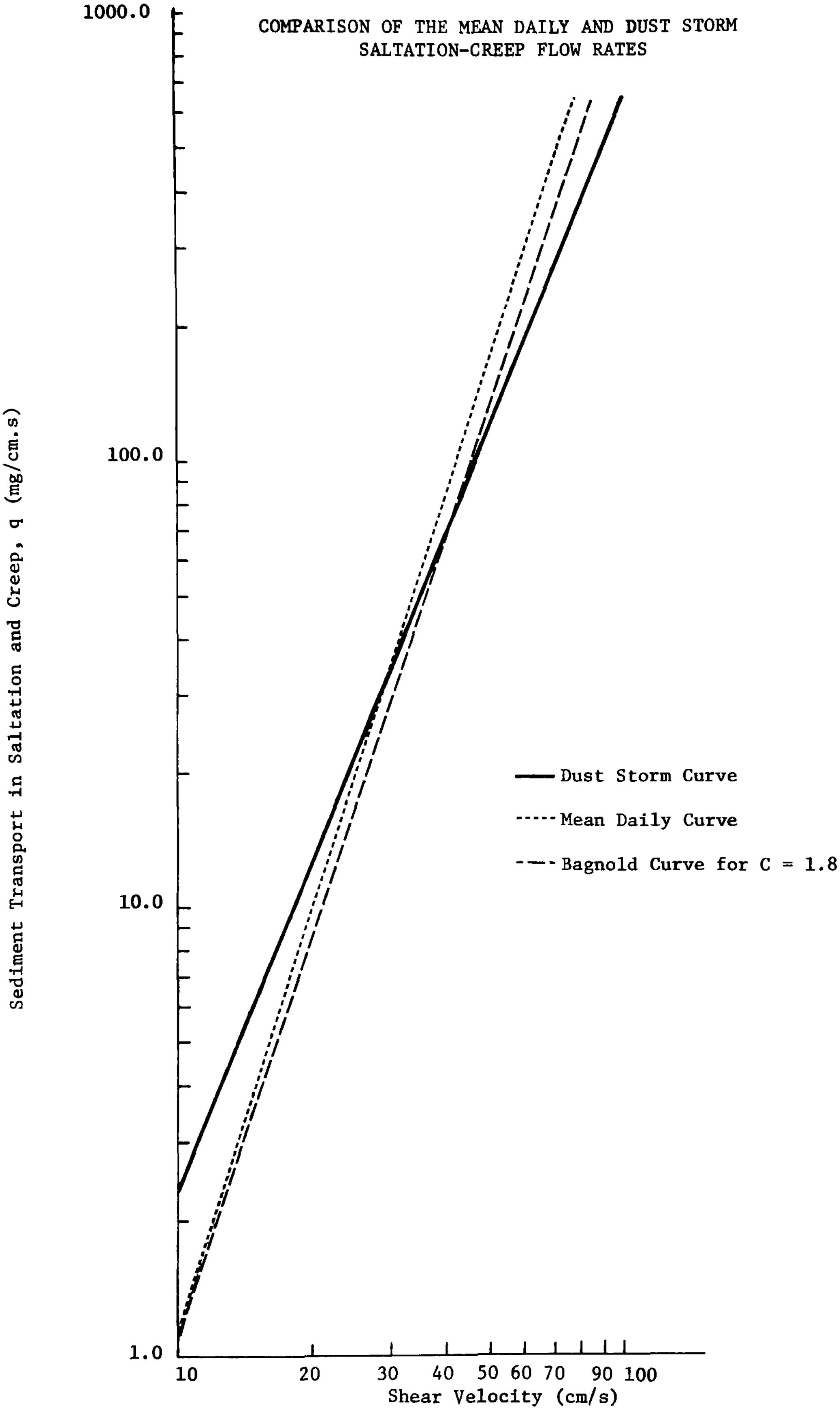

1000.0
COMPARISON OF THE MEAN DAILY AND DUST STORM
SALTATION-CREEP FLOW RATES
100.0
10.0
1.0
Sediment Transport in Saltation and Creep, q (mg/cm.s)
Dust Storm Curve
Mean Daily Curve
Bagnold Curve for C = 1.8
10
20
30
40
50
60
70
90
100
Shear Velocity (cm/s)

FIGURE 5.14

theoretical value of 3.0.

Although suspended sediment transport was not measured during the mean daily observations the lack of large plumes of dust indicate that the amount of sediment transported in suspension was considerably less than the amount transported during the dust storms. This probably accounts for the fact that the slope of the mean daily q against U_*' curve approximates Bagnold's (1941) theoretical value.

Despite the fact that a relatively large amount of sediment was transported in suspension (12.5 to 65.5 per cent) during the dust storms, the greatest proportion was transported in saltation (29.0 to 85.3 per cent) with a very small amount being moved in surface creep (1.3 to 3.6 per cent, Table 5.5). These values compare favorably with field observations made by Chepil (1945) over loam soils in southern Saskatchewan (Table 5.6). The slightly greater proportion of sediment transport in suspension and lower proportion in surface creep is directly related to the fact that surface sediments of the Slims River delta are somewhat finer textured than the soils investigated by Chepil (1945).

Grain Size Characteristics of Sediment Transported During the Dust Storms

Grain size analysis was carried out on all suspended sediment samples collected during the dust storms. This analysis was done on a Quantimet 720 in the Department of Geological Sciences, Brock University. The first three moments of the grain size distribution and the per cent volume in each size class are given in Table 5.7 and Appendix F.

TABLE 5.5

PROPORTION OF SEDIMENT TRANSPORTED IN CREEP, SALTATION AND SUSPENSION DURING THE DUST STORMS

Storm No.	Creep %	Saltation %	Suspension %	Total Flow, Q (mg/cm.s)
1	2.7	52.7	44.6	50.5
2	2.6	46.5	50.9	45.1
3	1.7	29.0	69.3	150.3
4	2.5	68.3	29.2	161.0
5	2.0	85.3	12.7	98.5
6	3.3	74.7	22.0	176.3
7	1.9	52.4	45.7	284.9
8	3.6	67.7	28.7	186.6
9	1.9	33.2	64.9	109.1
10	1.9	42.5	55.6	521.1
11	1.3	41.8	56.9	347.9
12	1.8	64.9	33.3	374.8
13	3.6	42.2	54.2	208.9
14	2.2	32.3	65.5	334.5
15	2.1	36.3	61.6	130.5
Mean	2.3	51.3	46.4	

TABLE 5.6

PROPORTION OF SEDIMENT TRANSPORTED IN CREEP, SALTATION AND SUSPENSION OVER DIFFERENT SOIL TYPES

	Creep %	Saltation %	Suspension %
Sceptre heavy clay	24.9	71.9	3.2
Haverhill loam	7.4	54.5	38.1
Hatton fine sandy loam	12.7	54.7	32.6
Fine dune sand	15.7	67.7	16.6

Size Distribution of Particles

Particle size (mm)--	0.83-0.48	0.42-0.25	0.25-0.15	0.15-0.1	0.1
Sceptre heavy clay	33.5	46.1	14.9	4.0	1.5
Haverhill loam	13.3	17.3	15.1	22.3	32.0
Hatton fine sandy loam	1.1	6.0	26.4	40.5	26.0
Fine dune sand	0.1	2.1	54.2	35.6	8.0

(after Chepil 1945a)

respectively.

In general the suspended sediments can be classified as moderately sorted, medium to coarse silts with strong positive skewness (Folk, 1966). Although the relatively small size of the suspended sediments is not unusual it does indicate the efficacy of wind as selective transporting agent if one considers the nature of source material in this particular case. As can be seen in Table 5.8 the mean size of the surface sediments range from coarse silts to very fine sands. Student's t tests between sediment sizes indicates that the suspended sediments are significantly finer than the surface sediments at the one per cent confidence level ($t = -6.09 < t$ for 0.01). Thus, it would appear that even though the mean size of the surface sediment is also relatively small the wind is able to preferentially remove the finer silts in suspension. Similarly the selective removal of the finer silts from the surface results in the suspended sediment being significantly better sorted than the surface sediments (mean standard deviation: surface = 1.35ϕ, suspended = 0.85ϕ, $t = 7.92 > t$ for 0.01).

From Tables 5.7 and 5.8 it can be seen that both the suspended and surface sediments are characterized by strong positive skewness. Friedman (1966) has suggested that positive skewness in sediments usually occurs when the flow of the transporting fluid is basically unidirectional. He also suggests that the largest particles which can be carried in saltation or suspension is governed by the competency of the transporting fluid. This critical limitation, however, does not seriously affect the finer particles in the distribution. This can be demonstrated if one considers a quantity of sediment with a symmetrical

TABLE 5.7

GRAIN SIZE ANALYSIS OF THE SUSPENDED SEDIMENT COLLECTED DURING DUST STORMS

Sample Number	Mean (ϕ)	Standard Deviation (ϕ)	Skewness
11*	4.14	0.73	0.90
12	4.66	0.57	2.46
13	4.80	0.71	1.84
14	4.91	0.90	2.17
15	5.24	0.90	0.98
16	6.11	1.00	1.25
17	6.28	1.13	1.07
21	4.23	0.87	0.95
22	4.30	0.84	0.75
23	4.39	0.78	0.50
24	4.43	0.80	0.93
25	4.91	0.71	1.42
26	5.73	0.67	2.19
27	6.41	1.19	0.96
28	6.97	1.98	-0.78
31	3.99	0.72	1.12
32	4.24	0.92	1.11
33	4.21	0.76	0.71
34	4.20	0.69	0.80
35	4.31	0.75	0.55
36	4.35	0.79	0.76
37	5.10	0.84	1.35
38	5.89	0.81	1.87

*Sample number 11
dust storm 1
height 1

	Ht. (m)
1	0.5
2	1.0
3	1.5
4	2.0
5	4.0
6	6.0
7	9.0
8	12.0

...continued

TABLE 5.7--continued

Sample Number	Mean (ϕ)	Standard Deviation (ϕ)	Skewness
41	4.29	0.88	0.85
42	4.31	0.92	0.63
43	4.48	0.80	0.84
44	4.78	0.76	2.40
45	4.90	1.01	1.90
46	4.85	0.63	1.28
47	4.90	0.74	1.95
48	5.80	1.28	0.25
51	4.41	0.94	1.22
52	4.45	1.09	0.88
53	4.91	0.65	1.57
54	5.08	0.91	1.46
55	5.00	0.83	1.34
56	5.35	0.88	0.95
57	5.47	0.90	0.87
58	5.93	0.76	1.46
61	4.75	0.67	2.12
62	4.44	0.89	0.81
63	4.50	0.88	0.89
64	4.71	0.59	2.08
65	4.82	0.63	1.52
66	4.87	0.69	1.90
67	6.00	1.28	0.03
71	4.19	0.99	1.24
72	4.16	0.89	1.30
73	4.16	0.81	1.00
74	4.33	0.90	0.91
75	4.35	0.94	1.07
76	4.62	0.95	0.75
77	5.13	0.94	1.39
78	5.44	0.88	1.10
81	4.38	0.94	1.04
82	4.54	0.93	1.20
83	4.88	0.76	1.97
84	4.79	0.72	2.15
85	4.79	0.69	1.97
86	5.14	0.90	1.41
87	5.29	0.90	1.33
88	5.32	0.95	1.20

...continued

TABLE 5.7--continued

Sample Number	Mean (ϕ)	Standard Deviation (ϕ)	Skewness
91	4.87	0.77	1.76
92	4.89	0.76	1.70
93	4.98	0.81	1.45
94	4.96	0.86	1.81
95	5.05	0.88	1.49
96	5.24	0.93	1.14
97	5.55	0.85	1.20
98	5.31	0.87	1.05
101	4.20	1.07	1.39
102	5.02	0.77	1.49
103	5.21	0.86	1.27
104	5.28	0.96	1.24
105	5.27	0.93	1.09
106	5.40	0.84	1.01
107	5.34	0.93	1.12
111	3.97	0.89	1.66
112	4.45	0.92	0.79
113	4.77	0.66	1.88
114	4.80	0.70	2.03
115	4.90	0.82	1.85
116	5.10	1.01	1.40
117	5.52	0.93	0.74
118	5.62	0.93	0.99
121	4.16	1.05	1.24
122	4.83	0.76	2.02
123	4.98	0.70	1.35
124	5.02	0.76	1.21
125	5.09	0.87	1.44
126	5.08	0.91	1.48
127	5.29	1.01	1.16
128	5.46	0.93	0.77
131	4.84	0.65	1.61
132	5.01	0.82	1.48
133	5.08	0.82	1.42
134	5.02	0.79	1.50
135	5.18	0.81	1.19
136	5.12	1.02	1.61
137	5.11	0.68	0.80
138	5.32	0.89	1.06

...continued

TABLE 5.7--continued

Sample Number	Mean (ϕ)	Standard Deviation (ϕ)	Skewness
141	4.39	1.03	0.86
142	4.92	0.77	1.77
143	5.09	0.83	1.28
144	5.28	0.76	1.64
145	5.05	0.73	1.12
146	5.46	0.78	0.98
147	5.86	1.03	0.74
148	5.68	0.91	1.06
151	4.55	0.90	0.39
152	4.95	0.71	1.53
153	5.04	0.84	1.45
154	5.19	0.86	1.16
155	5.06	0.69	1.18
156	5.34	0.88	1.13
157	5.80	0.73	2.08
158	5.38	0.90	1.47

TABLE 5.8

GRAIN SIZE ANALYSIS OF SURFACE SEDIMENTS

Sample Number	Mean (ϕ)	Standard Deviation (ϕ)	Skewness
14MY2*	4.29	1.10	2.27
14MY8	4.45	1.42	1.24
14MY9	3.98	1.35	0.83
21MY2	4.06	1.26	1.64
21MY8	4.60	1.12	0.12
21MY9	3.78	1.46	1.19
28MY2	4.41	1.57	0.45
28MY8	4.68	1.18	1.83
28MY9	3.69	1.58	1.47
03JN2	4.41	1.14	0.71
03JN8	4.69	1.23	1.47
03JN9	3.54	1.56	0.83
10JN2	4.43	1.28	-0.13
10JN8	4.56	1.23	1.77
10JN9	3.58	1.41	1.93
17JN2	4.25	1.24	0.82
17JN8	4.60	1.23	0.80
17JN9	3.63	1.30	2.08
24JN2	4.28	1.22	0.69
24JN8	4.65	1.33	0.35
24JN9	3.52	1.45	1.55
01JY2	4.34	1.57	0.48
01JY8	4.45	1.66	0.61
01JY9	3.73	1.67	1.62
07JY2	4.12	1.26	1.46
07JY8	4.80	1.42	0.73
07JY9	3.78	1.26	2.19

*14MY2 sample collected May 14, 1973
from grid intersection 2

distribution.

If this sediment is eroded by a basically unidirectional transport system (i.e. eolian or fluvial system), the grain size distribution of the initial sediment will be modified depending on the competence of the transporting fluid. If the fluid is only able to transport the finer grain sizes, the initial sediment will only have the fine tail of the symmetrical distribution removed and will therefore become negatively skewed. Similarly the frequency distribution of the transported sediment will lack a "tail" of coarse material which was too large to be transported by the available energy. Thus, the transported sediment will tend to be positively skewed.

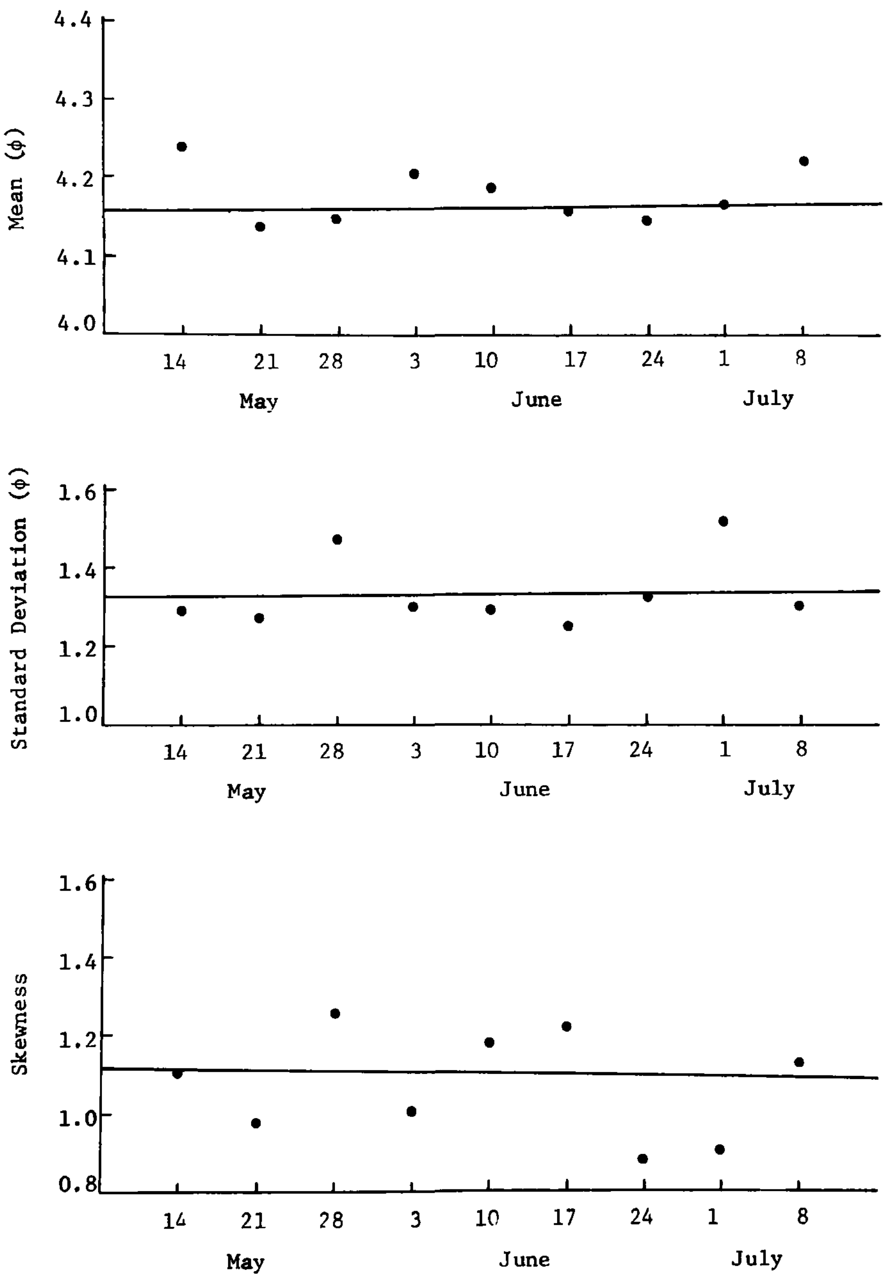
It is probable, therefore, that the strong positive skewness of the delta silts in this particular investigation is directly related to the selective transport of sediment by the Slims River in a manner similar to that just discussed. Since eolian transport systems are less competent than fluvial systems one might expect that any eolian sediment derived from a fluvial deposit would also tend to be positively skewed. It could be further argued that since the wind is a less competent transport system the suspended sediment would be more positively skewed than the initial sediments because of a greater degree of selectivity. A Student's t test between means of skewness values indicate that the suspended sediment is significantly more positively skewed at the 0.01 confidence level ($t = 2.79 > t$ for 0.01). Following the above argument this may also imply that the surface sediment should tend to become less positively skewed over time as long as no new sediment is added to the surface of the delta as a result of flooding by the Slims River.

The surface samples used in the above analysis were obtained by taking samples from the same three grid intersections each week (intersections 2, 8 and 9; Fig. 14.). In Fig. 5.15 the mean values of the first three moments for each week have been plotted against time. In each case the slope of the regression line is not significantly different from zero at the 0.05 per cent confidence level. This indicates that there was not a significant change in the grain size characteristics of the surface sediment during the observation period. It should be realized however, that this is in part due to the scatter of the data which results from the variability of the grain size distribution of the surface sediments.

It is thought that the reason for the lack of change in the grain size characteristics of the surface sediments may be related to the manner in which sediment is eroded from the surface. Fig. 5.16 shows the average frequency distribution for the surface sediments as well as average distributions for the sediment transported in suspension, saltation and creep during the dust storms. As can be seen from this figure and Table 5.9 the sediment transported in creep is characterized by slight positive skewness and the material transported in saltation by slight negative skewness. These types of distributions are similar to those observed during daily sediment transport (Table 4.1).

A series of Student's t tests (Table 5.10) were run to compare the means and standard deviations of the surface sediments and the sediment transported during the dust storms in saltation, creep and suspension. This table indicates that the mean sizes of the various sediments are all significantly different from one another. That is, the sediment

CHANGES OF THE FIRST THREE MOMENTS
OF THE GRAIN SIZE DISTRIBUTION OVER TIME



Mean

$\bar{\bar{x}} = 4.16 + 0.0117T$

$r = 0.29$

Standard Deviation

$\sigma = 1.31 + 0.0107T$

$r = 0.30$

Skewness

$s = 1.13 - 0.0118T$

$r = -0.23$

FIGURE 5.15

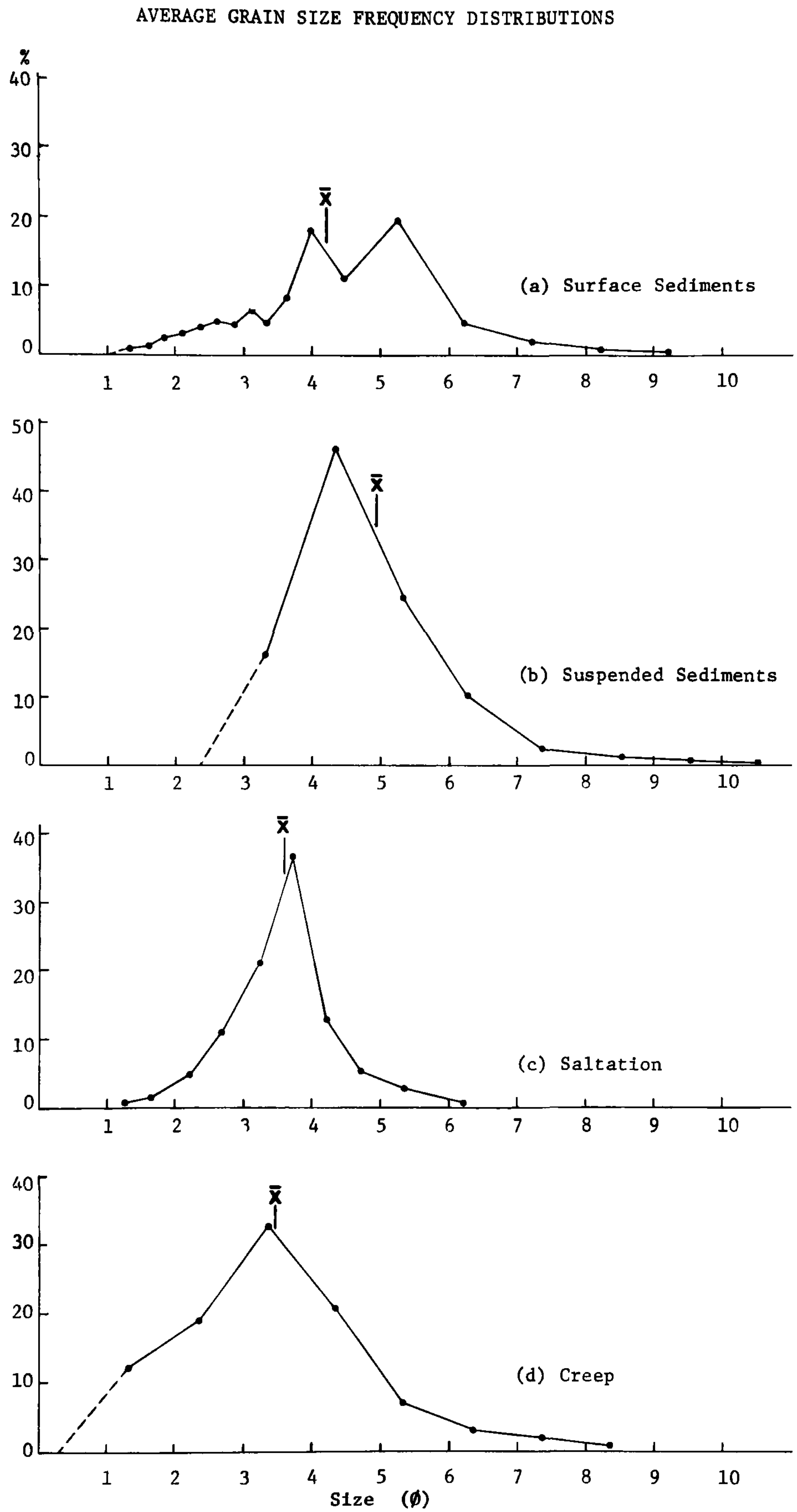

AVERAGE GRAIN SIZE FREQUENCY DISTRIBUTIONS
%
(a) Surface Sediments
(b) Suspended Sediments
(c) Saltation
(d) Creep
Size (ø)

FIGURE 5.16

TABLE 5.9

GRAIN SIZE ANALYSIS OF SEDIMENT
TRANSPORTED IN CREEP AND SALTATION
DURING DUST STORMS

Sample Number	Mean (ϕ)	Standard Deviation (ϕ)	Skewness
	Creep		
C01S*	3.76	1.66	0.82
C02S	3.34	1.29	1.07
C03S	3.72	1.76	0.71
C04S	3.50	1.31	0.83
C05S	3.73	1.00	0.72
C06S	3.88	0.98	0.91
C07S	3.57	0.95	0.20
C08S	3.34	1.90	0.69
C09S	3.42	1.55	0.92
C10S	3.31	1.46	1.15
C11S	2.93	1.51	1.17
C12S	3.45	1.65	1.01
C13S	2.83	1.32	1.04
C14S	3.69	1.14	0.23
C15S	3.84	1.54	0.65
	Saltation		
S01S	3.83	0.75	0.30
S02S	3.46	0.65	-0.76
S03S	3.72	0.80	-0.08
S04S	3.61	0.80	-0.23
S05S	3.73	0.97	-0.01
S06S	3.92	0.90	0.02
S07S	3.66	0.95	-0.05
S08S	3.43	0.55	-1.16
S09S	3.51	0.42	-0.65
S10S	3.42	0.95	-0.44
S11S	3.38	0.70	-0.42
S12S	3.54	0.44	-1.19
S13S	3.32	0.63	-0.11
S14S	3.67	0.77	-0.23
S15S	3.84	0.90	0.59

*C01S surface creep, dust storm 1.

TABLE 5.10

CALCULATED t VALUES

Comparison of Mean Sizes

	Creep	Saltation	Suspension	Loess
Surface	12.13*	11.07*	17.71*	8.01*
Creep		1.94**	10.26*	7.49*
Saltation			9.62*	7.48*
Suspension				1.22

Comparison of Standard Deviations

	Creep	Saltation	Suspension	Loess
Surface	1.47	22.93*	34.80*	3.49*
Creep		7.08*	10.53*	1.49
Saltation			2.66*	3.52*
Suspension				5.47*

Significant at: 0.01 confidence level*
0.05 confidence level**

transported in creep is significantly coarser than that transported in saltation, both of which are coarser than the surface sediments. Conversely the suspended sediment is significantly finer than the surface sediments and the material transported in saltation and creep.

Even though the mean sizes of the various sediments are significantly different, the distributions (Fig. 5.16) are not at all mutually exclusive but overlap considerably especially in the 3.0ϕ to 4.0ϕ size range. This overlap is most pronounced if one compares the creep and saltation frequency distributions. Thus, it would appear that sediment in this particular size range can be transported in either saltation or creep. This type of overlap is probably related to variations in wind and surface conditions during sediment transport and the fact that the surface sediments are themselves fine grained.

It is apparent that a large proportion of the coarse tail of the surface sediments is transported in creep and to a lesser extent in saltation. Similarly, the fine tail and modal sizes found in the surface sediments appear to be transported in suspension, saltation and creep. It is suggested, therefore, that because the surface sediments are fine grained all sizes found in the surface distribution will be removed over time by sediment transport in creep, saltation and suspension within the normal range of surface wind speeds found in the Slims River Valley. Since all the surface material can be removed one could suggest that the grain size characteristics of the surface would not change significantly over time.

One peculiarity found in this investigation is the fact that all the distributions are positively skewed except for saltation. This is somewhat unusual in that one would expect the saltation frequency distribution to be positively skewed because the sediment transport is also basically unidirectional (Friedman, 1966). This negative skewness might be explained if the transport in saltation is considered from another point of view. A fundamental difference exists between the initiation of movement and transport of sediment in creep to that in saltation and suspension. The grains moving in saltation and suspension receive their momentum directly from the pressure of the wind as soon as the grains are lifted from the surface (Bagnold, 1941). Grains moving in creep, however, are not lifted from the surface and receive the greatest percentage of their momentum from the impact of saltating grains. It is also of importance to note that grains transported in saltation are not usually lifted more than a metre above the surface (Bagnold, 1941; Chepil, 1945; Sharp, 1964). Conversely grains transported in suspension can be carried aloft several hundred metres (Chepil, 1945).

At the moment when a grain is lifted from the surface it can either be transported in saltation or suspension. If the gravitational acceleration acting on the particle exceeds the vertical acceleration of the wind, the particle will return to the surface, thereby having been transported in saltation. Conversely if the particle is kept aloft by the vertical component of the wind it is carried in suspension. The probability of a given particle being carried in either saltation or suspension will primarily be a function of the particle's mass and shape as well as the wind velocity and the surface conditions at the moment of entrainment.

At a given instant in time thousands of grains will be lifted from the surface if the wind velocity exceeds the threshold velocity. It is suggested that the size distribution of these particles at the moment of entrainment will tend to be positively skewed because the larger particles of the surface distribution will not be lifted into the air stream because of their mass. A certain percentage of the grains initially entrained into the air stream will be of a small enough mass to be transported in suspension at that particular wind velocity. The remaining particles which are too heavy will fall back to the surface and will therefore have been transported in saltation. It is also apparent that a large percentage of the grains transported in suspension will be derived from the fine tail of the distribution of grain sizes initially entrained into the air stream. Thus, the grain size distribution of the suspended material would be positively skewed because it would lack the tail of coarse grains which were too heavy to be carried into suspension. Conversely the remaining material transported in saltation would lack the tail of fines which was removed in suspension and would therefore tend to become less positively or possibly even negatively skewed. Tables 5.7 to 5.9 would tend to support this argument in that sediment transported in saltation is characterized by weak negative skewness and the suspended sediment by strong positive skewness.

The grain size characteristics of the suspended sediment collected during the dust storms were compared with those of the loess deposited just above the flood plain on the walls of the Slims River Valley. These deposits vary in thickness from approximately 0.5 to 2.0 metres, the thickest deposits being found near the terminus of the Kaskawulsh Glacier.

Nickling (1972) has shown that the loess deposits in this area are represented by a distinct size frequency distribution which differs somewhat from similar deposits just outside the valley in the Shakwak Trench. The eolian sediments in the Shakwak Trench, however, were primarily derived from the outwash deposits in the Slims River Valley. Nickling (1972) has suggested that the trench sediments with other fine grained sediment derived from the fluvial outwash of the many creeks which enter the southern end of Kluane Lake. It is also argued that differing wind conditions in the Slims Valley and the Shakwak Trench may also be partly responsible for the difference in the grain size characteristics.

The first three moments of the grain size distribution for fifteen loess samples collected in the Slims River Valley are given in Table 5.11.

Student's t test (Table 5.10) indicates that although the mean size of the suspended sediment and the loess are not significantly different the suspended sediment is significantly better sorted. It is thought that the poorer sorting in the loess deposits may be caused by the reworking of these sediments after their deposition.

Nickling (1972) has found that the local environment in which eolian sediment is deposited can affect the grain size characteristics of the sediment during the process of and subsequent to its deposition. For example, it was found that the loess deposits near the shore of Kluane Lake were more poorly sorted and in some cases considerably coarser than similar deposits a few hundred metres back from the shore.

TABLE 5.11

GRAIN SIZE ANALYSIS OF
SLIMS RIVER VALLEY LOESS DEPOSITS

Sample Number	Mean (ϕ)	Standard Deviation (ϕ)	Skewness
4	5.25	1.10	1.55
5	5.20	0.99	1.06
6	4.61	1.97	0.55
8	3.45	1.56	1.31
14	5.30	1.52	1.04
16	4.67	0.76	1.52
17	4.93	0.60	0.36
18	4.81	0.88	1.61
19	4.44	1.01	0.99
20	4.77	1.50	1.08
21	5.29	1.01	1.11
22	4.99	0.88	2.40
23	5.20	1.01	1.44
24	3.84	1.95	1.01

(after Nickling, 1972)

In these cases it appeared that during the deposition of the loess, fine sand derived from the Kluane Lake beach had been blown back from the shore a short distance and intermixed with the loessial material. Such intermixing was also found on several occasions near tributary creeks where downcutting had exposed till deposited prior to the loess.

It is felt that intermixing of this nature may be the reason why the loess deposits are not as well sorted as the suspended sediment. The loess deposits on the lower walls of the Slims River Valley often overlie till deposited during the Kluane Glaciation (Muller, 1967). It is possible that some intermixing could have taken place between the loess and exposed till as a result of eolian activity. Relatively fine grained sediment could also be derived from other sources such as the outwash fans of the many small creeks which flow into the Slims River.

Chepil and Woodruff (1957) found that in dust storms in Kansas and Colorado the mean diameter of the suspended sediment decreased as a power function of height. In order to investigate whether a similar relationship occurred in the Slims River Valley dust storms, the mean size of the collected suspended sediment has been plotted against height (Fig. 5.17). The mean particle size does appear to decrease with height but in a somewhat irregular manner. It is also evident that there is a great deal of variation from one storm to another. A similar situation was also encountered by Chepil and Woodruff (1957). Such variability is not surprising if one considers the variations in atmospheric conditions over the duration of a given storm and from one storm to another. In an attempt to arrive at a more general relationship the data for the fifteen storms was averaged and plotted (Fig. 5.18).

VARIATION OF MEAN SIZE WITH HEIGHT

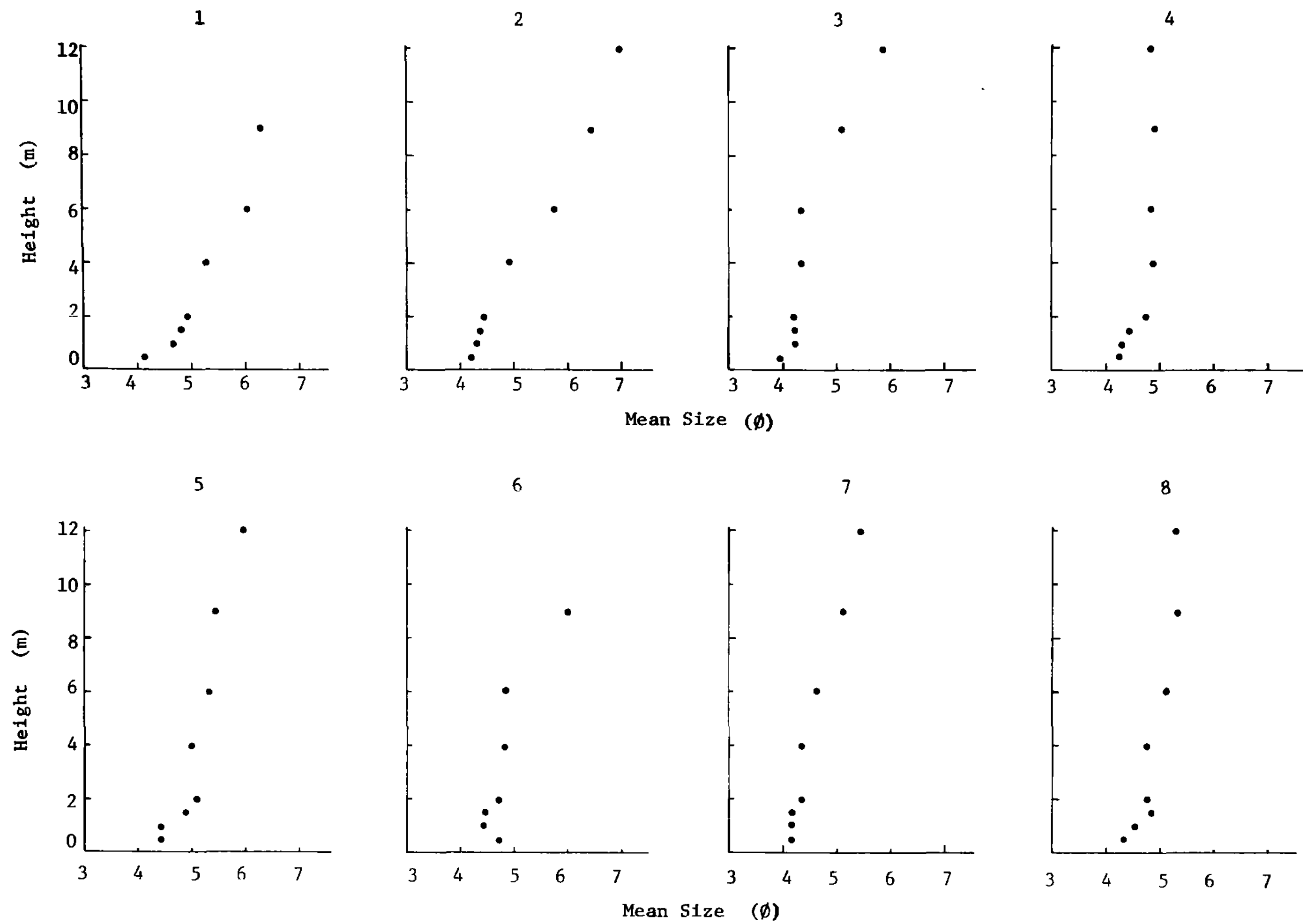


FIGURE 5.17

VARIATION OF MEAN SIZE WITH HEIGHT

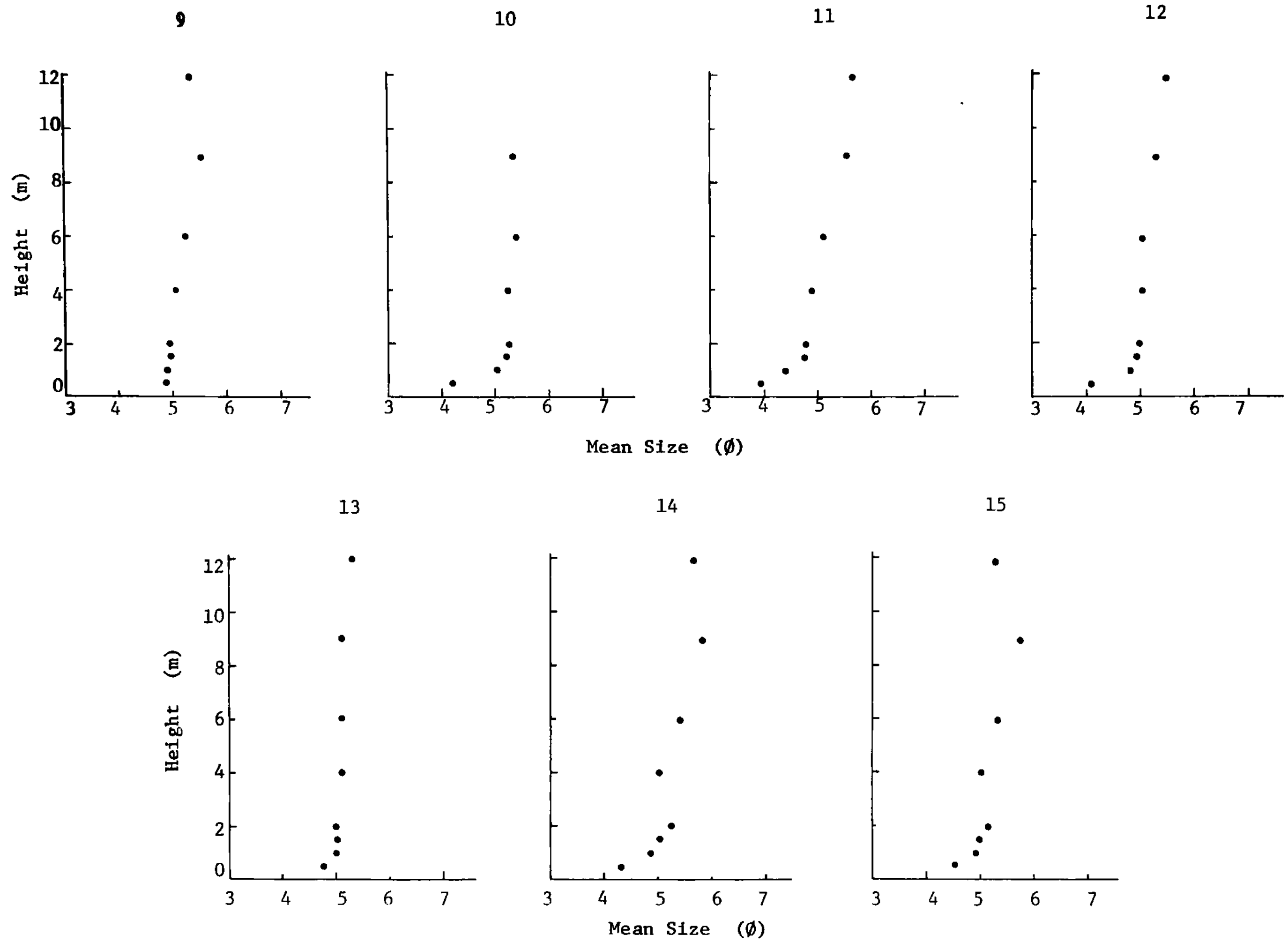


FIGURE 5.17 (continued)

VARIATION OF MEAN SIZE AND SKEWNESS WITH HEIGHT

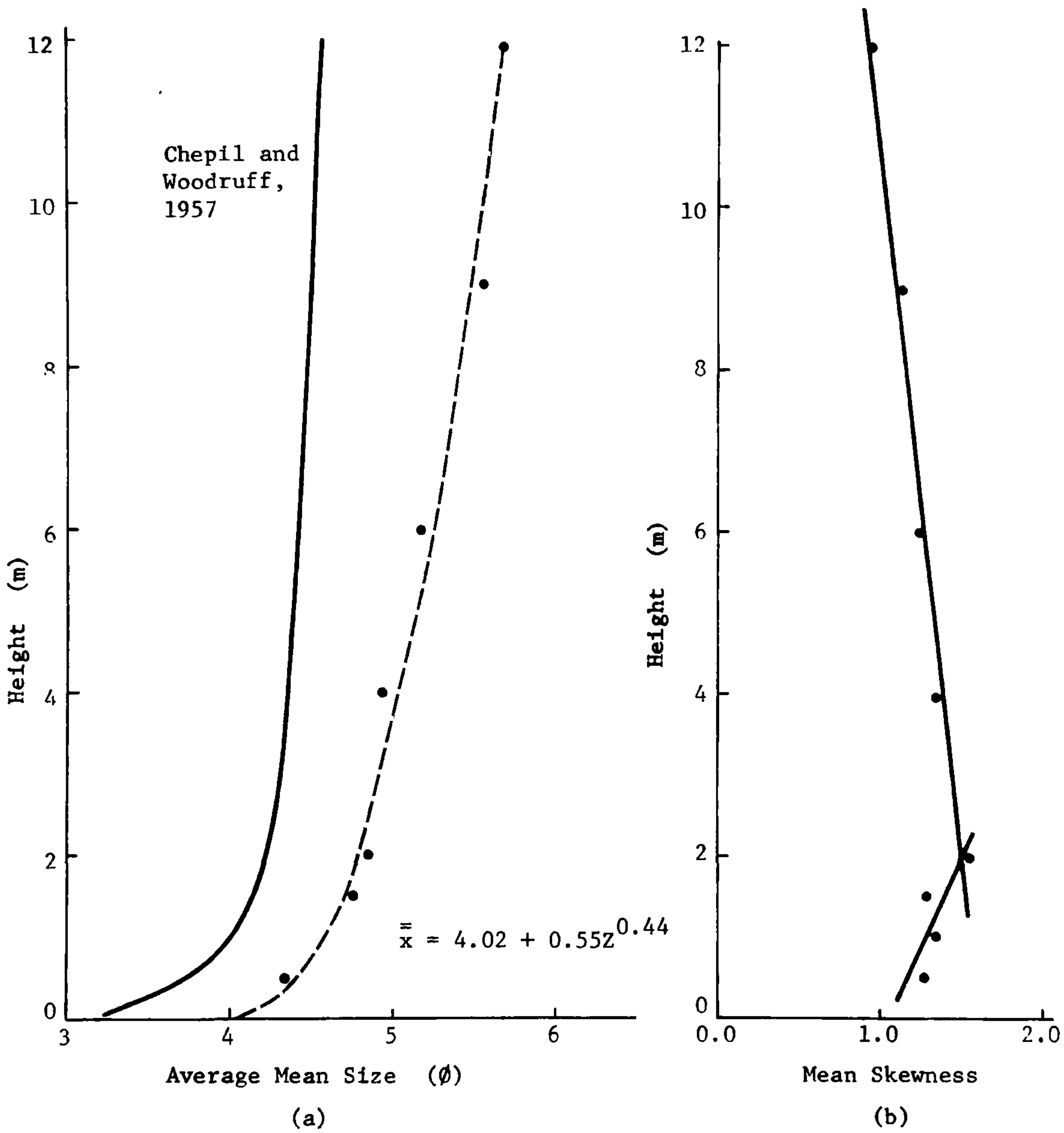


FIGURE 5.18

As can be seen, the data points can be fitted quite closely by the power function

$$d = 4.02 + 0.55\ H^{0.44}$$

where d = mean diameter (ϕ)

H = height (m)

The average size-height relationship found by Chepil and Woodruff (1957) for the dust storms in Kansas and Colorado is also given in this figure. Although the curves are not identical, the data in both cases are best fit by a power function. It is evident from a comparison of the curves that at any given height the suspended sediment transported in the Slims River Valley consists of smaller grains. It is also apparent that the size of transported sediment decreased much less rapidly with height during the dust storms in Kansas and Colorado. Several reasons for these differences can be given.

The size of the sediment transported at any given height is a function of many interrelated factors such as the wind velocity, degree of air turbulence and the grain size distribution of the eroding surface. For example, if the eroding surface is comprised mainly of very fine silt one would expect the average size of the suspended sediment transported above this surface to be as small as or smaller than the mean size of the surface sediments. Similarly, if a large amount of coarse silt was available from the surface and the wind competent enough to transport it in suspension, the size of the suspended sediment transported above the surface would be relatively coarser than in the previous case. As previously discussed (p. 139) the wind velocities measured during the dust storms in Kansas and Colorado were somewhat greater than those measured during the dust storms in the Slims River Valley. This may explain in part the somewhat smaller particle size of the suspended sediment transported in the Slims River Valley. It is also possible that the surface of the Slims River delta was able to supply a greater amount of finer silt to the air stream.

The less rapid decrease in mean grain size with height in the Kansas and Colorado dust storms may be related to the fact that suspended sediment in many cases had been transported a considerable distance before it reached the sampling point. This was not the case in the Slims River Valley Study. As suggested earlier (p. 141) the transport of the suspended sediment over a relatively long distance may cause the sediment to become somewhat more diffused. This appeared to be a possible reason why the suspended sediment concentrations were more uniform with height during the Colorado and Kansas dust storms. It might be argued, therefore, that because of the greater degree of mixing resulting from the longer transport distance the size of the suspended sediment might also tend to become more uniform with height.

The above relationship prompted further investigation to see if there were also significant changes in the second and third moments with height. The standard deviation and skewness values for the suspended sediment at each height for each dust storm have been plotted in Figs. 5.19 and 5.20 respectively.

The changes in standard deviation with height during each storm are somewhat irregular (Fig. 5.19). There is also a considerable degree of variation from one storm to another. In general, standard deviation appears to decrease from the surface up to approximately 1.5 metres

VARIATION OF STANDARD DEVIATION WITH HEIGHT

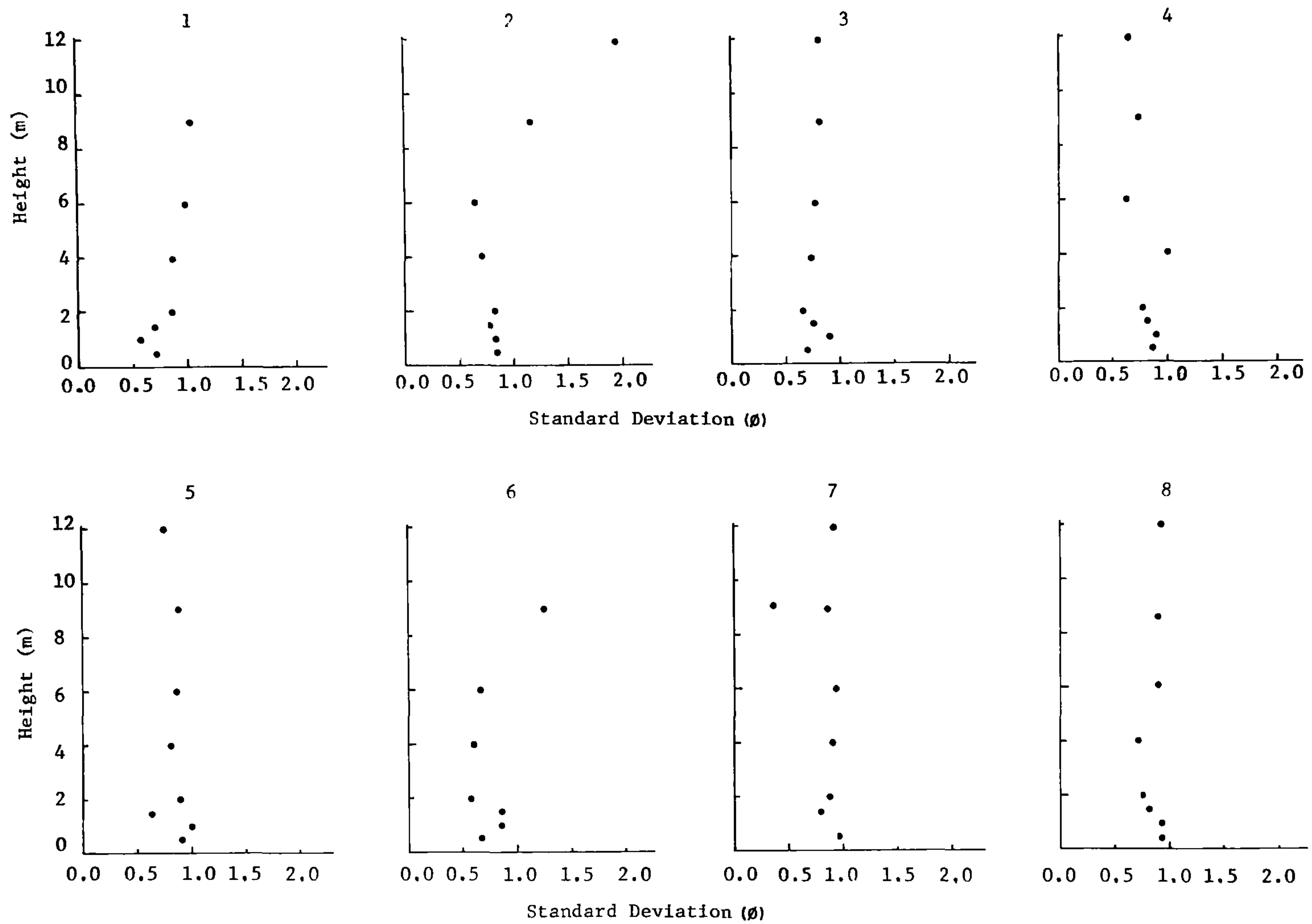


FIGURE 5.19

VARIATION OF STANDARD DEVIATION WITH HEIGHT

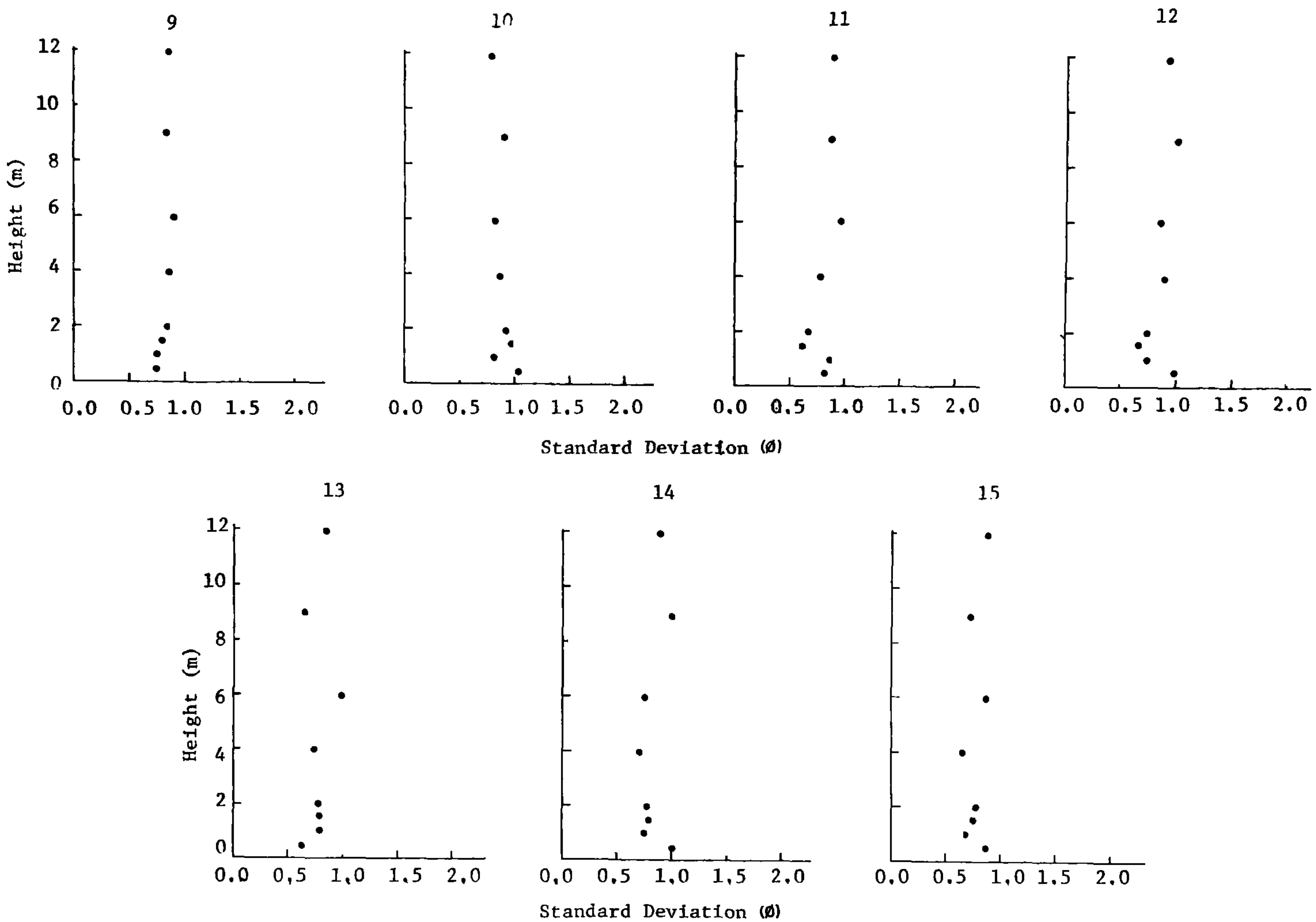


FIGURE 5.19 (continued)

VARIATION OF SKEWNESS WITH HEIGHT

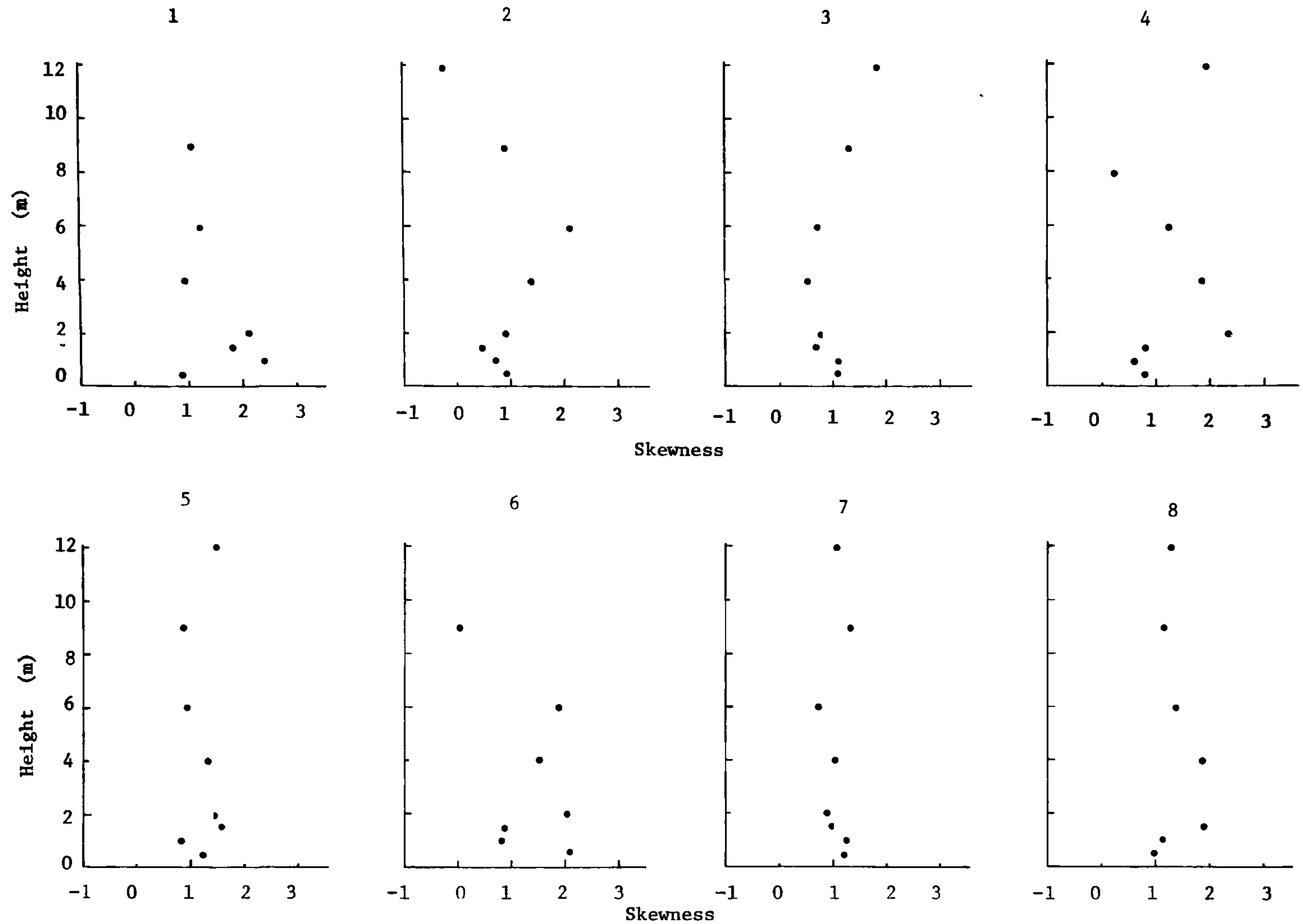


FIGURE 5.20

VARIATION OF SKEWNESS WITH HEIGHT

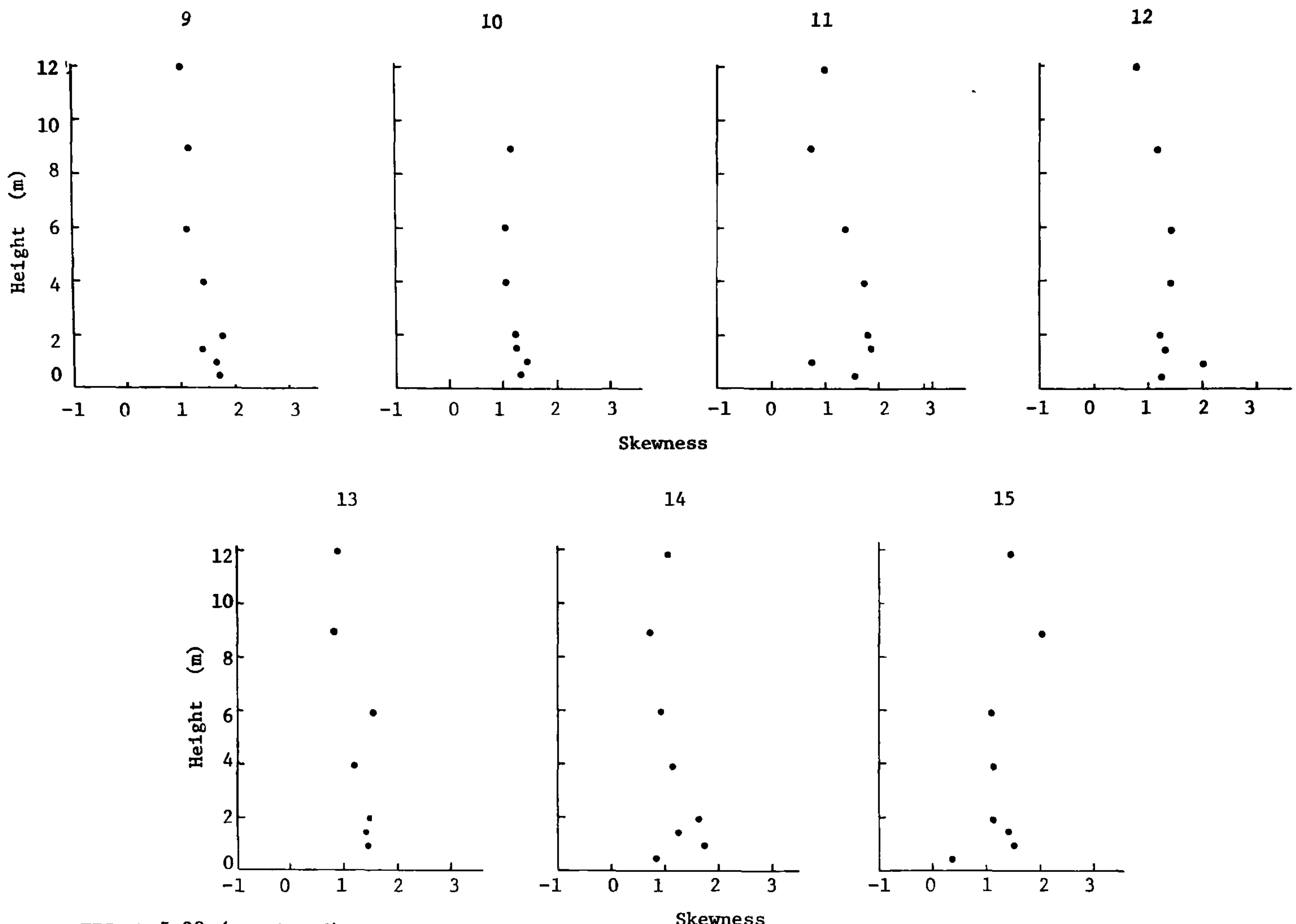


FIGURE 5.20 (continued)

after which it begins to increase. These general trends become somewhat clearer if the data for the fifteen storms are averaged and plotted (Fig. 5.21a). This figure indicates that the sorting of the suspended sediment above approximately 6.0 metres becomes increasingly more poorly sorted than sediment collected at 0.5 metres. Folk (1966) has shown that the standard deviation of different sediments can not always be directly compared because sorting (i.e. standard deviation) is a rather closely controlled sinusoidal function of mean size. Thus, in any body of data, the magnitude of the standard deviation is at least in part controlled by the magnitude of the mean. Gregory (1973) suggests that a more reliable method of comparing the standard deviations of data with varying means is the use of the coefficient of variation which can be expressed as

$$V = \frac{\mu}{\sigma} \; 100\%$$

where V = Coefficient of variation

μ = mean

σ = standard deviation

Thus this index allows for the relative comparison of the standard deviations of data groups by removing the effect of the magnitude of the means.

The coefficients of variation of the suspended sediment for each height were calculated and have been plotted in Fig. 5.21b. As can be seen from this plot the coefficient of variation increases in a linear manner above 1.5 metres. That is, above 1.5 metres the suspended sediment becomes increasingly more poorly sorted with height. Although

CHANGE OF STANDARD DEVIATION
AND COEFFICIENT OF VARIATION WITH HEIGHT

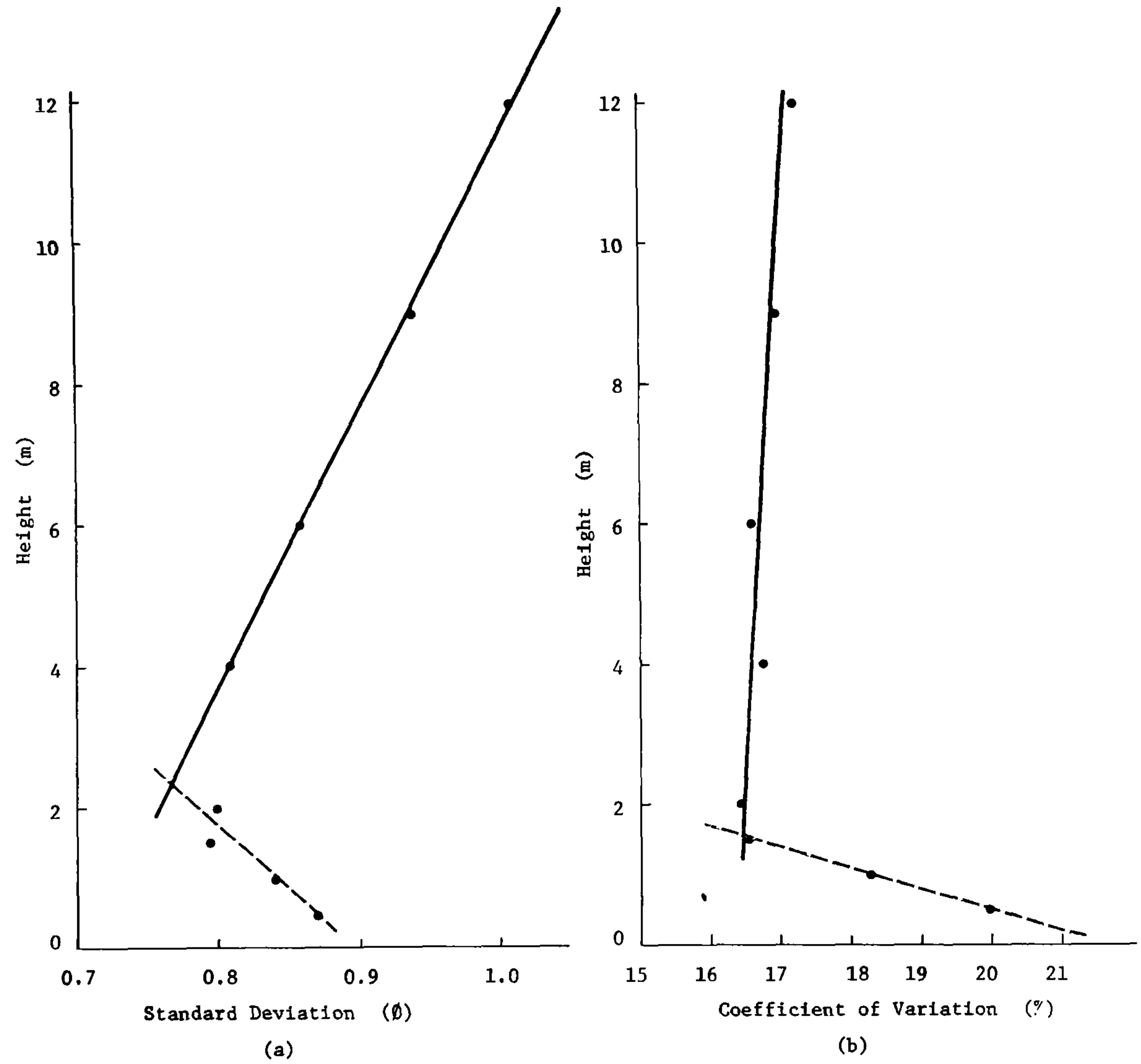


FIGURE 5.21

the slope of the regression line is significant, it is evident that the rate of change is very small. It is also apparent that the suspended sediment above 1.5 metres, in relative terms is considerably better sorted than the suspended sediment collected at 0.5 to 1.0 metres. It is also interesting that the sorting improves very sharply from the surface up to approximately 1.5 metres.

The overall shape of the coefficient of variation against height plot may demonstrate the selectivity of the wind as a transporting agent. When the wind blows across the delta surface a large number of grains can be entrained and lifted into the wind stream. Some of the grains will be too heavy to be transported in suspension and will immediately begin to fall back to the surface in saltation. However, the critical point between saltation and suspension flow for any given particle is not constant but changes from one moment to the next as a result of fluctuations in wind velocity and air turbulence. For example, a relatively large grain which is being carried in saltation may as a result of an increase in wind velocity or air turbulence move into suspension. Thus, in the distribution of grain sizes in the suspended sediment samples, there may be a number of relatively large grains which could have been transported in either saltation or suspension. This "tail" of relatively coarser grains would extend the range of grain sizes samples and result in a corresponding increase in the standard deviation (i.e. the sample would be more poorly sorted).

Bagnold (1941), Chepil (1945) and Williams (1964) have shown that sediment transported in saltation is usually not carried more than one metre in height. They have also shown that the mean size and amount of sediment transported in saltation falls off rapidly from the surface to 0.5 metres. Thus, the improvement in the sorting of the suspended sediment from the surface to 1.5 metres may result from a decrease in the mean size and number of grains which can be transported in either saltation or suspension.

The design of the suspended sediment samplers and the method of sampling may be responsible in part for the apparent improvement of sorting of the suspended sediment with height near the surface. During the dust storms sediment transport in true saltation and creep will occur between the surface and 1.0 metres. It is possible that some particles moving in true saltation would be collected in the suspended sediment samplers. This would add a coarse "tail" to the suspended sediment distribution making the near surface suspended sediment appear somewhat more poorly sorted. Since the amount of sediment transported in saltation decreases with height (most rapidly between the surface and 0.5 metres) the number of saltating grains collected in the suspended sediment sampler would also decrease with height. This may also cause an apparent improvement in sorting between the surface and approximately 1.0 metres.

The slight decrease in the sorting of the suspended sediment above 1.5 metres may be related to the fact that above this height almost all the sediment is transported in suspension. Folk and Ward (1957) have shown that in the case of very fine sediments there is a tendency for sediment to become more poorly sorted as the mean size decreases. They suggest the reason for this is that, as the mean size decreases, the transporting fluid even at relatively low velocities is more able to transport all particles in the size distribution of the sediment load. As a result there is a tendency for the grain size distribution of the transported sediment to become more uniform (i.e. more poorly sorted) as the mean grain size decreases. Thus, it is possible that the decrease in sorting of the suspended sediment with height (Fig. 5.21b) may in part be related to the fact that the mean size of the suspended sediment decreases with height (Fig. 5.18). The tendency for the grain size distributions to become more poorly sorted with height may also be enhanced by the mixing of the fine particulate load by air turbulence.

Some support for the above arguments may be found in the average skewness against height plot (Fig. 5.18b). If it is assumed that there is a "tail" of relatively coarse grains in the suspended sediment distribution which decreases from the surface to approximately 1.0 metres as a result of selective transport, there should be a corresponding increase in skewness with height. As can be seen in Fig. 5.18b, there is a tendency for the suspended sediment distribution to become increasingly more positively skewed from the surface to approximately 2.0 metres. Above 2.0 metres the skewness begins to decrease. This decrease in skewness above 2.0 metres indicates that the suspended sediment distribution tends to become more symmetrical with height. This may support the argument that the sorting above 1.5 metres becomes poorer because the suspended sediment distribution tends to become more uniform with height as a result of the decrease in mean particle size and turbulent mixing.

CHAPTER VI

SUMMARY AND CONCLUSIONS

The amount of sediment transported in creep, saltation and suspension by the wind was measured in the Slims River Valley, Yukon Territory. Although the amount of sediment transported at any given time varied considerably, the amount of sediment transported in surface creep usually accounted for less than five to eight per cent of the total sediment flow. This proportion is considerably lower than values presented by other investigators. The small proportion of creep flow in this investigation has been attributed to two principle factors: (1) the relatively fine texture of the surface sediments which can be readily entrained into saltation and suspension, and (2) the abundance of surface irregularities which impede the movement of surface particles in creep, but not in saltation and suspension.

Although the greatest proportion of sediment movement was usually in saltation, approximately fifteen to sixty-five per cent of the total sediment load was transported in suspension during the fifteen dust storms.

The data also showed that the mean particle size and the suspended sediment transport rate decreased with height. Each of these relationships can be expressed as a power function of height. The sorting of the suspended sediment improved from the surface up to approximately 1.5

metres, but above this height, the sediment became increasingly more poorly sorted. The improvement in sorting from the surface to 1.5 metres results from the selective removal of the finer sediments transported in suspension from the coarser particles moving in saltation.

Folk and Ward (1957) have suggested that in fluid transport systems, the sorting of fine sediments tends to decrease as mean size decreases because the transporting fluid is more able to transport all particles in the size distribution of the sediment load. Thus, the decrease in sorting of the suspended sediment with height above 1.5 metres probably results from the fact that the mean size of the sediment also decreases with height.

The mean size of the collected suspended sediment ranged from medium to coarse silt, which is similar to the mean size of the loess deposits located above the floodplain in the Slims River Valley.

The sediment transported in surface creep during the dust storms was significantly coarser than that transported in saltation. The third moment, skewness, showed the most striking difference between the grain size distributions of the creep and saltation samples. In general, the creep samples were characterized by positive skewness and the saltation samples by negative skewness. The positive skewness of the creep samples results from the selective removal from the surface of the particles small enough to be transported by the wind at a given moment in time. The tail of coarse grains (i.e. negative skewness) associated with the saltation samples is thought to be related to the fine mean size of the surface sediments as well as the nature of the transport.

At any given time, a large number of grains can be lifted from the surface if the shear velocity exceeds the threshold shear velocity. The size distribution of these entrained grains will tend to be positively skewed since the larger surface particles will not be lifted into the air stream because of their mass. However, some of the initially entrained particles will be small enough to be kept aloft in suspension at this particular wind velocity, with the remaining heavier particles falling back to the surface in saltation. Thus, the proportion of sediment which returns to the surface in saltation will lack the finer grains removed in suspension and will therefore tend to be less positively or even negatively skewed. Conversely, the positive skewness of the suspended sediment collected during the dust storms results from the removal of the coarser grains in saltation from the total sediment load initially entrained into the air stream.

The amount of sediment transported in saltation and creep (q) on the Slims River delta was significantly related to the shear velocity (U'_*). The mean daily observations indicated that the total sediment flow in saltation and creep varied approximately with the cube of shear velocity (b = 3.1074). This power relationship is similar to the theoretical models presented by Bagnold (1941) and Zingg (1952). However, it was also found that the rate of increase of sediment transported in saltation and creep (b = 2.4502) was significantly lower during the fifteen sampled dust storms. The lower exponent during these periods probably occurs because a greater proportion of the available sediment load was transported in suspension (Belly, 1964).

Despite the significant relationship between suspended sediment flow rate and shear velocity during the dust storms, the suspended flow rate appears to be more directly controlled by the amount of air turbulence as defined by the stability ratio. This is not surprising in that the fine suspended particles are kept aloft by turbulent eddies.

A large number of factors other than shear velocity can affect the amount of sediment transported by the wind. In the Slims River Valley, two of these appear to be of particular importance: (1) surface moisture content, and (2) the presence of soluble salts at the surface. Both of these factors tend to stabilize the surface by holding individual grains in place.

Woodruff and Siddoway (1965) have found from their field investigations that the rate of sediment transported in saltation and creep varies with the inverse of the square of surface moisture content. Similar results were also obtained from the mean daily observations taken in the Slims River Valley ($b = -12.2723$). That is, as moisture content increases, sediment transported in saltation and creep decreases.

In contrast, surface moisture content does not appear to significantly affect the amount of sediment transported in suspension during the dust storms. It was observed, however, that the dust storms only occurred when the surface moisture content was below approximately 3.5 to 3.8 per cent dry weight. The reason for this appears to be related to the fact that the pore size distribution of the delta sediments is strongly unimodal. It has also been suggested that the modal pore sizes are formed by a large percentage of the grains which are small enough to be transported in suspension.

The negative capillary pressures developed within the pore spaces hold together those particles in contact with the menisci. Thus, a large proportion of relatively fine grains will become more susceptible to entrainment and transport in suspension when the menisci retreat into the pores smaller than the modal pore size. This situation appears to occur when the surface moisture content falls below 3.5 to 3.8 per cent dry weight. Consequently, large dust storms will be generated only when the surface moisture content is below this range and sufficient fine sediment becomes readily available to the air stream.

The results of this investigation have demonstrated that the presence of soluble salts near the surface can affect the rate of sediment transported by the wind. The reduction of sediment transported in creep, saltation and suspension with increasing surface salt concentration is caused by the growth of salt crystals precipitated out of the pore water during the drying of the delta surface. The crystals located in the pore spaces bond together the surface particles in contact with them, thereby reducing the amount of sediment which can be entrained into the air stream. As the surface moisture continues to evaporate, soil water containing soluble material is drawn to the surface by capillarity. If evaporation continues, the newly derived soluble material is precipitated out with a consequent increase in the size and number of crystals bonding together the surface particles. The bonding effect of the crystals will only be removed or reduced if sufficient precipitation falls to leach the soluble salts from the surface.

Although the eolian transport of sediment is a common occurrence in the Slims River Valley, high saltation-creep flow rates and dust storms appear to be associated with a unique set of atmospheric and surface conditions. In general, high sediment transport rates occur when:

(1) shear velocity and atmospheric turbulence are high

(2) surface moisture content and surface salt concentration are low.

Field observations from this investigation indicate that these conditions are best developed on relatively warm, clear days following periods of heavy or extended rainfall.

The reason for this is that during the periods of precipitation, some of the soluble salts are leached from the surface of the delta, thereby reducing the number of grains which are bonded together by the precipitated salt. Visual observations also indicated that intensive precipitation tends to break up the surface, increasing the surface's susceptibility to wind erosion. As the surface of the delta begins to dry, a point may be reached when the shear velocity at a given moment is sufficient to overcome the forces which tend to hold the surface particles in place. When this occurs, surface particles will become entrained into the air stream and transported downwind in creep, saltation and suspension. If the surface continues to dry and the shear velocity remains constant or increases, there will be a corresponding increase in the sediment flow rate.

When the surface moisture content falls below 3.5 to 3.8 per cent, a large number of fine grains will no longer be held by surface tension effects and will therefore be more susceptible to entrainment. Shear velocities and air turbulence high enough to entrain particles of this size range will cause large quantities of fine grained sediment to be lifted from the surface in suspension and transported downwind in the form of dust clouds.

The work of this investigation has indicated the complex interaction of atmospheric and surface variables which affect the rate of sediment transported in creep, saltation and suspension. This field investigation also lends support to theoretical and empirical wind tunnel models presented by other authors. Although significant relationships have been shown to exist between the sediment transport rate and various atmospheric and surface variables, the complexity of this interaction creates problems which are difficult to assess in field situations. Therefore, extensive wind tunnel investigations will be necessary to clarify and set critical limits to these relationships.

APPENDIX A

MEAN DAILY OBSERVATIONS

		Mean Daily					
		Saltation Flow Rate (mg/cm.s)	Creep Flow Rate (mg/cm.s)	Saltation-Creep Flow Rate, q (mg/cm.s)	Shear Velocity (cm/s)	Surface Moisture Content (%)	Surface Salt Concentration (m.e./100g of soil)
May	12	4.4	0.8	5.2	-	4.88	30.4
	13	8.8	0.2	9.1	-	4.96	28.7
	14	2.9	0.5	3.4	-	5.01	29.2
	15*	1.6	0.2	1.8	5.64	4.93	34.9
	16*	0.3	0.5	0.8	12.50	13.92	28.3
	17**	0.8	0.8	7.6	21.81	22.74	26.9
	18**	23.9	1.2	25.1	7.42	5.36	26.4
	19	14.7	0.9	15.6	10.30	3.14	25.0
	20	6.1	1.6	7.7	8.23	3.26	23.1
	21	4.5	0.2	4.7	8.66	4.93	25.6
	22	2.1	0.2	2.3	9.57	2.40	28.9
	23*	2.3	0.6	2.9	7.05	7.62	31.8
	24*	-	-	0.0	4.19	16.13	27.2
	25**	-	-	0.0	5.62	17.42	24.9
	26	18.4	1.4	20.8	12.21	3.04	24.0
	27**	14.9	0.7	15.6	10.30	3.14	25.1
	28**	51.8	3.5	55.3	10.40	3.98	27.5
	29	46.0	0.4	47.4	9.45	2.81	22.1
	30**		Trap Repair				
	31	41.1	4.9	46.0	10.49	1.81	15.6
June	1**	25.3	2.8	28.1	15.18	2.08	17.1
	2*	6.2	1.4	7.6	10.46	3.30	17.9
	3	2.0	0.3	2.3	14.90	11.89	17.0
	4	0.9	0.1	1.0	7.51	9.63	19.8
	5	2.1	0.1	2.2	7.48	9.18	27.7
	6	2.3	0.4	2.7	7.44	3.51	23.9
	7	0.5	0.3	0.8	5.18	5.74	24.2
	8	0.2	-	0.2	3.81	5.50	23.9

...continued

APPENDIX A--continued

MEAN DAILY OBSERVATIONS

	Mean Daily					
	Saltation Flow Rate (mg/cm.s)	Creep Flow Rate (mg/cm.s)	Saltation-Creep Flow Rate, q (mg/cm.s)	Shear Velocity (cm/s)	Surface Moisture Content (%)	Surface Salt Concentration (m.e./100g of soil)
June 9**	46.9	5.4	52.3	10.42	3.93	16.2
10**	29.8	6.0	35.8	10.34	2.11	17.8
11			Trap Repair			
12*	0.2	0.4	0.6	10.20	23.84	17.2
13*	0.1	0.6	0.7	10.00	31.96	16.7
14*	-	-	0.0	14.35	17.43	15.4
15	0.3	0.1	0.4	7.08	10.91	17.1
16	2.3	0.4	2.7	5.51	5.87	18.0
17*	-	0.2	0.2	7.49	12.90	16.8
18*	0.3	0.8	1.1	13.60	13.64	16.1
19*	-	-	0.0	17.20	21.47	15.7
20*	-	0.1	0.1	11.10	25.16	14.0
21*	0.2	0.3	0.5	3.70	10.82	15.6
22**	20.1	2.4	22.5	9.24	4.10	15.2
23**	50.1	3.8	53.9	18.07	3.10	25.5
24	3.3	1.9	5.2	6.20	9.89	27.2
25**	41.7	2.1	43.8	8.89	3.29	27.0
26	5.1	0.3	5.4	5.93	5.00	32.8
27	6.4	0.8	7.2	5.93	4.14	36.9
28	0.3	0.1	0.4	1.90	19.34	22.6
29	4.2	0.2	4.4	4.14	4.57	20.5
30**	17.7	7.2	24.9	9.27	2.78	18.2
July 1**	39.7	1.2	40.9	9.72	4.08	16.4
2**	16.9	0.9	17.8	11.19	2.11	16.6
3	0.2	0.1	0.3	5.45	7.76	25.7
4	8.6	0.6	9.2	15.80	7.32	35.8
5	5.8	0.3	6.1	10.01	9.07	36.1

...continued

APPENDIX A--continued

MEAN DAILY OBSERVATIONS

	Mean Daily					
	Saltation Flow Rate (mg/cm.s)	Creep Flow Rate (mg/cm.s)	Saltation-Creep Flow Rate, q (mg/cm.s)	Shear Velocity (cm/s)	Surface Moisture Content (%)	Surface Salt Concentration (m.e./100g of soil)
July 6	2.4	0.1	2.5	7.08	5.28	46.4
7	1.6	0.2	1.8	4.15	4.33	41.7
8	7.0	0.6	7.6	14.02	5.02	59.9
9	4.1	0.8	4.9	7.88	3.21	53.4

*Days with greater than 0.025 mm of precipitation

**Days with dust storms

APPENDIX B

A METHOD FOR GRAIN SIZE ANALYSIS BY THE FALLING DROP TECHNIQUE

1. Remove organic matter from 10-15 grams of sample.

2. Rinse sample with distilled water and centrifuge to remove soluble salts.

3. Disperse soil with 5 ml of 50 gm/1 calgon solution (for predominantly silt and clay size samples, dispersing with a sonic dismembrator for two to three minutes is the best method.)

4. Wet sieve suspension through a 400 mesh (4.75 Ø) sieve collecting the material which passes through in a 500 ml volumetric flask.

5. Fill the volumetric flask to the 500 ml mark. Shake the flask thoroughly.

6. Transfer some of the suspension to a test tube being careful to get a representative sample. Place test tube in water bath of falling drop apparatus.

7. Sample with the micro-pipette from the test tube at the predetermined times and depths for the size classes required. (The times and depths are calculated using Stokes Law.)

8. Wipe the tip of the micro-pipette to remove any suspension on the surface.

9. Immerse the tip of the micro-pipette in the anisole and eject the suspension into it. (Care must be taken not to break the bubble of suspension while ejecting it.)

10. Time the fall of the bubble of suspension between the two lines scribed on the anisole cylinder.

11. Clean the micro-pipette by filling and emptying it three or four times in a beaker of distilled water. Place the point of the micro-pipette on a piece of filter paper to draw out any distilled water held in the tip.

12. Obtain the concentration of the sample from the calibration curve. The curve is constructed by plotting the known concentration of NaCl/water solutions against their falling time in anisole. Calibration curves should be constructed at the beginning of the sampling run. If more than a 0.2°C change in temperature occurs during

the experiment run, the apparatus should be recalibrated and a new curve drawn for subsequent samplings.

12. Per cent weight in each size class is calculated after making corrections for calgon content and original suspension concentration.

APPENDIX C

DAILY CREEP SAMPLES

PERCENT WEIGHT IN EACH SIZE CLASS

(CLASS MIDPOINT)

STN.	1.250	1.750	2.250	2.750	3.250	3.750	4.250	4.625	5.250	6.250
C02D	0.13	0.16	1.26	8.75	30.34	37.35	14.16	6.18	0.38	1.28
C03D	0.13	0.13	1.25	10.30	24.47	41.83	12.55	4.38	4.33	0.64
C04D	11.20	7.84	24.41	11.28	12.80	25.17	4.61	0.92	1.77	0.00
C09D	2.55	1.47	4.84	7.33	21.94	28.17	12.26	11.03	5.55	4.86
C11D	2.23	2.14	8.58	11.85	14.97	28.63	17.62	9.67	2.71	1.59
C12D	11.62	9.26	17.57	11.56	15.75	7.27	19.59	4.43	1.97	0.98
C14D	0.23	0.52	1.97	4.50	27.81	51.32	7.84	4.54	0.72	0.57
C19D	0.15	0.43	8.08	20.30	35.49	30.04	4.39	0.26	0.36	0.51
C20D	8.48	7.14	13.32	13.37	18.48	15.09	11.52	7.86	4.50	0.23
C24D	0.22	0.41	5.54	17.03	35.78	33.72	2.01	4.78	0.10	0.40
C25D	7.00	4.85	21.55	13.29	16.36	20.96	11.80	1.68	1.10	1.42
C26D	1.60	2.13	16.08	31.57	29.00	11.95	4.82	2.59	0.11	0.16
C27D	2.83	2.99	16.66	25.30	24.80	15.21	11.41	0.33	0.05	0.43
C29D	0.32	0.32	2.96	5.24	24.17	42.88	18.14	3.54	1.64	0.79
C31D	2.07	1.80	4.03	17.59	14.04	38.24	14.19	1.67	3.29	3.08
C46D	0.10	0.10	1.41	9.63	34.75	46.32	2.97	2.32	2.25	0.15
C47D	0.61	0.73	5.35	15.34	22.95	21.98	13.22	8.16	5.78	5.88

PERCENT WEIGHT IN EACH SIZE CLASS

(CLASS MIDPOINT)

STN.	1.375	1.625	1.875	2.125	2.375	2.625	2.875	3.125	3.375	3.625	3.875	4.250	4.625	5.250	6.250
C07D	3.12	0.63	1.34	1.42	3.08	7.75	6.63	13.27	6.59	7.93	5.45	17.03	8.91	12.28	2.57
C08D	0.00	0.00	1.67	0.20	0.90	3.62	5.47	14.26	11.40	14.22	21.17	15.47	3.77	7.57	0.28
C10D	0.95	0.20	0.41	0.48	1.03	3.21	4.39	11.88	8.93	16.50	14.45	20.72	6.37	10.46	0.00
C15D	0.95	0.23	0.48	0.43	1.05	3.20	4.00	11.93	9.15	16.29	8.74	26.94	6.24	8.50	1.87
C28D	0.00	0.00	3.70	1.10	2.64	8.25	10.97	21.41	11.94	9.78	11.47	10.37	3.43	4.62	0.31
C30D	1.06	0.17	0.41	0.41	0.96	3.11	4.16	11.36	8.84	16.29	11.04	24.98	7.42	8.03	1.73
C35D	0.93	0.18	0.51	0.37	1.01	2.88	3.93	10.97	9.04	16.40	3.66	32.20	7.19	10.37	0.35
C56D	0.93	0.17	0.38	0.48	1.18	3.72	4.93	12.52	9.69	16.20	14.01	19.50	5.82	10.24	0.23

DAILY SALTATION SAMPLES

PERCENT WEIGHT IN EACH SIZE CLASS

(CLASS MIDPOINT)

STN.	1.375	1.625	1.875	2.125	2.375	2.625	2.875	3.125	3.375	3.625	3.875	4.250	4.625	5.250	6.250
S02D	2.92	0.72	1.17	0.90	1.30	2.78	3.90	9.09	5.42	8.26	8.13	20.61	12.56	21.82	0.44
S03D	3.17	0.70	1.40	0.93	1.31	2.82	3.87	8.88	5.53	8.21	7.81	20.81	12.08	21.76	0.72
S04D	2.35	0.44	1.08	1.19	2.96	7.79	8.85	13.58	6.54	7.94	5.47	17.39	9.46	14.62	0.34
S07D	2.27	0.65	1.09	1.05	1.27	2.89	3.80	8.74	5.43	8.10	8.26	20.86	12.66	22.28	0.65
S08D	0.48	0.17	0.40	0.33	0.64	1.40	2.07	5.89	6.83	16.31	1.39	44.69	10.13	7.49	1.77
S09D	0.00	0.00	0.88	0.22	0.39	0.88	1.06	5.08	5.37	9.43	21.27	31.12	11.41	12.89	0.00
S10D	0.45	0.12	0.24	0.20	0.36	0.63	0.75	3.46	4.37	7.73	21.12	33.57	11.58	15.41	0.00
S11D	3.10	1.16	2.02	1.59	1.74	2.85	4.09	8.96	6.09	5.91	9.70	18.18	11.44	22.59	0.59
S12D	3.89	0.96	1.64	1.14	2.21	5.74	6.75	11.50	6.60	5.76	9.33	16.16	9.60	17.48	1.25
S14D	3.75	0.85	1.29	0.98	1.37	2.95	4.41	8.64	5.57	8.31	8.27	20.27	11.86	20.80	0.67
S15D	0.51	0.28	0.80	0.59	0.73	0.98	1.00	2.77	2.08	4.10	5.97	23.21	19.89	35.58	1.51
S19D	0.93	0.15	0.35	0.31	0.80	2.64	3.68	10.70	8.70	15.92	14.83	21.76	6.93	12.31	0.00
S20D	0.89	0.19	0.44	0.37	0.98	3.02	4.16	11.60	9.54	16.67	7.82	27.93	6.95	9.09	0.35
S24D	0.77	0.20	0.39	0.37	0.90	2.66	3.88	10.29	8.33	16.09	13.34	23.87	7.47	11.44	0.00
S25D	1.49	0.35	0.86	0.99	2.48	6.72	7.61	12.73	6.14	7.87	5.44	19.36	10.25	17.35	0.35
S26D	0.91	0.21	0.39	0.49	1.12	3.54	4.59	12.48	9.13	16.07	12.03	22.41	6.67	8.64	1.32
S27D	2.09	0.43	1.03	1.10	2.69	7.05	8.11	11.98	5.93	7.23	4.99	17.65	9.98	19.42	0.33
S28D	0.71	0.18	0.37	0.30	0.84	2.72	3.86	10.10	9.46	11.41	20.58	21.84	7.58	8.17	1.86
S29D	4.05	0.88	1.46	0.93	1.36	3.02	3.87	8.57	5.33	7.89	3.26	24.38	11.77	21.70	1.52
S30D	0.18	0.03	0.12	0.14	0.24	0.53	0.68	3.15	3.89	10.78	15.15	30.73	14.01	19.67	0.69
S31D	3.59	0.76	1.41	1.10	1.49	3.16	4.30	9.38	5.70	8.69	6.55	23.14	12.77	17.76	0.19
S35D	0.00	0.00	0.65	0.10	0.23	0.29	0.20	0.94	0.61	0.97	3.73	20.46	19.31	52.37	0.14
S46D	0.77	0.15	0.35	0.31	0.74	2.52	3.59	10.33	8.78	16.45	9.28	29.47	7.78	9.24	0.24
S47D	1.80	0.72	1.46	0.91	1.79	2.94	3.39	6.95	5.74	9.28	0.35	29.02	12.42	22.60	0.64
S56D	0.73	0.19	0.43	0.36	0.48	0.91	1.27	4.20	4.66	8.04	19.01	29.84	13.90	14.99	0.96

APPENDIX D

CREEP SAMPLES COLLECTED DURING THE DUST STORMS

PERCENT WEIGHT IN EACH SIZE CLASS

(CLASS MIDPOINT)

STN.	1.380	2.380	3.380	4.380	5.380	6.380	7.380	8.380
C01S	12.63	17.13	27.41	26.12	4.93	7.28	0.43	4.07
C02S	11.14	22.11	44.23	14.91	1.23	3.69	2.46	0.25
C03S	14.49	19.12	29.46	16.62	0.50	6.75	8.85	0.00
C04S	11.77	15.23	38.93	26.22	3.30	2.35	0.63	1.57
C05S	4.22	8.08	46.34	35.67	1.18	2.15	0.86	0.50
C06S	1.98	5.95	48.43	34.05	5.54	2.56	1.24	0.25
C07S	5.79	9.01	54.72	24.25	4.08	1.93	0.21	0.00
C08S	30.07	25.12	6.47	10.09	20.84	2.76	2.53	2.13
C09S	16.88	21.62	25.52	27.06	1.05	3.49	2.65	1.74
C10S	14.99	25.75	31.54	19.46	1.54	2.86	2.31	1.54
C11S	25.90	36.69	16.91	4.32	12.59	1.80	0.90	0.90
C12S	16.68	22.45	30.28	15.90	5.15	4.21	3.00	2.34
C13S	26.65	30.69	27.94	4.59	7.35	1.84	0.92	0.02
C14S	7.98	11.46	34.39	37.48	7.21	0.39	0.19	0.90
C15S	8.59	17.97	32.03	15.02	18.55	1.95	2.93	1.95

SALTATION SAMPLES COLLECTED DURING THE DUST STORMS

PERCENT WEIGHT IN EACH SIZE CLASS

(CLASS MIDPOINT)

STN.	1.250	1.750	2.250	2.750	3.250	3.750	4.250	4.625	5.250	6.250
S01S	0.09	0.17	0.93	12.64	19.98	26.98	22.12	7.77	8.80	0.52
S02S	1.54	2.31	5.01	8.80	20.96	55.08	2.89	2.25	0.77	0.39
S03S	1.16	1.31	4.04	7.69	21.73	30.84	19.94	7.43	4.65	1.21
S04S	1.51	1.44	6.32	8.39	23.05	31.03	18.08	4.82	4.92	0.43
S05S	1.81	1.67	6.66	9.87	21.37	23.27	10.90	12.70	10.16	1.60
S06S	1.08	1.30	3.08	8.23	14.27	26.08	23.53	12.05	7.00	3.39
S07S	1.41	1.69	8.78	12.04	18.42	24.05	13.37	8.31	11.67	0.27
S08S	0.68	0.81	5.93	13.30	18.64	55.60	4.69	0.17	0.17	0.00
S09S	0.11	0.17	1.50	7.78	34.83	49.37	5.94	0.10	0.20	0.00
S10S	5.15	3.54	8.92	12.50	15.35	27.18	15.55	9.54	1.76	0.51
S11S	1.94	1.82	7.77	13.75	18.28	51.02	2.58	1.24	1.08	0.54
S12S	0.16	0.72	1.47	7.70	26.18	58.52	3.91	1.34	0.00	0.00
S13S	0.76	1.09	8.74	14.70	33.14	35.16	3.01	2.15	1.08	0.16
S14S	1.06	1.12	3.84	11.57	19.85	30.75	21.81	4.07	5.76	0.17
S15S	0.75	0.64	2.96	9.70	18.32	33.64	19.78	1.22	8.55	4.43

APPENDIX E

SURFACE SAMPLES

PER CENT WEIGHT IN EACH SIZE CLASS

(CLASS MIDPOINT)

STN.	1.375	1.625	1.875	2.125	2.375	2.625	2.875	3.125	3.375	3.625	4.000	4.500	5.250	6.250	7.250	8.250	8.875	9.250
14MY2	0.08	0.82	0.08	0.25	0.41	1.23	1.14	3.59	4.66	10.78	39.54	22.06	8.91	3.02	0.69	0.49	0.33	1.92
14MY8	0.10	1.01	2.32	0.64	0.71	1.81	1.84	4.72	6.29	12.58	18.66	17.86	22.08	2.03	2.94	1.01	2.03	1.37
14MY9	0.07	0.67	0.74	2.48	4.43	8.86	9.73	11.21	8.32	9.80	10.20	1.68	22.72	5.37	2.69	0.67	0.34	0.01
21MY2	0.09	0.93	0.09	0.37	1.49	5.48	6.32	12.55	7.62	9.76	25.09	8.46	16.64	0.79	0.70	2.09	1.39	0.14
21MY8	0.08	0.84	0.25	0.42	1.00	2.43	2.26	3.51	5.69	5.02	21.33	9.45	36.72	8.57	1.59	0.75	0.08	0.00
21MY9	0.07	0.68	2.94	5.61	8.08	10.06	7.05	11.36	7.39	9.92	8.76	2.05	18.14	4.00	0.65	2.59	0.53	0.10
28MY2	0.07	0.68	5.90	2.98	2.64	3.73	2.98	6.24	5.83	6.51	11.12	5.56	33.20	3.73	6.44	0.68	1.02	0.68
28MY8	0.08	0.82	0.08	0.08	0.08	0.33	0.25	1.06	1.80	6.94	27.78	28.68	23.20	2.74	2.41	1.02	0.20	2.45
28MY9	0.07	0.72	6.61	6.91	7.63	10.39	6.38	8.58	7.44	10.19	11.57	6.08	6.09	5.73	1.97	1.61	1.15	0.89
03JN2	0.10	1.01	0.51	0.60	1.73	1.77	1.98	3.78	3.20	14.08	21.68	17.15	22.49	7.71	0.55	1.01	0.40	0.23
03JN8	0.08	0.82	0.08	0.16	0.49	0.33	0.66	1.32	1.73	4.53	35.66	20.86	20.30	6.59	2.26	1.85	1.32	0.95
03JN9	0.07	0.70	10.57	14.89	9.46	10.02	5.01	7.10	3.41	4.03	3.27	0.83	22.99	4.52	2.43	0.35	0.17	0.17
10JN2	0.11	1.07	5.15	2.04	1.83	2.15	1.93	3.54	2.68	6.87	11.28	19.13	33.62	4.83	3.22	0.27	0.21	0.05
10JN8	0.10	1.03	0.10	0.21	0.41	0.31	0.41	1.13	1.75	5.85	49.08	10.47	17.04	7.70	0.56	0.46	1.69	1.69
10JN9	0.08	0.82	2.54	5.16	12.94	9.50	8.68	10.07	5.73	7.37	18.18	7.86	4.55	2.42	1.02	0.61	1.11	1.35
17JN2	0.07	0.66	0.79	1.51	2.44	4.41	4.94	6.39	7.44	9.35	16.33	11.00	29.57	1.81	1.38	0.99	0.26	0.66
17JN8	0.10	1.00	0.95	0.67	0.67	1.18	1.15	2.45	3.16	13.57	15.08	21.47	27.82	5.36	2.46	1.70	1.20	0.00
17JN9	0.09	0.93	1.86	2.88	9.77	7.26	8.56	11.72	6.79	12.09	21.49	6.88	4.65	1.40	0.23	1.77	0.51	1.12
24JN2	0.07	0.69	0.69	1.79	2.34	4.27	4.27	6.13	5.17	7.17	18.40	20.06	21.47	2.83	3.86	0.21	0.10	0.48
24JN8	0.09	0.86	1.38	1.47	2.16	3.10	1.03	3.10	1.29	2.59	25.26	17.76	20.13	14.78	2.89	1.34	0.65	0.13
24JN9	0.06	0.57	8.92	8.52	8.58	10.34	6.87	8.80	5.62	5.28	10.39	11.02	10.62	0.77	0.43	1.42	1.08	0.71
01JY2	0.07	0.66	5.55	4.23	4.10	4.69	3.90	5.29	4.69	5.22	8.46	8.07	33.92	5.98	1.52	2.25	0.13	1.26
01JY8	0.07	0.68	0.48	1.71	2.32	8.32	10.78	9.01	7.03	9.35	2.32	3.62	22.51	11.69	7.47	0.99	0.91	0.76
01JY9	0.08	0.80	8.67	5.89	6.21	8.83	5.73	7.56	6.84	14.41	11.65	7.44	7.01	1.71	0.20	3.46	3.50	0.00
07JY2	0.10	0.99	0.69	1.78	6.13	1.88	2.37	3.75	3.66	5.63	47.23	10.08	8.05	1.98	2.57	1.93	1.19	0.00
07JY8	0.10	0.99	2.58	0.99	1.98	1.29	1.19	1.49	1.29	8.04	8.15	20.98	37.78	8.19	0.25	0.89	0.40	3.42
07JY9	0.10	1.01	0.33	1.09	4.48	5.57	6.84	11.76	9.35	27.40	14.40	8.88	0.15	4.87	0.70	1.31	1.06	0.70

APPENDIX F

SUSPENDED SEDIMENT SAMPLES

PERCENT VOLUME IN EACH SIZE CLASS

(CLASS MIDPOINT)

STN.	3.380	4.380	5.380	6.380	7.380	8.380	9.380	10.380
11	38.82	48.53	11.12	1.16	0.30	0.06	0.01	0.00
12	0.00	77.46	18.65	3.00	0.72	0.15	0.02	0.00
13	0.00	68.20	23.44	6.88	1.17	0.26	0.04	0.00
14	0.00	64.83	24.31	6.08	3.17	1.14	0.43	0.05
15	0.00	41.08	38.52	14.96	4.70	0.65	0.08	0.01
16	0.00	0.00	58.61	18.32	16.94	4.92	1.00	0.21
17	0.00	0.00	52.09	19.53	17.91	8.04	2.02	0.41
21	40.39	39.38	16.28	3.22	0.62	0.09	0.01	0.00
22	34.75	43.43	18.28	3.03	0.45	0.06	0.01	0.00
23	26.35	49.41	21.82	2.06	0.29	0.05	0.01	0.00
24	22.79	55.55	17.45	3.27	0.75	0.16	0.02	0.00
25	0.00	57.05	34.86	6.44	1.39	0.23	0.03	0.00
26	0.00	0.00	73.92	19.14	5.36	1.31	0.24	0.03
27	0.00	0.00	46.43	23.21	15.96	10.97	2.88	0.54
28	9.18	18.93	5.88	0.82	0.13	65.05	0.00	0.00
31	51.43	38.57	8.49	1.33	0.15	0.02	0.00	0.00
32	41.42	38.31	14.76	4.32	1.01	0.16	0.02	0.00
33	36.23	46.79	15.28	1.39	0.27	0.04	0.01	0.00
34	32.17	56.30	10.05	1.17	0.25	0.05	0.01	0.00
35	29.19	51.09	17.79	1.65	0.24	0.03	0.00	0.00
36	28.29	51.28	17.02	2.90	0.41	0.08	0.01	0.00
37	0.00	47.07	38.24	11.58	2.08	0.84	0.17	0.02
38	0.00	0.00	64.40	25.16	7.04	2.61	0.68	0.11
41	37.54	39.89	17.89	4.03	0.54	0.09	0.02	0.00
42	40.70	30.52	25.22	2.99	0.47	0.08	0.01	0.00
43	20.24	55.65	19.29	3.99	0.68	0.13	0.02	0.00
44	0.00	72.51	19.42	5.50	1.74	0.67	0.14	0.02
45	0.00	75.33	6.73	11.10	5.25	1.33	0.22	0.03
46	0.00	59.88	34.51	4.99	0.51	0.08	0.02	0.00
47	0.00	58.43	35.30	3.65	1.90	0.55	0.14	0.02
48	0.00	34.35	23.81	9.22	31.65	0.84	0.12	0.01
51	28.87	50.52	12.40	6.03	1.81	0.33	0.04	0.00
52	38.87	29.15	21.86	7.67	1.97	0.40	0.07	0.01
53	0.00	53.66	42.26	2.60	1.08	0.32	0.06	0.01
54	0.00	52.45	31.84	10.54	3.97	1.00	0.18	0.03
55	0.00	56.55	28.82	11.62	2.57	0.38	0.06	0.01
56	0.00	32.58	44.79	17.81	3.66	0.95	0.18	0.03
57	0.00	26.62	46.59	20.38	5.10	1.02	0.25	0.04
58	0.00	0.00	58.97	30.35	8.56	1.77	0.31	0.04
61	0.00	71.80	21.62	5.25	1.01	0.26	0.05	0.01
62	28.02	45.53	21.01	4.10	1.20	0.12	0.02	0.00
63	23.36	49.63	21.17	4.29	1.31	0.23	0.03	0.00
64	0.00	72.54	23.39	3.28	0.62	0.14	0.03	0.00
65	0.00	62.09	33.31	3.64	0.81	0.12	0.02	0.00
66	0.00	59.11	36.33	2.46	1.57	0.46	0.07	0.01
67	0.00	26.88	25.20	9.24	37.34	1.14	0.17	0.03
71	50.92	25.46	18.30	3.78	1.10	0.37	0.06	0.01
72	45.41	37.84	12.30	3.25	0.98	0.19	0.02	0.00
73	42.29	40.53	15.20	1.40	0.43	0.12	0.03	0.00
74	35.60	40.05	20.02	3.23	0.85	0.21	0.04	0.00

SUSPENDED SEDIMENT SAMPLES

PERCENT VOLUME IN EACH SIZE CLASS

STN.	(CLASS MIDPOINT) 3.380	4.380	5.380	6.380	7.380	8.380	9.380	10.380
75	35.64	40.09	18.93	3.69	1.25	0.33	0.06	0.01
76	22.50	42.19	27.42	6.02	1.41	0.37	0.08	0.01
77	0.00	50.57	31.61	12.38	4.08	1.07	0.25	0.04
78	0.00	24.73	52.54	16.81	4.24	1.35	0.28	0.04
81	33.55	41.93	18.35	4.55	1.24	0.31	0.06	0.01
82	21.62	51.34	20.27	4.26	1.78	0.63	0.09	0.01
83	0.00	62.00	30.44	4.71	2.16	0.56	0.11	0.02
84	0.00	69.89	22.41	5.65	1.49	0.47	0.08	0.01
85	0.00	68.60	24.56	5.26	1.25	0.29	0.04	0.00
86	0.00	47.19	36.71	11.63	3.00	1.20	0.24	0.02
87	0.00	34.86	47.93	11.44	4.02	1.44	0.28	0.04
88	0.00	36.62	42.35	14.59	4.45	1.60	0.36	0.03
91	0.00	64.38	25.35	7.80	1.96	0.43	0.07	0.01
92	0.00	61.51	28.83	7.57	1.59	0.38	0.10	0.01
93	0.00	56.77	30.32	10.24	2.07	0.52	0.08	0.01
94	0.00	60.57	27.26	8.23	2.71	1.01	0.20	0.02
95	0.00	53.13	33.20	8.72	3.87	0.99	0.09	0.00
96	0.00	41.64	39.04	13.18	4.82	1.19	0.12	0.01
97	0.00	16.21	60.80	14.95	6.36	1.36	0.29	0.03
98	0.00	33.18	48.11	13.27	4.34	0.98	0.11	0.01
101	52.83	23.11	16.92	4.80	1.50	0.71	0.12	0.01
102	0.00	49.89	40.09	7.57	1.67	0.65	0.12	0.01
103	0.00	39.55	44.50	11.59	3.07	1.08	0.19	0.02
104	0.00	39.46	41.10	12.13	5.50	1.50	0.29	0.02
105	0.00	39.70	38.46	16.75	3.66	1.14	0.25	0.02
106	0.00	26.43	51.77	17.62	2.82	1.12	0.22	0.02
107	0.00	34.09	44.03	15.80	4.35	1.40	0.30	0.03
111	61.44	23.04	12.00	2.58	0.70	0.18	0.05	0.01
112	29.94	41.17	23.39	4.15	1.05	0.25	0.03	0.00
113	0.00	69.18	24.56	4.97	1.11	0.15	0.02	0.00
114	0.00	67.30	25.84	5.33	1.11	0.31	0.10	0.01
115	0.00	64.31	24.12	8.71	2.03	0.65	0.16	0.02
116	0.00	57.25	23.85	10.54	6.88	1.33	0.14	0.01
117	0.00	25.94	43.77	22.90	5.95	1.22	0.21	0.02
118	0.00	18.03	52.98	19.16	7.24	2.30	0.27	0.02
121	56.72	17.72	19.05	4.93	1.18	0.32	0.07	0.01
122	0.00	67.31	23.37	7.25	1.43	0.47	0.15	0.02
123	0.00	49.74	42.74	6.12	0.97	0.32	0.10	0.01
124	0.00	49.89	38.98	8.83	1.95	0.31	0.03	0.00
125	0.00	49.10	36.83	9.78	3.16	0.96	0.16	0.01
126	0.00	52.43	31.95	10.55	3.76	1.09	0.20	0.02
127	0.00	43.40	32.55	16.95	4.98	1.67	0.41	0.04
128	0.00	29.47	41.75	22.11	5.26	1.21	0.18	0.02
131	0.00	61.18	33.46	4.18	0.93	0.21	0.04	0.00
132	0.00	53.31	34.80	8.33	2.94	0.49	0.11	0.01
133	0.00	47.77	39.24	10.13	1.96	0.68	0.20	0.02
134	0.00	51.02	37.84	8.61	1.75	0.59	0.17	0.02
135	0.00	39.45	45.79	11.80	2.10	0.66	0.18	0.02
136	0.00	54.10	28.74	9.93	4.73	2.02	0.44	0.04
137	0.00	38.76	51.22	8.91	0.88	0.18	0.04	0.00

SUSPENDED SEDIMENT SAMPLES

PERCENT VOLUME IN EACH SIZE CLASS

(CLASS MIDPOINT)

STN.	3.380	4.380	5.380	6.380	7.380	8.380	9.380	10.380
138	0.00	34.22	44.20	16.75	3.41	1.14	0.25	0.02
141	39.83	29.87	22.40	6.30	1.28	0.25	0.06	0.01
142	0.00	59.26	31.48	6.71	1.82	0.60	0.12	0.01
143	0.00	47.81	38.10	10.55	2.87	0.54	0.12	0.01
144	0.00	27.00	61.89	7.88	1.83	0.97	0.39	0.04
145	0.00	45.63	44.00	8.76	1.22	0.32	0.06	0.01
146	0.00	19.81	58.19	18.26	2.71	0.79	0.21	0.02
147	0.00	15.35	40.29	30.94	9.08	3.51	0.76	0.08
148	0.00	15.00	52.49	24.13	5.62	2.16	0.55	0.05
151	27.18	33.98	35.25	2.81	0.49	0.20	0.07	0.01
152	0.00	53.74	38.76	5.81	1.19	0.38	0.10	0.01
153	0.00	53.05	33.15	10.42	2.56	0.66	0.14	0.02
154	0.00	41.69	41.69	12.38	3.39	0.70	0.14	0.02
155	0.00	42.30	50.23	5.95	1.09	0.35	0.08	0.01
156	0.00	31.52	48.27	15.02	3.83	1.04	0.29	0.04
157	0.00	0.00	69.22	23.18	5.18	1.92	0.46	0.04
158	0.00	26.77	55.77	11.15	3.89	1.86	0.51	0.05

APPENDIX G

GRAIN SIZE ANALYSIS COMPUTER PROGRAM

```
$CARD LIST SINGLE HOL
C**   MECHANICAL ANALYSIS OF SEDIMENTS BY MOMENTS
C**   FIRST FOUR MOMENTS ARE CALCULATED
C**   STAT IS SAMPLE NUMBER
C**   NO IS THE NUMBER OF OBSERVATIONS (SAMPLES)
C**   MAXIMUM NUMBER OF SAMPLES IS 150
C**   NV IS THE NUMBER OF VARIABLES (SIZE CLASSES)
C**   MAXIMUM NUMBER OF SIZE CLASSES IS 20
C**   MIDMAX IS MIDPOINT OF COARSEST SIZE CLASS
C**   THIS PROGRAM ALLOWS FOR A VARIABLE INPUT FORMAT
C**   THIS FORMAT IS PUNCHED IN ON FOUR CARDS WHICH IMMEDIATELY
C**   PRECEDE THE DATA CARDS.
C**   CARD 1
C**   COL 1-3 NUMBER OF OBSERVATIONS (SAMPLES)
C**   COL 4-5 NUMBER OF  VARIABLES (SIZE CLASSES)
C**   CARD 2
C**   IN BRACKETS STARTING IN COL 1 MIDPOINT FORMAT
C**   (FLOATING POINT, E.G. (10F6.3))
C**   CARD 3
C**   IN BRACKETS STARTING IN COL 1 THE DATA FORMAT WHICH INCLUDES
C**   A SAMPLE INDENTIFIER. THE IDENTIFIER FOLOWS THE DATA FORMAT
C**   AND MUST NOT BE GREATER THAN FOUR SPACES, ALPPHAMERIC.
C**   E.G. (10F6.3,1X,A4)
C**   CARD 4
C**   CLASS MIDPOINTS (FORMAT FROM FORMAT CARD 2)
C**   CARD 4 FOLLOWED BY DATA
                                                       START OF SEGMENT
      DIMENSION X(20),F(20),F1(20),ARRAY(150,20),FSUM(150),KSTAT(150)
      DIMENSION A(150,20),FMT(20),FMT1(20)
      DIMENSION CPCT(150,20),CUMP(150)
      READ (5,111) NO,NV
  111 FORMAT (I3,I2)
      READ (5,112) FMT1
      READ (5,112) FMT
  112 FORMAT(20A4)
      NCT=0
      CALL HEAD1
      DO 707 IJ=1,NO
  707 FSUM(IJ)=0.0
      READ(5,FMT1)(X(I),I=1,NV)
      DO 10 N=1,NO
   11 READ(5,FMT)(F1(I),I=1,NV),STAT
      KSTAT(N)=STAT
      DO 909 II=1,NV
  909 FSUM(N)=FSUM(N)+F1(II)
      F(1)=(F1(1)/FSUM(N))*100.0
      CPCT(N,1)=F(1)
      DO 910 JJ=2,NV
      F(JJ)=(F1(JJ)/FSUM(N))*100.0
      CPCT(N,JJ)=CPCT(N,JJ-1)+F(JJ)
  910 CONTINUE
   41 SUMP = 0.0
      DO 20 I=1,NV
   20 SUMP= SUMP + X(I)*F(I)
C**   SUMP IS SIGMA FX
      SUMFR = 0.0
```

```
      DO 40 I=1,NV
   40 SUMFR = SUMFR + F(I)
C**   SUMFR IS SIGMA F, IDEALLY 100 PERCENT
      AVX = SUMP/SUMFR
C**   SVX IS ARITHMETIC MEAN IS SIGMA FX DIVIDED BY SIGMA F
C**   AVX IS FIRST MOMENT
      SUMPS = 0.0
      DO 60 I=1,NV
   60 SUMPS = SUMPS+F(I)*(X(I)-AVX)**2
C**   SUMPS IS SIGMA F(X-MEAN)SQUARED
      SMOM =SUMPS/SUMFR
C**   SMOM IS SECOND MOMENT
      SUMPT= 0.0
      DO 80 I=1,NV
   80 SUMPT =SUMPT+F(I)*(X(I)-AVX)**3
C**   SUMPT IS SIGMA F(X-MEAN)CUBED
      TMOM =SUMPT/SUMFR
C**   TMOM IS THIRD MOMENT
      SUMPFR = 0.0
      DO 100 I=1,NV
  100 SUMPFR =SUMPFR+F(I)*(X(I)-AVX)**4
C**   SUMPFR IS SIGMA F(X-MEAN) TO FOURTH POWER
      FMOM =SUMPFR/SUMFR
C**   FMOM IS FOURTH MOMENT
      STDEV=SQRT(SMOM)
      SKEW = TMOM/STDEV**3
      CURT = FMOM/STDEV**4
C**   STANDARD DEVIATION, SKEWNESS AND KURTOSIS
      IF(SKEW)12,13,13
   12 SKEW=SKEW-0.005
      GO TO 14
   13 SKEW=SKEW+0.005
   14 AVX=AVX+0.005
      STDEV=STDEV+0.005
      CURT=CURT+0.005
      IF (NCT.GE.50) CALL HEAD1
      IF (NCT.GE.50) NCT=0
      PRINT 2,KSTAT(N),AVX,STDEV,SKEW,CURT
    2 FORMAT(7X,A5,4F11.2)
      NCT=NCT+1
      A(N,1)=AVX
      A(N,2)=STDEV
      A(N,3)=SKEW
      A(N,4)=CURT
      DO 928 L=1,NV
  928 ARRAY (N,L)=F(L)
   10 CONTINUE
      CALL HEAD2(X,NV)
      NCT1=0
      DO 808 JK=1,NO
      PRINT 809,KSTAT(JK),(ARRAY(JK,L2),L2=1,NV)
  809 FORMAT(1X,A5,1X,10(F6.2,1X))
      NCT1=NCT1+1
      IF (NCT1.GE.50) CALL HEAD2(X,NV)
  808 IF (NCT1.GE.50) NCT1=0
      PRINT 601
  601 FORMAT ("1")
      NCT2=0
      CALL HEAD3
      DO 924 JO=1,NO
      IF (NCT2.GE.50) CALL HEAD3
      IF (NCT2.GE.50) NCT2=0
      PRINT 925,KSTAT(JO),(CPCT(JO,JV),JV=1,NV)
  925 FORMAT(1X,A4,1X,18(F6.2,1X))
      NCT2=NCT2+1
  924 CONTINUE
   90 STOP
      END
```

SEGMENT

START OF SEGMENT

```
      SUBROUTINE HEAD1
      PRINT 48
   48 FORMAT("1",1X//////)
    3 FORMAT(6X,53H STATION       MEAN      STANDARD    SKEWNESS    KURTOSIS
     1 )
    4 FORMAT(38H          NUMBER                     DEVIATION /)
    5 FORMAT(1H1,6X48HPARTICLE SIZE ANALYSIS OF SEDIMENTS  (PHI UNITS)/)
      PRINT 5
      PRINT 3
      PRINT 4
      PRINT 49
   49 FORMAT(/)
      RETURN
      END
```

SEGMENT

START OF SEGMENT

```
      SUBROUTINE HEAD2 (X,NV)
      DIMENSION X(NV)
      PRINT 48
   48 FORMAT("1",14X,"PERCENT VOLUME IN EACH SIZE CLASS"/)
      PRINT 49
   49 FORMAT(23X,"(CLASS MIDPOINT)")
      PRINT 50,(X(II),II=1,NV)
   50 FORMAT(1X,"STN.",2X, 8(F6.3,1X)/)
      RETURN
      END
```

SEGMENT

START OF SEGMENT

```
      SUBROUTINE HEAD3
      PRINT 58
   58 FORMAT("1"/)
      PRINT 59
   59 FORMAT("1",30X,"CUMULATIVE PERCENT"/)
      RETURN
      END
```

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