THE GEOGRAPHY OF DUST STORMS

Volume 1

BY

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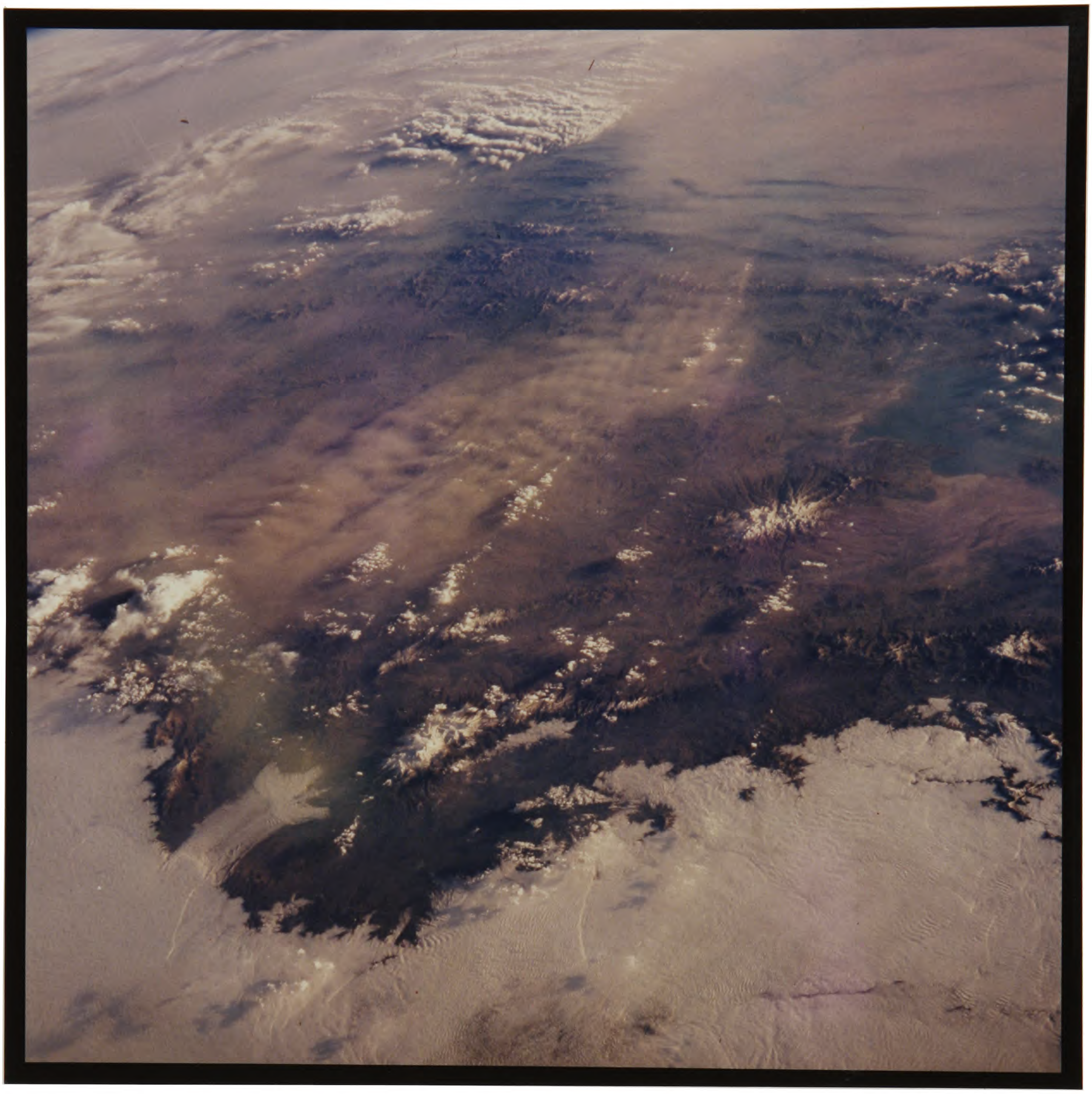

Frontispiece: Space Shuttle photograph of a dust storm on the floodplains of the Amu Darya, northern Afghanistan (9 October 1984). Dust is raised from the alluvial spreads & loess coverings that are widespread in the arid & semi-arid plains of Afghan Turkestan. Dust storms occur on more than 30 days a year on the upper reaches of the Amu Darya and are often associated with the so-called **Afghan Wind**. (see chapter 6).

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SHORT ABSTRACT

THE GEOGRAPHY OF DUST STORMS

By Nicholas J Middleton, Hertford College, Oxford.

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Dust storms have a great many environmental implications in the world's dry lands where they are particularly common. Four main classes of dust event are identified and defined: dust storms, dust haze, blowing dust and dust devils.

The geography of dust storms is analysed in each of eight major world regions: Africa, the Middle East, South-west Asia, Europe and the USSR, China, Australia, North America and Latin America. Terrestrially observed meteorological data and data from remote sensing platforms are employed to identify the major source areas in each region, their seasonality, diurnal patterns of activity and trajectories of long-range transport.

Among the important controls on the frequency distribution of dust storm activity are the meteorological conditions that generate dust-raising winds, and a number of meteorological systems commonly cause dust storms in all global regions. These include low pressure fronts with intense baroclinal gradients, pressure gradient winds between moving or stationary air masses, katabatic winds and

convectional cells.

The nature of the surface upon which deflation occurs is also important; typical dust-producing geomorphological units include alluvial spreads, lacustrine deposits, desert depressions, loess deposits and reactivated fossil dunes.

Dust storm activity is prone to considerable variation. The seasonal characteristics are explicable with reference to the meteorological systems generating dust, the state of ground cover, particularly vegetation, and the effects of seasonal rainfall. Substantial variations also occur from year to year, and land use and climatic variations can substantially affect their occurrence.

LONG ABSTRACT

THE GEOGRAPHY OF DUST STORMS

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Dust storms are typical features of the world's dry lands where readily available sources of fine grained particles are often sparsely protected by vegetation and susceptible to entrainment by strong winds. Their environmental significance is varied, representing large volumes of erosion and deposition and contributing to a number of geomorphological processes such as the formation of loess, desert varnish and duricrusts. The consequences for human populations include disruption to transport and communications, the transmission of diseases and atmospheric pollution.

Four classes of dust event are identified and defined in accordance with international meteorological observing standards: dust storms, dust haze, blowing dust and dust devils. Methods for their investigation are by examination of terrestrially observed meteorological data, data from remote sensing platforms, ground surface turbidity and aerosol monitoring and analysis of long-term deposition in deep-sea sediments and ice cores. This thesis employs the first two methods and relates the findings to the results of

investigations based on other approaches.

A dust storm is internationally defined as a dust-raising event that reduces visibility to below 1000m, and use of the 'dust storm day' (defined as a day on which a dust storm occurs) has enabled a quantitative assessment of dust storm activity in major world regions. Meteorological data are widely available, but care needs to be exercised in their use. Data problems include those of observer error, of distinguishing between dust and other causes of visibility obscuration, daily frequency of observations, sparsity of stations in remote regions and problems of duration, areal and volumetric extent of individual events. A number of remote sensing platforms have proved useful in monitoring dust events, providing a measure of the reliability of meteorological observations, confirmation and identification of source areas and a true indication of the scale of dust-raising and transport.

Using these data the geography of dust storms is analysed for each of eight major world regions. The analysis includes identification of major source areas and their geomorphological nature, their diurnal and seasonal characteristics, the meteorological conditions favourable for dust generation, the controls on the dust storm systems, such as rainfall, vegetation, wind speed and human activities, their effects on human populations, changes in dust storm frequency over time and the major trajectories of long-range transport.

The Sahara is the world's largest area of dust storm activity, producing as much as half of the global atmospheric load of soil dust. Major sources are located in southern Morocco, south-west Algeria, southern Mauritania and northern Mali, the Chad Basin, eastern Algeria and northern Libya, and north-east Sudan. The important dust-generating meteorological systems in these areas are the passage of low pressure systems and local convective cells. The Horn of Africa appears to be an important source for dust transported over the Arabian Sea in a large-scale flow identified on remote sensing imagery that is part of the south-west monsoon circulation, but lack of dust storm frequency data has prevented a quantitative assessment of this area for comparison with other North African regions.

In the Middle East the main dust sources are identified on the alluvial plains of lower Mesopotamia and the deserts of Syria and northern Saudi Arabia. In these areas dust-raising winds are generated by airflow converging between quasi-stationary pressure cells and low pressure frontal passage.

Major sources of dust-raising activity in south-west Asia are found in the Seistan Basin at the convergence of the borders of Iran, Afghanistan and Pakistan, and the plains of Afghan Turkestan. In these regions topographically induced airflows entrain material from desert depressions, alluvial and loess deposits. In northwestern India a closer understanding of the dust storm system has been gained by tracking a number of typical events using daily SYNOP data. Pressure gradient dust-raising winds, associated with western disturbances during the pre-monsoon period, are largely responsible for dust events in the Thar Desert and this material is often advected in light winds southeastwards down the Ganges and northeastwards towards the Himalayas while some material follows a recurved transport path towards the upper Indus plains. Dust-raising events down the Gangetic plain are more commonly caused by thunderstorm downdraft winds, also characteristic of the pre-monsoon season.

In the USSR dust storms are a feature of the southern European agricultural areas, the steppes of Kazakhstan and western Siberia and the deserts of Central Asia. In these regions dust is commonly raised by strong winds associated with the passage of low pressure fronts and topographically induced flows.

Chinese areas of major dust storm activity are located in the arid and semi-arid north of the country; in the Taklimakan Desert and Gansu Corridor, the Gobi Desert and the Loess Pateau. The passage of low pressure fronts in the spring is the typical dust-raising system in these parts of Asia.

The dry lands of Australia and the Americas are quantitatively less important world dust storm regions. In Australia maximum dust-raising occurs in the arid areas of the Simpson Desert and Lake Eyre Basin, where winds associated with low pressure fronts and local thunderstorms raise alluvial material and sediments from desert depressions. In North America dust storms occur in the semi-arid prairies of the Great Plains and the Canadian Prairie provinces as well as the cold arid areas of northern Canada and Alaska. Material from agricultural lands is raised by typical meteorological systems that include frontal passage, stationary low pressure systems and thunderstorm downdraft winds. In Mexico the thunderstorm downdraft is the most important dust-raising mechanism, the thunderstorms being associated with cold air oubreaks from the north in spring. Maximum dust storm activity occurs at Torreon and Mexico City. In Argentina dust is raised from point source salt lakes and alluvial fans on the Puna de Atacama by the prevailing upper westerlies, and by low pressure systems and pressure gradient winds on the Pampas and Chaco. Data were not available to assess the dust storm systems of the Atacama Desert.

Diurnal patterns of dust events are related to the proximity of observations to dust sources. Typically, dust-raising is at a maximum during the afternoon when winds are most turbulent due to solar heating.

Dust storms are distinctly seasonal in their occurrence in all areas. In many regions they are characteristic of the dry season, but in some areas the meteorological systems that bring seasonal rainfall also raise dust from surfaces that quickly dry in high daytime temperatures. The relationship between dust storms and rainfall is not simple. Correlations between annual numbers of dust storm days and annual rainfall totals are weak. The role of precipitation, and its effects on soil moisture properties, vegetation and ground surface stability have been related to specific geomorphological environments. Fairly strong relationships are found, however, in areas where decreasing rainfall trends over several years are accompanied by increasing dust storm activity. Wind characteristics are difficult to measure, but monthly dust storm frequencies at many world stations show good correspondance with monthly Drift Potential (Fryberger 1979) which incorporates a threshold velocity for particle movement. The importance of wind is best related to generating meteorological systems.

Variations in dust storm activity also occur on the timescale of decades, and although major world dust-producing regions are identified it is noted that considerable variability occurs. Such variability is in many cases related to human activity on a number of spatial scales. The draining of Lake Texcoco, to the north-east of Mexico City, left a ready source of dust storm sediments that have affected the city and particularly the International Airport adjacent to the lake. A steady decrease in dust storms recorded at the airport since the 1950s, however, is related to a gradual increase in rainfall over the region, the extention of the urban area and a project specifically set up in the early 1970s to stabilize the former lake bed. Similar regional dust sources have appeared at Owens Lake and Mono Lake in southern

California due to urban water use by the Los Angeles conurbation.

On a much larger scale the Dust Bowl of the 1930s on the Great Plains of the USA and the Virgin Lands scheme on the Soviet steppes in the 1950s resulted in significant increases in dust storm activity due to inappropriate agricultural techniques and poor land management. During a prolonged period of drought in the Sahel region of North Africa in the 1970s and early 1980s dramatic increases in dust storm activity have occurred in the major source regions of southern Mauritania and the Chad Basin and significant new areas of high frequency have appeared in the inland delta of the Niger River in Mali and in the Sudanese Sahel. This enhanced dust output is related to a prolonged period of below average rainfall and the actions of an expanding human population in the semi-arid environment. Human activities that have exacerbated the effects of the drought in this region include vegetation stripping by overgrazing, overcultivation and wood-cutting. In addition to an increased dust output from existing sources, new material has been raised from a variety of palaeofeatures in the Sahel, including reactivated fossil dune fields.

Study of the world distribution of dust storms reveals a number of characteristic dust-producing geomorphological environments. These include alluvial plains and fans, sediments in desert depressions, lacustrine sediments and loess coverings. Similarly, a number of meteorological systems on the meso and synoptic scales are commonly responsible for the generation of dust-raising winds in world regions. These include convective cells, topographically affected airflows, frontal systems, cyclogenic systems and pressure gradient winds between moving or stationary air masses.

In areas where dust storms are regular features of the climate they have received a variety of local names, and these are presented in tabular form, indicating the areas affected, typical seasonality, directions of flow and the generating meteorological systems.

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CHAPTER ONE : INTRODUCTION

1.1 STUDY OF DUST

This thesis is about dust storms. It aims to find where they happen on earth, why they occur and how they affect human populations and the environment in which they live. It is, therefore, a geographical study, involving elements of geomorphology, meteorology, climatology and ecology.

A dust storm is the blowing of small soil particles into the air from a dry, unprotected ground surface. It is particularly common in the world's arid and semi-arid regions where vegetation is often sparse, although they also occur in more temperate climates during dry periods. Dust events range in size from the dust devil, measuring just a few metres across, to events that can cover hundreds of thousands of square kilometres. Dust raised in a storm can be transported over thousands of kilometres from its source; such long-range transport of dust is one of the phenomena that has been fully recognised with satellite imagery of the earth's surface, and exploration of the solar system has revealed that dust storms occur on a much greater scale on the planet Mars.

There is a wide variety of environmental implications associated with dust storms. Human activity, for example, may be severely affected; Herodotus recounts how the Persian army led by Cambyses was caused to disappear wholly in a dust storm in North Africa while travelling to Ammon (quoted in Greaves 1880). In Yugoslavian Macedonia the demise of the village of Stobi in AD 600 coincided with a gradual accumulation of wind-blown desert dust that completely buried the settlement (Folk 1975). The reduction in visibility caused by dust storms is a considerable hazard to modern air and road transport, the presence of atmospheric dust can interfere with radio wave propagation, and dust storms can transport diseases and radioactive particles from desert nuclear testing grounds.

Perhaps the most widespread hazard to human populations is in agricultural areas, where dust storms represent a large volume of soil erosion. Soil surfaces in arid regions show a high concentration of nutrients in the surface layer, as moisture transport is generally towards the soil surface. If the surface layer is removed, therefore, productivity is decreased. In addition, the desiccating effect of dust winds may reduce crop yields and damage young plants. Nutrients and organic material removed from a soil and transported away, however, will be trapped in vegetation or in the oceans outside the erosion area, and this material contributes to productivity in both terrestrial and marine ecosystems (Lundholm 1979).

There are a great number of other environmental implications besides the hazards posed to man. The importance of desert dust in certain geomorphological situations is now being recognised; perhaps the most fundamental aspect of desert dust deposition is the recognition of peri-desert loess deposits, but deposited dust is also important in the formation of duricrusts and desert varnish and the action of salt weathering among others. Dust deposited in the oceans over geological timescales is proving to be a valuable indicator of past climatic conditions, while dust resident in the atmosphere is important as a source of natural pollution and has effects on meteorological processes and thus, perhaps, climate itself. The range of environmental implications is indicated in table 1.1, with examples of recent work in these fields.

Human actions can affect the output of dust into the atmosphere, and has done so through the history of human occupation of the planet. As Lucretius (c100 - c55 BC), in 'On the nature of the Universe' notes:

> Take the earth first. Part of it, parched by incessant sun and trampled by the tread of many feet, exhales a vapour and flying clouds of dust which strong winds scatter and comingle with the air.
> (Penguin Classics 1951, p178).

With an intensifying use of the world's dry lands by an expanding world population the human impact has become more important as an agent of dust storm generation in certain areas. Such impact is derived from land use practices that result in widespread removal of protective vegetation, destabilization of ground surfaces or the creation of erodible surfaces by drainage. Some examples of these human impacts are shown in table 1.2. The scale of potential impact was illustrated during the Dust Bowl years in the American south-west and again during the Virgin Lands scheme in the steppes of the USSR in the 1950s. Most recently, interest has been focused on the Sahel where wind erosion and dust storms are intimately related to the problems of drought and desertification.

Given the wide range of effects dust storms have on environment, and thus their importance to the disciplines of geography, geomorphology, meteorology and climatology, sedimentology and environmental studies in general, it is surprising that the study of this phenomenon is in its infancy. One of the earliest scientific investigations of dust storms was carried out by Baddeley (1860) in northwestern India, and the work conducted by Chepil and his co-workers after the Dust Bowl era in the USA made significant advances to our knowledge of the wind erosion system. However, in the last fifteen years or so there has been a considerable increase in interest as exhibited in papers by Idso (1976a), Goudie (1978, 1983) and Coudé-Gaussen (1982, 1984a) and the volumes edited by Morales (1979a) and Péwé (1981). In the same period a considerable volume of literature has been produced on extraterrestrial dust storms, particularly on Mars: the works of Greeley **et al** (eg 1974,

1981) and Iversen **et al** (eg 1976) are noteworthy in this field.

Due to the early stage of dust storm research in general, therefore, combined with the fact that some areas of the world have been the subject of much greater scientific effort than others, there are a number of basic gaps in our knowledge of the global geography of dust storm activity. This thesis is designed to go some way towards filling these gaps.

1.2 AIMS

The framework of this thesis is based on a set of specific aims. These are as follows.

1. To map the global frequency distribution of dust storm occurrence, thus identifying major world dust source areas, and giving some quantification to their level of activity.

2. To analyse the seasonal and diurnal patterns of dust storm occurrence from region to region.

3. To identify the meteorological systems that generate dust storms and transport dust on a regional basis.

4. To identify the geomorphological nature of dust storm

source areas.

5. To evaluate the effects of specific meteorological controls such as rainfall and wind speed on dust storm generation.

6. To pay specific attention to the role of human activities on dust storm generation.

7. To analyse changes in dust storm activity over time in relation to 3 to 6 above.

8. To attempt a global classification of common dust-generating meteorological systems and dust-producing geomorphological units.

9. To obtain a global perspective on long-range transport of dust in relation to sources identified in 1 above.

1.3 ORGANISATION OF THESIS

Chapter two reviews the theory of wind erosion and the sources of dust in desert soils, it outlines the working definitions of dust events and describes the data used and the problems encountered in the study of dust storms.

Chapters three to eleven are divided into major world regions of dust storm activity for which analysis of the dust storm systems has been made. These are Africa (two chapters), the Middle East, South-West Asia, Europe and the USSR, China, Australia, North America and Latin America (one chapter each). These chapters include reviews of the literature on dust in these areas and analysis according to the aims outlined above; by necessity the depth of analysis in any particular region varies according to the quantity and quality of data and information available. At the end of each regional chapter a series of conclusions are drawn on the dust storm systems of the region.

In chapter twelve conclusions on dust generation on the global scale are drawn: outlining the major dust-generating areas; relating dust events worldwide to a characteristic suite of meteorological systems and geomorphological units, and describing the major controls on the dust storm system including the actions of human populations. In addition, a comprehensive tabulation of the world's named dust-bearing winds is presented, indicating the regions of generation and areas affected, their seasonality, and the meteorological conditions causing their generation.

CHAPTER TWO : THEORY, DEFINITIONS, METHODS & DATA

2.1 WIND EROSION THEORY

The structure of the wind near the ground surface, and the nature and structure of the surface are interrelated factors. A moving fluid such as wind or water exerts three types of pressure on a soil particle at rest: **impact pressure (+ve)** on the windward area of the particle; **viscosity pressure (-ve)** on the leeward area of the particle; and **static pressure (+ve)** on the top of the particle caused by the Bernoulli effect. In many texts the forces of lift and drag are described in terms of these three pressures, drag on the top of the particle being due to the pressure difference against its windward and leeward sides, and lift is caused by the decrease in static pressure at the top of the particle compared to that at the bottom. In general, these forces of drag and lift act against the cohesive forces holding a soil particle in position, the surface roughness, and the weight of that particle to initiate particle movement at some threshold wind velocity.

Wind velocity is the only widely-used practicable field measurement of wind force, but a wind capable of initiating particle movement is always turbulent, the flow being characterised by eddies moving at variable velocities and in all directions. The limitations of the wind velocity measure are revealed by Wilson and Cooke (1980) who quote a number of measures of turbulence to illustrate the variety of conditions that may be experienced given the same mean wind velocity. Warren (1979) points out that both lift and drag are probably greatly affected by turbulence. Intense eddying that develops in winds over rough granular beds can produce occasional 'jets' of violent activity inducing wild fluctuations of velocity and pressure on the bed. Momentary values of velocity may be greatly influenced by these fluctuations, and since shear (drag) varies as the second power of velocity the effect on entrainment is marked.

Experimental studies of threshold velocities for particle movement have used simple soil systems consisting of loose, monodisperse (all particles the same size) and similar particles (eg Bagnold 1941, Chepil 1951, Greeley **et al** 1974). If the first particles moved do so in direct response to the pressure of the wind -at a **fluid threshold velocity-**, however, most later grains are moved by means of a ballistic transfer of kinetic energy by already saltating grains -at an **impact threshold velocity.** Figure 2.1 shows the relationship between grain size and fluid and impact threshold velocities with the characteristic modes of aeolian transport. This graph shows that if loose particles are available in the soil, 100 micron particles require the lowest fluid threshold velocities for initiation of movement, but once movement has begun smaller sized particles may be moved by lower velocity winds through particle collisions. Although natural soils are rarely characterised by loose monodisperse particles, observations show that a small, highly erosive spot, from which soil particles may be removed by direct wind pressure, is often sufficient to initiate erosion over an entire field (Chepil 1946).

Experimentation suggests, therefore, that dust sized particles (<0.08mm in diameter according to Bagnold 1941) are entrained either by high-velocity winds that are capable of disturbing the fine particles that usually form a highly cohesive, aerodynamically smooth surface, or by ballistic impact from saltating sand-sized grains. Certainly, the role of saltating sand grains is thought to be important in initiating dust entrainment in many environments, by breaking up surface crusts and dust aggregates (Gillette 1979, McTainsh 1985a).

Idealised wind erosion systems are, of course, far removed from the true system of erosion that occurs on terrestrial soils. Not least of the complicating factors are the range of particle sizes found in the real world and the presence of non-erodible elements on surfaces that absorb wind momentum and protect surfaces from erosion. When and where wind erosion occurs is determined by the mutual interaction between the elements of wind erosivity and surface erodibility; and the movement of particles, indicating a departure from the stable condition, may be initiated by a change, positive or negative, in one or more of the variables shown in figure 2.2. In the field these elements of erosivity and erodibility change through time and space, at varying rates and differing scales, so that the relationship between these variables is in a constant state of flux. In order to ease description, it is appropriate, firstly to review the role of wind, followed by that of soil, the sources of dust in desert soils, the threshold velocities for dust entrainment and the affects of ground surface characteristics.

2.1.1 WIND

Wind moves soil particles by virtue of its energy, and the availability of a wind's energy to promote entrainment is related to a large number of atmospheric variables in an extremely complicated manner. As such, the details concerning wind velocity, turbulence, gustiness, shear forces, humidity, and temperature, and their interrelationships in the wind erosion system are little known, but some comments can be made on work done in relation to wind characteristics to date.

Perhaps the most important of these wind variables is velocity, although the limitations of this measure have been mentioned above. Wind tunnel and field measurements have shown that the rate of soil movement is proportional to the third power of mean wind velocity, and Skidmore and Woodruff (1968) applied this relationship in their assessment of wind erosion forces for 212 locations in the USA. For each location three measures were calculated: the *magnitude* of wind erosion forces; the prevailing wind *direction* , and the *preponderance* of wind erosion forces in the prevailing wind erosion direction. The magnitude is calculated for each month in each of sixteen compass points and the total magnitude for each month is given by summing the magnitude from each direction, giving a measure of the relative capacity of the wind to cause soil erosion. The prevailing wind erosion direction is based upon the ratio of forces parallel and perpendicular to the prevailing direction, its final value being a measure of the preponderance of the prevailing direction.

A major deficiency of this technique is that the wind velocity measures include all winds, irrespective of whether or not they exceed the threshold velocity for particle movement at a particular location. This problem is overcome in an index developed by Fryberger (1979) to express wind forces in terms of their potential for sand transport. The index is based on Lettau and Lettau's (1978) equation for the rate of sand drift, from which an expression is derived for the annual or monthly rate of sand drift:

$$Q \propto V^2(V-V_t)t$$

where Q= annual or monthly rate of sand drift

V= wind velocity at 10m height

V_t= impact threshold wind velocity at 10m height

t= length of time wind blew, expressed as a percentage of a year's observations

Fryberger uses an impact threshold velocity of 12 knots. The results are expressed in vector units and the vector unit total for any wind summary is proportional to the sand moving power of the wind at the station of record, and is known as the **Drift Potential (DP).** The applicability of DPs as an appropriate measure of dust-raising potential depends on threshold velocities for dust-raising. In a number of works by Gillette and his co-workers field studies of such thresholds have been undertaken, and these are reviewed below (see section 2.1.4).

A deficiency common to both Skidmore and Woodruff's and Fryberger's indices is that the measures fail to take account of all weather conditions that might be associated with winds from a particular direction. Thus, for example, the prevalent wind might typically bring rainfall while most of the erosion is caused by less frequent but drier winds from another direction. The validity of this point will be assessed in this study as meteorological systems associated with dust-raising in any particular location are to be identified.

The relative humidity of a wind will effect its turbulence characteristics, the threshold velocity of entrainment for a particular soil type and its capacity for transport. Knotterus (1980) has investigated the relationship between threshold wind erosion velocities and the relative humidity of the air. He observed that

> ....under the influence of the relative moisture content of the air, the wind velocity, and the time of exposure, sometimes changes occur in the soil surface; those changes are evidently the cause of the different levels at which the critical wind velocity manifests itself for the same soil.
> (Knotterus 1980, p 531).

Many studies show that soil drifting increases with length of exposed area. A wind may start to entrain material at the windward end of a field and continue to increase its load until it can carry no more if the exposed area is long enough. As a wind continues it may still pick up material but it will also deposit some of its load since the carrying capacity is finite. The maximum rate of transport of a wind with a specific friction velocity is very similar for all soils, although the exposed length needed for a wind to reach load capacity depends upon the erodibility of the particular soil. The more erodible the soil, the shorter the distance to reach carrying capacity. The distance required for the maximum load to be picked up by a strong wind of 18m/s at 10m above the ground varies from less than 55m for a structureless fine sand to more than 1500m for a cloddy medium-tec tured soil (Chepil & Woodruff 1963).

2.1.2 SOIL

Soil erodibility depends largely on the mechanical stability of the soil, which is defined by Chepil and Woodruff (1963) as the resistance of a dry soil to breakdown by a mechanical agent such as tillage, force of wind or abrasion from wind-blown materials. This mechanical stability is dependent upon the size, density and shape of its individual particles, and most soils consist of individual particles held together by various forces as aggregates or clods of varying sizes. Smalley (1970) points out that the mechanical stability of a soil depends largely on these forces of inter-particulate cohesion in the soil system. Perhaps the most important binding agents for these soil structures are the ratio of sand, silt and clay particles within them, their soil moisture properties and the presence of cements such as salts and those associated with decomposing organic matter.

Chepil and Woodruff have investigated the relative effectiveness of silt and clay as binding agents, which 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 100 per cent the silt fraction creates more clods, but clods are softer and more readily abraded 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 clay, 40 to 50 per cent silt, and 20 to 40 per cent of sand. (Chepil & Woodruff 1963, p262).

It is only dry soil particles that are readily erodible by wind since soil moisture promotes particle cohesion and thus restricts erodibility (Bisal & Hsieh 1966). Also, soils differing in texture require different percentages of moisture to resist initiation of soil movement at specific wind velocities. Experimental work by Bisal and Hsieh using the erodible fractions of three soil types confirm the general rule first suggested by Chepil (1955a) that the higher the proportion of silt and clay in a soil, the greater is the production of clods and the lower is soil erodibility. The moisture content of a soil at any particular time is in turn determined by the properties of that soil and by particular weather conditions.

Chepil (1956) found that the erodibility of a soil decreases as the square of the soil moisture increases, upto 15 atmosphere percentage (the amount held at a suction equivalent to 15 atmospheres pressure which is approximately equal to the permanent wilting point) where no erosion occurs.

Water also affects the wind erosion system in the form of raindrop impact, which may form surface crusts, normally consisting of silt and clay sized particles, with coarser particles left loose on the surface (Chepil & Woodruff 1963). These loose particles are easily dried and may be moved by the wind soon after rainfall has ceased, contributing to the initial stages of wind erosion, breaking down the crust by abrasion and enhancing further drying. Rainfall may also reduce erosion through its effect on plant growth (see section 2.1.5).

A variety of cements can be identified that decrease a soil's erodibility, one of the commonest in the desert landscape is salt. Salt may act to combine particles and is also present in the form of hard crusts that cover large areas of the desert surface.

The combined effects of added sodium, calcium and magnesium chloride salts, rainfall intensity and duration on the wind erodibility and mechanical strength of a sandy soil have been evaluated by Lyles and Schrandt (1972). Losses from soils treated with salts were less than untreated soils, and those treated with sodium chloride were significantly less than soils treated with other salts. Sodium chloride treated soils increased in mechanical strength following rainfall and drying much more than did non-saline soils or those treated with other salt combinations, regardless of rainfall intensity or duration. Otherwise, soil loss following low intensity rain was always less than that following high intensity rain for the salt-duration combinations studied, and except for short duration rains at low intensity, rainfall duration did not significantly affect soil loss.

A variety of cements are produced from the breakdown of organic material by micro-organisms. These cements are derived from the decompostion products of plant residues, the decomposer micro-organisms and their secretory products. Together they serve to bind particles together thus improving soil structure. Chepil (1955b) found that between one and six per cent organic matter added to a soil in the early stages of decompostion (less than one year) led to enhanced clod production and decreased erodibility. Over a period of four years, however, there was a decline in clod production and consequent increase in erodibility, as initial cementing materials change, lose their cementing properties and become brittle when microbial activities diminish. The microbial fibres also disintegrate in time and a high proportion of medium-sized water-stable aggregates develop which are highly susceptible to wind erosion (Troeh **et al** 1980). This point is particularly relevant in areas susceptible to prolonged drought, where several years of below-average rainfall will indirectly result in less organic matter reaching a soil surface through its effects on vegetation growth. The process may be exacerbated in areas where widespread removal of vegetation by human populations occurs, as in the Sahel particularly since the 1960s through overcultivation, overgrazing, excessive burning of rangelands and the chopping of wood for fuel. Destroying vegetation also has a more direct effect on wind erosion by subjecting exposed soil surfaces to increased aeolian stress (see section 2.1.5). Other ways in which human action may remove vegetation cover from soil surfaces to varying degrees of permanency are shown in table 1.2.

A number of other processes can be identified as contributing to soil disaggregation. Chepil (1954) investigated the effects of adding calcium carbonate to a soil, and concluded that the general outcome of such additions was to weaken soil structure and increase erodibility. Sandy soils provide an exception to this rule, however, since they display little structure anyway, so that the addition of calcium carbonate, comprising mostly silt-sized particles, acts as a weak cement, reducing erodibility.

Gillette **et al** (1982) have also noted the mechanism that increases the erodibility of soils associated with 'salt blisters'. Working on threshold friction velocities in the Soda Dry Lake, Mojave Desert, California they observed blister-like domes on the salty-clay crust. Several of these domes had split open, exposing loose, dry soil which was highly erodible.

A number of other mechanisms that increase the erodibility of soils related to predominantly clay-based surface materials have been recognised. Blackwelder (1946) and Gillette **et al** (1982) have observed cracking of clay into pellets as small as one to five millimetres in diameter, aggregates small enough to be eroded by wind. The peeling of thin clay crusts by curling has been observed by Yaalon and Ganor (1979) in Israel, Young and Evans (1986) in Nevada and by Gillette **et al** again in the Mojave Desert. Gillette and co-workers suggest that this phenomenon occurs following an infrequent heavy rainfall event, when water may accumulate on the low-permeable surface of a dry lake. Wave action disturbs the soil and suspends some of it, and during calm conditions coarse particles settle, leaving fine clay in suspension. When the water has evaporated a clay film is deposited which will curl upon drying due to mineralogical shrinking. The accumulation of clay and organic material as a foam on the leeward side of a lake by wind/wave action might also occur after infrequent heavy rainfall, and after drying this foam is highly erodible.

The presence of lunette dunes on the lee sides of desert depressions in many parts of the world, and their often substantial portion of fine grained silts and clays confirms the susceptibility of depression surfaces to dust production (Bowler 1973). It has been suggested (Yaalon & Ganor 1979) that this type of action (wetting followed by shrinking) is responsible for much production of desert dust.

2.1.3 SOURCES OF DUST IN DESERT SOILS

Having reviewed the characteristics of soil that affect its erodibility it is appropriate to identify the sources of dust-sized particles found in desert surfaces. Such work also has important bearing on the debate concerning the relative importance of glacially derived and desert derived loess deposits (see eg Pye 1984). The suggestion that glacial grinding is the only important mechanism capable of physically comminuting quartz (Smalley 1966, Smalley & Vita-Finzi 1968) has been challenged in recent years by an increasing body of evidence put forward by desert geomorphologists to show that significant quantities of silt may be formed in the hot desert environment.

The effectiveness of salt weathering to cause static breakdown of quartz grains has been demonstrated both in the laboratory (Goudie **et al** 1979, Pye & Sperling 1983) and in the field (Goudie & Day 1981), and the susceptibility of feldspars and micas to salt damage seems to be even greater than quartz (Pye & Sperling 1983). Other weathering processes may also be operative on the desert fringe as indicated by the high silt content of stabilized dunes compared with active dunes (this may be more than 30 per cent silt in the former [Goudie **et al** 1973] whereas the latter normally contains less than 2 per cent). Pye (1984) considers that not all of this difference can be attributable to inputs of airborne dust, and suggests that other mechanisms for in situ grain breakdown in areas where no salts are present may include chemical weathering along lines of structural weakness (Nahon & Trompette 1982), crack-tip stress corrosion processes and crushing beneath the overlying sand column. Breakdown due to frost weathering may also be a contributing process in some desert areas.

Grains are also comminuted during transport. In desert areas where glaciers are currently active (eg the Karakorams of northern Pakistan), or have been active in the past, the subglacial processes of grinding and crushing produces silt-sized quartz grains. Recent work by Sharp and Gomez (1986) suggests that crushing is the more important of the two. Attrition also reduces quartz grains during fluvial and glacio-fluvial transport, and perhaps most important in arid areas Whalley **et al** (1982b) have shown that aeolian transport, even at very low particle velocities, produces silt-sized debris. These processes of grain comminution are listed in table 2.1 with examples of recent work, both experimental and in the field.

A variety of geomorphological environments contain silt-sized debris that is available for deflation. Coudé-Gaussen (1984a), whose work is largely based on the Sahara, has attempted to categorise desert surfaces that are highly favourable for producing dust:

Les cuvettes de sebkhas - dried out salt lakes of internal drainage, the surface of which is disrupted and rendered mobile by salt cystallisation;

Les épandages d'oueds - wadi sediments containing silt and the floodplains of great rivers like the Niger;

Les surfaces de fech-fech - powdery areas derived from ancient lake muds or on certain argillaceous rocks;

Les takyrs - desert clay soils with polygonal desiccation cracks;

Les affleurements rocheux meubles - outcrops of rocks like unconsolidated Neogene fine-grained sediments.

Yaalon (1986) proposes that in the Sahara only areally small but specific desert environments are dust producing. These include flood deposited debris on alluvial fans and in broad wadis and terminal depressions that receive periodically removed material from desert slopes. He considers that widespread stable geomorphic surfaces on interfluves (regs) and sand seas (ergs) are not significant sources and supplies of dust.

The importance of such geomorphological units will be assessed by looking at the locations of meteorological stations that show high levels of dust storm activity in relation to their geomorphological setting, and by investigating the patterns of dust haze revealed on satellite imagery as indicated in section 1.2 above.

2.1.4 THRESHOLD EROSION VELOCITIES FOR NATURAL SOILS

In recent years Gillette and his co-workers have attempted to determine threshold erosion velocities for a variety of natural desert soil types. Working mostly in the Mojave Desert, Gillette **et al** (1980) tested 38 different sandy soils using a portable wind tunnel. Several different geomorphological settings were chosen such as playas, alluvial fans and aeolian features, and three main factors were identified as increasing the threshold velocity . These were decreasing proportion of sand, increasing size of dry aggregates of the soil, and increasing fraction of soil mass larger than one millimetre in diameter. The different types of soil surface were ranked according to the increase in threshold velocities from the most erodible to the least erodible: disturbed soils (except disturbed heavy clay soils); sand dunes; alluvial and aeolian sand deposits; disturbed playa soils; skirts of playas; playa centres; and desert pavements (alluvial deposits).

The investigation also found that the size distributions of saltating grains were practically the same as the size distributions of the loose aggregates of the surface soil from which the saltating grains were generated. Further, a definite trend of increasing threshold velocity with larger mode of the aggregate size distribution of soils was found, threshold velocity increasing approximately as the half power of the size of the mode of the aggregate size distribution.

The data compared favourably to Chepil's (1951) data for threshold velocity versus monodisperse particle size, lending confidence to the idea that the mode of the size range gives an erosion response similar to that for monodisperse particle size. Thus, the threshold velocity for a particular soil can be predicted if the mode of the mass-size distribution is known for the loose particles present on the surface, although the polydisperse nature of natural soil systems will make such prediction limited.

For undisturbed crusted soils the thickness of crust was found to be an extremely important parameter. No undisturbed clay crust soil with thickness greater than 5mm was eroded by velocities less than 2.5m/s. In a disturbed condition, caused by driving a three-quarter-ton truck over the soil, clay-crusted soils were effective in resisting wind erosion to varying degrees according to the mode of the mass-size distribution. For soils with more than 90% by mass of sand, disturbed threshold lay between friction velocities of 0.2 and 0.6m/s.

In a more recent paper (Gillette **et al** 1982) the relation of soil composition and threshold velocity for undisturbed and disturbed soils having high clay content were investigated. The two most convenient field parameters relating to threshold velocity were found to be thickness of crust and modulus of rupture of the crust. They found that crusts of relatively small modulus of rupture (greater than 0.7 bar) and crust thickness of 7mm to 3mm were effective in resisting erosion for all but extremely high winds. Disturbed crusts having modulus of rupture before disturbance greater than two bar, with thickness less than 19mm did not ec perience significant wind erosion, yielding a very thin surface layer of fine material that eroded at friction velocities less than 0.5m/s. When the small amount of material (a few grams per m^2) was eroded away by the wind the underlying crust was erodible. Crusts that had modulus of rupture less than one bar were totally disintegrated by truck tyres, and the threshold friction velocity was less than 0.45m/s for the powdery residue. The threshold velocities of disturbed soils were found to vary inversely with the severity of disturbance and to be proportional to the modulus of rupture times thickness squared for modulus of rupture greater than one.

The modulus of rupture of the soil was found to be related to soil composition but was shown to depend mostly on clay content. The strength of the crust was found to increase with a higher percentage of exchangeable sodium and calcium carbonate proportional to the percentage of clay, whereas organic materials seemed to make no difference to crust strength in the desert soils tested, although recognition of biological effect was not possible with the methods used. Gillette (1982) has incorporated the data from the work described to produce a rough generalisation of threshold velocity relative to desert geomorphology (fig 2.3).

2.1.5 GROUND SURFACE CHARACTERISTICS

It has been shown that threshold wind erosion velocities are determined by the availability of loose particles at the soil surface, so that threshold velocity will be increased by the effect of aerodynamic partitioning of wind stress by non-erodible elements, such as pebbles and larger objects and vegetation. Wind erosion will continue on a surface until a sufficient number of non-erodible elements are uncovered to provide direct cover and shelter to remaining erodible grains. This situation may alter should the wind change direction or velocity, but for a given direction and velocity the point at which the cover is sufficient to prevent particle movement continuing or starting is called the 'critical surface barrier ratio'. This ratio is defined as the distance between the non-erodible barriers divided by the height of the barriers (Chepil & Woodruff 1963), and its value was found to vary between 4 and 20 on cultivated soils, depending on the wind shear velocity at the time and the threshold shear velocity of the erodible fractions.

Further investigation into the effects of non-erodible elements in the absorption of wind stress has been carried out by Marshall (1971) who found that for a configuration in which the silhouette area of non-erodible roughness element per floor area (lateral cover or Lc) was greater than 0.1 most of the wind's momentum was absorbed by the elements, leaving little stress available for movement of the erodible fractions lying between the non-erodible elements.

A similar effect for Lc >0.1 was observed by Lyles and Allison (1976) while studying the threshold velocity of sand between non-erodible dowel rods. Marshall (1971) also noted that slender elements with a prominent upper edge are more protective than broad, rounded elements. Thus, it seems that even a relatively sparse ground cover of non-erodible elements will provide a large amount of protection from fine particle entrainment, although there may be a stage at which a few non-erodible elements may affect the wind flow so that more erosion will occur than if no elements were present (Logie 1982).

The effect of non-erodible elements is most obvious on so-called 'wind-stable' surfaces such as stone pavements. Stone pavements occur widely in environments with little vegetation, especially hot deserts, and generally act to protect otherwise potentially wind-erodible surfaces such as residual weathering mantles or alluvium (Cooke 1970). Deflation can occur from desert pavements, however (Chepil & Woodruff 1963), and it may be that the 'wind-stable' surface, in reducing the high frequency/low intensity deflation events, allows the build-up of a reservoir of fine material that is susceptible to removal during a violent wind storm as Bagnold (1941) suggests.

Surface ridges, produced by tillage, effect the quantity of a soil that is eroded. This effect is dependent upon the height and lateral frequency of ridges, their shape, orientation to the wind direction, and their proportion of erodible to non-erodible elements. The most effective orientation of ridges is at right angles to the erosive wind.

The factors tending to reduce the rate of soil flow over ridges as compared to a smooth surface appear to be the reduction of the average wind velocity for some distance above the ground and the trapping of entrained particles on the leeward side of ridges (Chepil & Milne 1941). At the same time, however, ridges cause an increase in wind eddying and a greater wind velocity and consequently a greater wind erosive potential, at the crests of the ridges. Chepil and Milne confirmed that the gross effect of the former pair of factors was always markedly greater than that of the latter.

Ridges do not protect a surface from erosion per se, their effect decreases as the wind moves parallel to them, and if ridges are composed wholly of erodible grains then there will be no soil protection. But if ridges are composed of both erodible and non-erodible fractions, as they commonly are, the erodible fractions are moved from ridges to furrows where they are trapped, and the ridges soon become stabilised with a mantle of soil aggregates too large to be moved by wind.

Vegetation influences the nature of wind erosion in several ways. The quantity (proportion of ground surface covered) and quality (height, density and flexibility) of vegetation governs the extent to which a surface is exposed to erosion and the degree by which surface roughness is increased. These properties will of course vary with vegetation type and , for a given type, according to the time of year. The latter point is particularly important where sown crops are concerned, and the length of time for each crop to grow sufficiently to protect soil against erosion is another important factor. Also, the orientation of crop rows affects the nature of wind erosion in the same way as described above for ridges. In general, the taller the crop, the finer the vegetative material and the greater its surface area the more wind velocity is reduced. Chepil and Woodruff (1963) suggest that grass offers one of the most efficient protective covers, grass that is easily bent in the wind being less protective than a structurally stronger type.

Vegetation also stabilises soil structures through its root systems, and vegetative decay adds organic matter to the soil. Plant litter is important in protecting the soil surface, as it both reduces wind velocity and traps eroded material, as well as by contributing organic cements. With this understanding of the ways in which vegetation affects the erodibility of a soil surface it follows that should the vegetative cover be removed from any area, that surface will become more susceptible to wind erosion and dust storm generation. Such removal may occur over large areas through natural processes, particularly in semi-arid and drought-prone regions, and in many instances human and animal populations may also destroy vegetative cover as mentioned in section 2.1.2 (table 1.2).

Topography may have a number of effects on wind erosion. Over long slopes, short slopes not exceeding 1.5%, and level land the velocity gradient and friction velocity are reasonably constant for a given wind (Troeh **et al** 1980). Over relatively short slopes wind shear is greatest on the upper part of windward slopes, a local variation noted by Bagnold (1941) while examining wind shear above a sequence of sand ripples. Chepil **et al** (1964) examined the consequences of this local concentration of wind shear in knolly terrain, and showed that for windward slopes less than about 150m in length soil loss increases rapidly with both increased slope and distance towards the top of the knoll. Variations in local topography will influence the nature of wind erosion in any particular area, so that the presence of knolls and hollows for example is likely to affect the variables influencing erodibility, lower soil moisture being likely on the higher parts of knolls due to better drainage, and so on (Wilson & Cooke 1980).

Larger scale topography may have affects on a potentially wind-erodible area, so that the configuration of a valley for example may induce the transport of fine sediments to valley bottoms where they are susceptible to locally increased wind speeds as airflow is channelled by the funnelling effect of the valley (the Venturi effect). Topography also plays a part in the formation of particular meteorological systems that produce strong winds capable of wind erosion and dust storm generation. Thus, the intense solar heating of wide flat desert landscapes during daylight hours induces convective activity and the production of turbulent wind flow, and mountain and valley slopes may induce katabatic flow that locally increases wind speeds.

2.2 DEFINITIONS

Before outlining the data and methods employed in this thesis it is necessary to define some of the terms that are to be commonly used. Dust present in the atmosphere can be defined in terms of its source, composition, and grain size. The material that comprises dust storms and other dust events is very largely derived from terrestrial soil surfaces and is hereafter referred to simply as **dust,** or **aeolian dust,** indicating the process by which it is introduced into the atmosphere. Estimates of global dust contributions to the atmosphere vary widely: Peterson and Junge (1971) suggest 500 million tons a year, whereas Schutz's (1980) estimate is up to ten times that figure. The main regions where aeolian dust originates are largely coincident with the world's major deserts, and the chemistry of the dust is dominated by silica (SiO_2), the great bulk of which is made up of quartz (Goudie 1978). Other quantitatively important components include Al_2O_2, Fe_2O_3, CaO, MgO, K_2O, water and organic matter, although locally some components may be more important, depending on the nature of the source region.

The grain size of dust is <0.08mm according to Bagnold

(1941), but a sub-division can be made depending on the distance that the material, once entrained, is transported. Dust that is carried a few kilometres to less than 100km is generally between 0.005mm and 0.05mm. Material that has the potential for long-distance transport is generally smaller than 0.02mm in diameter (Gillette 1979), and can remain suspended in the troposphere as an aerosol, often at high levels, for many days, sometimes more than a week. Jackson **et al** (1971) note that dust of this type is between 0.002mm and 0.01mm, and Saharan dust collected over the Caribbean by Prospero **et al** (1970) was more than 98% smaller than 0.01mm in diameter. Cumulative frequency curves of the major grain size groups of desert dust are shown in figure 2.4, with a comparative curve for a typical loess. Long-range transport of dust is reviewed in detail in individual regional chapters, but table 2.2 shows some of the available data. Satellite monitoring programmes and calculations of atmospheric dust loadings indicate that the Sahara is the most important source area for long-range transport, while the deserts of China and Mongolia are a second major source.

Any instance in which dust is present in the air will be referred to as a **dust event,** and all dust events are part of the group of phenomena known to weather observers as lithometeors - an aggregate of very small particles most of which are solid and non-aqueous, more or less suspended in the air. The following terms are employed to define dust events generated by aeolian processes, and these definitions conform to the international standards of the WMO.

DUST STORM : Large quantities of dust raised into the air by strong, turbulent winds, reducing surface visibility to below 1000m.

DUST HAZE : A suspension of dust in the air which is not being entrained at the point of observation. Thus, the dust may have been raised at that point prior to the time of observation, or at a considerable distance from the observer. Visibility may or may not be reduced to below 1000m.

BLOWING DUST : Dust raised into the air to moderate heights above the ground reducing visibility at eye level (1.8m above the ground) but not to below 1000m. For synoptic observing purposes the dust is raised by wind, but blowing dust may also be generated by human activity such as the passing of motor vehicles.

DUST WHIRL (or DUST DEVIL) : A localised vortex in which dust is raised in a spiralling column of air with an approximate vertical axis. It seldom travels far, is characteristically of short duration, and is associated with very unstable air and strong convection currents.

2.3 METEOROLOGICAL SYSTEMS GENERATING DUST STORMS

In accordance with the aims outlined in chapter 1 an attempt will be made in this thesis to produce a comprehensive classification of meteorological systems that cause dust events in all parts of the world by building up a picture of systems that operate in the major world regions. A review of the literature on dust events reveals a number of common meteorological systems that cause dust-raising through their generation of strong and turbulent winds. Broadly, these can be divided into mesoscale and synoptic scale systems. The most comprehensive attempt to classify the dust-raising systems in any one area has been made for the Great Plains (Henz & Woiceshyn 1980) and this is shown in table 10.1. This classification will serve as a basis for some initial comments on dust-raising systems.

The dust devil is the smallest scale feature which occurs in most dryland regions and in more temperate latitudes during dry periods. It raises dust in a near vertical spiralling column of air generated in a very unstable boundary layer that is usually just a few metres in diameter but may be up to several hundred metres high. These systems may travel across the landscape, but seldom last more than half an hour and often die out in a much shorter time.

The **Haboob** is a storm caused by dust raised in the downdraft of cold air from a thunderstorm cloud. The term originates in the Sudan where such storms are frequent. These storms are characterised by an advancing dust wall which is generated by the strong winds at the gust front some distance before the advancing cumulonimbus clouds. This gust front represents the forward edge of the cool, moist air mass that has descended below the thunderstorm cloud base and entered a significantly drier layer of air and is thus cooled, causing acceleration downward. This air arrives at the ground surface at high velocity and spreads horizontally, whirling up clouds of dust. As such, therefore, the arrival of a thunderstorm downdraft dust storm is characterised by a sharp drop in temperature and rapid rise in pressure with a sharp increase in wind speed and often a change in direction, as well as the sharp drop in visibility due to the dense dust cloud.

These storms occur in many parts of the world, and the term **Haboob** is used in the literature as a generic term for such events. Such use, however, can be misleading. Meteorologists recognise two major types of thunderstorm, the single cell, and the severe thunderstorm (or squall line) that comprises a series of more or less joined up individual cells. Both systems can cause dust storms, the single cell being more limited in its areal extent and duration. Strictly, the **Haboob** is a single cell thunderstorm downdraft storm, but the term has been assigned to indefinite types of dust storm in the Sudan (Morales 1979b) and other regions. Thus, the term 'thunderstorm downdraft dust storm' is preferred in this thesis. Where information is available the type of thunderstorm system is indicated.

The effect of topography on wind flow has been mentioned in section 2.1.5 above. In areas where mountains or other steep slopes are present airflow may be locally increased by katabatic effects, and also by channelling through valley constraints, that may be further enhanced beneath temperature inversions.

Dust-raising occurs along the leading edge of fronts as they pass over erodible areas by high velocity winds that accompany intense baroclinal gradients, and are further enhanced by a dynamic transfer of momentum into the boundary layer. These storms are often characterised by a dust wall, usually on a larger scale than the thunderstorm downdraft.

The cyclonic dust storm system is divided into four components by Henz and Woiceshyn. Strong winds may be caused at the surface by the generation and transfer of momentum as the boundary layer becomes adiabatic and surface pressure gradients intensify after a low level jet forms near the top of the boundary layer within the developing cyclone. Strong surface winds may also be generated by momentum transfer beneath an upper level jet, and the surface cyclone itself may generate dust-raising winds by cyclonic gyre as the cell intensifies and winds sweep out in all directions around the cyclonic centre.

2.4 METHODS AND DATA

Study of individual dust events and long-term dust storm frequencies involves four main methods of investigation that may be used independently or in conjunction: terrestrially observed meteorological data; data from remote sensing platforms; aerosol monitoring; and reconstructions from deep-sea and ice cores.

2.4.1 TERRESTRIALLY OBSERVED METEOROLOGICAL DATA

Dust events are recorded at meteorological stations using the WMO SYNOP codes that delimit a range of dust events by visibility, whether dust is raised locally or advected from afar, and some element of event scale (see fig 2.5). Although summaries of SYNOP observations can be prepared from WMO data banks, the cost (US$45,000) was beyond the budget of this thesis. Tracking of a number of dust events has been carried out using SYNOP data, but data for the world wide frequency distribution analysis were obtained from national meteorological services. These data were obtained in a variety of forms, and the only compatible world wide data base available is in the form of **dust storm days,** defined as a day on which a dust storm is recorded. In the large majority of countries this involves a reduction of visibility to below 1000m, in accordance with international standards and the definition used in this thesis.

The term **dust storm** has, however, been employed using other definitions. The Japanese Meteorological Agency define a **dust storm** as a dust haze or dustfall from Asian sources (known as **Kosa**) with visibility < 10km (Arao & Ishizaka 1986). Other authors have used different visibility limits for dust-raising events: Oliver (1945) < 700m; Orgill and Sehmel (1976) < 7miles (about 11.3km); Péwé **et al** (1981) < 800m, and Jauregui (1973) does not use a visibility limit. In this thesis only the data from Argentina do not comply with the 1000m visibility limit. The data in that case do not incorporate a visibility limit, and as such is a more subjective data set and not strictly comparable with data from other world regions.

Data have been collected from a large number of countries where dust events occur. In some countries, however, dust events are not recorded routinely at meteorological stations (eg Chile, Mexico). In Mexico, therefore, I have collected data from observations made at airports in the country, but this was not possible for other countries in this category. In other cases data were unavailable for essentially political reasons, as in the case of the USSR and Mongolia, or due to internal political unrest (eg Chad). The data and information used in this thesis were made available by the national meteorological authorities listed in the Acknowledgements.

Dust storms are recorded by meteorological observers using fixed objects as visibility targets in all countries of the world where such phenomena occur, and such observations thus introduce an element of observer error and subjective decision making. A subjective element may also be present in distinguishing between dust related visibility deterioration and that due to other causes such as fog, smoke and anthropogenic pollutants. The effects of atmospheric pollution was particularly obvious while collecting data at Mexico City International Airport where wastes from industry and motor vehicles maintain a more or less permanent haze over the city. Thus, although in this case records are kept of all atmospheric dust events that reduce visibility to below 11 miles, only the data for <3 miles were used.

As with all meteorological data there may also be changes in observational procedures. Hulme (1985) illustrates this point in the Sudan where he has noted ephemeral objects such as trees and electric lights being used at Sudanese stations. There may also be a subjective element involved in making the distinction between locally raised dust (**dust storm** sensu stricto) and that blown from elsewhere (**dust haze** sensu stricto) and this distinction may be blurred by using the term **dust storm day.**

A dust storm day also has important drawbacks in that it does not indicate the duration of dust events at a resolution finer than the 24-hour period, indeed a dust storm day may include more than one dust storm within that period (in areas for which more detailed data were available information on event duration has been included). In addition, no indication is given of the areal or volumetric extent of dust raised (in some cases these parameters are given by the study of individual events).

The availability of data may present problems, particularly in remote areas where station densities are low and where observation frequencies during each day are limited. It seems certain that mesoscale dust storms, such as those caused by a thunderstorm downdraft, may go unrecorded in many areas of arid and semi-arid terrain where stations are completely absent. Synoptic scale events, too, may go unrecorded by surface observers but may be identified on imagery derived from satellites (Wolfson & Matson 1986) or Space Shuttle missions (see plate 2.6 and section 2.4.2). Comparison of Space Shuttle imagery with surface station reports during a large dust event over West Africa showed that dust events may be reported by less than 25% of the surface stations in the dust pall area photographed by the astronauts (Middleton **et al** 1986). Remote sensing imagery has already proved an invaluable tool in tracking dust palls across oceans, where surface stations are few and long-range dust transport may occur in a particular layer of air above the surface (eg Prospero & Carlson 1981 for Saharan dust over the Atlantic, Iwasaka **et al** 1983 for central Asian dust over the Pacific).

Despite these problems, meteorologically observed dust storm days provide a very useful indication of the world-wide distribution of dust storm activity, and enable analysis of the seasonality of dust storms. From this data base more detailed local investigations have been undertaken in areas for which data were available, to identify the diurnal variability of dust-raising and relate it to synoptic processes.

Major global sources, identified using dust storm day data, have been confirmed in many instances with remote sensing imagery, and tracking of individual dust events from such sources has been conducted using surface station SYNOP reports and remote sensing imagery. As well as the present weather codes for dust events referred to above, SYNOP data also include other simultaneous meteorological readings, so that in combination with synoptic charts these codes give some reliable indications of the areas of dust-raising, dust transport and synoptic scale generating systems. Thus, for example, if the onset of the dust storm is accompanied by a sharp drop in temperature and change in relative humidity then this would suggest a dust storm associated with the arrival of a different air mass to that previously at the station, which may represent a frontal or thunderstorm downdraft dust storm.

In addition to the observer problems noted above, others become clear. Mesoscale dust events are not easily identified using the SYNOP codes due to low station network densities and low observation frequencies. Indeed, some events may also be recorded wrongly. This problem is perhaps most accute in the case of thunderstorm downdraft dust storms. Although there is a specific code number for these systems (ww = 98), it has been noted that the dust wall generated is located often some tens of kilometres in advance of the thunderstorm cloud. Thus, the arrival of the dust storm is not necessarily immediately associated with a thunderstorm, the dust itself may obscure the towering cumulonimbus clouds, and the strong winds may drown peals of thunder.

My own analysis of daily SYNOP data for northwestern India and Mexico, both areas where thunderstorm dust-raising occurs, tends to confirm these suggestions in that the codes used appear to under-represent the occurrences of these systems.

The tracking of dust events using the WMO network of surface stations was one of the areas for further research recommended by the workshop on Saharan Dust held in Gothenburg in 1977 (Morales 1979a). To date, just two other studies of this nature have been conducted, both in North Africa (Bertrand 1977, Morales 1979b). In this thesis I have carried out similar studies in north-west India.

2.4.2 REMOTE SENSING

Various kinds of remote sensing imagery obtained from Earth orbit offer a good opportunity for calculating the source regions, areal extent and trajectories of major dust events. Useful images are collected by three types of orbiting platforms: geostationary meteorological satellites (eg GOES, Meteosat, GMS, SMS); polar orbiting satellites (eg NOAA-n series, DMSP series); and the Space Shuttle orbiters. Each type of system has its merits and weaknesses when the data are used to attempt to monitor dust outbreaks (Middleton **et al** 1986).

Meteorological satellites in geostationary orbit keep a constant watch over continental-scale regions of the Earth. Though their data transmission is continuous, the spatial resolution is typically quite low (4 to 20km), and their visible and thermal radiometer bandwidths are broad. The thermal instruments are designed to detect the extent of relatively cold cloud tops contrasting against warm ocean and land surfaces. As a result, warm tropospheric dust palls often create insufficient temperature contrasts to be detected by the thermal radiometers on board meteorological satellites. For example, the Meteosat-2 thermal infrared image in plate 2.1 gives no indication of atmospheric turbidity over the eastern Mediterranean on 17 April 1985 at 0600 GMT whereas the STS astronaut photograph taken at 06:16:04 GMT on the same day (plate 2.2) shows a dust pall over the Nile delta and Sinai.

The low spatial resolution of the visible data from these platforms often makes dust identification difficult. In many cases, the disappearance of high-contrast terrain features usually detected by these sensors is the signal of a dust event, rather than an obvious well-defined dust plume seen on the imagery. Only the largest and densest dust systems are well-documented by geostationary satellites such as Noyalet (1978) with Meteosat over West Africa, Gurka (1977) with GOES-1 over the USA, and Kästner **et al** (1980) with SMS-1 over West Africa and the Tropical North Atlantic. This last study is presented as a 16mm filmstrip which animates a sequence of images of a dust event over the period 27 July to 4 August 1974, tracking material to the Caribbean using images in the visible part of the spectrum. This team of scientists has developed a method for determining the atmospheric turbidity (see section 2.4.3) caused by dust in the atmosphere over the ocean using these visible images (Jobner **et al** 1980).

The Advanced Very High Resolution Radiometer (AVHRR) on board the NOAA Polar Orbiter series, and similar instruments on the DMSP series image large regions from low Earth orbit. Though the AVHRR is in continuous operation and can be monitored by ground stations within its footprint, only limited recorder time is available to collect data from remote parts of the world for retransmission to ground stations either in the USA or at a number of other receiving stations. This lack of recording capacity, coupled with the sun-synchronous polar orbits, prohibits constant monitoring of regions with frequent dust activity, limiting image collection to one or two orbital passes a day. The AVHRR may be operated with one or four kilometres spatial resolution and has visible, near-infrared and thermal radiometers with medium bandwidths. Though the presence of dust is not always unambiguous in the thermal channels, the visible and near-infrared radiometers permit precise determination of dust extent and trajectory during daylight passes. Given its 2200km swath width and approximately 2000km scene length, the Polar Orbiter AVHRR is capable of providing detailed information on pan-regional dust events (eg the sequence shown in plates 2.3-2.5).

The most impressive orbital imagery comes as a result of the training of STS astronauts to recognise and photograph dust events with hand-held cameras. Their visible colour film images may achieve 20m spatial resolution in nadir views and can provide stereoscopically-overlapping frames for three-dimensional analysis. During STS missions, surface station reports and meteorological satellite data are monitored for dust events and messages are transmitted to the astronaut crews to alert them of orbital passes over relevent areas. Conversely, the astronauts often report and photograph dust events not recorded by other data sources: plate 2.6 shows a dust plume blowing northwards from an area north-west of Tokar in north-east Sudan in October 1984. The pall was identifiable on other images as far as the eastern Mediterranean (plate 2.7), but a survey of the meteorological observations made in the area showed no reports of any dust events as the pall was located over an area with very few observatories.

STS missions offer only periodic regional coverage, however, due to launch schedules, orbital parameters and conflicting crew activities. Nevertheless, astronaut photography has provided new information on dust dynamics and a means for evaluating the success of other data sources for providing information concerning the occurrence, scale and trajectory of dust events. Thus, photography can be compared to simultaneous visibility and thermal radiometer data gathered by geostationary meteorological satellites, as well as surface station reports.

The photographs may eventually lead to new means for calculating the mass budget of dust events. By examining the astronaut dust images exposed at a known time with a given sun elevation, it may be possible to derive a measurement of relative optical depth in relation to clear, stable areas of known surface characteristics, such as the ocean. In combination with an increasing knowledge of the vertical structure of dust plumes the chemical composition of dust particles and their size distribution, such a measurement would yield the minimum dust density required to completely obscure surface features. If such a measurement can be obtained, the imagery from orbital remote sensing platforms will be capable of providing additional information on dust event density to complement that available from satellite data sources (Jobner **et al** 1980), as well as trajectory and areal extent.

2.4.3 GROUND SURFACE TURBIDITY MEASUREMENTS & AEROSOL MONITORING

In recent years a number of authors have monitored dust events using measures of atmospheric turbidity and sampling of aerosols. Atmospheric turbidity is defined as the extinction of solar radiation by suspended particles that are large enough with respect to the wavelength of light; that is particles with radii from about 0.1 to 10 microns. Thus, such a measure is one of atmospheric dust concentration. A number of turbidity studies have been undertaken over North Africa (eg Jaenicke 1979) and across the northern Equatorial Atlantic as part of the Global Atmospheric Research Program (GARP) Atlantic Tropical Experiment (GATE) (Prospero **et al** 1979). More recently a network of sunphotometers has been set up across the Sahara and Sahel to systematically monitor dust output (d'Almeida **et al** 1983), and McTainsh (1980) working in northern Nigeria has demonstrated a relationship between dust deposition, solar radiation (as measured by turbidity) and horizontal visibility. When monitoring aerosol transport over very long distances, however, other effects on turbidity such as a change of air mass over a monitor, may mask the effects of suspended dust (Rahn **et al** 1981).

Aerosol sampling may be in the form of deposition (wet or dry) or by using an air pump with replaceable filter. Results from the latter are often expressed as Total Suspended Particulates (TSP) in microgrammes/m^3. Goudie

(1978, table 6) presents data on dust deposition for individual events which may be of very high magnitude. The 1903 Saharan dust fall over England, for example, is estimated to have deposited 9.8 x 10^6 tonnes of sediment (Mill & Lempfert 1904). When expressed per unit area rates may reach several hundred tonnes/km^2. Monitoring over longer periods allows estimates to be made of annual deposition rates (table 2.3). These rates seem to be of the order of 50 tonnes/km^2/yr. In terms of mm/1000yr, Goudie shows that

> ....it becomes apparent that when compared with rates of fluvial erosion rates of aeolian accummulation (themselves representing rates of aeolian deflation elsewhere) may be of a similar order of magnitude. (Goudie 1978, p300).

For the Sahara, Goudie derives a wind erosion rate of about 20mm/1000yr from the dust accumulation rate in the North Atlantic which compares with recent continental dust deposition rates estimated for Kano (116mm/1000yr, McTainsh 1980) and Ndjamena (42mm/1000yr, Maley 1980). Monitoring of Asian dust deposited in precipitation over the Pacific, however, suggests that estimates of the dust contribution to ocean sediments based on extrapolation from terrestrial measurements may be invalid. While over land the largest particles are scavenged most efficiently by rain, there is no clear evidence of particle size dependency for scavenging in marine areas (Buat-Menard & Duce 1986).

The dust contents of air in major dust storms may range from 10^2 to 10^5 microgrammes/m^3 (Goudie 1978) while average concentrations over major world oceans are much lower (less than 1.0 microgrammes/m^3) except those adjacent to large deserts where concentrations are at least an order of magnitude higher : Mediterranean (4.29); Indian Ocean (4.76); with the highest mean concentration over the tropical North Atlantic (14.2) according to the most extensive study conducted by Prospero (1979). During large, long-range dust events these mean values are exceeded, so that concentrations during peak months at Cayenne, for example, were 29 and 23 microgrammes/m^3 in 1978 and 1979 respectively (Prospero **et al** 1981). It is important to note that aerosols may not be sufficiently characterised by a single parameter such as particle concentration, as Jaenicke and Schutz (1978) found aerosols over the Cape Verde Islands had a particle concentration comparable to that of 'clean air' but a total mass similar to 'polluted air' (as defined in The Study of Man's Impact on Climate 1971).

Material collected in dust sampling programmes is processed in a variety of ways, including analysis of elemental and chemical composition. A commonly employed technique for determining sources of atmospheric dust is the aerosol-crust enrichment factor, defined for an element x by:

$$EFx = \frac{(x/Al)_{aerosol}}{(x/Al)_{crust}}$$

where x/Al refers to the ratio of concentrations of elements x and Al in the crust and in the aerosol. Aluminium is used as a reference element for a number of reasons: it is easily determined by neutron activation even at very low concentrations; easily determined by other analytical techniques (eg by atomic absorption); it has very few specific pollution sources, and it is generally free of contamination during sampling (Rahn **et al** 1981). Although soil, rather than crust, ought to be taken as a reference material, few data are available for its elemental composition on a global basis. Other crustal reference elements commonly used are Si and Fe. Some of the other methods for determining source regions of dust transported over long distances are summarised in table 2.4.

2.4.4 DEEP-SEA CORES/ICE CORES

The variation in aeolian action over geological timescales can be determined from deposits in ocean sediments and ice caps. Many of the indicators of source regions used in aerosol collection are also employed to identify the sources of material in ocean and ice cores (ie grain size distribution, chemical and elemental composition, the presence of plant remains etc). In ocean sediments the structure of the sediments may also preclude deposition by water currents (see eg Eriksson 1979). When the deep-sea sediment record is used in conjunction with aerosol collection from ships annual mass budgets can be calculated as shown in figure 2.6 for transatlantic dust events derived from the Sahara.

Identification by these methods of variations in rates of accumulation during geological time have made a useful contribution to our knowledge of past climates. The evidence for pre-Quaternary climates is reviewed by Goudie (1983), and recent additional work has been completed by Rea **et al** (1985) who have analysed dust components in North Pacific cores spanning the past 70 million years. There seems to have been some considerable variation in dust sedimentation during the Quaternary itself. Thompson (1977) indicates a period of high dust deposition at the last glacial maximum (about 18,000 years BP) from analysis of the Camp Century Ice Core in Greenland. This conclusion is confirmed by Sarnthein and Koopmann (1980) from their investigation of ocean sediments off the West African coast, sedimentation rates in this area during this period being about double what they are today. These conclusions fit in with the idea of this most recent glacial maximum being a period of aridity as well as cold, and with the vastly increased area covered by deserts in the tropical latitudes (Sarnthein 1978).

Athough deep-sea sediments clearly contain a detailed record of varied aeolian inputs that must be related in some way to past climate over the continents, however, the interpretation of such records is not necessarily straightforward (Prospero 1985). The relationship between aridity and dust storms is not a simple one. Goudie (1983) has shown that the correlation between dust storm frequency and rainfall is very weak, and that dust storms are , in fact, at a maximum in regions with 100 to 200mm annual rainfall. It has also been noted above (section 2.1.3) that water may be important for the production of large quantities of fine particles. Further, the relationship of dust storm frequency on the continents to the rate at which material is transported over long distances is also complex. Prospero concludes that the interpretation of deep-sea core records will be established by further work in this field:

> ...together with studies of the present-day relationship between weather and processes of dust mobilization and transport. (Prospero 1985, p280).

This thesis will attempt to make a significant contribution to this latter area of research.

CHAPTER THREE : AFRICA

3.1 INTRODUCTION

Wind-blown dust in the Sahara, the world's largest desert, is referred to in some of the earliest written texts. Herodotus recounts how the people of the country of Psylli in North Africa were buried and destroyed in a dust storm (quoted in Greaves 1880), and the Atlantic off the coast of West Africa was known as the 'Mare tenebrum' (Sea of Darkness) in classical times due to the remarkable reduction in horizontal visibility caused by atmospheric dust. A long record of dust deposition is available in the peri-desert loess deposits, particularly in North Africa and the Middle East, and in the deep-sea sediments of the Atlantic and Mediterranean.

In recent years research has focused on the quantity of material blown from the Sahara, its origins and ecological impact. Since the 1960s, drought in the Sahel and the associated famine and ecological catastrophes, epitomised by the process of desertification, has led to renewed interest in dust generation as an agent of soil erosion and its effects on climatic change. This interest stimulated the Workshop on Saharan Dust, held in Gothenburg in 1977, which focused on the three processes of mobilization, transport and deposition (Morales 1979a).

Rough estimates show that the Sahara contributes between 60 and 200 million tons of soil dust to the troposphere each year according to Prospero and Carlson (1972), while Schütz **et al**'s (1981) figure is 260 million tons to the Atlantic alone. Estimates by d'Almeida and Jaenicke (in press) for the whole Sahara are 630 and 710 million tons in 1981 and 1982 respectively, with 60% of this mass moving southwards to the Gulf of Guinea, 28% westwards to the equatorial North Atlantic and 12% northwards to Europe. These estimates suggest that the Sahara may contribute as much as 50% of the total soil dust emissions into the troposphere, making the Sahara the largest area of dust storm activity in the world. Relatively, the deserts of southern Africa are much less important. Some of the material from the Sahara is regularly transported across the Atlantic to the Caribbean, North and South America, to Europe and the Middle East, to the Soviet Union and south-west Asia. Clearly, this amount of material has important effects on erosion, sedimentation, meteorology, climatology and ecology at local, regional and continental scales.

3.2 NORTH AFRICAN CLIMATE & EFFECTS ON DUST-RAISING

In order to obtain an understanding of the generation of dust storms at the local and continental scale in North Africa it is appropriate to outline the climate and dust-generating systems that prevail in the region. Broadly, the raising and transport of dust in the southern Sahara and Sahel can be explained with reference to the prevailing winds over the region, dust winds being essentially monodirectional in general accordance with the north-east/south-west flow of the **Harmattan.** In the northern Sahara, by contrast, the effects of depressions that cross the Sahara and the Mediterranean basin play an important role, so that dust winds are multidirectional. This discussion is facillitated, therefore, by a division into the southern and northern Sahara.

The climate of the southern Sahara and its Sahelian fringe is very largely dominated by the seasonal movement of the Intertropical Convergence Zone (ITCZ) or Intertropical Front (ITF), a low pressure zone that actually consists of a string of low pressure centres (Dubief 1979). The ITCZ demarcates the boundary between a warm, moist maritime equatorial air mass to the south and a very dry, thermally turbulent air mass to the north which in winter is cool in the Sahara but warm in the southern Sudan and in summer is always very hot. The ITCZ is situated over the Gulf of Guinea in winter and migrates northward to a maximum position in summer when it reaches about 20^{o}N. The main dust-raising season in the southern Sahara is during the winter when the ITCZ is over the Gulf of Guinea, and the area is affected by the dry air mass north of this zone. This air mass is described by Dubief (1979) as flowing from the belt of high pressure that extends from the Azores across north Africa to south Asia. The relevant wind is a weak northerly at source but gradually turns eastwards and increases in force with decreasing latitude. This is the **Harmattan** wind that raises much dust in the Bodélé depression of the Chad Basin between Bilma (Niger) and Faya Largeau (Chad) and transports this material towards the south-west. An important effect on the dust-raising capacity of the **Harmattan** is the influx of cold air outbreaks into North Africa after the passage of mid-latitude depressions across southern Europe and the Mediterranean. The development of conditions conducive to dust-raising associated with these cold air outbreaks has been investigated in detail by Kalu (1979, 1982) who suggests that the development of high wind speeds at low level over the Bilma/Faya Largeau area is generated in the following sequence.

Cold polar air, usually associated with cold fronts, intrudes into North Africa in the form of upper troughs from mid-latitudes on the easterly side of a mid-latitude blocking high pressure cell situated over 0^0 to 5^0E longitude. These cold air intrusions in winter induce a mesoscale pressure surge arising from intense low level anticyclogenesis west of the intruding mid-latitude trough. This pressure surge is associated with strong low level easterly winds, particularly at ground level where wind speeds of more than 30knots can be observed (a low level 'jet'). With such a wind speed enough instability is usually generated to raise large quantities of dust and keep the dust particles airborne for a considerable length of time. The turbulence associated with this low level jet is probably also affected by local relief, so that katabatic effects south-east of the Tibesti massif and funnelling of the airflow between Tibesti and the Ennedi massifs will increase its turbulence characteristics as in the 'Venturi effect' suggested by DeSouza **et al** (1971).

Kalu (1982) believes that during winter this series of events leads to dust-raising at some location in the Bilma/Faya Largeau source area and about 24 hours after the pressure surge stage a dust front will arrive in northern Nigeria. Certainly the role of a rapidly developing strong pressure gradient in the development of dust haze was recognised by Burns (1961) although his forecasting technique was based on reporting of dust-raising from meteorological stations in the source region. Kalu's approach overcomes the fact that stations are few in the source region.

Further west, over Mauritania and the easten Atlantic the true trade winds prevail; blowing from the north the whole year round they are cool, moist winds of maritime tropical air from the Azores high pressure zone. These are not the winds that generate dust-raising activity in Mauritania, however. The main dust-raising winds are associated with depressions that track from east to west, or that develop in the lee of the Atlas mountains and sweep dust from Mauritania out over the Atlantic by cyclonic gyre. These low pressure systems are covered in more detail below.

Dust-raising winds in the northern Sahara are largely controlled by depressions of various origins (Dubief 1979).

1. Mediterranean depressions skirt the Mediterranean coast of North Africa particularly in winter, their trajectories being related to the position of the Mediterranean Polar Front. Their dust-raising action is particularly effective when they remain stationary for some days in a certain spot on the coast or in the eastern Mediterranean where they affect dust-raising in Libya and Egyptian deserts and transport material to Israel and the Levant (Yaalon & Ganor 1979).

2. Atlantic depressions are also connected to Polar Front depressions and usually only cause dust storms in winter when they approach the Sahara by the south of Morocco.

3. Secondary south Moroccan depressions occur when Atlantic depressions move north-east at the level of the Iberian Peninsula. They develop at the secondary front between dry, tropical Saharan air and the humid tropical air of the Azores high (the Saharan Front defined by Petitjean (1924)). Originally these were designated 'Saharan depressions' by Queney (1936 quoted in Dubief 1979) but as they track in a northeasterly direction from Mauritania and south Morocco across the Maghreb towards Libya and the eastern Mediterranean basin Coudé-Gaussen (1982) prefers to call them 'Mauritano-saharan depressions' (eg Oliva **et al** 1983, for 12 March 1982 by remote sensing).

4. Sudano-saharan depressions arising along the ITF (thus in the Sudan area) are, more simply, tropical

depressions. They develop when the polar or pseudo polar air mass, in conjunction with modifications of the upper atmosphere, meets a vast wave of the Polar Front. Their trajectories stretch across the Sudan from east to west and then move towards the north or north-east describing an arc, and are swept along by the upper westerlies, crossing the Sahara to the Mediterranean coast where they continue more or less linked to the Polar Front depressions. Their most frequent trajectories are from Senegal to Morocco in autumn, from the loop of the Niger River to the Gulf of Sirte in spring and, to a lesser degree, particularly in May and June from the north of Nigeria to Lower Egypt where they are known as **Khamsin** depressions. These **Khamsin** depressions may reach the Levant bringing substantial quantities of dust from Egypt, and also cause the wind in the Sinai and Negev known as **Sharav** . In Libya such depressions usually give rise to the **Ghibli** wind, and in Morocco they cause the **Chergui** and **Sahel** dust winds. The opposing directions of these two Moroccan winds are derived from the relative position of the arriving fronts to the Atlantic secondary south Moroccan and Mauritano-saharan depressions. When the sudano-saharan fronts evolve in the south of south Morocco the dust-raising winds blow from the north or north-east (**Chergui**), but when the fronts pass below south Morocco or north of the region the winds blow from the south-west (**Sahel**).

In late spring and early summer easterly winds prevail across the whole of the southern Sahara at all levels, and a zone of relatively high pressure is found at 500 to 200mb. At this time the Sudano-saharan depressions move only from east to west as a squall line consisting of a group of thunderstorm clouds with a longitudinal extent of 300-500 km. The leading edge of these disturbances is characterised by the gust front with wind speeds sometimes in excess of 30m/s, occasionally generating a dust wall several hundred metres high (Tetzlaff & Peters 1986). When these systems are very active they generate strong sand and dust storms across the whole of the southern Sahara and the Sahel (eg the dust outbreak between 5 and 9 June 1967 monitored by ESSA satellite imagery in Dubief (1979) when two depressions in rapid succession moved from the top of the Niger River loop across Senegal to the east Atlantic, generating hot dry dust-raising east to northeasterly winds associated with the warm sectors of the depressions).

To the east, dust storms in the Sudan are generated by three types of systems as summarised by Morales (1979b). In winter and early spring they are caused by cold air intrusions that penetrate into the Sahara on the rear side of mid-latitude depressions tracking eastward across the Mediterranean. These cold air outbreaks move in a southerly direction preceeded by a more or less marked cold front. When they are connected with a major pressure rise this creates a strong pressure gradient behind the cold front which produces violent northerly dust-raising winds. During the summer season dust storms form in the monsoon air south of the ITF, between about 15°N and 9°N, and are caused by thunderstorm downdrafts. These downdraft storms are either relatively shortlived from isolated thunderstorms (the classic **Haboob**) or are more extensive and longlived in connection with clusters of thunderstorms (convective systems) coupled to synoptic scale areas of convergency and divergency in the atmosphere.

3.3 SAHARAN DUST SOURCES

An annual frequency map of sand and dust storms over large parts of North Africa has been presented by Dubief (1952), while a number of other authors have attempted to identify the various sources of dust that is raised in the Sahara and is often transported over long distances, using remote sensing data. By far the largest quantities are carried out over the tropical North Atlantic, but substantial quantities are also blown to the Mediterranean and over Europe, to the Middle East and over the Arabian Peninsula and the Indian Ocean.

Dubief shows a number of high sand and dust storm frequency regions using data for the period 1925-50, although no visibility limit is mentioned. The Mauritanian coast, from Port Etienne to Nouakchott, shows an average of 80 sand/dust storms a year; and a broad area from this coast eastward as far as the Air mountains has 40 a year. Eighty is also the annual average at the centre of an elongate zone sandwiched between the Anti Atlas and Atlas mountains and the Hamada du Dra in Morocco. This area is largely coincident with the Dra River that drains these mountain massifs. East of this zone 40 is the annual average in localised areas just north of the Grand Erg Occidental, on the Grand Erg Oriental and in a west-east band from south of Tripoli towards the Egyptian border.

For material carried over the Atlantic, the most fundamental distinction to be made is in terms of seasons of activity. McDonald's (1938) haze maps are the first comprehensive indication of a seasonality of transport (fig 3.1), showing that during the summer (June, July, August) maximum dust haze occurs off the West African coast between 15° and 25°N, in spring (March, April, May) between 10° and 20°N and in winter (December, January, February) between 5° and 15°N. Summer also shows a definite maximum over the Mediterranean coast from Tunis to Casablanca, and over the Red Sea and Indian Ocean between 10° and 20°N. These seasonal variations in dust haze over the sea adjacent to the dry lands of North Africa may be due to seasonal variations in source areas.

Most authors seem to agree that an important source is located in the area roughly between Bilma (Niger) and Faya Largeau (Chad), which is the source of the **Harmattan** dust plume, prevalent during the winter months. It is certain that this is the source of dust over the Gulf of Guinea during the winter and this material may be carried across the

Atlantic to reach South America (this will be discussed at some length in section 3.3.2). An alternative source for the winter activity over the North Atlantic has been suggested by Prospero and Carlson (1981) from analysis of satellite imagery, seasonal variations in atmospheric turbidity and the character of the wind field over West Africa. They propose that the principal sources are in the deep Sahara in Mauritania, northern Mali and central Algeria. Bertrand **et al** (1979) also identify the Mauritanian coast as one of two distinctly dusty zones south of the Sahara (the other being Bilma/Faya Largeau) where large quantities of material are exported over the Atlantic.

A major source for the summer transatlantic dust transport towards the Caribbean is thought to be the dune system stretching from the coast of Mauritania and the Western Sahara into central Algeria west of the Hoggar massif (Prospero & Carlson 1981). Mainguet (1983), however, identifies southern Morocco near the Algerian/Mauritanian border as the source for this flow. Mainguet's work is based on the daily analysis of Meteosat images for a nine month period; the other three major Saharan sources she identified were Bilma/Faya Largeau, the northern Sudan and Egypt/Libya. Dust from Egypt and Libya is regularly transported eastwards to the Levant (Yaalon & Ganor 1979).

A complete frequency distribution map of dust storm activity in North Africa has not been presented. This is partly due to the lack of a complete data base, and partly because many countries south of the Sahara have experienced a dramatic increase in dust storm activity dating from the late 1960s. Thus, the calculation of long-term averages would be somewhat misleading. This increase in activity is analysed in chapter 4. It has been possible, however, to construct a map of dust storm seasonality over North Africa which is shown in figure 3.2. This map shows the primary season of dust storms. The season is a span of three consecutive months in which dust storm frequency is highest. In some cases there is a month overlapping two seasons, so that February, for example, is included in both spring and winter, but this was a degree of flexibility considered to be desirable to obtain the broad-scale pattern of seasonality shown. In addition to this map figure 3.3 was prepared showing the frequency of dust storm days in the northern Sahara (Morocco, Algeria, Tunisia, Libya, Egypt).

Figure 3.2 goes some way towards making a division into major North African source regions, and these regions correspond broadly to some of those suggested by the above authors. From the foregoing discussion the sources of major dust storm activity in the Sahara are divided into three broad-scale regions according to the areas of long-range transport of dust. These are sources for Atlantic transport, sources for Mediterranean, European and Middle Eastern transport, and sources for transport to the Arabian Peninsula and Indian Ocean.

3.3.1 SOURCES FOR ATLANTIC TRANSPORT

The study of Saharan dust dates from the first serious scientific observations made by Darwin (1846) and Ehrenberg (1862) whose works were largely based on records taken from ships off the west coast of Africa. In the second half of this century some pioneer works were produced on the mainland of Africa, particularly by Dubief (1952, 1959) who has been concerned with the spatial distribution of sand and dust storms and their relationship to the climate of the region. There have also been investigations in northern Nigeria (eg Burns 1961, Hamilton & Archbold 1945) on the **Harmattan** dust haze and more recently by McTainsh (eg 1980, 1984) in the same area. A major thrust in research since the mid 1960s has been back in the areas worked in by Darwin and Ehrenberg. A large volume of work has been undertaken on the long-range transport of Saharan dust across the Atlantic using aerosol monitoring programmes, turbidity measurements and remote sensing (eg Delany **et al** 1967, Carlson 1972, Prospero & Carlson 1981).

Coudé-Gaussen's works on the continent of Africa have gone some way towards redressing the balance between long-range dust transport and analyses closer to the terrestrial source areas (eg Coudé-Gaussen 1982, 1984a&b). Nevertheless, there is still a surprising lack of detailed knowledge on the exact locations of mainland source areas and their seasonal characteristics of generation for the large quantities of material that are raised and carried away from North Africa across the Atlantic. With this background in mind the present study is timely in its concentration on the dust-raising system of West Africa. The following sections refer to the areas in figures 3.4 and 3.5, showing locations referred to and their geomorphological nature.

3.3.1A MAURITANIA & MALI

Dust storm data were obtained from the Mauritanian Service Météorologique, in tabular form of monthly percentage frequencies of dust storm occurrences for 11 stations over the period 1968-77, and for Nouakchott the monthly number of dust storm days from 1960 to 1984. These data were supplemented by similar tables presented by Hinds and Hoidale (1977) for stations in Mauritania and Mali. Other data were not available from Mali. Although dust storm frequencies rose substantially after the late 1960s (see chapter 4) the seasonality of dust-raising remained constant over the two periods. The dust storm frequency data used in this section for Mauritania are those for the period 1968 to 1977.

Some indication of the dust-generating synoptic systems has been given in section 3.2, but a brief review of the climate of this part of the western Sahara is appropriate to analyse more closely the seasonality of dust storms in Mauritania.

Surface wind circulation in the western Sahara and Sahel is largely determined by interactions between the ITCZ, the Azores high pressure cell and the Saharan high pressure cell. In winter, the Azores high is nearest the African coast and the Saharan high is strongest, while the ITCZ is at its southernmost position over the Gulf of Guinea (Breed **et al** 1979). Outward flow from these pressure systems results in winds from the north-east or east over much of the western Sahara. By June the Saharan high has been replaced by a thermal low in the central Sahara and the Azores high is at its strongest and further from the coast, resulting in more northerly winds at most stations. During July to September, when the ITCZ is at its northernmost position, stations south of about 18°N receive southwesterly monsoon winds that bring the seasonal rainfall to the area. The northward progression of the ITCZ , being the northern boundary of the moist maritime equatorial air mass, is reflected in the seasonal rains being greater, and beginning earlier in the season at the more southern stations in Mauritania (fig 3.6). The effect of this seasonal rainfall is that dust-raising activity is largely confined to the first six months of the year (January to June), with stations north of about 20°N experiencing slightly longer and later active periods, continuing into August.

Mauritania is sharply zoned in terms of wind energy. Fryberger (1980) has analysed Drift Potentials for sand transport and shows that high annual DPs occur along the north coast (Nouadhibou about 2230 vector units) and low DPs in the south near the Senegal River (eg Rosso, 155; Kiffa, 75) with contours of equal drift potential extending roughly north-east to south-west across the country. This pattern, in combination with decreasing rainfall totals as latitude increases might suggest that dust storm frequency would increase northwards from the Senegal River, but this relationship is complicated by local geomorphological situations and wind direction. Below are presented a number of monthly analyses of dust frequency, with their relation to rainfall and DPs for selected stations, and an appraisal of local surficial geology, to explain dust storm frequency and seasonality.

Figure 3.7 shows monthly dust frequency, rainfall and DPs for three stations: Bir Moghrein in the arid north; Atar in the centre of the country; and Boutilimit in the Mauritanian Sahel. The increase in DP and decrease in rainfall with increasing latitude is clearly shown. At Boutilimit the dust season is during January to June, when DPs are highest, and dust frequency is lower from June to October when the rains arrive and DPs decline. At Atar dust-raising reaches two maxima, during February and again in July, corresponding to two peaks in monthly DP. Dust frequency declines from August to November, in response to the rainy months of August and September and low DPs in October and November. The dry salt lake of Sebkha de Chinchane acts as a dust source locally where aeolian erosion of gypsiferous clay alluvial deposits on a large scale has been demonstrated (Grove & Warren 1968). At the northernmost station, Bir Moghrein, DPs are the highest of these three stations, and are particulaly high from January to June/July, while annual rainfall is very low, with the maximum month (September) receiving less than 20mm, being at the northernmost extent of the seasonal movement of the ITCZ. Despite these high wind/low rainfall conditions dust frequency is negligeable due to the lack of appropriate dust-producing geomorphological units in this area that largely comprises rock and gravel desert plains.

Figure 3.8 shows monthly dust storm frequency, rainfall and DPs for two coastal stations. Nouadhibou (Port Etienne) experiences the highest DPs in Mauritania with a definite peak in the period April to July. Despite these high DPs they seem to be poorly related to dust frequency which peaks in December, February, May and July and is at a low in August and September when the small annual rainfall falls.The fairly high dust storm frequencies might be expected to be greater still given the high DPs and low rainfall, but Nouadhibou's location may go some way to explaining this apparent lack of correlation. The town is situated on a spit on the Atlantic coast, and although this sandy spit represents a readily available dust source the prevailing winds at Nouadhibou are northerly, and thus blow largely from the Atlantic. Thus, despite their speed they will be moist and may not always blow across erodible land surfaces. Further down the coast, at Nouakchott, DPs are very much lower, rainfall is greater than at Nouadhibou but still low, but dust-raising is more active. Again dust storms are most common from January to June and decline in the rainy months of July, August and September, but there seems to be little relation to DP. Nouakchott, however, is surrounded by wind erodible geomorphological units: dune fields; desert depressions and former coastal lagoonal deposits such as the Sebkha de Ndramcha.

The synoptic situations that generate dust-raising over Mauritania thus operate north of the ITCZ and are often associated with the effects of depressions tracking across the area or north into the Mediterranean. The effects are to accelerate the prevailing winds in the north-east sector. These synoptic situations are summarised below.

1. Influxes of cold air from mid-latitude depressions that track west to east across the Mediterranean. The influx occurs around the high pressure ridge extending eastwards from the Azores over North Africa, a characteristic pressure distribution in the early months of the year. An example of this type of dust storm is tracked by Noyalet (1978) using Meteosat.

2. Sudano-saharan frontal depressions moving along the ITCZ from west to east in spring. A dust outbreak of this type is shown in Dubief (1979, fig 1.14).

3. Westward-moving dry severe thunderstorms. An example is shown by the Space Shuttle image in plate 3.1, taken in February 1982, showing the 120km dust wall advancing towards

Nouakchott.

4. Strong summer winds generated in the extremely unstable lower layers of air created by the intense solar heating of the land, which typically attains daytime temperatures of 60° to 65°C (Prospero & Carlson 1981). As a consequence a very deep isentropic mixing layer develops, which may extend up to 500mb. The high velocity, gusty winds associated with this atmosphere generate high dust loads which are carried to the top of this mixing layer where they may be transported across the Atlantic as a Saharan Air Layer (Carlson & Prospero 1972). From the study of satellite imagery Prospero and Carlson (1981) point out that dust storms generally appear to develop very rapidly during mid to late morning, which they suggest indicates that the dust generation may be associated with the coupling of the convective layer with an upper level jet.

Thus, by analysing the rainfall, wind regime and geomorphological situation at each station in Mauritania, and by taking the regional climatology into account the dust-raising characteristics of the country can be adequately explained. The situation has been further complicated in Mauritania since the late 1960s due to the combined effects of drought and human activities in causing a widespread loss of vegetation cover, particularly in the southern Sahelian zone. This removal of the savannah vegetation has increased outputs from existing dust sources and uncovered new sources. Important in this respect are the widespread palaeodunefields that spread across the Sahelian zone (Grove & Warren 1968), some of which have been reactivated since the onset of the drought, resulting in a dramatic increase in dust storm activity in Mauritania (see chapter 4).

South of Mauritania, in Senegal, Buroleau (1937) notes that minimum visibility due to dust at Dakar is one to two kilometres, and rarely below one kilometre. The seasonality of dust haze is from November to May. Picq (1936), however, notes two situations that produce visibility less than one kilometre in Senegal, when easterly or southeasterly winds blow from the interior bringing dust or when a **Brume Sèche** occurs, which it frequently does during the dry season (November to April). Monthly mean dust concentrations at Dakar are at a maximum in March and a minimum in July and August (Prospero & Carlson 1981).

Analysis for Mali is not as comprehensive due to a lack of data, but a general indication of the dust storm system can be gleaned. The country can be divided into three broadly west-east zones which reflect three different dust sources each with a distinct seasonality of activity (fig 3.2).

In the south and west of the country, roughly south of a line joining Menaka and Nioro du Sahel, the area is in the path of the **Harmattan** dust plume (see section 3.3.1B below). Rainfall in this zone increases southward from 250mm to more than 1000mm, so that locally raised dust is not common. Dust events are, therefore, largely haze, and are at a maximum from December to March (see fig 3.9 for Bamako and Mopti).

The zone north of this area has a drier climate and includes the inland delta of the Niger River. This zone is little affected by the **Harmattan** dust plume from northern Niger and Chad as blowing dust frequency is low during December to February, it is protected by the Air massif in central Niger. Dust frequency is generally higher from March to May at Timbouctou (fig 3.10) and it seems likely that dust events at this station are derived from desert sources to the north-east. These may be the longitudinal dunes of Azaouad and/or the seasonally dry water courses of the Tilemsi Valley draining the Adrar mountains. The inland delta of the Niger itself has certainly become a significant dust source since the onset of the most recent prolonged drought period as irrigated agriculture has been forced off the land while the river has been at low levels (see chapter 4).

Gao, on the Niger River, is located approximately at the northernmost edge of the area affected by the **Harmattan.** As figure 3.11 shows, blowing dust events are at a maximum in February, when DPs are low (perhaps suggesting a dust haze or **Harmattan**) and again in May and June when material is perhaps derived from more local sources such as the Tilemsi Valley as DPs are at a maximum.

The northernmost station for which data were available is Tessalit, and here it seems a different dust regime again is in operation (fig 3.10). Dust events are at a maximum in June and July, in accordance with the season of activity further north in southern and central Algeria (fig 3.2). Dubief (1979) suggests that dust is raised in this region by Saharan disturbances.

As in all parts of the western Sahel, annual rainfall totals decline from the southernmost of the three identified dust zones in Mali with distance northward. The diurnal variations of dust haze in the three zones seem to confirm this tentative classification (fig 3.12). Mopti experiences a more continuous 24-hour constant dust haze compared to Timbouctou which shows a more definite maximum during daylight hours, suggesting a location closer to the dust source.

The southern fringes of the Sahara have undergone considerable climatic change in the later stages of the Quaternary, and it seems likely that some dust sources in Mali are related to former more pluvial periods. In the dunes of the Erg In Koussamene, 200km NNE of Gao, for example, Coudé-Gaussen **et al** (1983) have shown that extensive winnowing has occurred of clays from a dry Holocene palaeolake. With distance downwind from this dry lake bed the colour of the dune sands changes from grey at the lake bed to yellow due to gradual quartz enrichment as clayey aggregates in the dunes are fragmented and winnowed out by the mechanical action of quartz grains.

3.3.1B NIGER, CHAD & NIGERIA

Dust storm data for Niger and Chad were impossible to obtain, but data on the number of days with thick dust haze (visibility $< 1000m$) recorded at eight stations in Nigeria were supplied by the Nigerian Meteorological Department for the period 1955 to 1979. Some stations are missing data for a few years in the late 1960s due to the civil war. Otherwise, some data for Niger are presented in tabular form by Hinds and Hoidale (1977).

Dust events occurring in these countries, and in many others to the south-west, are very largely associated with the **Harmattan** dust wind, which has its source in the Chad Basin (fig 3.5). Thus, dust-raising occurs in Niger and Chad, but in Nigeria very few dust storms occur, dust events in that country being in the form of **Harmattan** haze. One of the earliest accounts of the **Harmattan** is by Dobson (1781) who noted the fine material that was

> ...deposited on the grass, the leaves of the trees, and even on the skin of the negroes, so as to make them appear whitish.
> (Dobson 1781, p 47).

The **Harmattan** season is from October to May, and the periodic emissions of large quantities of material from the former lake bed and subsequent transport as a thick dust haze to the south-west is a distinctive feature of the climate of West Africa, and is known as **Brume Sèche** in the French speaking areas. Gouault (1937) suggested that the most frequent area of dust-raising is between the Ennedi and Tibesti massifs, including the depression known as Bahr el Ghazal, while Hamilton and Archbold (1945) suggested the Bodélé Depression as a more specific dust source. Evidence presented by Bertrand **et al** (1979) seems to confirm that dust is raised in the broad area between Faya Largeau and Ndjamena, while much material is also entrained to the west in the region of Bilma (Niger) in the Ténéré sand desert where interdune depressions are occupied by many palaeolake deposits (Hall **et al** 1971). The material of the Bilma/Faya Largeau plain is alluvial on the surface and the subsoil consists of light fine and loose clay particles also of alluvial origin, which appear to be derived from the Tibesti mountains transported by intermittent streams from these highlands (Wilson 1971).

Geochemical comparison of **Harmattan** dust collected in northern Nigeria with lake deposits from Borkou Lake show a broad similarity (McTainsh & Walker 1982), and further support for a lake deposit source is provided by diatoms found in dust from Ibadan (Hamilton & Archbold 1945). McTainsh and Walker identified two likely feeder zones of deposits into the Chad Basin. The Chari and Logone Rivers drain large catchments of humid tropical rainforest to flow into Lake Chad, and also the Yobi River from northern

Nigeria. Confirmation of the Bahr el Ghazal as a potential dust source is provided when it is realised that this depression has carried overflow from Lake Chad to the Bodélé Depression at times of high lake levels. McTainsh and Walker suggest that this intermittent fluvio-lacustrine transport process may have been competent to supply the clay fraction of Bodélé Depression lake deposits, but equally the Bahr el Ghazal itself may be a dust source as Gouault suggested.

The other feeder zone put forward by McTainsh and Walker is again the highlands of Tibesti and Ennedi, and the high quartz content of **Harmattan** dust may be derived from the Erg du Djourab which occupies the Bodélé Depression. The windflow analysis of the Sahara presented by Wilson (1971) further suggests that the Bilma/Faya Largeau source area is constantly replenished with new dust deposits from the east, with the formation of silt-sized quartz grains occurring during aeolian transport. McTainsh and Walker feel that these dunes may represent an important entrainment mechanism for fine particle emission at impact threshold velocities. More than one source region for **Harmattan** dust is proposed by Whalley and Smith (1981) as a possible explantion for the wide range of grain roundness characteristics identified in dust samples taken in northern Nigeria.

The synoptic meteorology of the emission phase from the Bilma/Faya Largeau area has been studied most recently by Kalu (1979) who has shown that the high wind speeds over this area are created by the development of an induced mesoscale phenomenon of pressure surge arising from intense low-level anticyclogenesis west of an intruding mid-latitude trough. This has been known to be associated with large-scale intrusion of 'cold outbreaks' from the middle latitudes into the tropical atmosphere in winter in the form of upper troughs, moving usually eastward along the Mediterranean latitudes. Kalu proposes a critical wind velocity of 30knots (about 14 m/s) for dust to be raised in the area, although gentler winds may be sufficient. Observations at Faya Largeau show a threshold wind speed of 6 to 8m/s (Gouault 1937). The direction is almost invariably from the north-east, and the turbulence characteristics of these strong winds are probably increased by funnelling effects between the Tibesti and Ennedi massifs and katabatic effects off the Tibesti (Kalu 1982).

At Faya Largeau dust is most commonly raised between 0500 and 1000 local time and dust storms usually cease between 1400 and 1800, although they may continue for two or three days (Gouault 1937). Figure 3.13 shows a similar peak in activity in the early morning at Bilma. Visibility at Faya Largeau during a storm is generally between 200m and 500m, but may be as low as one metre.

Dust raised in the Bilma/Faya Largeau source area is transported in a WSW direction across the savannah regions of West Africa and out over the Gulf of Guinea and Atlantic. Bertrand **et al** (1979) suggest that the limits of the area affected by the **Harmattan** dust plume are marked to the north at a line approximately linking the Air mountains and Mopti (Mali), and to the east by a line joining the Ennedi mountains to Douala (Cameroon). This northern limit corresponds well to the seasonal divide in dust storm activity shown on figure 3.2. To the east Pélassy (1984) has detected **Harmattan** aerosols at Yaounde. Further west **Harmattan** dust haze may be experienced as far north as Dakar where dust haze is at a maximum from November to May (Buroleau 1937), although this station is probably more affected by sources in southern Mauritania and Mali. Further discussion of long-range transport of **Harmattan** haze is presented below (see section 3.3.2).

3.3.1C MOROCCO, ALGERIA & WESTERN SAHARA

The wide expanse of desert areas in Morocco south and east of the Atlas mountains and the Saharan areas of south and west Algeria appear from satellite imagery to contain important sources for dust generation. The frequency distribution of dust storms in North Africa (fig 3.3) shows an area of high frequency across central Algeria and into northern Libya. This region is almost exclusively one of spring time maximum (fig 3.2), and the north-west/south-east dividing line between this area and that of summer time maximum in dust storm activity can be regarded as a good approximate boundary between sources that lead to Atlantic dust transport and those from which transport occurs to the

Mediterranean, Europe and the Middle East. This chapter, therefore, covers the south and west of Algeria which has the summer maximum, the north and east of the country will be covered in section 3.3.3A.

It is important to note at this stage that this frequency map may not show all regions in Morocco and Algeria that are important dust sources. Recent data were unavailable for large expanses of the Sahara, and it is in some of these areas that monitoring of satellite imagery suggests the major sources of dust storm activity are located. Thus, there are no recent terrestrial dust storm data to confirm the dune systems of south-west central Algeria and the Western Sahara as major dust sources as proposed by Prospero and Carlson (1981). The only exception is the lone Algerian station of Tindouf, situated in an area of seasonal floodplains and sebkhas surrounded by the stony Hamada du Dra, which averages two dust storm days a year. Dubief's (1952) map, however, although based on few data points in these dune areas, shows them to average 20 sand and dust storms a year. Some work has been done on the aerosols in a storm at Tanezrouft (about 500km south of Adrar) in southern Algeria by Coudé-Gaussen (1981), who found the dust to be largely composed of quartz (66%) and muscovite (24%) from local sources.

In the same way, Mainguet's (1983) identification of southern Morocco as a major dust source cannot be confirmed with recent terrestrial meteorological data. The most likely source of dust storm activity in this area is the floodplain of the Dra River that drains from the Anti Atlas mountains south-east of Agadir, and indeed Dubief's (1952) high frequency region is coincident with this floodplain, where annual average rainfall is less than 100mm. The high resolution of the Space Shuttle photograph shown in plate 3.2, taken in October 1984, pinpoints a large-scale dust event in this location, with dust being raised along a 200km stretch of the floodplain of the Dra River and immediately to the south in a 150km long basin, divided from the floodplain by a ridge, the Jbel Ouarkziz, west of the Hamada du Dra. The total area covered by this event is about 18,000km^2. The upper reaches of the Dra River is the region in which the southwesterly **Sahel** and the northeasterly **Chergui** dust winds blow (Mainguet 1980).

Data were available for Ouazazate, situated on the headwaters of the Dra, south-east of the High Atlas in an area of Miocene deposits. An annual average of 4.1 dust storm days are recorded with annual average rainfall of 119mm. Activity is at a maximum from January to August when rainfall is low (fig 3.14), the rainier months of September to December showing a distinctly lower dust storm frequency. The annual variation in frequency of dust storm days from 1951 to 1980 shows a distinct peak in activity from 1970 to 1975. Although rainfall was below average in 1971, 1973, 1974 and 1975, other periods of below average rainfall in the 1960s produced less remarkable peaks in dust storm activity. This either reflects the generally weak relationship between rainfall and dust-raising in this area or the arrival in the early 1970s at some sort of wind erosion threshold after intermittent below average rainfall in the 1960s.

North of the High Atlas, Marrakech is the only other Moroccan station for which recent data were available where significant dust-raising occurs. Here the 247mm annual average rainfall occurs very largely from October to May, and dust storm activity is at a peak from July to September, during the dry months (fig 3.15), the annual average number of days being 5.0. The interannual variation (1951-80) shows a distinct peak from 1952 to 1955, reflecting low rainfall in the period 1951 to 1955. A minor peak in dust storm activity in 1961/62 seems to be related to rainfall of about half the annual average in 1961, and a decreasing trend in rainfall from 1972 to 1980 is mirrored to an extent by a minor rise in dust storms over the same period after generally higher rainfall and minimal dust storm activity from 1963 to 1971.

Other stations north of the Atlas show little dust storm activity (fig 3.3), although minor dust sources may occur, such as in the coastal dunes at Cap Sim, north of Agadir, where Coudé-Gaussen **et al** (1982) show that calcareous dust is produced from the disintegration of shells by saltating quartz grains blown by the prevailing NNE/SSW trade winds.

Very little data or information were available for the Western Sahara (formerly Spanish Sahara). Hinds and Hoidale (1977) show that Dakhla (Villa Cisneros) on the coast recorded an average four dust storms a year during 1950 to

1967, and 49 events of blowing dust (visibility <11km). The season of maximum activity is from February to August, with a just discernable summer maximum. The country is very largely arid; potential dust sources include the dune expanses particularly near the coast, the expansive drainage network of the mainly dry Saguia el Hamra in the north and some sebkhas. The dominant winds are the prevailing northeasterly trades. The **Irifi** is a dust wind that blows from the south-east, however, and is similar in meteorological origin to the **Sahel, Khamsin** and other North African dust winds caused by tracking depressions across the area or to the north. Morales (1946) notes that the **Irifi** can reduce visibility on occasions to 10 metres, and the dust it carries causes the conjunctivitis that is common among the nomads of the country. The wind can last for several days and may blow at speeds of up to 17m/s.

3.3.2 TRANSPORT ACROSS THE ATLANTIC

It has been noted above that the transport of thick dust haze off the West African coast has been observed by seamen since classical times, and that McDonald (1938) has shown a distinct seasonal shift in the latitude of maximum dust export. The intercontinental scale of some of these dust events has only been fully appreciated with the use of satellite imagery after various researchers found evidence of North African dust in aerosols as far away as Florida (Junge

1956) and Barbados (Delany **et al** 1967), and more recently at Cayenne, French Guiana (Prospero **et al** 1981). Analysis of satellite imagery over long periods indicates that the dust transport zone off West Africa is between 10° and 25°N (Prospero & Carlson 1972) or further north to 27°N (Mainguet **et al** 1980). This northern limit should, however, be extended to 29°N to incorporate the southern Moroccan source identified by Mainguet (1983), while north of 30°N dust loadings over the Atlantic are much lower due to the protection afforded by the Atlas mountains (Chester & Johnson 1971a). To the south, McDonald's maps show an area of dust haze off the coast of the Gulf of Guinea at about 5°N.

Although McDonald divides the seasonality of maximum dust transport into three, with a southward latitudinal shift from summer, through spring to winter, most texts that deal with transatlantic dust transport identify just two main distinct seasons of transport with characteristic paths as determined by satellite imagery and aerosol monitoring prgrammes. These are summer dust transported to the Caribbean and winter dust transported to the north-east facing coast of South America. For these characteristic seasons a variety of sources have been suggested, as outlined in section 3.3.1 above.

It should be noted, however, that these are areas of most frequent transport during a particular season; at any time during summer, spring and winter thick dust haze may be present off the West African coast across the whole north-south extent of the dust transport zone from about 10°N to about 29°N. For example, in February 1898 ships off the West Arican coast from 9°N to 29°N reported thick dust haze (Scott 1900), and in June 1975 Mainguet **et al** (1980) monitored a dust outbreak that passed between the Canary Islands (28°N) and Freetown in Sierra Leone (9°N). Furthermore, the transport paths and transatlantic destinations of dust, although characteristic of certain seasons, should not be considered as exclusive. Prospero **et al** (1981) indicate that a dust outbreak that emerged from the West African coast on 20-22 March 1978 transported material to both Cayenne (maximum dust concentration = 130 microgrammes/m^3 on 25 March) and Barbados (maximum dust concentration = 107 microgrammes/m^3). These facts complicate attempts to discern whether summer transport is largely from different sources to winter transport and whether spring transport is from different sources again or from a combination of summer and winter sources.

Much of the research into Saharan transatlantic dust transport has been concerned with the summer dust to the Caribbean. An aerosol monitoring station set up by Parkin **et al** (1967) has been taken over by Prospero (1968) and has been operational since (Prospero & Nees 1977, 1984) stimulating considerable interest in Saharan dust. This research has produced data on typical dust mineralogies (Chester & Johnson 1971a&b, Glaccum & Prospero 1980), particle-size characteristics (Schütz & Jaenicke 1974, Emery **et al** 1974) and organic constituents (Lepple & Brine 1976), and models of Saharan dust transport (Prospero & Carlson 1972 & 1981, Jaenicke & Schütz 1978). Studies of winter dust, by comparison, are relatively few and far between, which is surprising given the quantity of research from mainland Africa on the winter **Harmattan** dust flow (see section 3.3.1B).

It is an attractive proposition to relate the **Harmattan** system to the winter transatlantic dust flow, and McTainsh (1986) has made a tentative start to confirming this relationship. The most immediate evidence of a correlation is from the haze frequency map over the tropical Atlantic (McDonald 1938) which matches up with dust plume direction evidence in West Africa. Further support is provided by analysis of dust mineralogy. Paquet **et al** (1984), reviewing the studies of dust over the Atlantic off the West African coast, show that the typical clay mineralogies of dust over the Atlantic grade from a relatively high illite/low kaolinite concentration at the northernmost end of the West African dust export zone (about 30^{o}N) to reduced illite and increasing kaolinite and smectite concentrations towards the equator, these more southerly dusts being predominantly kaolinite dominant. This latitudinal variation is also confirmed by the studies of the dust component of Atlantic Ocean sediments (Biscaye 1965, Delany **et al** 1967, Folger 1970, Goldberg 1971).

The clay mineralogy of **Harmattan** dust deposited in

Nigeria is also kaolinite dominant (50-60%), with lesser smectite (20-30%) and minor illite (5-10%) (Pickering 1980) which correlates with the winter dust over the Atantic. Comparison of mineralogies of **Harmattan** and Cayenne dust is complicated by minerological differences in settling velocities which become increasingly important with longer range transport, but the beige colour of dust collected at Cayenne is comparable to the Munsell colour notation of Kano dusts (McTainsh 1986).

McTainsh also compares mean monthly dust concentration in surface air at Cayenne (Prospero **et al** 1981) with mean monthly dust deposition at Kano from October 1978 to April 1979, showing a fairly close relationship. A similar comparison, for the total aerosol collection period at Cayenne (December 1977 to December 1979) with monthly number of days with thick dust haze at two stations in northern Nigeria is shown in figure 3.16. In a previous paper McTainsh (1980) has shown a clear relationship between monthly number of days with thick dust haze and the monthly deposition at Kano. In figure 3.16 the relationship between Nigerian dust and dust loads at Cayenne is not so good. For the winter of 1978/79 the relationship is close to McTainsh's, excepting February 1979 when just two days of thick dust haze were recorded at Kano and none at Maiduguri, with high dust concentrations at Cayenne. The lack of correlation for January 1979 McTainsh suggests reflects the time-lag between the onset of dusty conditions at Kano and Cayenne 6500km downwind. It is during the period from December 1977 to April/May 1978 that the relationship seems to be weak however. December 1977 saw seven days with thick dust haze at Kano and nine at Maiduguri, the peak month of activity for that winter, and yet this month shows particularly low concentrations at Cayenne. January 1978 shows maintained high numbers of dust haze days at Maiduguri (9) and a decrease at Kano (3), yet concentrations at Cayenne are still very low though slightly higher than in December. By February 1978 dust activity at both stations in Nigeria is in decline, but at Cayenne the concentation shows a slight increase over January. In March and April, however, Cayenne recorded greatly increased dust loads (about five times the January concentration) but during these two months combined Kano recorded no days with thick dust haze and Maiduguri just one.

The data for two full years monitoring at Cayenne show that the month of maximum aerosol concentration is March, with a seasonal maximum from about February to May and minimum in December and January. It has been noted above that the **Harmattan** blows from October to May with a maximum from December to February. Although comparison of the data from Nigeria and Cayenne does not preclude the possibility of **Harmattan** dust reaching South America, from evidence over the winter of 1978/79, it does suggest that the Chad Basin, if it does contribute to Cayenne aerosols, is not the only source, and is probably not the most important one if the relative seasonalities of maximum activity are indicative.

The seasonal maximum at Cayenne is much more closely aligned with the seasonal maximum in dust storm activity in southern Mauritania (see section 3.3.1A). Indeed, an aerosol monitoring programme at Dakar, in operation since 1974, shows that maximum aerosol concentrations are always recorded in March (Prospero & Carlson 1981) and Prospero and Carlson also indicate that the sources of winter dust (Cayenne) may be in 'the deep Sahara in Mauritania, northern Mali and central Algeria' (1981, p682). Further, they suggest that **Harmattan** dust must be very largely washed out in the regions of high precipitation in the ITCZ over the Gulf of Guinea and tropical Atlantic.

Although these arguments do not disprove the possibility of **Harmattan** dust reaching Cayenne, they suggest that **Harmattan** dust is not the only source of aerosols at Cayenne, and the possibility still remains that little, if any, material from the Chad Basin is transported as far as Cayenne. Although high dust loads characterised the period from February to April 1979 in both northern Nigeria and Cayenne, as shown by McTainsh (1986) and confirmed here, these were also months of high dust storm activity in southern Mauritania as shown by the data for Nouakchott in figure 3.16. This much can be stated usefully from the Nouakchott data, but it is unwise to attempt further correlation between dust storm frequency on land in Africa and aerosol concentration across the Atlantic. This is because the nature of the dust storm data makes it impossible to distinguish between events on a scale, both horizontally and vertically, large enough to lead to transatlantic transport, and those of a more local nature. The dust haze data for northern Nigeria, however, represent events where material is being transported over long distance in higher level winds (Kalu 1982).

Thus, it seems that transatlantic transport towards Cayenne, commonly referred to as 'winter dust' should be termed 'spring dust', and the evidence is slightly in favour of sources located in the dunefields of Mauritania and northern Mali, with a spring seasonal maximum.

The 'summer dust', which arrives in Barbados, where monthly maximum concentrations are in June, July and August, seems likely to originate in south-central Algeria (as analysis of SMS-1 satellite data during GATE suggested), but also from the source identified in section 3.3.1C which is the alluvial plains of the Dra River in southern Morocco, in both areas dust storm activity is at a maximum in summer. Comparison of clay mineralogies of dust sampled in the atmosphere over northern Algeria, from rock sources similar to those in southern Morocco, with dust off the coast in this region show a broad similarity (Paquet **et al** 1984). The material is high in illite, with chlorite and palygorskite also in high concentrations. This material is also thought to be transported occasionally to the British Isles, and for some events the sources have been confirmed with satellite imagery (eg Bain & Tait 1977).

3.3.3 SOURCES FOR TRANSPORT TO THE MEDITERRANEAN, EUROPE & THE MIDDLE EAST

Substantial transport of dust also occurs from sources in North Africa northwards to be deposited over the Mediterranean Sea and Europe, and eastwards to the Levant and the Middle East. Loÿe-Pilot **et al** (1986) estimate the Saharan dust input to the north-west Mediterranean Basin to be 3.9 million tonnes a year, while Yaalon and Ganor (1979) have calculated that about 25 million tonnes of material reaches the East Mediterranean Basin each year, most settling in the Mediterranean Sea. Computed trajectories of dust particles show that sources for eastern transport are partly local (Sinai and Negev) and partly carried from the Libyan and Egyptian deserts.

Many instances of dust falls, red or 'blood' rains and coloured snow have been documented over Europe and have been tabulated by Goudie (1978) and Mainguet (1983). Such material often reaches the southern European countries, and on occasion may travel as far as Sweden (eg 26 February, 1896), the Ukraine (April, 1960) and the Caucasus (March 1964). Sources for such transport, in the cases where they have been identified, seem to be largely in North Africa, in areas of Algeria, Tunisia and Libya. Bucher and Lucas (1975) estimate deposition rates over the Pyrenees to be 18 to 23 mm/1000yr, and Loÿe-Pilot **et al** (1986) calculate a rate of 10 mm/1000yr over Corsica. Dubief (1952) shows two areas of more than 40 sand/dust storms in northern Algeria, one located just north of the Grand Erg Occidental, the other on the Grand Erg Oriental.

3.3.3A NORTHERN & EASTERN ALGERIA, TUNISIA, LIBYA & EGYPT

Figure 3.3 shows the frequency distribution of dust storm days based on available data for periods during the 1950s, 1960s and 1970s. This illustrates an area of high dust storm frequency across central Algeria and into northern Libya. The highest frequency is 28 days a year at In Amenas, situated on the northeastern fringe of the Erg Issouane. This high frequency area is one in which rainfall is minimal and highly variable (eg Timimoum, 14mm a year; In Salah, 9mm; In Amenas, 26mm; Ghadames, 27mm). The maximum season across this area is in the spring months, while nearer the Mediterranean coast the season of maximum activity is in places bimodal, with a maximum in spring and autumn, and also in places with a maximum in autumn (fig 3.2).

The sources of dust in these parts of North Africa seem to be related to ergs, with local sources including sebkhas and other desert depressions. In northern Algeria Paquet **et al** (1984) show that atmospheric dust is high in illite and chlorite and is principally derived from sebkha surfaces on the high plains including Chott Chergui south of Oran. Other saline depressions include the Sebkha de Ben Ziane in the

Oran region where the activity of current processes producing fine particles is evidenced by lee side lunette dunes (Boulaine 1954), and the Quaternary evaporites of the Sebkha el Melah north-west of Adrar, on the southwestern fringe of the Grand Erg Occidental (Paquet **et al** 1984). Further east, the bajada surface south of the Aurès mountains in the Biskra region, aggraded during the late Pleistocene by braided streams and mudflows, is estimated by radiocarbon dating to have been deflated by one to four metres in 2000 years (Williams 1970).

The meteorological systems that generate dust storms in North Africa are largely related to the movement of low pressure cells (see section 3.2). It has been noted that spring is the time of maximum dust-raising over much of North Africa, and this is the season of the hot, dusty desert winds that blow from the south or south-east. These winds blow in front of the eastward moving depressions, that are often frontal, which usually take a preferred track near the North African coast over the desert or the southern Mediterranean. **Scirocco** is the name used for these winds as they blow over the Mediterranean Sea and its coast, and they are known by a variety of other names locally: **Chehili** or **Chili** in the southern desert areas of Algeria and Tunisia; **Guebli** in the northern departments of Tunisia and of Algeria where they may be exacerbated by a foehn-type effect as the wind descends the northern coastal plain; **Ghibli** in Tripolitania, and **Khamsin** in Egypt. The effects of these winds can be severe for local populations. Their desiccating nature can destroy harvests in a few hours, and the sudden onset of the winds may cause fatal cases of heat stroke and brings on attacks of 'Cafard' or temporary madness in Algeria (Naval Intelligence Division 1943).

These winds are characterised by very low visibility, Banoub (1970) suggests less than 200m is typical in Egypt, intense temperatures and very low humidities. They may blow for two or three days as the low pressure system moves along its track (Lunson 1950), but rarely last for more than 12 hours at any one place (Banoub 1970). As the depression passes an area, dust storms cease as the wind veers to the north-west and decreases in speed, although these northwesterly winds may themselves cause relatively minor dust storms (eg Flower 1936a, Durward 1936). Banoub (1970) notes that the **Khamsin** continental tropical air mass usually has an almost dry adiabatic lapse rate and is convectively unstable with strong convection up to great heights, especially during the afternoon. In the vicinity of the depression and its cold front these conditions may produce cumulonimbus clouds with cloud bases usually above 3000m, and precipitation from these clouds frequently evaporates before reaching the ground. These dry thunderstorms may cause severe local dust storms from their downdraft winds. Such localised thunderstorm downdraft storms, associated with cold frontal passage, also occur in northern Libya (Johnston 1952).

Dustfalls over Cairo, Helwan and the northwestern coast of Egypt have been monitored by Kolkila (1975), who related the mineralogy of material to source areas which he identified. These were the sands and gravels of the Gebel Ahmar, the upper Eocene clayey gypsiferous deposits of the Gebel El-Mokattam, Maadi marls, the Helwan sandy shales and areas in the northwestern coastal hills of Egypt. Dust deposition at all sites in north Egypt was at a maximum from March to May, with the maximum month being April. These are the months of the **Khamsin** . Dust deposition rates at Cairo, presented from a number of authors, range from 114 to 410 tonnes/km^2/yr.

The relationship of monthly dust storm frequency to Drift Potential and rainfall for two Algerian stations is shown in figure 3.17. Both In Salah and Ouargla are situated in desert areas of simple sand sheets and streaks (Breed **et al** 1979); In Salah to the east of the Erg Chech, just south of the Plateau du Tademait, and Ouargla north-west of the Grand Erg Oriental. In Salah records minimal rainfall (9mm a year), and Ouargla 53mm a year. Highest dust storm frequencies are recorded at In Salah in spring (March to May), and March is a month of very high DP. Fairly high dust storm frequency is recorded in the summer months, with a minor peak in DP in July. From August to December low dust storm frequency seems to be a reflection of low DPs. The significant dust during the summer months reflects In Salah's position on the approximate boundary of seasonal maxima shown in figure 3.2, south of which are areas with a summer maximum in dust-raising activity. Ouargla, by contrast, is too far north to reflect any summer peak in DP, although a secondary peak in dust storm activity is discernable in this season. The main dust-raising season, however, is also in the spring months, corresponding to high DPs.

Figure 3.18 shows a similar monthly comparison for Sebha, Libya. The station records 11.6 dust storm days a year and is situated on the eastern fringe of the Sahra Awbari sand sea of northwestern Libya, to the east is an area of rock or gravel desert plain (Breed **et al** 1979). Annual average rainfall is less than 10mm. Monthly dust storm frequency is fairly closely related to monthly DPs; dust storm activity is at a maximum from February to June when DPs are high, for the rest of the year dust storm frequency is low and DPs are low, except for a minor peak in October.

The role of large-scale human activities on the dust storm system in North Africa during the war years has been noted by Oliver (1945), who documented a significant increase in dust-raising activity in Maryut, Egypt during the North African campaign of the early 1940s. Before the second World War three to four dust storms a year was the average level of activity, but the widespread disruption to desert surfaces by tank and troop movements and the other activities of war resulted in a rapid rise from 1939, the annual number of storms in 1941/42 being 51. Oliver also notes the desert's capacity for recovery from this widespread disruption, however, so that by 1944/45 the annual number of dust storms had returned to four.

3.3.4 TRANSPORT TO THE MEDITERRANEAN, EUROPE & THE MIDDLE EAST

The transport of dust from North African sources across the Mediterranean to reach southern Europe occurs regularly, perhaps at least once a year. Bucher **et al** (1983) have analysed records on dustfalls over Europe between 1546 and 1980 identifying 98 cases during this period, with nearly 30% occurring in March and a secondary maximum in activity in the autumn months, although they suggest that this number of events is likely to be an underestimate. For the period 1926-34, Combier **et al** (1937) note 11 cases of 'pluie de boue' (mud rain) over southern France and Italy, also largely occurring in the spring, and Prodi and Fea (1978) monitored dust falls over the Alps from 1968 to 1978, recording 34 events.

The sources of this material vary, but Combier **et al** show Algeria to be the most common source for southern France and Italy, with Morocco and Tunisia the other source regions occasionally involved. Algerian sources are the most oft-quoted in Mainguet's (1983) table of dustfalls over all Europe, and the impression gained from this list for transport to the Iberian peninsula and France is that northwestern Algeria is the common source (eg Champollion 1965, Bucher **et al** 1983). Sources in Libya also generate dust that is transported to Malta (Ehrenburg 1849), the Tyrrhenian Sea (Chester **et al** 1984) and the Italian peninsula (Prodi & Fea 1979) as well as further north to the Alps (Ehrenburg 1849). Dust from the Egyptian deserts caused 26 cases of significant visibility deterioration in 10 years from 1953 at Cyprus, in one instance visibility was as low as 1500 yards (Gordon & Murray 1964). Dust may also reach the Greek Islands and the Pelopennese from Sudan (plate 2.7).

Longer range transport seems to occur less frequently. Zamorskii (1964) documents an event in March 1962 when material from Tunisia and Libya reached the central Caucasus after three or four days transport. The event of March 1901 deposited dust over much of Europe as far north as the Baltic. In this case the dust cloud covered about 800,000km^2, and deposited an estimated 1.8 x 10^6 tonnes on Europe, 3 x 10^6 over the Mediterranean and 0.5 x 10^6 tonnes over Tunisia and Libya (Hellman & Meinardus quoted in Nalivkin 1983).

The documented cases in which Saharan dust reached the British Isles this century are shown in table 3.1. Ten dust falls were recorded in the nine years from 1977 to 1985, while just four events were recorded in the first 77 years of the century. Whether this apparently dramatic increase in dust events reaching the British Isles is real or simply a reflection of increased interest and awareness of such events by observers is open to debate. Indeed, despite this current interest, not all dustfalls over the British Isles are reported in the literature. The April 1984 event over south-west England and south Wales, for example, has not been commented on in any scientific paper. A survey of Monthly Weather Reports submitted to the Meteorological Office at Bracknell revealed the following comments under Miscellaneous Phenomena for the 22nd:

Cwmbargoed - Cars with traces of red dust.
Bristol Weather Centre - Brown dust in rain.
Exton - Pink hail.
Downside Abbey - Red rain.
Hartland Point - Sandy coloured rain.

The sources of material for transport to Britain, where they have been identified, seem to be largely in Algeria. Areas that appear more than once for events noted in table 3.1 are south and west of the Hoggar mountains and in north-west Algeria in the lee of the Atlas mountains - In Salah and Bechar. For the 6th March 1977 fall over Northern Ireland and the Isle of Skye, Tullett (1978) shows two possible trajectories, one indicating a source just to the south-west of the Hoggar and one further to the south, in northern Niger. Kalu (1982) shows that during the period 28 February to 5 March 1977 there was continuous dust-raising from the Bilma/Faya Largeau source, creating a thick **Harmattan** dust haze over much of West Africa (Kalu 1982, fig 5.3) which continued for three weeks after the 28th. After 3 March, there was also widespread development of dust-raising across the Sahel as far west as Timbouctou and Segou. If the material reaching Northern Ireland and Skye took six days to arrive from the Sahara, therefore, as Tullett suggests, it may be that this material originated further east than Tullett calculates, in the Chad Basin.

Wheeler (1986) has noted that many of the dust falls reaching Britain are transported in the middle troposphere in conjunction with a high pressure cell located over western Europe, providing an arc of clockwise circulation from north-west Africa across the eastern Atlantic, to approach the British Isles from the south-west. Wheeler notes that the following events were associated with such a pressure distribution: 1903, 1968, 1977, May 1979, November 1979, January 1981 and September 1983. Analysis of surface synoptic charts for the other dustfalls shown in table 3.1 shows that most of those cases he did not include were also associated with anticyclones over western Europe. These are the events of 1902, 1982, January 1983, April 1984 and April 1985. The event of November 1984 was unusual in that it was associated with a quasi-stationary low pressure system over the Bay of Biscaye (Wheeler 1985, 1986).

The contribution of wind-blown dust from North Africa to present sedimentation in the Mediterranean has been indicated by several studies (eg Tomadin 1974, Tomadin **et al** 1984). Over the longer term dust accumulation has been indicated by deep-sea coring (eg Eriksson 1979), analysis of peridesert loess in Tunisia (eg Coudé-Gaussen 1984b) and the aeolian contribution to soils of the Mediterranean area, with an emphasis on the Levant by Yaalon and Ganor (1973) and Yaalon and Dan (1974). More recently Rapp (1983) has looked at the aeolian contribution to terra rossas in the Iberian Peninsula and southern France.

3.3.5 SOURCES FOR TRANSPORT TO THE ARABIAN PENINSULA

Dust raised in north-east Africa is transported across the Red Sea to the Arabian Peninsula. This material is derived from lower Egypt, Sudan, Ethiopia and the Horn of Africa. Of the dust flows emanating from the Sahara this is the least studied of all, although the occurrence of dust storms in the Sudan has been widely investigated, and this area is covered in detail below using past studies. Dust storms in Ethiopia and the Horn Africa are less well documented, and are covered in detail in section 3.4.

3.3.5A SUDAN

North-east Sudan is one of the five world areas identified by Idso (1976a) where dust storms are especially intense, and this impression is confirmed by Goudie (1983, fig 3) who shows that highest dust storm frequencies are recorded at Dongola (20.2 days a year), Khartoum (18.4), Kassala (11.0) and Karima (10.6). Although Goudie's map shows that dust storm frequency decreases towards the southern

Sahelian zone, where average annual rainfall is greater than in the arid north of the country, the correlation between dust storms and annual average rainfall for 12 stations in the Sudan was very poor. This lack of correlation becomes clearer when it is realised that many of the dust storms in the Sudan are generated south of the ITCZ, in the moist air masses, and that some dust-raising typically occurs during the wettest months that are June to September.

Figures for annual dust storm frequency in the Sudan, especially those in the Sahelian areas, should be treated with some caution, however, as the occurrence of dust-raising has significantly increased since the late 1960s at a time of widespread drought and desertification. This situation is covered in detail in chapter 4, in the present section I shall outline the seasonality of dust storms in Sudan by relating them to the generating meteorological systems.

Several authors have described the different sorts of dust storms occurring in the Sudan (Farquharson 1937, Freeman 1952, Morales 1979b), and these can be summarised into three main types. The instability type is associated with the development of isolated cumulonimbus clouds south of the ITF during the advancing monsoon or early monsoon period (May, June and July). This is the **Haboob** (from the Arabic word 'habb' meaning 'to blow', Sutton 1925), caused by the thunderstorm downdraft. They are mesoscale summer storms that generally occur between about 12^{o}N and 15^{o}N, south of this zone the ground is generally too wet and north of it the monsoon air is usually too shallow to allow cumulonimbus clouds to develop (Morales 1979b). Freeman (1952) gives a description of a well-developed downdraft dust storm, from when it begins to form on the edge of a thunderstorm cloud.

> At first dust is blown upwards from the ground and resembles smoke from a line of fires on a windy day, with dense columns swirling upwards. As the haboob develops the columns merge to form a billowing wall of dust. The leading edge bulges forward to form a nose and the upper surface slopes back as in well developed cold fronts. The dust soon rises to the level of the cloud base, about 5,000 - 6,000 feet, and later extends higher still. Haboobs near Khartoum are usually reddish in colour, but they may be yellow or black according to the type of sand in the locality.... The wind is light and variable ahead of the haboob. Its arrival coincides with a sudden increase of wind and the initial squall is often severe. The direction of the squall is usually backed from the earlier prevailing wind. Almost instantaneously the visibility falls from good to some hundreds of yards. The dense blowing sand is most unpleasant, and, in spite of all precautions, everything indoors becomes coated with a layer of fine dust. There is a rapid rise in pressure of 1 or 2 mb and a drop in temperature of 2° or $3^{\circ}C$. There is little change in wet-bulb temperature. (Freeman 1952, p6-8).

Plate 3.3 shows the characteristic dust wall arriving at Khartoum. **Haboobs** usually occur in the afternoon hours, when the lower atmosphere is convectively unstable, and most storms at a station last for an hour or less (Freeman 1952).

Southerly dust storms also occur during the wet season: the advancing monsoon or early monsoon period. They develop south of the ITF due to the steepening of the pressure gradient for southerly winds or as a result of heavy thunderstorm activity over a large area. Morales (1979b) suggests that the causes of this heavy thunderstorm activity may be the existence of widespread areas of convergency and divergency in the lower layers of the atmosphere, possibly coupled to the waves in the easterly air current in these latitudes. Thus, although this is a thunderstorm downdraft type storm, typically blowing from the south, it results from a severe thunderstorm or series of thunderstorm cells.

The third type of dust storm occurring in Sudan is a winter/early spring storm that is generated in dry air masses north of the ITF. This storm moves southwards, often marked by a cold front, and is connected to cold air outbreaks that penetrate from the Mediterranean in the wake of depressions tracking eastwards along the coast of North Africa or the Mediterranean. A dust storm of this type has been tracked through the Sudan by Morales (1981).

3.3.6 TRANSPORT TO THE ARABIAN PENINSULA

Some transport of material in **Haboobs** occurs over the Red Sea (Schroeder 1985), but the major portion of dust carried eastwards is taken by the severe thunderstorm downdraft storms or the winter storms. Grigoryev and Lipatov (1974) have monitored an event that raised dust over a large zone in north-east Sudan, from the Nile to the Red Sea coast in June 1971. Dust raised reached 3000m at the Red Sea coast where it was carried in a southwesterly current towards the Arabian Peninsula. Dust flows from the Sudan and the Horn of Africa appear to be part of the south-west monsoon circulation that operates to the south of the ITF in summer, when it is at its northward extreme (see also chapter 5).

3.4 EASTERN & SOUTHERN AFRICA

The dry lands of eastern and southern Africa are relatively unimportant sources of dust storm activity when compared to the arid lands of North Africa. The low frequency of these phenomena is partly responsible for the general lack of data and other information. From the available sources the following areas are covered:

1. Ethiopia, Djibouti and Somalia.
2. Uganda, Kenya and Tanzania.
3. Angola, South Africa, Botswana and Zimbabwe.

3.4.1 ETHIOPIA, DJIBOUTI & SOMALIA: THE HORN OF AFRICA

The Horn of Africa is an arid area that comprises the coastal belt of Eritrean province in Ethiopia, Djibouti, the Somali Republic, the Ogaden (south-east Ethiopia) and parts of north-east Kenya. This is the region of the Somali-Chalbi desert. Generally, the climate of the area is little known and meteorological data are few and far between. In Ethiopia dust storms are not recorded at meteorological stations (Tesema, pers. comm. 1984) and in the Somali Republic dust storm data are also unavailable (Mahad, pers. comm. 1984). Nevertheless, some comments can be made on dust-raising in the Horn of Africa from the available literature and some other sources.

In the north-west of Ethiopia dust may be raised locally by systems arriving from neighbouring Sudan, which also bring material already entrained in Kassala and Nile provinces. Morales (1979b) has tracked a southward moving dust storm associated with an outbreak of cold air behind a cold front that reduced visibility to less than 500m at a number of stations in Sudan, and was at this intensity as it crossed into Ethiopia at the Blue Nile, no doubt to be dissipated as it reached the Ethiopian Highlands. Thunderstorm downdraft dust storms also occur along the Ethiopia-Sudan border. Vallely (1985) for example, describes the effects on a refugee camp near Kashim el Girba.

On the dry eastern lowlands of Ethiopia the northeasterly continental winds from the Arabian Peninsula bring dust haze during November to April before the south-west monsoon sets in (Tesema, pers. comm. 1984) but it is during the period of the south-west monsoon (about May to September) that the major local dust-raising activity occurs in the Horn of Africa. This activity is most important on the coastal lands and plains adjoining the Red Sea and the Gulf of Aden and can be largely described with reference to three local dust-raising winds.

The **Karif** is an offshore wind that blows from the south coast of the Gulf of Aden. The wind is at its strongest from mid July to mid August, and its strength is related to the height of the interior highlands as it is a partially katabatic airflow. Although it is not known whether the **Karif** blows on all parts of the coast, as reliable data are available only for the station of Berbera (CIA 1965), it is probably felt some distance east of Berbera but less likely to the west where the highlands recede from the coast. There is no mention of the wind in the Djibouti records. The **Karif** is invariably from the south-west and blows chiefly at night, usually setting in suddenly at about 2100 LST, increasing in force to reach a maximum at 0800 to 1000 LST. The average velocity is 10 to 20 knots, although it often reaches 35 to 45 knots and occasionally 50 knots. Although it is usually replaced with a sea breeze during the late morning and afternoon, the wind may blow incessantly for three or four days. It is characterised by high temperatures (40° to 46°C in daytime 38°C at night) and generally raises a great deal of dust and sand, reducing visibility considerably (Brooks 1920).

A hot, dry dust-raising wind commonly experienced in the Republic of Djibouti is the **Khamsin**, a squally northwesterly that blows from May to September. It is not as frequent at the capital of Djibouti as at Obock, 50km to the north. The **Khamsin** rarely blows in the mornings, commonly setting in at 1300 LST and continuing until 2000 LST to midnight. Occasionally it may blow for three or four days, and may be very violent, with wind speeds exceeding 50 knots.

In contrast to the **Karif** and **Khamsin** the **Saba** wind is always cool and blows from the west or south-west, often with gale force. The **Saba** also blows during the summer months and is frequently accompanied by drops of rain, thus generally stirring up less dust than the **Karif** or **Khamsin.** The **Saba** usually blows in the morning hours (0800 to 1100 LST).

In addition to these winds on the coastal plains, dust devils are common over hot inland areas and dust storms also occur in the Danakil desert, especially during the summer months. In winter, from October to April, local topography affects airflow at Dire Dawa, so that northerly winds bring dust from the Danakil (Griffiths 1972).

Some records on the mean frequency of dust days are given by the CIA (1965). Although no visibility limit is used, the stations at Djibouti and Gardo (Somali Republic) seem worst affected. Griffiths (1972) reports that at Djibouti dust storms reducing visibility to less than 1000m occur on average three times a year, mainly in June and July. The regional airflow across this region during the south-west monsoon is generally to the east, south of the seasonal position of the ITCZ that crosses the Red Sea at about 16°N. It seems likely that material entrained into this flow is transported across the Red Sea to the Arabian Peninsula and the Gulf of Aden to the Yemen and Oman coasts and the Arabian Sea. This flow will also pick up material raised in the southern regions of the Arabian Peninsula creating the seasonal maximum in dust haze over this ocean during June, July and August (McDonald 1938). This pan-regional dust flow has been confirmed by the use of satellite imagery by Grigoryev **et al** (1971) and more recently with Space Shuttle images (plates 3.4 & 3.5). The dust flow from the Eritrean coast and Djibouti across the Red Sea and the Gulf of Aden is clearly shown in plate 3.4, taken in June 1985, and plate 3.5, taken on the same day, shows plumes emanating from mountain valleys on the Somali coast adding to the dust flow. Some of this material may be transported as far as the Makran coast of southern Pakistan (Wells, pers. comm. 1985), as plate 3.6 indicates.

3.4.2 UGANDA, KENYA & TANZANIA

Minor dust-raising activity occurs in semi-arid areas of Uganda, Kenya and Tanzania. In parts of north-east Uganda and the Lake Rudolph region of Kenya dust may be raised on plains that are often bare of vegetation, especially in drought years: dust haze and storms occurring commonly during April, May and June (US Hydrographic Office 1943). Dust devils and high atmospheric dust loads are also common in the Lake Baringo area of Rift Valley province in northern Kenya, again particularly during drought years such as the first four years of the 1980s (Bowyer-Bower, pers. comm. 1986). Similarly, dust-raising occurs on seasonally dry alluvial plains of rivers in semi-arid areas, as shown by Reid and Frostick (1984) at Tulu Bar, northern Kenya. In Uganda Visser (1964) found particularly high nitrate levels in rainwater that he attributed to atmospheric soil dust from northeasterly and southwesterly land winds.

An early example of dust storms caused by desertification occurred on the semi-arid savannahs of Ugogo in central Tanzania during the latter half of the last century (Christiansson 1981). Accounts of the region towards the end of the nineteenth century differ markedly from earlier reports, which seldom mentioned any lack of vegetation or degraded areas, although dust devils were obviously a common feature. Descriptions of the area dating from the 1890s, however, depict Ugogo as a treeless country with frequent occurrence of bare soils. The destruction of the natural vegetation was largely due to overgrazing by cattle, particularly along the caravan routes through central Ugogo. This denudation, in combination with a series of dry years during the 1890s, meant that land was prone to wind erosion, and accounts from this period frequently mention dust storm activity:

> Extensive areas are covered with thin steppe grass only, and some are totally devoid of vegetation, so that strong winds can easily transport the sand away.
> (Stuhlmann 1891, p53).

> It should be noted that in Ugogo the winds are generally blowing fiercely, whirling large amounts of red dust into the air, thus causing eye infections.
> (Herrmann 1892, p193).

3.4.3 ANGOLA, BOTSWANA, ZIMBABWE & SOUTH AFRICA

In Angola minor dust-raising activity occurs along the arid coastal strip that stretches northwards to about 12°S. Hinds and Hoidale (1977) show that Moçamedes (or Mossamedes), which records 56mm rainfall a year has an annual average of 0.1 dust storms and nine dust events with visibility below 11km. This activity occurs during the driest months of April and May and August to November with a diurnal maximum from 1300 LST during the afternoon hours. This blowing dust is probably associated with the **Berg Wind,** a dust-raising airflow common along this coast of southern Africa (see below). Further north and inland, Henrique de Carvalho records 10 occurrences of blowing dust a year, largely concentrated in June, July and August, with a night-time maximum from 1900 to 0400 LST.

In Botswana dust storms are rare (Gower, pers. comm. 1984), although they occur in areas of particular susceptibility such as the Makgadikgadi Pans (Goudie, pers. comm. 1986). In Zimbabwe wind speeds of the required strength are infrequent. Any dust storms that do occur in Zimbabwe are mainly caused by thunderstorm squalls; Torrance (pers. comm. 1984) recalls just two dust storms in Harare in the last 30 years, both associated with thunderstorm downdrafts and lasting half an hour. A 10-year search (1971-80) of WMO 'present weather' codes (see fig 2.5) for a number of stations in selected months revealed 10-year totals of two dust storms at Harare, five at Bulawayo and zero at Beitbridge. October is the month of most activity, being the start of the wet season when any thunderstorms tend to be more severe than in other months. This activity is most typical in the afternoon hours.

McDonald's (1938) haze frequency maps (fig 3.1) show an increased frequency off the coast of southwestern Africa during autumn which is attributabe to airborne dust from the Namib and Kalahari deserts. Dust storms are not recorded at meteorological stations in South Africa, but an aeolian transport on the scale of the North African dust sources can be excluded reasonably from a number of indirect lines of evidence. Not least of this evidence is just the fact that these phenomena are not recorded along with other meteorological elements, in itself an indication of their relative infrequency. The pattern of haze distribution is oriented in a northwesterly direction due to the southeasterly trades which carry the dust out to sea. This pattern, however, is not reflected in the deep-sea sediment distributions of kaolinite and illite, so that Schütz (1980) suggests that a considerable aeolian transport can be excluded. Indeed, aerosol measurements from this area of the eastern Atlantic seem to confirm this assumption, as dust concentrations were only two times that in undisturbed marine air (Chester **et al** 1972, Aston **et al** 1973). Isotope investigations by Biscaye **et al** (1974) of Kalahari dust over the Atlantic also seem to exclude this region as a significant source of airborne dust, although Parkin **et al** (1972) detected aerosols largely composed of quartz off the Brazilian coast, the small size of which (none larger than 9 microns) they suggested indicated transport in the south-east trades from the Namib. Nevertheless, the general conclusion that the southern African deserts are not significant dust sources is further supported by the lack of appropriate large-scale dust-producing geomorphological units in the region (Goudie, pers. comm. 1985).

Despite this conclusion, a number of comments should be made on dust-raising in this part of Africa. A warm, off-plateau partially katabatic airflow that blows from the east and north-east on the coast of southwestern Africa is known as the **Berg Wind** , which is often characterised by high dust loads. The **Berg Wind** blows in all the seasons but is particularly prevalent from May to August when it blows between four and six times a month (Knox 1911). Höflich (1984) reports that the wind is experienced along the entire arid coastline south of about 15°S, its duration is often just a few hours but it may blow for several days, at times reducing visibility to 30m. The wind is very squally by nature and may reach Bft7 at Lüderitz and 30 knots at Swakopmund. The coastal fogs common in this part of the world are occasionally yellowed by atmospheric dust.

Further south, at Cape Town, where dust events are uncommon, Groves (1961) describes a severe dust storm in June

1961 caused by a secondary depression associated with a frontal trough. Visibility was reduced to 1000 yards by dust that had been blown from freshly ploughed agricultural land at Swartland, north of Cape Town.

Inland, dust storms occur in the arid and semi-arid regions of the subcontinent. These areas include western, central and northern parts of the Cape Province; central and southern parts of the Orange Free State; and the southwestern Transvaal. Activity in these regions is more likely in spring and summer months when convection is at a maximum (Keyworth, pers. comm. to D. Excell 1984). In the western interior of South Africa dust storms are often associated with thunderstorms (Schulze 1972). Jacobs (1964) describes a thunderstorm downdraft storm at Upington in February 1964 that struck from the north-west with winds gusting upto 42m/s. A similar storm associated with convective activity struck Kimberley in the afternoon of 19 November 1983 when visibility was reduced to 100m with a westerly wind averaging 16m/s (Keyworth, pers. comm. to D. Excell 1984).

3.5 CONCLUSION

Although the Sahara has long been recognised as the world's largest area of dust storm activity, and much work has been carried out in recent years on long-range transport of Saharan dust particularly over the Atlantic Ocean, the source areas of Saharan dust storm sediments are not well known (Schütz 1980). This chapter has attempted to make progress in this direction.

Saharan dust sources are divided according to the long-range paths commonly followed by their material. These source areas have been tentatively specified using terrestrial dust storm frequency data, with data and observations reported in the literature on satellite, turbidity and aerosol monitoring programmes. The main source areas identified are shown with the principal directions of long-range transport in figure 3.19.

Dust transported across the Atlantic is derived from a variety of sources. These include the alluvial plains of the Dra River in southern Morocco (summer maximum in acivity), the dune fields of south-central Algeria (summer maximum), the dune fields of central Mauritania and northern Mali, including sebkhas and other interdune depressions (spring maximum), and the lacustrine and fluvial sediments of the Chad Basin (winter maximum).

Dust transport from these areas is in predominantly monodirectional winds, in accordance with the prevailing northeasterly **Harmattan.** Dust is raised by the effects of cold outbreaks entering the regions from mid-latitudes and disturbances (both depressions and thunderstorm squall lines) that track westwards across the southern Sahara. In all areas the effects of strong solar heating enhances turbulence in the lower troposphere, encouraging dust-raising winds, and heat lows developed in this way may sweep material out over the Atlantic by intense cyclonic gyre.

Material raised from all of these areas may be transported over long distances in the upper wind regimes, characteristically as a distinct layer of dusty air. Dust transported in the latitude of Barbados (usually referred to as 'summer dust') probably originates largely in the southern Moroccan and south-central Algerian sources noted above. Material transported to Cayenne, commonly referred to as 'winter dust', is more accurately described as a spring flow, as atmospheric concentrations at that monitoring station (although for just a two-year period) are at a maximum in March. Evidence is slightly in favour of the important sources of this dust being in Mauritania and northern Mali. Although dust from the Chad Basin is carried over large parts of West Africa in winter, it is thought to be largely washed out over the Gulf of Guinea.

In the northern Sahara dust storms are very largely attributable to the multidirectional winds associated with the passage of depressions across the Mediterranean region. Dust storm frequency data indicate that the most intense area of activity in North Africa is in a broad band stretching from central Algeria eastwards to the Gulf of Sirte. This is an area of ergs with numerous desert depressions and sebkhas. The season of dust storm maximum in this area is during the spring months when material is transported north over the Mediterranean and Europe and eastwards to the Levant, occasionally reaching as far as the European areas of the

USSR. The number of reported instances of Saharan dust reaching the British Isles has been more than one a year since 1977, as compared with just four events recorded in the first 77 years of this century, but this may simply reflect a recently increased awareness of such phenomena in this country.

In the eastern Sahara, the region of north-east Sudan is an area of high activity, where dust storms are generated by cold air outbreaks moving southwards from Mediterranean latitudes and the thunderstorm downdraft storms that may be single cell systems or synoptic scale events associated with severe thunderstorms. Material from this area is transported across the Red Sea to the Arabian Peninsula.

The Horn of Africa appears to be the source of material that is carried in a large-scale flow over the Arabian Sea and the Arabian Peninsula as far as the Makran coast. Dust storm frequency data from this part of Africa are sparse, but remote sensing attests to the scale of this flow, which is part of the south-west monsoon circulation.

The dry lands of southern Africa do not appear to be a significant source area of dust storm sediments, partly due to the lack of widespread dust-producing geomorphological units in the area. Quantification of this region is impossible, however, due to the lack of data, but some long-range transport occurs over the South Atlantic Ocean.

The variation in dust storm activity in various parts of Africa can be understood with reference to meteorological parameters. In Mauritania dust storm activity occurs in the dry season, when wind speeds are high, and apparently poor relationships between dust storms and wind speeds at individual stations are understood with reference to wind direction and the proximity of stations to dust sources. In the Sudan dust-raising occurs in the 'rainy' season, which is understood with reference to the meteorological systems during this season.

The scale of long-range transport of dust from the Sahara has been indicated by remote sensing and ground monitoring programmes, and the history of such export is recorded in deep-sea sediments and areas of peri-desert loess. Such loess deposits have been studied in the Levant (Yaalon & Ganor 1973), southern Tunisia (Coudé-Gaussen 1984b) and northern Nigeria (McTainsh 1984).

Investigations of the dust sources in the Sahara suggest that specific small geomorphological units are the most important dust-producing areas (Coudé-Gaussen 1984a, Yaalon 1986). For some areas this chapter identifies dust sources with a fair degree of accuracy, such as the alluvial plains of the Dra River, southern Morocco, and the lacustrine and alluvial deposits of the Chad Basin. In other areas, however, identification of such units has been more tentative, such as the ergs of central Mauritania, Algeria and Libya. This lack of detail results from the broad-scale approach of the present thesis. Within such large-scale sand sea regions some likely specific sources of dust storm sediments have been indicated, such as desert depressions, sebkhas and palaeo-river channels. Further investigation, on the local scale, is required to identify more accurately the dust-producing units within the broad-scale regions noted here.

It should be noted that some of the most arid areas of the central Sahara do not appear to be major dust sources. This situation may be a reflection of the small number of meteorological stations located in such remote regions, but satellite monitoring of the Sahara confirms the relative lack of activity in some of the most arid regions. Although mechanisms for the production of dust-sized material in arid lands have been recognised (eg attrition during aeolian transport of sand grains and salt weathering - see section 2.1.3) such less important arid areas may be regions of relative depletion after long-continued wind erosion and the formation of relatively stable surfaces.

CHAPTER FOUR : THE SAHEL - EFFECT OF DROUGHT & DESERTIFICATION ON DUST STORMS

4.1 INTRODUCTION

Evidence from deep-sea coring projects shows that the output of aeolian dust from North Africa to the North Atlantic has varied greatly since the last glacial maximum (about 18,000 years BP), reflecting major changes in temperature, aridity and wind regimes over the continent (Parmenter & Folger 1974, Sarnthein & Koopman 1980). In recent times the ability of human actions significantly to affect the African dust regime became clear during the tank battles of the North African campaign in World War II (see section 3.3.3A). Since the late 1960s, a severe drought has gripped the Sudano-Sahelian zone to the south of the Sahara, and continues to this day. Numerous analyses of below average rainfall over a wide area have appeared (eg Tanaka **et al** 1975, Lamb 1982, 1983) and this drought period has been characterised by a marked increase in dust-raising activity across the whole region.

It seems likely that this increase in dust storms is due partly to the effects of a prolonged period of below average rainfall and partly to the effects of increasing population pressure and poor land use practices which have exacerbated the effects of the drought. Human action may add to the depletion of vegetation cover brought about naturally by drought conditions or, conversely, overuse of a land surface that goes unnoticed during high rainfall years may be made evident by pluviometric deficiency. In this chapter I shall present the evidence for an increased dust output from the Sahel during the drought, and assess the relative importance of decreased rainfall and the actions of human populations in generating this increase. The environmental implications of this enhanced dust output will then be examined, including the possible effects on climate.

4.2 THE SAHEL

In a geographical introduction to the Sahel Grove (1978) describes the region as

> ...essentially a phytogeographical unit, a zone of grassland, scrub and thornbush passing northwards into the Sahara and southwards into savannah.
> (Grove 1978, p407).

The most commonly used limits of the Sahel are the annual isohyets, although there seems to be no overall concensus in the specific isohyets used, the area being variously delimited within a rainfall maximum of 750mm and minimum of 100mm. Since the effects of the drought and the actions of local populations have severely modified the phytogeographical nature of the Sahel in localised areas, a deliberately approximate definition of the region is adopted in this chapter. Thus, the Sahel is here defined as lying approximately between latitudes 10°N and 20°N and includes parts of Mauritania, Senegal, Mali, Burkina, Niger, Chad, Nigeria, Sudan, Ethiopia and Somalia. Figure 4.1 shows the distribution of stations referred to in this chapter.

The geomorphology of the Sahel exhibits a range of palaeofeatures that reflects the climatic fluctuations that have occurred in late Quaternary times. Much of the Sahel is characterised by fossilized dune fields that were active at the last glacial maximum (Sarnthein 1978). These fields include the Qoz of Kordofan, the ancient erg of Hausaland, the dunes of the inland Niger Basin, the Ferlo of Senegal and the Trarza in southern Mauritania.

4.3 EVIDENCE FOR INCREASED DUST STORMS WITH THE ONSET DROUGHT

The first quantitative documentation of an increased dust output from the Sahel with the onset of drought was suggested by Prospero and Nees (1977). They monitored variations in the turbidity of the lower atmosphere at Barbados as affected by the African Dust Plume (Carlson & Prospero 1972) and found that during the severe drought period 1972-74, mean aerosol concentrations were three times those of pre-drought levels, that is before 1968. Continued monitoring at Barbados shows a renewed increase in aerosol concentration in 1982 and 1983 in response to recent severe North African rainfall deficiencies (Prospero & Nees 1984). Although these findings indicate that dust output from North Africa may have increased significantly with the onset of drought, they cannot be regarded as conclusive for a number of reasons.

Firstly, the increase in dust concentrations at Barbados may reflect changes in large-scale global circulation patterns. Newell and Kidson (1984) suggest that the lower tropospheric easterly jet over West Africa may have been stronger during the drought, which may have increased the amounts of material carried over long distances. Further, the relative lack of rainfall during a drought period would mean that smaller amounts of airborne dust would be washed out, thus leaving more for potential long-range transport. Perhaps most importantly, however, is the realisation that the main source of material reaching Barbados is not from Sahelian latitudes but from areas further north, as shown in chapter 3. The increase in dust monitored at Barbados may have been a reflection of increased dust-raising activity in North Africa and figure 3.14, for example, shows that Ouarzazate experienced a marked peak in dust storm activity during the early 1970s. It should be noted, however, that direct correlation between terrestrial dust storm frequency and long-range transport should not be made since the atmospheric systems generating storms are not necessarily those conducive to long-range transport.

Some evidence has been presented from North Africa however. Bertrand **et al** (1979) show the variation in annual numbers of days with dust haze from 1947 to 1978 at Bilma, Zinder and Niamey, although no visibility limit is indicated for their data. At Bilma the number of hazy days rose steadily from 20-30 in 1955 to more than 200 in 1973. At Zinder and Niamey a rising trend is discernable from the mid-1960s. At all three stations a significant rise is indicated in 1968/69. Investigation of rainfall at Niamey, Zinder and Agadez shows that the period of significantly increased dust haze accompanies a period of pluviometric shortage. The linear correlation coefficient between rainfall at Agadez and days with dust haze at Bilma is -0.65; this becomes -0.88 if dust haze is correlated with the rainfall of the preceeding three years.

These stations are in the area affected by the **Harmattan** dust haze. Further south and west the evolution of dust haze at Bobo-Dioulasso (Burkina) and Bouaké (Ivory Coast) is estimated by the number of hours of visibility reduced to less than 5km amd 10km for two periods: 1957-69 and 1970-74. At Bobo-Dioulasso the average number of hours of dust haze at these visibilities increased by a factor of between 3.0 and 4.5 between the two periods, and between 2.0 and 2.5 at Bouaké. The seasonality of dust haze was also modified. Dust haze tended to appear earlier in the season and finish later, and the months of maximum activity changed from February at

Bobo-Dioulasso for 1957-69 to December in 1970-74.

Further evidence from Niger is presented by Heusch (1980). At Tahoua, in the Ader Dutchi massif, the mean annual number of hazy days for the period 1965-74 was 54, but the frequency varied from nine days in 1965 to an average of 82 days a year from 1971 to 1973.

In northern Nigeria, hazy days with visibility <3000m at Samaru generally increased as rainfall declined from 1962 to 1973 (Adetunji **et al** 1979), and observations of severe dust haze (visibility < 1000m) at three-hourly intervals at four stations during the period 1970-76 showed a marked increase in frequency in 1972 and 1973 when the drought was particularly severe (Kalu 1982). Also in Nigeria, Adefolalu (1984) shows that the highest numbers of cases of respiratory problems associated with dust were recorded during the drought spell 1970-75, and the effects of **Harmattan** dust on human health in Nigeria has been on the increase since 1976, and may become serious in the immediate future in southern parts of the country which have hitherto been little affected.

The unusual strength of the **Harmattan** in recent years is also suggested by reports that in 1983 dust reached Libreville, the capital of Gabon, a few kilometres north of the Equator, for the first time in 25 years (Morrison 1983).

The above evidence from the literature on the Sahel drought refers to that part of the region affected by the dust source of the Chad Basin. I now present data confirming the increase in dust storm activity from this source and showing similar increases for the other major dust source regions in the Sahel as outlined in chapter 3. The main data obtained are from Mauritania, Nigeria and Sudan, with additional data, personal observations and press reports from Senegal, Mali and Chad.

4.3.1 NIGERIA & CHAD

Further data for Nigeria were obtained from the Nigerian Meteorological Department for eight stations mainly in the north of the country. These data, which are shown in figure 4.2A-H, indicate the variation in annual totals of severe dust haze days (visibility $<1000m$) and rainfall for the period 1955-79 and the average monthly dust day totals for two periods: 1955-65 and 1969-79, with the change in average annual rainfall between the two periods. Several stations are lacking data for 1966, 1967 and 1968, when civil war disrupted normal life in Nigeria.

All stations show a period of decreased rainfall at some time during the late 1960s/early 1970s, a relative recovery in rainfall totals in 1974 and 1975, with some stations experiencing a second low rainfall period in 1976 and 1977. At some stations (Yelwa, Gusau, Sokoto and Nguru) this rainfall pattern is mirrored by an upward phase in the number of dust days recorded, that falls off again as rainfall rose in 1974/75. At these stations the average monthly totals of dust days for the two periods do not greatly differ, although average rainfall totals were significantly lower during the second, or drought, period (1969-79) at all stations.

The pattern is somewhat different at Maiduguri and Yola, to the east of the country. Both stations show an increase in the number of dust days during the early 1970s, and the frequency of dust haze remains generally high after this time despite the relative recovery in rainfall totals during the mid-1970s. Thus, for these two stations the average number of dust days during 1969-79 is significantly greater than in 1955-65, by a factor of two at Maiduguri, and a factor of three at Yola for the maximum month in each case.

At Potiskum, high dust haze day frequencies for 1970-76 do not show any obvious correlations with corresponding annual rainfall, and dust frequency falls sharply in 1978/79. Thus, average monthly dust haze days are just slightly increased during 1969-79 as compared to 1955-65. Katsina, by contrast, shows a marked decline in the number of dust haze days in 1972/73 when rainfall was lowest, although the average frequency of dust haze days is two times greater in the maximum months during 1969-79 compared to 1955-65 due to high dust day frequencies in 1970/71 and 1977/79, when rainfall was not markedly deficient.

Several stations show a relatively minor peak in dust around 1960, with Nguru showing a very marked peak in that year, with more haze days in this year than at any time during the drought period. This peak does not appear to be particularly related to below average rainfall.

It should be noted that since the dust haze at Nigerian stations is caused by material entrained to the north-east and blown by the **Harmattan,** the local rainfall totals are not directly related to dust haze, except by the amount of wash-out they represent. Thus, the relationship between dust haze and local rainfall is not a simple one, and although the general deficiency in rainfall totals from about 1968 occurred throughout the Sahel, the stations in Nigeria furthest from the dust source - Yelwa, Gusau, Potiskum and Yola - do not show such obvious pluviometric control on the number of dust haze days. At Sokoto and Maiduguri the year-to-year variations in rainfall are more closely related to the variation in days with dust haze. Closer proximity to the dust source does not necessarily produce a closer relationship betwen the two parameters however, as seen at Katsina, and at Nguru, although there seems to be a fairly obvious increase in dust haze frequency in the early 1970s, the peak during the early 1960s does not show any obvious relationship with rainfall. Nevertheless, although the relationship between rainfall deficiency and dust haze frequency is not always obvious or consistent at stations in Nigeria, all stations analysed show some degree of increased dust haze frequency during the period of the first ten years or so of the current Sahel drought.

Data from Niger and Chad, in the major source region of the **Harmattan,** the Chad Basin, were unobtainable. However, some qualitative evidence for an increase in dust-raising at Ndjamena, the Chad capital, during the drought was found in a press release from Agence France Presse:

> La sécheresse qui en l'espace de dix ans a fait avancer le désert de 200 kilomètres a aggrave du même coup l'intensité et la fréquence des vents de sable. (Agence France Presse 1985, FLT2).

4.3.2 MAURITANIA, MALI & SENEGAL

Data from Mauritania were supplied by the Service Météorologique, Nouakchott, and for Senegal by the Agence pour la Sécurité de la Navigation Aerienne en Afrique et a Madagascar (ASECNA), Dakar. The variation in frequency of annual dust storm days and annual rainfall totals for Nouakchott is shown in figure 4.3. The increase in dust storm days after 1968 is dramatic. Low rainfall totals of 48.1mm in 1970 and 17.9mm in 1971 represented just 32 and 12% respectively of the 1949-67 average, and can be seen as the main onset of the drought. The number of dust storm days increased markedly from 6 in 1970 to 65 in 1974 before a reasonably high annual rainfall of 190.6mm in 1975; dust storm activity declined to 25 days in 1976 and 27 days in 1977. In 1977, however, the rainy season brought just 2.7mm of precipitation, making it the driest year since records began in 1931, and dust storm activity rose to 55 and 61 days in 1978 and 1979 respectively. The total dropped to 33 dust storm days in 1980 after a relatively heavy rainfall in 1979, but rose to an unprecedented 85 days in 1983 and 79 days in 1984 as rainfall again diminished.

The quality of data from Mauritania enables a better assessment of the magnitude of change in the quantities of material raised than is possible for Nigeria above. Figure 4.4 shows the average monthly percentage frequency of dust storms by hour at Nouakchott for two periods: 1949-67 (data from Hinds & Hoidale 1977), representing the pre-drought period, and 1968-77, covering the first ten years of the drought. The differences in frequencies between the two periods are marked. The frequency during the peak time of activity, 1200 to 1500 local time, in the drought years being on average nearly seven times that of the pre-drought period during the dustiest months of January to April. The increase in terms of dust storm events between the two periods is smaller, being just over three times more frequent during the drought years. Similar comparative data were available for seven other stations in south and central Mauritania; table 4.1 summarises these data to show the increase in average diurnal frequency per cent from the pre-drought period to the drought years. Combining these results shows that, on average, dust-raising activity in the Mauritanian Sahel was more than six times greater during the drought as compared with the pre-drought years.

Coherent rainfall data covering these periods were available only for Nouakchott, which indicate that the average for the 1968-77 period was 51% of the pre-drought average. Vermeer (1981), however, shows that for the entire Mauritanian Sahel precipitation after 1971 was considerably lower than normal, and that the drought continued throughout the decade. At Rosso, for example, the 1935-72 annual average rainfall was 284mm, while the average for 1968-72 was 149mm (UN 1977). Recent reports on sub-Saharan rainfall (eg Lamb 1982, 1983), that have included data for some of these Mauritanian stations, also indicate that the drought that started in 1968 persisted strongly throughout the 1970s, and, indeed, still continues. It seems clear, therefore, that the quantity of material raised by wind has greatly increased during a period of substantially reduced rainfall in the Sahel of Mauritania.

It is interesting to compare the increase in dust storm activity in central and southern Mauritania with the northernmost stations of Bir Moghrein and F'Dérick. Figure 4.5 shows the average monthly percentage frequency of dust storms by hour at these stations for the same two periods. At F'Dérick the frequencies are also higher during the 1968-77 period, the change of average hourly frequency per cent of dust storms is +206%, or a doubling in frequency. At Bir Moghrein, however, dust storm activity is reduced during the 1968-77 period, and the change of average hourly frequency per cent of dust storms is negative, the average 1968-77 values expressed as a percentage of 1957-67 values being

+29%, representing a reduction in activity by about three times. These northern stations record very low average annual rainfall, Bir Moghrein <50mm and F'Dérick 59mm. Rainfall figures over the two periods were not available for these two stations, so that it is not possible to state that these changes occurred during a drought in the arid north of the country, but the relatively small increase at F'Dérick and the decrease at Bir Moghrein serves to emphasise the large increases in the semi-arid south, which did experience drought conditions in the 1970s. It is also worth noting that the term 'drought' is slightly less meaningful in the more arid north of the country.

At Nouakchott, further investigation of the dust storm/rainfall relationship is shown in figures 4.6 and 4.7. Figure 4.6 compares dust storm frequency to the previous year's rainfall (note that the dust storm season at Nouakchott is primarily in the first six months of the year, before the onset of the rainy season, as shown in section 3.3.1A), the linear correlation coefficient is -0.53 (28% explanation). The relationship between dust storms and antecedent rainfall is stronger, however, when annual dust storm days are compared to the average annual rainfall over the previous three years (linear correlation coefficient = -0.75, with 56% explanation). This fairly strong relationship is similar to that found by Bertrand **et al** (1979) referred to above for Agadez and Bilma in Niger.

In Senegal, dust haze from the more arid areas to the north and east is more common than dust storms generated locally. Data for annual frequencies of dust haze and rainfall at Dakar are shown in figure 4.8. Fairly closely related trends in these two parameters are apparent. Low rainfall in 1968 marks the beginning of an increasing trend in the numbers of dust haze days, peaking in 1970 and staying high while rainfall was low, until a relative recovery in rainfall in the period 1973-75, during which time dust haze days declined in frequency. After 1975 rainfall again declined, with a concomitant increase in dust haze.

In Mali, where the River Niger sank to its lowest recorded level in 1984, data were unavailable, but reports reaching this country in the media attest to the increase in dust haze and storms that has occurred since the onset of the recent drought:

> ...every hour of every day of the week I spent in Mali....a giant dust cloud hung in the air, covering the country. It was Mali's precious topsoil, turning into its shroud.
> (Lean 1985, p14).

4.3.3 SUDAN

Data from the Sudan were taken from the Sudan Meteorological Department Annual Meteorological Reports, with some additional material supplied direct from the department itself. The data refer to five stations across the Sudanese Sahel. Data in the yearbooks are given as monthly frequencies of dust storms recorded at 0600 GMT, 1200 and 1800, so that a record of an event at 0600 and at 1200 may refer to a single storm that lasted for six hours or more, or to two separate events. Thus, the graphs in figures 4.9A-D show annual rainfall totals with annual totals of recorded dust events (ie the summation of frequencies for 0600, 1200 and 1800 GMT). From 1975 most Sudanese stations started making an extra visibility reading at 0000 GMT, and some of the data presented below have been published without taking this additional observation into account (see Middleton 1985a, Hulme 1985, Middleton 1985b). Some of these data, therefore, were inflated - those for El Fasher, El Obeid and Khartoum - but those for Aroma and Tokar were unaffected by the extra reading. Nevertheless, the general trends indicated in Middleton (1985a) are still evident in the adjusted data (omitting the 0000 GMT reading, thus maintaining consistency before and after 1975) as shown in figures 4.9A-D and table 4.2.

The annual dust storm frequency and annual rainfall totals for El Fasher, El Obeid and Khartoum (Fig 4.9A-C) show a marked rise in dust storm activity dating from the late 1960s/early 1970s. Particularly low rainfall in 1972 and 1973 at El Fasher, for example, was followed by a distinct rise in dust storm frequency, peaking in 1974, falling in 1975 and 1976 after high rainfall in 1974, but remaining at increasing levels after that year as annual rainfall remained for the most part below 200mm. The zero dust storm reading for 1979 followed the wettest year in the cental Sudan (1978) in the last 20-25 years (Trilsbach & Hulme 1984), although particularly high rainfall was not evident at El Fasher itself.

The variation in dust storm frequency and rainfall at Tokar has been smoothed using three-year running means because of the high variability of both parameters. Until the mid-1960s the trends of dust storm frequency and rainfall are broadly similar, with roughly simultaneous periods of low precipitation coinciding with few dust storms, confirming Olsson's (1983) observation that the correlation between dust storms and rainy days at Tokar is 'positive and rather high' (Olsson 1983, p32). After the mid-1960s, however, this correlation changed during a prolonged period of low rainfall lasting until the early 1970s, during which time dust storms were of consistently high frequency.

Further analysis of the Sudanese Sahel was carried out using the relatively gross variable of mean numbers of dust storm observations occurring during pre-drought and drought years (table 4.2). Here the averages for 1950-67 and 1968-78 are shown with the increase between the two periods, and corresponding decrease for rainfall. No mean increase in dust storms is shown for all five stations as this would lead to biasing by the stations that had few storms previously and now experience many more storms, but still relatively few compared with other stations (eg El Fasher). Nevertheless, the pattern is clear; for five stations across the Sudanese Sahel dust storm activity during the drought has markedly increased while rainfall averaged 70% of the pre-drought average.

It is not possible, using these data, however, to state whether the increased dust storm activity is in the form of more dust storm events, or a similar number of events of longer duration.

4.4 DROUGHT OR DESERTIFICATION?

Correlations between mean annual rainfall and dust storm frequency have been shown to be poor (see Goudie 1983 and selected chapters of the present work). The relationship between drought and dust production is complex. However, the evidence presented above, showing an increased dust output from the Sahel during a prolonged period of drought, indicates that rainfall may play an important role in the occurrence of dust-raising activity. The theory behind the importance of rainfall to the wind erosion system has been discussed in chapter 2, and it appears that variations in annual totals are more significant than annual averages.

The effects of a reduced rainfall on a soil's wind erodibility can be exacerbated by the actions of human populations, or conversely, misuse of semi-arid ecosystems may go undetected over a period of high rainfall but be accelerated and made evident by drought. Such misuse has a variety of forms; overgrazing reduces vegetation cover, and livestock trample and disaggregate soil structure, organic litter and soil humus which becomes more susceptible to aeolian transportation. Overcultivation can lead to decreased soil fertility, poorer soil structure and ultimately to a reduction in protective vegetation cover. Burning of agricultural lands also reduces vegetation cover, as does the felling of trees and shrubs for fuel, to feed livestock or to clear land for cultivation. These actions form some of the main elements of the desertification process and may all increase the potential wind erodibility of a soil, by reducing its mechanical stability or decreasing its protective covering. The relationships between drought, human action, desertification and dust storms are summarised as a model in figure 4.10.

The theory relating desertification to dust storm activity is sound, and increasing numbers of dust storms has been suggested as a feasible indicator of desertification (Reinig 1977, McTainsh 1986). Relatively soon after the onset of the Sahelian drought, and the local recognition of a phenomenon usually referred to as desertification, workers recognised that human action was degrading the environment in the Sahel by burning, excessive woodcutting, overgrazing and overcultivation, and causing increased erosion by wind and water during the drought (eg Delwaulle 1973, Michon 1973). Numerous subsequent studies have given credence to these initial observations (eg Rapp 1974, UN 1977). The fact that drought and desertification seem to be linked makes the assessment of the relative importance of each in the wind erosion system difficult. The term 'desertification' is itself difficult to define, and has been adopted into popular parlance and generally overused and misused as a result (Warren 1985). If desertification is taken to be 'the spread of desert-like conditions in arid or semi-arid areas, due to man's influence or climatic change' (Rapp 1974, p3, although Rapp actually prefers the term desertization) then care must be taken to make the distinction between natural changes in the desert fringe ecosystem that are in response to the occurrence of drought - a normal characteristic of semi-arid climates - and more permanent changes that are due to human action and/or drought and/or some form of climatic change. Whether or not the widespread degradation of the Sahel by an expanding human population is permanent can only be assessed in the long-term, and even then the relative role of the drought and possible long-term changes in climate will make such an assessment difficult. Change is further complicated by the possible feedback effects on climate that have been suggested since the beginning of the most recent Sahel drought (see section 4.5). Nevertheless, large-scale increases in dust storm activity have occurred in the Sahel at the same time as a prolonged drought and at a time of widespread ecological degradation in which an expanding human population has played a significant part. The relationship between increasing dust storms, drought and desertification can only be assessed adequately by specific field evidence which will now be reviewed.

4.4.1 THE CHAD BASIN

Evidence of desertification in the Chad Basin is difficult to obtain. In the first years of the drought Depierre and Gillet (1971) noted that vegetation had been destroyed by tree cutting for firewood in Chad and the clearing of trees along river banks had led to lowering of water tables and drying up of streams, while the overgrazing of cattle, trampling and burning was exposing soils to erosion and destroying humus. UNEP (1985) also highlights the indiscriminate use of bush fires, especially in the south of the country, which aggravates the situation, and the political instability that has hampered many of the attempts to control desertification.

The combination of this overexploitation with drought has resulted in widespread loss of vegetation on the southern fringes of the Sahara in Chad, as indicated by the personal observations of Father Faure, Vicaire General of Ndjamena:

> Il n'y a plus rien pour arrêter le vent....les arbustes qui formaient la milleure protection ont disparu depuis que le désert do a nos portes.
> (Agence France Presse 1985, FLT2).

The desiccation of the recent drought has certainly had a profound effect on the level of Lake Chad, leaving the northern portion of the modern lake completely dry by 1982, and considerably lowering levels in the southern portion. There are reports that agriculturalists have moved into areas that had standing water in the 1950s and 1960s (Wells, pers. comm. 1985) and it seems that some of the most recently exposed areas of former lake bed may have become sources for present dust storm sediments as indicated in plate 4.1.

Thus, it appears that during the current Sahel drought the Bilma/Faya Largeau dust source has experienced an increased dust output and that the region south-west of the dust source, in the semi-arid Sahel zone, has also been contributing to increased **Harmattan** dust loads as vegetation has been degraded by natural and human action. The Chad Basin has experienced considerable fluctuations in climate over the Quaternary, and its geomorphology is characterised by a number of potential dust-producing units such as widespread palaeodunes and closed depressions (Grove & Warren 1968) as well as the lacustrine spreads of the former lake bed.

4.4.2 MAURITANIA, MALI & SENEGAL

Reports from Mauritania indicate that desertification problems are:

> ....the result of severe degradation of rangelands due to overgrazing and overstocking, and the excessive cutting of woody vegetation especially for fuelwood. The prevailing dry conditions of the 1970s had the effect of amplifying this already serious problem which in addition to encouraging herdsmen to cut woody vegetation to provide fodder, later forced them to concentrate in the southern parts of the country and Senegal....Sand encroachment and shifting dunes have affected many areas, in particular around urban centres and settlements.
> (UNEP 1985, p64).

Vermeer (1981) shows that the severity of drought in the Mauritanian Sahel during the 1970s was greatest in the west-central region, the zone which during good years is the area of maximum opportunity for pastoralists. Between 1959 and 1968 cattle numbers increased from 1.25 million head to 2.3 million, resulting in severe overgrazing in the Sahel (Gornitz & NASA 1985), but the degradation of the pastures, in combination with the desiccating effects of the drought, saw a 55% reduction in the number of cattle between 1968 and

1973 due to the reduced carrying capacity of the land (Vermeer 1981). In an attempt to maintain surviving animals herdsmen topped shrubs and trees to provide feed, especially for the more drought tolerant sheep and goats whose numbers were not as severely affected. However, these:

> ....futile, last-resort efforts served only to remove the meagre remaining vegetation and permitted additional deflation of soils by dust storms and sand storms. (Vermeer 1981, p293).

The widespread degradation of the Sahelian savannah vegetation in southern Mauritania has been noted on imagery from various remote sensing platforms over the period 1965 to 1985, showing a severely degraded strip about 200km in width running across the southernmost portion of the country (Wells, pers. comm. 1985). This zone of degradation is indicated in figure 4.11 with the areas of drought severity during the 1970s as determined by Vermeer (1981) and the mean increase in dust storm frequency per cent for stations in Mauritania as discussed in section 4.3.2 above. Although increases in dust storm activity have occurred during the drought in central Mauritania, on the northern fringes of the Sahel zone, figure 4.11 shows that the stations located in the Sahel, in those areas that suffered from severe drought and also from widespread desertification (both through natural and human causes) have recorded significantly greater increases in dust storm activity. These Sahelian areas, as figure 4.12 shows, are characterised by extensive palaeodunefields and other geomorphological environments containing dust sized sediments that have become zones of reactivation and dust export, representing a major extension to the already important dust sources in the dunefields of central Mauritania.

The increase in dust storms affecting Nouakchott has stimulated a green belt project that was initiated in 1975, aiming to protect the city from windblown dust and shifting dunes. Although the project has been hampered by the prolonged drought, UNEP (1985) reports that the survival rate of a range of native and exotic plant species has been high (around 50% for **Euphorbia balsamifera**) which has contributed greatly to the stabilization of the project area. Although dust storm frequency has continued to increase since 1975 perhaps the situation may be ameliorated as the green belt becomes effective, although such approaches to combat desertification are not generally well thought-of in many academic circles (Warren 1984, Mabbutt 1985).

The desertification problems of Mali have also been aggravated by the severe drought since the 1970s, and amplified by:

> ....the degradation of range and arable lands through overgrazing, excessive collection of wood for fuel and other domestic purposes, and much reduced fallow of

cultivated areas. Droughts have forced livestock to concentrate in the southern regions, increasing existing pressures. The development of a network of water supplies for stock in the form of wells and temporary surface dams has aggravated desertification problems further. Attempts to increase food and cash crop production through irrigated agriculture, in particular along the Niger, have not only introduced the problem of salinization but have also encroached on grazing lands. With poor cultivation methods, the combined effect has been a general degradation of grazing areas leading to severe wind and water erosion, sand encroachment especially in the southern regions, shifting sand dunes, and general desiccation.
(UNEP 1985, p62).

This desiccation and desertification has been particularly severe in the inland delta of the Niger River; Barth (1978) fears that the agricultural potential of the inland delta is being seriously threatened as the alluvial basins have been transformed into major sources of dust storms. Indeed, satellite monitoring of large-scale dust storms by Grigoryev and Kondratyev (1981) confirms that the alluvial spreads of the Niger in Mali have become a major zone of dust export.

In Senegal Avenard and Michel (1985) report that Dakar has begun to experience an increasing number of locally-generated sand and dust storms that very rarely occurred before 1970. Rural-urban migration, that has intensified during the drought years, has led to increased human pressure on the land around the satellite town of Pikine. Destruction of vegetation has destabilized the semi-fixed dunes around Pikine, and these dunes are now subject to increasing deflation.

4.4.3 SUDAN

In the Sudan the environmental degradation associated with human activities in the areas around settlements was recognised by a Soil Conservation Committee set up by the Sudan Government in 1942:

> Town and village peripheries generally are deteriorating rapidly due to overcultivation, overgrazing, to cutting trees for firewood, or to combinations of these activities. Important examples are Tokar town, Khutsum, Omdurman and Khartoum North, El Obeid and Nahud. (Sudan Government 1944, quoted in Rapp 1976).

In more recent times this localised exploitation of land around settlements in the Sudanese semi-arid zone has been accelerated and exacerbated by drought, and this environmental degradation is apparent on satellite images and air photographs as distinctly blurred, bright patches (eg

Adams & Hales 1977). In the case of more intensified land use these areas may merge into larger zones in the classic pattern of 'desert advance'.

At El Fasher, in Darfur, increasing frequency of dust storms has been directly related to the expansion of millet cultivation onto marginal land (Ibrahim 1978). Two air photographs are compared, 14 years apart, of the same area 20km north of El Fasher. The first, taken in 1954, shows millet fields covering about 24km^2 of Qoz palaeodunes, and in 1968 the same area has lost virtually all these fields and the dunes have been reactivated. Although Ibrahim related increasing dust storms at El Fasher to the expansion of millet cultivation and the decrease in rainfall, Olsson (1983) has shown his data to be dubious. Nevertheless, with a larger and more complete data base on dust storms at El Fasher than Olsson shows, the increase in activity has been proven (fig 4.9A) and although Ibrahim's statistics may be questionable, the air photographs show conclusively the type of environmental degradation that has occurred in the area. Indeed, Babiker (1982), in an article on desertification in Sudan, calls El Fasher 'the most notable example of a wind-eroded area.'(Babiker 1982, p72).

Other large towns in the Sudan where it has been observed that the **Haboob** has become more frequent in recent years, include Khartoum and El Obeid (Bakhit & Ibrahim 1982), both mentioned as suffering severe degradation in the 1940s. Khartoum has experienced the problems of urban expansion with associated tree felling, and a 'total degradation of the vegetation cover has taken place' (Babiker 1982, p74) resulting in soil deterioration and accelerated aeolian erosion. **Acacia** shrubs that were common around the city in the mid-1950s could only be found 90km south of the town 17 years later, leaving a 'man-made desert' on the clayey soil (Rapp 1976, fig 8.5.5).

Contrasting this increase in dust-raising that has been exacerbated by human action, there is some evidence to suggest that the initiation of the Gezira irrigation scheme south-east of Khartoum may have acted to depress the number of dust storms occurring over the city. The Gezira grew largely in the 1920s and 1930s on the alluvial clay plains of the Blue Nile, and the suggestion that it may have had an effect on dust storm activity at Khartoum is made by Fleming (1953) who notes that:

> Older inhabitants in Khartoum are convinced that present-day haboobs are not nearly so heavily sand-laden as those of earlier years.
> (Fleming 1953, p27).

This impression is reinforced by El-Bushra (1976) who shows that the frequency of **Haboobs** at Khartoum in the period 1916-29 was two or three times greater than during 1949-55, and these data become more significant when it is noted that the 1916-29 period was one of generally higher than average rainfall at Khartoum and stations on the Blue Nile, while 1949-55 was one of below average rainfall (Trilsbach & Hulme 1984). Although this evidence lends some support to the idea that the Gezira irrigation scheme reduced the incidence of storms from this region, the suggestion cannot be confirmed without a directional analysis of dust storms at Khartoum. Although Sutton (1931) shows that about 25% of **Haboobs** at Khartoum approached from the south-east (for 1916-29), more recent directional data for the low frequency period were not available.

Despite the marked increases in dust storm activity at El Fasher, Khartoum and El Obeid the annual frequencies at these stations, and most others in Sudan, are not great. The town of Tokar, in the eastern part of the Red Sea hills region, is an exception to this general rule. During the pre-drought period (1950-67) it averaged 37 dust storms a year and during the drought (1968-78) 47 a year. Dust storms are largely caused by thunderstorms with very high local wind speeds occurring before the onset of rain. The town is situated on a large alluvial fan at the end of a long, deep valley in the Red Sea hills. The climate is very much dependent on local topography and the valley acts so as to funnel winds, thereby yielding a higher percentage of days with winds above critical threshold velocity to entrain the concentration of fine material in the valley bottom. The early reports of the Sudanese Soil Conservation Committee, as mentioned above, included Tokar amongst the towns severely affected by land degradation. It can, perhaps, be argued that while this degradation had only a minor affect on the dust storm/rainfall relationship during the normally erratic inter-annual rainfall regime, over a prolonged period of drought these exploitative land practices have become a major factor governing the generation of dust storms in the Tokar valley.

There is a considerable body of evidence attesting to widespread desertification in the Sudan, and it appears to be closely related to an increase in dust storm activity. Although some of the degraded areas on the desert fringe may not necessarily represent long-term change in ecology, as much as a short-term vegetational response to drought, there is no doubt that severe degradation of vegetation has been affected by human activities. In some areas the clearance by human action is fairly permanent, such as the destruction of **acacia** shrubs to the south of Khartoum. As in other Sahelian areas noted above, the geomorphology of parts of the Sudanese Sahel is dominated by palaeodunes such as the Qoz of Darfur and Kordofan, and it is these areas that have become reactivated and dust-producing during the drought years.

4.5 CONCLUSION

It has been shown that a considerable increase in dust storm activity has occurred throughout the Sahel during the most recent period of intense and widespread drought. During the same period there is also a considerable body of evidence indicating that human actions have played a significant role in destroying vegetation and destabilizing surfaces in the Sahel. Evidence of documented increases in dust-raising activity occurring as a direct result of human misuse of land is not as clear-cut however. The problem of assessing whether loss of vegetation on a particular piece of land is as a result of reduced rainfall or due to the actions of human populations is an important one in this context. Nevertheless, there is evidence of significant increases in dust storm activity at settlements that have experienced severe desertification due to overexploitation of the environment. Examples include Boutilimit in Mauritania (Gornitz & NASA 1985), and several settlements in Sudan: Khartoum (Rapp 1976); El Fasher (Ibrahim 1978) and Tokar (Sudan Goverment 1944).

Although increases in dust storm activity could be attributable to other meteorological factors, such as increased wind speeds, gustiness, turbulence, alterations in circulation patterns and a higher frequency of dust storm-generating meteorological systems, it has not been possible to assess the importance of these factors due to a lack of information. Despite these problems, with a knowledge of the effects of a reduced rainfall on the wind erodibility of a soil surface, it seems reasonable to suggest that drought in the Sahel, in conjunction with poor land use practices that have been noted in many areas, is a major influential factor in the observed increase in dust storm activity in the region. If human-induced desertification is locally important in affecting increased dust storm activity, therefore, it is important to note that the increases in dust storm frequency, being based on data observed in centres of population, may not necessarily reflect increases of comparable magnitude over all areas of the Sahel.

As far as the sources of this increased dust output are concerned there is a combination of greater volumes of material being raised in existing dust source areas with evidence of major new dust sources being created during the drought. An enhanced output has occurred from the Chad Basin, as the areas of dust-raising have been extended due to loss of vegetation in the basin and falling of levels of Lake Chad. In Mauritania important new sources of dust storm sediments have been activated on the palaeodunes in the south of the country and in localised areas in Senegal. Field observations and satellite monitoring suggest that the alluvial spreads in the area of the inland delta of the Niger River have become a major new source of dust storm activity. In the Sudan, Sahelian settlements have experienced increased dust storms on recently denuded palaeodunes of the Qoz and increased activity in pre-existing dust source areas such as in the Tokar region.

The 'fixed' dune systems of the Sahel are relatively rich in fine grained sediments (eg Tricart & Brochu 1955, Barby & Carbonnel 1972, Talbot & Williams 1978, Chamard & Courel 1979) which can be assumed to be in part an accumulation of dust blown from active sources caught in the vegetation of the stabilized dunes, and in part resulting from in situ particle breakdown. In many cases this fine material forms a hard coherent crust on the dune surface, but devegetation in drought periods and through human clearance exposes such crusts to agents of breakdown that may include the trampling of animals and the ploughing of soils for cultivation, and thus the fines become available for aeolian entrainment. Alternatively, these fines may in places be washed out of fixed dunes as they are subject to gullying, depositing the fine material in interdune depressions in pools and marshes and/or alluvial fans (eg Talbot & Williams 1978, Barby & Carbonnel 1972, Barth 1982). The effects of drought and human action on such deposits is to desiccate and devegetate them leaving ready sources of fine material for aeolian suspension.

In addition to formerly fixed dune systems acting as sources for dust storm sediments there are other modern and ancient geomorphological units rich in dust sized material. These include closed depressions such as in the Chad Basin, alluvial spreads of rivers such as the Niger and lacustrine sediments of former areas of lake bed in the Lake Chad area.

There are many environmental consequences of the dramatically enhanced aerosol output from the Sahel, perhaps the most immediate of which is the loss of soil and nutrients from agricultural lands. There may also be significant effects on weather and climate due to the presence of dust in the atmosphere. Dust is important to the atmospheric radiation budget, and has direct effects by changing the fluxes in a cloud-free atmosphere by scattering and absorption of radiation, and indirectly by modifying the optical properties, particularly the albedo of clouds (Junge 1979). Perhaps the most significant meteorological effect of these influences for the Sahel is their enhancement of stability in the planetary boundary layer. The base of the Saharan Air Layer is defined by a persistence of a temperature inversion that is still recognisable on meteorological soundings from the Caribbean and in Miami (Carlson & Prospero 1972). The dust plume over northern Nigeria is also often associated with a pronounced temperature inversion (Kalu 1979, 1982), and although this night-time phenomenon is a climatological feature resulting from radiation cooling from the ground surface, the presence of dust in the atmosphere may reinforce the feature and feedback on surface conditions. Such stability has important consequences by its inhibition of convection and potential rain-making processes over the Sahel.

Large-scale dust-raising activity is postulated by

Bryson and Baerreis (1967) to be responsible for a major aridification of north-west India, due to the basic mechanism of mid-tropospheric subsidence, and Idso (1981) feels that there is considerable evidence to validate their theory.

In addition to the stabilizing effect on convective processes, the presence of large amounts of atmospheric dust may further inhibit the development of convective systems by supernucleation. Some of the oblique views of the Sahel provided by the Space Shuttle astronauts have shown very poor development of cumulonimbus clouds within a dust plume area. The presence of large quantities of dust in the atmosphere over longer periods may modify the type of rain-producing clouds and thus the size of rain drops according to Maley (1982). Decreased rainfall may also result indirectly through a reduction in the quantities of biogenic ice nuclei (important for rain drop formation) reaching the Sahelian atmosphere. Reductions in these quantities have come about due to the loss of vegetation coverings by natural processes during the drought but more particularly due to widespread overgrazing before the main onset of the drought (Schnell & Brine 1975).

It seems, therefore, with the knowledge of the increased dust output from the Sahel indicated here, that such dust-induced climatic feedback processes are as worthy of attention as those due to albedo changes (Charney 1975, Idso 1977, Wendler & Eaton 1983) and soil moisture properties (Walker & Rowntree 1977).

The accumulated evidence presented in this chapter confirms the observations of Mainguet (1984) who notes that the Sahel is currently becoming an area of reactivation and export from its legacy of Quaternary sand dune and other accumulations of dust sized sediments. This phase, that is so closely linked to drought, human actions and suffering in the area, may also be acting to prolong itself.

CHAPTER FIVE : THE MIDDLE EAST

5.1 INTRODUCTION

The Middle East, an area characterised by aridity and continentality, is here defined as comprising the territories of the Arabian Peninsula countries, with Syria, Lebanon, Jordan, Israel and Iraq. Some reference is also made in this chapter to Iran, but for the most part that country is covered in chapter 6.

The Arabian Peninsula has been identified by Idso (1976a) as one of the five major world regions where dust storm generation is especially intense, and Prospero (1981) reports that a major zone of dust haze can be observed in the Arabian Sea during June, July and August (see McDonald's haze maps: fig 3.1).

The data for this chapter were derived from a number of sources: Hinds and Hoidale (1975), International Aeradio plc (IAL), who have a number of meteorological observers contracted to governments on the Arabian Peninsula, and some national meteorological services. Hinds and Hoidale's data cover periods of varying length, but all for more than seven years, they are in the form of numbers of dust events (both dust storms and blowing dust with visibility <11km), and also show the diurnal variation of the occurrence by month. The data from IAL are for 10 years or longer, dating from the early 1970s and are as dust storm days. Data from the various meteorological services, in dust storm days, cover the following periods. Israel: 10 years from 1961; Jordan: seven years from 1961; Syria: 20 years or more from the late 1950s, and Oman: four years from 1980.

Figure 5.1, therefore, gives a general indication of the distribution of frequencies in a region in which information is relatively sparse. The analysis of seasonality below is very largely based on data for dust storms, although in areas where events of such intensity are rarely experienced, data for blowing dust have been used. The investigation into diurnal patterns of occurrence was conducted using the data for blowing dust.

5.2 FREQUENCY, DISTRIBUTION & SEASONALITY

Figures 5.1 and 5.2 show the distribution of annual frequencies of dust storms and blowing dust respectively in the area. For both visibility classes, the highest number of episodes occurs on the alluvial plains of southern Iraq and Kuwait. Nasiriyah has the largest average number of occurrences of both dust storms (33 a year) and blowing dust (208 a year). This part of the Middle East region was identified by Grigoryev and Kondratyev (1980) as one of the major world source areas of dust.

To investigate the seasonal variation of dust-raising events, figure 5.3 was constructed to show the percentages of stations with a maximum frequency of dust events during a particular month. The figure shows a bimodal distribution, with 28% of stations experiencing maximum monthly dust storm frequencies in March, and 22% in July. During the months of March, April, May, June and July 85% of stations experience maxima. A map was also drawn to show the distribution of primary seasons of dust storm occurrence (fig 5.4). Here, the season is a span of three consecutive months in which dust storm frequency is highest. The map shows a number of areas of different seasonality, and characteristic monthly percentage frequency distributions of occurrence of the areas are depicted in figure 5.5.

Further analysis will be by areas, divided according to the initial patterns indicated above :

1. Lower Mesopotamia.
2. The Arabian Peninsula.
3. Northern Areas.

5.2.1 LOWER MESOPOTAMIA : THE SHAMAL

The alluvial plains of the rivers Tigris and Euhrates are the source of the most important dust flow in the Middle East. Material is raised in this region and transported in a southeasterly direction across Kuwait, the eastern coastline of Saudi Arabia and over the Arabian Gulf, reaching as far as Dubai and Sharjah (plate 5.1). Although this activity occurs at all times of the year it is at a maximum during May, June and July when mean wind speeds are at a maximum in southern Iraq (Jawad & Al-Ani 1983), so that the major part of the lower Mesopotamian plains and the west coast of the Arabian Gulf as far south as Qatar experience the same primary season of dust events (figs 5.4, 5.5 for Baghdad).

The prevailing northwesterly wind that carries the dust is known as the **Shamal** ('North' in Arabic), or **Simoom** in Kuwait. The **Shamal** occurs as a result of the Gulf area being strongly influenced by the heat low over Pakistan and Afghanistan and a consequent trough to the lee of the Zagros mountains. The topographical features of central Saudi Arabia, with gradual upsloping terrain to the west and the presence of a semi-permanent high over northern Saudi Arabia tend to create an enhancement of the low level flow, particularly below about 1500m (Membery 1983). Grant (1983) shows the strength of this flow in July to be 10-12m/s over Kuwait and southern Iraq at 900mb (about 1km). The **Shamal** may raise the dust itself when strong solar heating creates turbulent flow and strong convectional upcurrents in an unstable boundary layer, and activity is, therefore, greatest during the afternoon hours (see fig 5.7 for Basrah). Conversely, meteorologists in the Gulf area suggest that the raising of the dust in southern Iraq, and northeastern Saudi Arabia is often caused by strong winds resulting from rapid pressure rises behind eastward moving troughs of low pressure that track from the Mediterranean across Turkey or northern Iraq (Aspin, pers. comm. 1983; Jackson, pers. comm. 1984).

The dust, once raised, is then advected down the Gulf by the prevailing **Shamal.** When the **Shamal** is strongest (May, June and July) it may also generate local storms in the Gulf area.

Total Suspended Particulate levels can be very high during **Shamal** episodes, Khattack (1982) noted values of nearly 3000 microgrammes/m^3 during intense **Shamals** in Dhahran, while the mean value during a five-month monitoring period was 339 microgrammes/m^3. Membery (1985) reports that rapid deterioration in visibility occurs between about 0700 and 1000 local time at Kuwait, and usually continues until 1600 local time. Maximum dust haze at Bahrain, some 400 to 500km downwind, is usually experienced around 0400 (Houseman 1961), and at Sharjah minimum visibility usually occurs around dawn (Jackson, pers. comm. 1984).

Plate 5.2, taken on 16 June 1983, shows plumes emanating from southern Iraq and crossing Kuwait and north-east Saudi Arabia to the Arabian Gulf. This image was taken midway through the most intense phase of long-range transport of the **Shamal** in 1983, as indicated by a year-long collection of airborne dust in Qatar, just north of Doha (Wimpol 1984). The programme sampled at 1.5m above ground level using a simple collection method by each of the four main compass point directions. During the five-week period from 1 June to 7 July, 5147 g/m^2 dry weight was collected from all directions, and the clear dominance of **Shamal** dust loads was indicated by the directional breakdown of these samples: 67% from the north; 24% from the west; 7% from the east, and 2% from the south. Powell (pers. comm. 1985) confirms this impression but notes that some material during this period was derived from local sand and dust storm events. Williams (pers. comm. 1984) notes that the north of the Qatar peninsula has a generally hard, rocky surface, with areas of loose, fine sand and dunes mostly confined to the southern half of the country, south of Doha.

The variation of monthly dust storm frequency with rainfall totals and DPs at Dhahran on the Gulf coast of Saudi Arabia is shown in figure 5.6A. There is a clear peak in dust events in June at the same time as a peak in DP, with low dust storm activity occurring during months of low DP and/or seasonal rainfall, although rainfall totals are very low. This apparently clear relationship is also evident using annual totals (fig 5.6B), when years of high dust storm activity correspond to years with high total DP (from Fryberger **et al** 1984) and low annual rainfall. Since Dhahran derives a large proportion of its dust events from southern Iraq, however, this apparent correlation should be treated with some caution. Some confusion may be incorporated into the data, in that when thick dust haze (as defined in this thesis) arrives at stations in the Gulf, with high wind speeds (Aspin quotes 17 knots or more for the UAE) then the event will be recorded as a dust storm, even though the material may not necessarily have been entrained locally. Thus, the dust storm day data for Dhahran may include both locally derived dust storms (from dunes, sebkhas and other desert surfaces) and also **Shamal** thick dust haze.

A comprehensive study of dust and dust storms in Kuwait is presented by Safar (1985). Kuwait is affected by both locally raised dust and **Shamal** material, and Safar notes that dust in the atmosphere (as dust storms, rising and suspended dust) is present in the atmosphere of Kuwait for 25% of the daytime throughout the months from April to August. Dust storms occur on an average 27 days a year or for an average 128 hours a year. Khalaf and Al-Hashash (1983) note that the local name used generally for dust and sand storms in Kuwait is **Toze.**

Occasionally the Gulf may also be affected from the south-east by dust raised in Iran and Pakistan (Otterman **et al** 1982), and a minor maximum in dust activity occurs in March in southern Iraq and Kuwait, when dust is raised by southeasterlies blowing from the Arabian Gulf which commonly develop in winter and spring during the passage of Mediterranean depressions. In April in Kuwait and southern Iraq, thunderstorm clouds usually appear in the afternoon and during the night hours, and are occasionally accompanied by severe dust storms during which visibility may drop to zero (Al-Najim 1975, Khalaf & Al-Hashash 1983). In an unusual circumstance, a severe thunderstorm originating over northern Saudi Arabia moved southeastwards down the Gulf in May 1983 as a 'front' or squall line, producing dust storms in the strong winds accompanying it as far south as Abu Dhabi (Membery 1985). The system was characterised by unusually strong winds, a pressure surge of 1.5mb in minutes and a very rapid drop in visibility.

5.2.2 ARABIAN PENINSULA

The larger part of Saudi Arabia, the Trucial States excepting Qatar, and the western half of lower Mesopotamia experience dust storm activity throughout the year, but a maximum occurs during the spring months (March, April and May). The percentage frequency distribution for Jeddah (fig 5.5) is typical of stations in this area. Dust haze in these parts is caused by a variety of mechanisms; locally squalls, thunderstorms and dust devils raise material, and the western side of the area may be affected by dust haze advected from the Sahara. Aspin (pers. comm. 1983) reports that the majority of dust events at Abu Dhabi, Dubai and Sharjah are as haze brought by the **Shamal,** following the passage of Mediterranean depressions across southern Iraq. A minimum visibility of 100-200m is not uncommon in severe cases of this kind at these stations.

Synoptic scale dust storms may also be generated by cyclonic gyre in a surface heat low brought about by intense solar heating of desert surfaces. Wolfson and Matson (1986) show an example of such a comma-shaped storm in north-east Saudi Arabia, although this particular storm was not detected by the sparse network of ground surface meteorological stations. This storm only became visible on thermal satellite imagery at night-time, as the ground surface cooled faster than the dust cloud top. During daylight the system was virtually undetectable on visible and thermal bandwidths as the albedo of the dust cloud was very similar to that of the ground surface.

A summer maximum (June, July & August) of dust storm activity occurs in the region that comprises the south-east-facing coast of the Arabian Peninsula and the southern half of the Red Sea coast (see fig 5.5 for Gizan). Dust storm activity is largely concentrated in six months of the year. Little dust haze is experienced from October to March, the season of the north-east monsoon, although from mid-December to mid-March the dusty **Belat** wind blows off the hot, dry interior of the Rub' al Khali. Grant (pers. comm. 1986) suggests the **Belat** may be associated with southward moving fronts that pass Masirah every two or three days at this time of the year.

The climatology of this region is not well known, but **Khamsin** winds blow on the coast during the south-west monsoon season (June to September), often raising clouds of red sand and dust during daylight hours. Land squalls are not infrequent during this time, and these squalls raise thick banks of dust haze, especially where the coast is low (Keeton 1928). Russell (1879) described a sand storm at Aden in July 1878 when clouds of black, yellow and orange sand obscured visibility for nearly an hour, commencing at 1730 local time with a westerly wind:

As the storm now burst upon us, we ran inside and closed the doors. Almost immediately the room became so dark that we could not see a yard before us.
(Russell 1879, p48).

The dividing line on figure 5.4, between this seasonal area and the remainder of the Arabian Peninsula, is closely aligned with the position of the zone of intertropical convergence over the Middle East during the summer months. Grigoryev **et al** (1971) have identified a large-scale dust/sand flow across this south-east coast of Arabia from Zond 7 satellite imagery taken in August, and it appears that the dust haze may originate in the arid and semi-arid areas of the Horn of Africa.

It has been shown that the Horn of Africa experiences maximum dust-raising activity in the summer months (see section 3.4.1), and it is suggested that this flow, charged with material entrained over arid eastern Africa, continues over the southeastern coast of Arabia where it picks up more sand and dust. This is the origin of the haze observed over the Arabian Sea in the vicinity of this coast during June, July and August (fig 3.1). This suggestion is given some validity by Grant's (1983) analysis of vector mean winds over the region for July, which shows a strong flow at 900mb (about 1000m) from the Horn of Africa northeastwards towards the Indian coast at about 20^{o}N. Further support is given by

Space Shuttle photographs over this region which clearly illustrate this flow in operation (plates 3.4 to 3.6).

Data provided by the National Meteorological Service of the Sultanate of Oman reveals a little more information on the dust system of this area of the Arabian Peninsula. Table 5.1 shows the hourly weather observations for a part of 26th April 1981 at Seeb International Airport (Muscat). During the morning visibility is generally good (20km) and winds of around 10 knots are blowing from the NNE or north-east, while pressure is gradually falling. At 1150 local time a thunderstorm is recorded (Present Weather code = 17) and at 1215 a sudden severe reduction in visibility occurs (from 10km at 1150 to 500m), caused by dust raised in the thunderstorm downdraft (Present Weather code = 98). Winds have shifted to blow from the south-west at 30 knots, with maximum gusts of 40 knots. Twenty minutes later the wind has switched once more and is reduced to 18 knots, the visibility has improved to 10km and the present weather codes (95) record a thunderstorm with some rain at the time of observation . At 1350 there is a second severe downdraft dust storm, again associated with a sudden reduction in visibility, switch in wind direction and increase in speed. For this second storm pressure and temperature readings were taken, showing a fall in temperature and rise in pressure, in the classic **Haboob** sequence as described by Freeman (1952)- see section 3.3.5A. An hour later pressure has continued to rise while temperature is still falling, visibility is greatly improved and wind speed has dropped, with another change in direction, to blow from the WSW.

Surveys of the daily meteorological observations for days with dust storms in Oman show that thunderstorm downdraft dust storms also occur at Buraimi and Masirah Island. At Masirah a particularly prolonged dust storm reduced visibility to <1000m for 29 hours on the 1st and 2nd July 1981 with wind speeds gusting to more than 40 knots from the SSW, and at Thumrait on 27 March 1982, a storm blowing from the south with speeds greater than 44 knots reduced visibility to zero for eight hours.

Thunderstorm downdraft storms may also occur over the Yemeni countries; Spurr (1927) reported a dense dust storm off Perim Island (in the straits between Djibouti and Yemen) that struck in the afternoon from the north, reducing visibility to <1000m for 10 minutes with winds of force eight, with no rain but several flashes of lightning from cumulonimbus clouds.

A factor considered to be of importance in the generation of convective activity over the Arabian Peninsula in non-summer seasons is the effect of the subtropical jet streams over the Middle East. Taha **et al** (1981) mention the lower tropospheric convergence associated with upper tropospheric divergence around the jet stream which generates thunderstorms, dust and sand storms. Fisher (1978) quotes instances in the southern Arabian deserts, in spring, of small living creatures such as fish and frogs being carried upwards with sand and dust in the convective currents, to descend later in 'mud rain'.

5.2.3 NORTHERN AREAS

A bimodal maximum of activity characterises an area that includes the Negev, Jordan, western and northern Iraq, and the northernmost part of Saudi Arabia. Here maxima occur in spring and winter (see fig 5.5 for Turaif). The area can be thought of as a corridor since dust storms are largely caused by depressions moving eastwards from the Mediterranean, generating rain, dust and thunderstorms. These depressions are characteristic of spring (March & April) and winter (December & January), and this seasonal pattern is maintained from Be'er Sheva in the northern Negev (Katsnelson 1970) to Mosul in northern Iraq (Coles 1938). The low frequency of dust storms in northern and western Iraq during June, July and August as compared with a maximum of activity at Baghdad and southwards during this period is due to the pressure distribution. As Coles (1938) explains, northern and western Iraq are situated in a region of relatively high pressure at this time, just outside the influence of the Iranian low, so that strong northwesterly winds and maxima of summer dust storms are therefore confined to an area roughly south-east of a line between Ramadi and Kirkuk.

Flower (1936b) studied sand devils as observed at stations in Palestine, Trans Jordan and Iraq. He noted that their season of maximum occurrence is from June to September in these areas, with a diurnal maximum during daylight hours. The highest mean number of sand devils was 18 at Amman. Data are also given on direction of rotation, dimensions and duration of events.

The months of spring are the primary season for the remaining areas of the Levant. Dust is transported into the area in the south by **Khamsin** depressions. The dusty desert winds associated with these depressions are known as **Sharav** in Israel, although this term is generally used for hot, dry conditions (it is derived from an old Hebrew word meaning 'heat of the land') and the pre-frontal dust wind is just one of the meteorological situations that may bring them on (Winstanley 1972). Winds blowing from the south-west, south or south-east also raise dust locally, due either to thermal lows developing or intensifying over Africa or a deepening of the Sudanic trough (Joseph **et al** 1973). Occasionally dust storms may occur in the Negev from the east (Yaalon & Ginsbourg 1966). TSP levels during **Sharav** conditions and sand storms have been monitored by Kushelevsky **et al** (1983) at Be'er Sheva who show them to be particularly high (890+/-250 microgrammes/m^3), and this material is often rich in sulphates (Mamane **et al** 1980, Nativ **et al** 1985).

Three types of dust in the atmosphere were identified in Syria by Combier (1935): sand haze caused by **Khamsin** weather in the spring; dust devils occurring especially in summer and sand storms that occur in desert regions all year round. In the eastern desert of Syria the main period of activity is slightly later, in May, June and July. Sivall (1957) reports that the number of different fever diseases and especially neuroses reaches a marked maximum at the American Hospital in Beirut during days when the dusty desert winds are blowing.

5.3 METEOROLOGICAL SYSTEMS GENERATING DUST EVENTS

From the above discussion, the meteorological systems that generate dust-raising in the Middle East can be classified as follows.

1. Convective weather systems - single cell and severe thunderstorms.
2. Passage of frontal systems.
3. Pressure gradient winds associated with more or less permanent pressure systems.
4. Cyclonic gyre from intense low pressure systems.

5.4 DIURNAL VARIATION OF DUST EVENTS

A number of distinctive diurnal patterns of dust events were discerned from the summaries of three-hourly observations for blowing dust, and these patterns were classified into four types: daylight maximum; afternoon maximum; 24 hours constant, and random. Examples of these four diurnal frequency types are shown in figure 5.7.

The daylight maximum in dust occurs in many locations all across the area, and is slightly more common than the afternoon maximum. Riyadh, Saudi Arabia, illustrates this diurnal variation. Highest dust-raising activity occurs during the daylight hours when intense solar heating of the ground surface creates a high degree of turbulence and very strong pressure gradients locally, producing a range of small-scale dust events. The afternoon maximum in activity is illustrated by Basrah, Iraq. This pattern of diurnal variation occurs when turbulent mixing is more pronounced than at other times of the day and the atmospheric boundary layer is deep.

The 24 hours constant dust activity is illustrated at Aden, People's Democratic Republic of Yemen (formerly South Yemen), and is characteristic of stations on the Arabian Peninsula's south-east-facing coast. It has been noted above that the lack of climatological data for this area makes detailed analysis of small-scale regional patterns difficult. However, the suggestion that this area is affected by the large-scale dust flow from eastern Africa in conjunction with the south-west monsoon is supported to some degree by this diurnal pattern. The random diurnal pattern is illustrated at Tehran, Iran. This diurnal variation in activity lacks any significant regular peaks in dust frequency and is to some extent a reflection of the small scale of the activity.

5.5 VARIATIONS THROUGH TIME

Some data covering a sufficiently long period were avaiable for three stations located in the area largely affected by the **Shamal** dust flow. Figure 5.8 shows the variation in annual numbers of days with visibility obscured to <1000m by dust at Kuwait (1962-84), Bahrain (1946-83) and Doha (1962-83). The graphs seem to indicate a gradual decline in frequency at the two Gulf stations. It should be remembered, however, that in these readings there is no distinction made between dust raised locally and that advected from other areas (largely from northern Saudi Arabia and southern Iraq in these two cases), so that it is impossible, from these data alone, to draw any conclusions as to possible decreasing dust storm frequency on either the west coast of the Arabian Gulf or the lower Mesopotamian Plain. Indeed, at Kuwait, nearer to the northern sources, but also affected by local dust-raising, no trend is discernable. Nevertheless, it is worth noting that in south-east Iraq some large areas on the Iranian border have been flooded since the start of the Iran/Iraq war, which may reduce **Shamal** dust loads in future.

Another factor that has had an effect on atmospheric dust loads, particularly on the Arabian Peninsula, is the industrial development that has taken place since the mid-1970s. Development of industry and the concomitant expansion of urban areas destabilizes desert surfaces producing increased local dust-raising during the construction phases, and once established, industrial pollutants contribute to atmospheric dust loadings locally (Behairy **et al** 1985). Although such activities are highly localised and often on a small scale with respect to the vast areas of desert unaffected by human activities, they become more significant when it is remembered that meteorological stations are, for the most part, situated within or on the outskirts of settlements. Thus, industrial and construction activities may take on a disproportional significance in their local effects on dust events as recorded by meteorological observers.

5.6 LONG-RANGE TRANSPORT

The most important source for long-range transport in the Middle East is the alluvial floodplains of the Tigris and Euphrates in the area of lower Mesopotamia in southern Iraq, the source of the **Shamal** dust flow (Khalaf **et al** 1985). The dust fallout over Kuwait and southern Iraq can be described as quartzitic calcareous sandy silt (Khalaf **et al** 1985) and it is clear that **Shamal** dust is an important contributor to sediments in the norhwestern part of the Arabian Gulf (Sugden 1963, Khalaf & Al-Hashash 1983, Al-Bakri **et al** 1984). Dust fallout rates over Kuwait monitored by Khalaf & Al-Hashash (1983) vary between mean monthly maxima of 1002.7 tonnes/km^2 (July 1979) and minima of 9.8 tonnes/km^2 (November 1979) and estimates of sedimentation rates in the northern Arabian Gulf, based on fallout over Kuwait, are about 1mm/yr (Khalaf & Al-Hashash 1983, Foda **et al** 1985).

The contribution of **Shamal** dust to sedimentation in the Indian Ocean is more difficult to assess, as suspended dust in this area is also derived from the Arabian Peninsula and eastern Africa. **Shamal** dust certainly reaches the UAE states of Dubai and Sharjah where it is the major cause of dust events (Aspin, pers. comm. 1983). It is interesting to note, however, that these stations experience dust event maxima in spring (March) as shown in figure 5.4, despite the fact that the **Shamal** is strongest and most dust-laden in the summer months. This observation is confirmed by Jackson (pers. comm. 1984) who states that:

> The 40 day Shamal (May/June/July) affects much of the Gulf between Kuwait and Abu Dhabi but it is less noticeable in the northen Emirates of Dubai, Sharjah and Ras Al-Khaima.

This apparent inconsistency has a number of possible explanations if Aspin's observations on the importance of **Shamal** dust are accepted (the northern Emirates also experience dust events from local sources and from the Makran coast). It is possible that **Shamal** dust loads are largely deposited between Qatar (with a maximum in June/July) and the northern UAE states, but this seems unlikely as no mechanism for such large-scale deposition is apparent. Perhaps a more feasible exlanation involves a shift in the axis of flow of material transported in the **Shamal.** Grant (1983) shows the major vector streamline at 900mb in July to flow almost directly down the Gulf coast and continue as a northwesterly inland at about 24°N, 52°E, thus passing well to the south of Dubai and Sharjah. No similar data for March are available, but it may be that this July maximum flow, blowing from lower Mesopotamia from about 330°, represents a shift from an axis of flow closer to 310° earlier in the year which would result in large concentations reaching the northern Emirates in March even though those concentrations are not as large as in summer.

The other important long-range dust flow identified in the Middle East has its origin in eastern Africa and crosses the south-east-facing coast of the Arabian Peninsula. This flow is contributed to by dust raised in this part of the Arabian Peninsula and this flow seems to be the source of the seasonal maximum in dust haze identified by McDonald (1938).

Although there have been a few studies of aerosols over the Arabian and Red Seas (Cailleux 1961, Aston **et al** 1973, Sadasivan 1978, Prodi **et al** 1983) none have been conducted during the seasons of maximum dust flows as indicated here. Future investigations of aerosols during these seasons would make significant contributions to the identification of specific source areas.

Dust from Arabia may be transported over much greater distances if Prospero's (1981) speculation that material in the upper westerlies reaches the South China Seas is correct. Material from the Middle East is also carried to the central Asian states of the USSR in depressions that travel across Iraq and Iran (see chapter 7). In addition to material from the Horn of Africa reaching the Arabian Peninsula, dust is transported across the Red Sea from Sudan and Egypt, from northern Libya and Egypt to the Levant and from the Makran coast of Iran and Pakistan to areas on the Gulf coast.

5.7 CONCLUSION

Virtually the whole of the Middle East is subject to dust storm activity, a conclusion to be expected as the region is very largely arid or semi-arid. The most intense area of dust-raising occurs on the alluvial floodplains of the Tigris and Euphrates in lower Mesopotamia, southern Iraq. This material is transported down the Arabian Gulf by the prevailing northwesterly **Shamal** wind with a strongest dust-moving season in the summer months. The **Shamal** is a pressure gradient wind that flows between a high pressure cell located over north-east Saudi Arabia and a low pressure trough, that is part of the heat low over Pakistan and Afghanistan, to the lee of the Zagros mountains. The **Shamal** itself may raise material in lower Mesopotamia or transport dust raised by passing frontal lows that cross the northern part of the area. These depressions, originating in the eastern Mediterranean, are the most important dust-raising system in the north, from the Levant to northern Iraq, where they are most common in the spring and winter.

Other significant meteorological systems that generate dust storms are convective cells that produce thunderstorm downdraft storms in Iraq, Saudi Arabia and Oman. Some of these thunderstorms are associated with lower tropospheric convergence due to upper level divergence around the subtropical jet stream over the Middle East.

For the most part, dust storms are characteristic of the spring and summer months in the Middle East when rainfall is low and wind speeds are high. Observations show that the highest frequencies of airborne dust occur during daylight hours , with some areas experiencing an afternoon maximum.

The most important long-range transport from the area is the **Shamal** dust flow which makes a substantial contribution to the sedimentation of the Arabian Gulf. Material is also transported across the north of the area, to the Asian states of the USSR, by depressions tracking from the eastern Mediterranean. Dust is transported into the area from the Makran coast and northeastern Africa, across the Red Sea. The dust haze experienced off the south-east coast of Arabia from June to August is related to a large-scale dust flow that appears to originate over the Horn of Africa, and is part of the south-west monsoon circulation.

CHAPTER SIX : SOUTH-WEST ASIA

6.1 FREQUENCY, DISTRIBUTION & SEASONALITY

Data for the countries of south-west Asia, here defined as Iran, Afghanistan, Pakistan and India, were obtained from national meteorological yearbooks with additional information supplied by the national meteorological organisations. For all stations frequencies of dust storm days were averaged from at least seven year's data, although unfortunately the data periods were not all concurrent. The geographical distribution of stations used is shown in figure 6.1; only those specifically mentioned in the text are labelled.

Figure 6.2 shows the distribution of average annual numbers of dust storm days throughout the region. In the west a number of high frequency areas can be identified: around Hamadan in western Iran (34.4 dust storm days a year), Abadan and Dezful on the lower Mesopotamian plain (43.1 and 40.0 respectively), Sabzevar in north-east Iran (21.0), Yazd in central Iran (24.0), and on the south-east coast of the Arabian Gulf around Bandar Abbas (23.1) and Jask (27.3). The highest frequencies in the region occur at the convergence of the common borders between Iran, Pakistan and Afghanistan. Zabol, in Iranian Seistan, has the highest annual frequency in the region of 80.7 dust storm days. In northern Afghanistan on the plains of Turkestan a high frequency area is centred on Chardarrah (46.7), Kabul records an average of 20.0 dust storm days a year, Sardi Ghazni and Ghazni record 26.3 and 19.3 respectively, and Jalalabad (24.4) is situated on the western limb of a high frequency area that extends across the upper Indus plains of Pakistan to Jhelum (18.9). Fort Abbas in Pakistan (17.8) and Ganganagar in northwestern India (17.0) constitute a high frequency region in the north of the Thar desert.

The seasonality of dust-raising activity is illustrated by table 6.1 which shows percentage frequency of dust storm days by month. This indicates that spring and summer months are the time of greatest activity. Detailed analysis of frequencies and related meteorological and pedological influences are now presented on a country by country basis.

6.2 IRAN

Dust storms occur in all parts of Iran. In the north few storms are experienced on the coast of the Caspian Sea due to high precipitation, annual average rainfall reaching 1400mm in some places; two or three dust storm days a year take place in north-west Iran (Azerbaijan) occurring largely during the months from July to October. An area of high frequency stretches from Hamadan in the west to Ghavzin and Tehran in the east, the main season of activity here being April to July in association with eastward moving depressions originating in the Mediterranean. South of this area Abadan, on the lower Mesopotamian plain, is affected by the prevailing **Shamal** wind, a northwesterly that blows material down the Arabian Gulf reducing visibility as far as Jask in south-east Iran. The primary season for the **Shamal** is May to July (see chapter 5 above).

North-east of Abadan and south of the Elburz Mountains the Dasht-i-Kavir or Great Salt Desert experiences one or two dust storm days a year, the aggregating effect of salts on available erodible material perhaps being a factor in the relatively small number of storms here. Yazd, in the central desert of Iran, records a maximum of activity from February to July. Fookes and Knill (1969) estimate the wind velocities of dust storms at Yazd to reach 17m/s, storms may last for up to 24 hours and can reduce visibility to 20m. Loess covers several hundreds of square kilometres of the inter-montane desert plateau in this region.

Relatively high dust storm frequencies are recorded in the mountain valleys of north-east Iran, on the border with Turkmenistan (USSR), the season of maximum activity being April to September. Lieutenant-Colonel C. E. Yate described a dust storm that prevented him from leaving Kuchan in the valley of the Atrek River for seven days in September 1894, the storm blowing during daylight hours, down the valley from the east and up the valley from the west on alternate days.

> One could neither read nor write. Everything was so quickly covered with dust that one could hardly tell

> writing-paper from the table-cloth. It was impossible to make a pen travel on paper, and often the only place where I could find refuge was in bed. I was told these storms were of frequent occurrence. (Yate 1900, p178).

A foehn-type wind, called **Garmsil,** meaning hot wind or hot storm, blows from these mountains northward over Turkmenistan, where it often raises large quantities of dust (see chapter 7 below).

The Afghan frontier region is the area affected by the **Sad-ou-bist bad** (Wind of 120 days), a northerly wind that can blow at speeds of over 28m/s for days on end, strips vegetation of leaves and can cause extensive structural damage to buildings by corrasion from its load. Colonel Maitland, a member of the scientific body accompanying the Afghan Delimitation Commission in northern Afghanistan in the mid-1880s, described the wind as experienced in north-west Afghanistan, at Herat:

> In May, June and July the wind is very strong, being always highest after sunset, and if there should be rain in the hills to the north or a shower in Badghis, it increases almost to a hurricane. The wind is felt far to the south and blows with unabated fury in Sistan. (Maitland 1891, p527).

Statistics show that the Seistan Basin is the area of highest dust storm frequency in south-west Asia, where Zabol records an average 80.7 days a year, and the seasonality at this station confirms Maitland's observation that May, June and July are the most active months. Maximum winds at Zabol occur in late May and June, though winds are still strong in July (Grant 1983). A review of the literature on the basins of the south-east Iranian highland has been combined with his own personal observations by Stratil-Sauer (1952) to show that the **Sad-ou-bist bad** occurs in the same or similar form in these areas. On the northern border of the salt pan of Hamum-i-Mashkel and in the desert of Lut similar summer storms can be found which, though not so strong and lasting, occur also in other basins of east Iran, although a direct connection between all these violent air movements was not found. He suggests that these summer storms, developing from the north and north-west winds predominant in the rear of the great Iranian-Indian summer depression, are produced in precisely limited areas by large air masses precipitating into the overheated basins where the air has become thinner, and are thus caused by morphological conditions. Conversely, these storms themselves influence the morphological conditions by their strong erosion of the region's sediments, which have accumulated as the sandy desert of south-east Iran.

Analysis of the available upper air data over the region shows that there appears to be a:

> ...strong stream of air flowing towards the south from Cardzou in Soviet Turkmenistan through the gap in the chain of mountains to the north-west of Herat, along the corridor of low-lying desert (mostly below 1000m in altitude) to Zabol, then decelerating after crossing the line of hills to the north of Nokkundi.
> (Grant 1983, p4).

South-west of the Lut Desert, Kerman is protected by a small mountain range and is not consistently affected by the **Sad-ou-bist bad,** Beckett and Gordon (1956) report that strong winds from the south or south-west, generated by the passage of depressions, cause fierce dust storms here during July to September.

A further region of high frequency (over 20 dust storm days a year) is on the southern coast from Bandar Abbas to Jask, this region is affected by **Shamal** dust haze as well as local storms in the desert interior. The season of maximum activity lasts from February to July.

The general name for the hot, dry wind of the deserts and semi-deserts of Iran and Arabia is **Simoun** (Poison wind), and dust storms are often raised by it. A vivid description of a **Simoun** over southern Persia is given in the work of Schaubert (1897). Djavadi (1965) notes that the local name in Baluchistan is **Badé Alvar.**

6.3 AFGHANISTAN

Extensive dust-raising activity occurs on the lowlands of the north, south and east, areas that are primarily composed of Quaternary deposits, including alluvial fans, playas and sand dunes, with widespread loess coverings locally. In the mountains of the Hindu Kush dust storm activity is generally at a much reduced level, and in some places they are not experienced at all.

The highest frequencies of dust storm occurrence are to be found on the desert plains of Seistan and Registan in southern Afghanistan. Deshoo, in Helmand Province on the south side of the Dacht-i-Margo, has the highest recorded average annual number of dust storm days in this region, 32.6, and Bust records 30.1. The area receives less than 200mm of rainfall a year. Salem amd Hole (1969) describe the soil at Bust as a sandy loam with a discontinuous desert pavement. Darweyshan, further downstream on the Helmand River, is in an area of silty clay loam alluvial soil and records 18.2 dust storm days a year. The main season of activity here is the dry season, beginning in March or April and continuing until August (fig 6.3A). The area along the Iranian border, from Herat to the Seistan basin is in the path of the **Sad-ou-bist bad** that blows from late May to mid-September (see above). The trajectory of this wind describes an arc, blowing due south from Herat and around the southern margins of the Dacht-i-Margo to rage from the south at Darweyshan. Its erosive power is great and clearly defined, a distinct zone of desertified land 1.5 to 3km in width can be easily identified in the surrounding landscape along its trajectory (Visaie & Massaudie, pers. comm. 1984). In Seistan houses are built with dead walls facing the wind. Dust devils also commonly occur on the plains of southern Afghanistan, Humlum (1959) describes them as rising in the midday heat, measuring up to several metres across and travelling with a velocity of 14 to 28m/s or more, often disappearing as soon as they are formed.

North of the Hindu Kush, from Herat in the west to Faizabad in the east the dry season starts somewhat later, dust-raising activity occurring largely from June to October (fig 6.3B&C). Within this area the plains of Turkestan, roughly from Andkhoy to Faizabad, experience particularly intense activity, a maximum being on the Chardarrah-Kwaja Ghar stretch of the Amu Darya (Oxus). Turkestan is largely semi-arid or arid, for the most part receiving less than 250mm in annual precipitation, and the surficial geology is mainly composed of alluvium and loess sheets that extend northward into the Soviet states of Turkmenistan, Uzbekistan and Tadjikstan. These southern Asian states also experience high dust storm frequencies (see chapter 7 below).

Strong dust storms over the southern part of central Asia are often associated with the **Afghan Wind,** a southwesterly that blows from the plains of Afghan Turkestan and is particularly common on the upper course of the Amu

Darya in Termez. A graphic account of its effects in Termez is given in section 7.2.5.

Relatively fewer dust storms are experienced on the lowlands of the east than those of the north and south. From Mokur northeastwards the area receives more than 250mm in annual precipitation, excepting a small region around Jalalabad. The dust storm season starts in April/May and continues until August near the Pakistan border where a narrow strip of the country comes under the influence of the Indian summer monsoon. In this area monsoonal air is affected orographically by the Spin Ghar mountains producing a very dry, warm dust-carrying airflow down the Nangarhar valley called the **Borow Wind.** This wind blows from the south-west with a speed of 3 to 7m/s, and its effect in accelerating evapotranspiration causes considerable losses to rice production in the valley.

The dust storm season at Kabul continues until December (fig 6.3D), 20.0 dust storm days occurring in an average year in an area of silty loam desert soils (Salem & Hole 1969). During the summer months of June, July and August low pressure prevails over the Kabul basin and the basins of Parwan and Kapisa to the north, while a high pressure cell is established over the Salang mountains of the Hindu Kush still further north. The resulting pressure gradient produces a cold, dry northerly that is often associated with dust storm activity, this wind is known as the **Parwan Wind.** The effect on agriculture in the region is again adverse, reducing yields of the fruit trees in the orchards, although farmers traditionally use the force of the **Parwan** to thresh their crops.

Less dust-raising occurs in the mountains of the Hindu Kush. This situation reflects a combination of high precipitation (up to 1000mm in places) and a relative lack of widespread fine grained material such as the loess of the lowlands, as well as the relatively few meteorological stations in the remote mountain areas.

Visaie and Massaudi (pers. comm. 1984) report that generally dust storms are characteristic of the afternoon hours in Afghanistan, between 1230 and 1630 LST (0800 to 1200 GMT) when winds attain their maximum velocities.

6.4 PAKISTAN

Dust storms occur in all parts of Pakistan, largely during the hot season that prevails from April to June. Ahmad (1964) divides the country into four climatic regions that provide a good framework for an analysis of dust-raising activity in Pakistan.

The sub-tropical coastal strip (arid marine) includes the stations of Pasni and Ormara and the Indus delta south-east of Hyderabad. The area receives less than 175mm rainfall a year, and dust storms occur largely from April to July. Dust plumes from the western part of this coastal plain and on into south-east Iran can be clearly identified on satellite imagery during this period extending southward over the Arabian Sea (Grigoryev & Kondratyev 1981). Poor visibility due to this haze off the Makran coast is worst during May (Indian Met. Dept. 1931). Dust storm frequencies are of the order of two to three, a maximum occurring at Umarkot (6.4 a year).

The sub-tropical continental lowlands (extreme climate, arid with summer rains) includes the whole of the plains of Pakistan except the coastlands. Rainfall here is predominantly monsoonal, arriving in late summer, and excluding the northern sub-montane part it totals less than 250mm annually. The soils are largely derived from the alluvium of the Indus; the **Wariasi** of the Sind, a loose grey sand, and its allied light soils that disintegrate virtually to powder, produces the notorious Sind dust of the lower Indus plains. In the north of this area the loessic silts of the Potwar plateau are also prone to dust storm generation.

Perhaps the first comprehensive studies of dust devils and dust storms on the upper Indus plains were carried out by P.F.H. Baddeley during the years 1847 to 1853. Baddeley, a surgeon in the Bengal Army, published several papers that were brought together in a book entitled 'Whirlwinds and dust storms of India' in 1860, and subtitled 'An investigation into the law of the wind and revolving storms at sea with an addendum containing practical hints on sanitary measures for the Europen soldier in India'. This three-volume work contains a wealth of observations and theory: Baddeley followed dust devils on foot, horseback and by horse and cart in order to observe their nature, and he conducted experiments on their electrical properties using copper wire and a gold-leaf electrometer. He observed dust storms (calling them 'tornadoes') as a development of several large whirlwinds moving together, formations that Idso **et al** (1972) prefer to call 'lobes'.

Summer on the upper Indus plain is also characterised by high temperatures and an often dry, dust-laden atmosphere with light winds. These are so-called **Loo** conditions, and they may last for several days. **Loo** is a popular term, without any strict meteorological definition but refers to the dust haze conditions downwind of a source area. The haze may be particularly uncomfortable physiologically for local inhabitants due to its associated high temperatures, particularly at night (Joseph, pers. comm. 1986). Occasionally the hot weather is broken by thunderstorms and hot weather depressions which take the form of violent dust storms. These localised systems often lead to light rains after the dust storm has passed, and the sharp drop in temperature is stimulating, if short lived (Spate 1954). From his descriptions and investigations of electrical properties it is clear that it was these thunderstorm downdraft dust storms in which Baddeley was largely interested .

The sub-tropical continental highlands (cold, snowy winters, mostly arid with mainly winter and spring rain) include the mountains to the west and north of the Indus plain. To the west two to five dust storm days a year are typical. In the north dust storms occur on the arid lower slopes of the Karakoram mountain valleys. Climatic data are few for this area, however the members of the International Karakoram Project witnessed seven severe dust storms in six weeks from 12 July 1980 at Aliabad in the Hunza Valley (Goudie **et al** 1984a). Strong winds and gusts due to topographic influences and valley channelling entrain the fine material of the outwash plains in this mountain region.

The sub-tropical continental plateau (extreme climate, very arid) receives less than 125mm rainfall a year. This is the very arid desert of northwestern Baluchistan, bordering on Registan in Afghanistan, and Seistan on the Iranian/Afghan border. This area experiences dust-raising throughout the year, the peak season being from March to August. Nokkundi and Dalbandin record 36.5 and 29.1 dust storm days a year respectively, the highest values found in Pakistan. Some of these dust-raising winds are strong northerlies generated by steep pressure gradients in the lee of tracking depressions moving eastwards under the influence of 'western disturbances' (see section 6.5.1 below).

Snead (1968) reports that for southern Pakistan as a whole dust storms are most frequent in the afternoon hours, although larger, less local storms associated with an eastward moving depression occur at any hour of the day and sometimes at night.

6.5 INDIA

Dust storms are most common in northwestern India, in the arid and semi-arid states of Rajasthan, Haryana and Punjab, but they also occur on the Ganges plain and in the northern cold arid areas of Jammu and Kashmir. On the Indian peninsula dust-raising may occur in a few isolated areas during drought periods (eg on the Deccan Plateau in the area between Aurangabad with 3.0 dust storm days a year, and Nizamabad, 4.0 days) but normally the climate is too wet to enable storms to take place. Dust storms in India are highly seasonal, taking place very largely during the pre-monsoon months of April, May and June (table 6.1).

In Rajasthan maximum dust storm frequency occurs at Ganganagar, which with 17.0 dust storm days a year is the highest in India, and frequencies decrease south and westward, Bikaner records 8.0 days a year and Jodhpur 6.0 days. Dust storms occur down the Ganges plain. New Delhi records 8.0 days a year, Kanpur 5.0 and Allahabad 8.0, although the very hot conditions of the upper Ganges become ameliorated with distance southeastwards and the dust storms are associated with heavier showers in this direction (Singh 1971).

The importance of the alluvial soils of the Ganges in providing material fine enough to be entrained by winds in this area is paramount, and it is thought to be the source of sheets of late Pleistocene loess found in the valleys of the

Son and Belan rivers south of Allahabad (Williams & Clarke 1984) which is also being reworked today by aeolian processes. Figure 6.4 shows a clear concentration of dust-raising activity on these fine sediments of the Gangetic plain and the desert soils and sands of the Thar Desert.

A small area immediately north of the Ganges delta, from Jamshedpur in the west to Krishnagar in the east, experiences three to four dust storm days a year on the alluvium. This area experiences a general deficiency of rainfall from January to mid-June, and the dust storms are a manifestation of a markedly unstable atmosphere during the summer months due to three characteristic air streams. Dust storms in this part of India are largely generated by thunderstorm downdraft winds. These thunderstorms are known locally as **Nor'westers** or **Kal-Baisakhis** ('Doom of the Baisakh month') (Mull **et al** 1963).

6.5.1 METEOROLOGICAL SYSTEMS GENERATING DUST EVENTS IN NORTHWESTERN INDIA

Two distinct synoptic situations are commonly responsible for dust-raising in north-west India and both are related to the easterly movement of 'western disturbances'. A western disturbance is a low pressure zone or trough either at the surface or in the upper westerly wind regime, north of the subtropical high pressure belt, and such systems moving across Iran and Soviet Turkestan affect the Indian subcontinent north of 30^{o}N. Weak circulations, called induced lows, may simultaneously develop over central parts of Pakistan and Rajasthan and move E-NE wards (Rao 1981). The two situations commonly created by these induced lows are the creation of an area prone to thunderstorm generation, where dust storms are caused by the dry thunderstorm downdraft, and the setting up of a steep pressure gradient where strong winds may cause dust storms in dry areas. For all dust storms the Indian Meteorological Department has adopted three classes according to their intensities: light; moderate and severe (details are given in table 6.2).

6.5.1A THUNDERSTORM OR CONVECTIVE DUST STORMS

Dust is raised and advances as a thick wall of dust at the turbulent gust front that preceeds the main cumulonimbus clouds of the thunderstorm. In north-west India these storms are known as **Andhi** ('blinding', from the Sanskrit word 'Andha' meaning blind man). As the name suggests, the arrival of such a storm is characterised by a dramatic fall in visibility, often to below 100m, and also by a fall in temperature (by as much as 10^{o} to 15^{o}C) and a rise in humidity, while wind speed increases and changes direction (Joseph **et al** 1980). Examination of past charts shows that all spells of convective activity in northwestern India and west Uttar Pradesh and adjoining areas are associated with a western disturbance or its associated induced lows over

Pakistan, north-west India and adjacent areas. The western disturbances and induced lows may be seen either as a closed low or only as a trough; in the low levels of the troposphere a trough may extend eastwards or south-eastwards from the induced low, and this trough region is a potent field for thunderstorm activity (Srinivasan **et al** 1973). The effect of the western disturbance is to cause incursions of humid air of maritime origin at all levels, making the formation of convective clouds possible in the normally very unstable pre-monsoon air of northwestern India. The exact mechanisms governing whether such convective activity will produce a thunderstorm or a dust storm are not clear, the idea proposed at one time that dust storms are associated with less moisture than thunderstorms has been disproved by Bhalotra (1951, 1955) who shows that humidity values for both systems at all levels are similar.

At Delhi, **Andhis** seem to be more commonly associated with 'severe thunderstorms' (or squall line thunderstorms) as opposed to ordinary or cellular thunderstorms (Joseph, pers. comm. 1986). These systems can be identified on radar by the long bank of cloud from which the downdraft dust storm is generated. They are typically 200km long, with the dust front situated some 30km in advance of the cumulonimbus clouds. Severe thunderstorms commonly form when wind speed increases rapidly with height, and are generally predominant in India north of about 20°N due to the steep wind gradient up to the westerly jet. These dust storms, therefore, last less than an hour at any one station, although the severe thunderstorm itself may be relatively long-lived (5 to 10 hours) moving with a typical speed of 60km/hr. Convective dust storms in northwestern India are thus mesoscale phenomena. Joseph **et al** (1980) have classified **Andhi** into four types from the nature of variation of visibility and wind speed.

6.5.1B PRESSURE GRADIENT DUST STORMS

The pressure gradient dust storm, by contrast, is typically a synoptic scale feature, with dust-raising and dust transport occurring over large areas and often continuing for several days. Dust-raising winds occur when the pressure gradient is strong, and this situation typically occurs when an induced low moves east/northeastwards from west Pakistan to the northern Thar Desert producing a strong pressure gradient to the south of the low where the isobars are oriented east-west. Thus the most common area for pressure gradient dust-raising is in Rajasthan. Such lows usually have a large pressure defect from normal and are associated with strong upper winds in the very low levels (upto 1.5km).

The Indian Meteorological Department uses the term 'steep pressure gradient' for a pressure gradient of 1 to 1.5mb per degree of latitude and 'very steep pressure gradient' for more than 1.5mb per degree of latitude. When the pressure gradient is steep, according to this definition, and/or the lowest level upper winds are of the order of 30 knots or more, strong dust-raising winds may be expected over the area (Srinivasan **et al** 1973). Dust-raising winds may start in the morning and continue throughout the day, but their intensity reaches a maximum in the afternoon/evening, at the time of the maximum temperature epoch, when the superadiabatic lapse rates close to the ground favour the raising of dust. Dust may remain suspended in the atmosphere for upto a few days and generally is transported eastwards or northeastwards by the pressure gradient winds. To the south-east and east of the low pressure system, however, the pressure gradient is generally weaker and thus the winds are slight. Dust transported in this way, arriving at areas to the east and north-east of the dust source in light winds is the cause of conditions known as **Loo** (see section 6.4 above). **Loo** is typically experienced to the east and north-east of Rajasthan, in Delhi and down the Gangetic plain, as far as Bihar to the east (Singh 1971), although it is also felt over Rajasthan itself when dust is raised to the west on the Indus plains. In meteorological terms, therefore, **Loo** is a dust haze, and indeed it is often recorded as such by observers due to the low wind speeds involved because a dust event is only recorded as a dust storm in India when dust reduces visibility to below 1000m in association with a wind of Beaufort 4 (about 5.5m/s) or greater (see table 6.2).

6.5.2 RELATIONSHIP OF DUST STORMS TO METEOROLOGICAL ELEMENTS

The influence of rainfall on the wind erosion system and the generation of dust storms has been indicated above, although the relationship between dust storm frequencies and mean annual rainfall is not a simple one as correlations for many of the world's arid areas have indicated. Dust-raising is highly seasonal however, in south-west Asia it has been shown that storms are largely characteristic of months when rainfall is low (note that this is not universally the case eg Arizona, Sudan). Of the countries in south-west Asia India displays the highest degree of seasonality, dust storms being very largely concentrated in the months immediately preceeding the onset of the Indian monsoon (table 6.1). Thus, it was decided to make a closer analysis of the meteorological parameters affecting dust storms during the months of April, May and June in this area. India was also thought to be a good area to conduct further study as it has been shown that dust storm activity is largely concentrated on the alluvial plains of the Ganges, so that the effect of differing soil types on the relationships between other elements of the wind erosion system might be relatively minimised.

Data were taken from from the Indian Weather Review Monthly Weather Reports for 38 stations situated on the alluvium of figure 6.4, over the six-year period 1952-57.

Figure 6.5 shows a plot of the mean number of dust storm days during April, May and June in this period versus the mean rainfall recorded in these months. A Spearman rank correlation coefficient of -0.37 was obtained for these data. A similar correlation of dust storm frequency versus average wind speed during these months also showed a poor relationship, the coefficient being -0.21.

Further analysis was undertaken using the Wind Erosion Climatic Factor developed by Chepil **et al** (1962), who showed that to a first approximation wind erosion is directly proportional to the cube of the wind velocity and inversely proportional to the square of the effective soil moisture, as discussed in Chapter 2 above. Thus,

$$C = \frac{V^3}{(PE)^2} \times \frac{100}{2.9}$$

where C is the wind erosion climatic factor, V is mean annual wind speed in mph and PE is Thornthwaite's measure of soil moisture. The 100/2.9 factor was used to express C values in the midwest USA as percentages of a standard station at Garden City, Kansas. This factor was not used here as it did not seem relevant to this particular study. Thornthwaite's moisture index is an annual measure that can be defined solely in terms of temperature and precipitation:

$$PE = \sum_{1}^{12} 115\left(\frac{P}{T-10}\right)^{1.111}$$

where P is monthly precipitation in inches and T is mean monthly temperature in degrees farenheit. In the present example, however, the moisture index used covers just three months (April, May and June) so that

$$PE_{(AMJ)} = \sum_{1}^{3} 115\left(\frac{P}{T-10}\right)^{1.111}$$

Chepil **et al** showed that C values could be used to predict the wind erosion hazard in the midwest with reasonable precision and Yaalon and Ganor (1966) showed an accurate correlation between C values and arid and semi-arid zones in Israel. Dust storm frequency, being a measure of actual wind erosion, was thus expected to bear a close relationship to C values. A correlation of mean C values and mean dust storm day frequency for April, May and June for 34 Indian stations on alluvium for the period 1952-57 produced a poor relationship however, the linear correlation coefficient was -0.12. This result seems to suggest that the wind erosion climatic factor is a poor indicator of actual wind erosion events.

A number of suggestions can be forwarded as to why the correlations found are poor. As far as the relationship with rainfall is concerned, it may be more successful to correlate dust storms with some measure of antecedent soil moisture conditions, so that the annual or three-monthly rainfall before the onset of the dust storm season may bear closer relationship to dust storm frequency in April, May and June. The average wind speed correlation is also, perhaps, a little simplistic. A more realistic treatment of V would be to use only those velocity measures that exceed a threshold value for sediment transport on the Gangetic plain, that largely depends upon the soil surface type (see chapter 2). Unfortunately, available data were not of sufficient quality to enable this sort of refined investigation. Both these observations can also be made of the wind erosion climatic factor, and indeed, the PE measure is not as sensitive as would be desirable. In addition, dust storm frequency is not as accurate a measure of actual wind erosion as one would like. The duration of a dust storm is important here; an event that lasts for several hours will obviously represent a much larger amount of eroded material than one that is only a few minutes in duration. However, due to the nature of the data, both events would be represented equally as a dust storm day.

In addition to these problems concerning the data and the nature of the indices used, there are other considerations that need to be mentioned. Although according to figure 6.4 all the stations used in these correlations are situated on alluvial soils of the Ganges, at the local scale there will of course be variations in this category of surface covering, so that particularly vulnerable soil sub-types may act as dust sources locally but not be evident on a map of this scale. In addition the use to which a particular soil is put may be important as a factor affecting its wind erodibility from place to place.

An alternative explanation may be derived from the spread and extent of individual dust-raising events. A dust storm is recorded at a station in India when visibility is reduced to below 1000m with a wind speed greater than Bft4. It is possible that some distinction within individual events is concealed in the statistics between dust raised locally and material blown to a station from a distant source. Thus it may be that one or more distinct dust sources are responsible for visibility reducing dust events over the entire Gangetic plain. If this was indeed the case it would not be unreasonable to expect a poor correlation between dust storm frequency at a location and the corresponding rainfall, wind speed and soil moisture properties if the dust was not actually raised at that location but blown from a distant source. Further investigation, involving the tracking of individual dust events has thus been undertaken to evaluate the importance of this last suggestion.

6.5.3 DUST EVENT TRACKING IN NORTHWESTERN INDIA

It has been noted above that the **Andhis** are a mesoscale phenomenon, and as such, attempts to adequately track their movement were not successful given the low density of frequent-interval observatories in the region. Pressure gradient dust storms, however, are synoptic scale features, and thus tracking is possible using surface station reports and synoptic maps.

Four pressure gradient dust events were tracked using daily surface station observed data and synoptic maps, both for 0830 Indian Summer Time, IST (0300 GMT) and 1730IST (1200 GMT). Maps of events tracked show the surface pressure distribution and the station wind speed and direction and present weather codes (ww) relating to dust in the atmosphere (see fig 2.5 and section 2.4.1 for a discussion on SYNOP codes). To complement this information station visibility is indicated by visibility isopleths at intervals of 4.0km, 3.0km, 2.0km and 1.0km as decoded from present visibility values (VV) (see fig 6.6). These data were obtained from the Indian Meteorological Department at Pune. Four dust events are mapped, illustrating typical dust-raising conditions with differing durations and transport paths. Figure 6.7 shows the geographical distribution of stations used in this mapping exercise.

20-22 MAY 1965 (fig 6.8)

On 20 May 0830IST a shallow low pressure system is located over western Pakistan and dust is being raised by NNWly winds at Nokkundi in Baluchistan where the pressure gradient is steep on the southwestern side of the system. The pressure gradient over the rest of Pakistan and northwestern India is very gentle and winds are light and variable. By the late afternoon the low has moved northeastwards and deepened; winds are still light over north-west India and much of Pakistan but dust-raising is now occurring to the south-west of the low over much of Baluchistan from the NNW and at the coastal station of Pasni on the Makran coast. By 0830IST on 21 May the low has deepened considerably (24-hour pressure change is -8mb, and the pressure departure from normal is -8mb at the centre). In Baluchistan the dust-raising winds have developed into a full dust storm, the wind now from the north with an increased speed of 35 knots, while on the Makran coast the winds have subsided and haze is recorded at Pasni. In the Thar Desert and upper Indus plains winds have picked up as the pressure gradient has steepened and dust is being raised at Khanpur and Lahore in east Pakistan and at Jammu in northwestern India. Down the Ganges plain Delhi is experiencing a dust haze and Bareilly dust-raising, while the strongest winds from the south-west are blowing at 20 knots causing a dust storm at Phalodi in Rajasthan.

The low continues to deepen during the 21st so that by 1730IST the 24-hour pressure change is -10mb and dust-raising is occurring over virtually all of Pakistan south of about 33°N and much of northwestern India due to the steepening of the pressure gradient and solar heating during the day. In Baluchistan dust winds are still from the north-west, in the rear of the pressure zone, while over southern Pakistan and Gujarat winds are mainly westerlies of 10 to 15 knots and blowing dust at the stations or bringing it in suspension from further west. In Rajasthan the winds are generally stronger and from the south-west and are causing dust storms, the visibility isopleths show that maximum dust-raising is in an area that describes a horseshoe shape around the low pressure centre. Across the upper Indus plains to Peshawar thick dust haze is being transported from the south and west, while in the northern Thar dust storms are caused by southerly winds at Ganganagar and south-easterlies at Lahore. This last point is particularly interesting as most texts referring to the pressure gradient type dust-raising make no reference to a southeasterly element in a predominantly westerly flow.

By the morning of the 22nd the low has moved south-east and filled, pressure gradients are much reduced, winds have calmed and much of the suspended dust has settled except over a number of stations in Rajasthan where visibility is ¾4km with light winds. At Jodhpur dust is still being raised at this time, but at Barmer and Ajmer the suspended dust represents a **Loo** event.

7-12 MAY 1964 (fig 6.9)

A low pressure system, centred over the Seistan Basin north-west of Baluchistan at 1730IST on 7 May, arrives over the upper Indus plain at 1730IST, 8 May. Its arrival causes pressure gradient dust-raising in an area of Rajasthan broadly elongated west-east from Barmer to Jaipur, with a 20 knot wind blowing a dust storm at Phalodi. By 0830IST on 9th May the dust haze has been blown ENEwards as the low pressure area has moved over the northern Thar, the haze has reached Hissar and Delhi as suspended dust with light winds or calms (Loo) and Gwalior where there are light northwesterly winds.

By the afternoon of the 9th the situation has changed; the low has become more localised and the dust haze has drifted northwards affecting Ludhiana and Lahore (pressure departure from normal at the centre of the low at this time is -6mb). On 10 May the maximum dust haze is centred over the northern Thar with a tongue being advected down the Ganges plain with light winds. The 4km isopleth has reached Gwalior. By the afternoon of the 10th this pattern of haze has shifted eastwards and the maximum haze is located over Hissar as the low pressure system becomes elongated to the south-east. In north-west India pressure gradients are not steep enough to produce dust storm strength winds, although at 1730IST over Baluchistan Nokkundi and Dalbandin are recording dust-raising winds from the north-west of upto 20 knots. During 11th and 12th May the low fills slightly and retreats towards the south-west, pressure gradients and winds are still low but there remains a considerable amount of suspended dust in the atmosphere over the Thar, while the transport of material down the Ganges plain appears to have been much reduced. As the low retreats, winds in the vicinity of Ganganagar and Bikaner veer from S/SSEly at 0830IST 11th May to southeasterly on 12th May, blowing suspended material, with dust still being raised at Ganganagar and Jodhpur, towards the north-west to the stations of Lahore, Montgomery and Multan, producing a recurved dust transport path. Also during these two days afternoon winds at Leh are sufficient to produce localised dust storm activity in Kashmir.

25-28 JUNE 1953 (fig 6.10)

This dust event, which caused particular problems for communications and of livestock suffocation in Rajasthan, has been analysed by Roy (1954). On the 25th a low pressure zone over the Punjab has been induced by the passage of a western disturbance, and this low continues to develop into a deep depression which by 1730IST, 26 June has a pressure departure from normal of -12mb near Montgomery. The very steep pressure gradient set up to the south of this low causes strong southwesterly dust-raising winds that prevail from 0830IST, 26 June to 1730IST, 27 June, with dust storms being particularly fierce over the desert areas of Rajasthan. As the low deepens over the 24 hours from 1730IST on the 25th, sporadic areas of dust-raising can be identified by the visibility isopleths, with material being advected as haze northeastwards. By the late afternoon of the 26th dust is being raised over an area that extends from Karachi in the south-west to the north-east corner of the Thar Desert. At this time, and for the following 24 hours, the main area of dust-raising activity is in the Rajasthan desert, particularly in the Jaisalmer-Phalodi-Bikaner-Jodhpur area. Roy (1954) reports wind speeds exceeding 40 knots on the 26th and 27th for this area, and the dust storms caused considerable hazards by burying livestock and rail tracks in sand moved by the dust-raising winds.

The charts for the 27th June clearly show the dense dust haze being blown northeastwards from Rajasthan, with a definite southeasterly component carrying material down the Ganges plain, so that on the 0830IST chart the 2km isopleth reaches Bareilly, Mainpuri and Agra. At 1730IST Kanpur and Lucknow report widespread dust in suspension not raised at or near the station, reducing visibility to below 4km, brought by westerly or north-westerly winds of less than 10 knots. The main movement of material, however, is towards the north-east and the suspended dust movement is clearly tracked on the 28th by the visibility isopleths, away from Rajasthan as wind speeds subside while the depression weakens and retreats. To the north-east, the suspended dust was reported on the 27th and 28th at a number of hill stations such as Mussoree (visibility <500m 0830IST on the 27th) Simla and Jammu. Reports from these stations indicate that this material is not locally derived, indeed Roy points out that heavy rain fell between the 22nd and 24th June over a large area north-east of a line roughly joining Delhi and Ambala.

7-10 APRIL 1983 (Fig 6.11)

Two western disturbances are located over Pakistan, one over Baluchistan and a second over northern Pakistan, Jammu and Kashmir and Punjab, on April 7 1983, 0830IST. Pressure gradient winds from the south-west over Rajasthan are 10-12 knots raising dust at Barmer and Jaisalmer, with haze recorded at Bikaner. No meteorological charts were available for 1730IST on April 7, but observations at stations in Rajasthan indicate that winds have strengthened to 15-20 knots at Barmer, Phalodi and Ajmer, probably due to a steepened pressure gradient as the low over Baluchistan has moved eastwards, causing dust-raising and dust storms with a large area of reduced visibility. The large-scale dust-raising has built up during the day as shown in plate 6.1, a Space Shuttle photograph taken at about 1200IST on April 7 over western Rajasthan.

By 0830IST on the 8th the low over the Thar has filled out and winds have weakened over Rajasthan, though the atmosphere is still heavily dust-laden. Much of the material raised in Rajasthan has been advected eastwards and widespread haze is reducing visibility to <4km over the upper Ganges plain with light winds (**Loo**). By 1730IST there is widespread dust haze reported at Bahraich, Lucknow and Allahabad, and this haze continues to travel slowly down the

Ganges in very light winds. On 9 April 0830IST this suspended dust has reached stations not shown on the map - Patna (visibility <4km in haze), Kanpur (visibility <2km by dust in suspension) and Varanasi (visibility <1km in haze).

6.5.4 DISCUSSION

The foregoing analyses of dust events have been selected from a systematic study of synoptic charts during such events, and illustrate the typical area of dust-raising during pressure gradient type dust storms in northwestern India, with typical transport paths of dust once raised. From long experience of forecasting dust events, the Indian Meteorological Department state that 'Rajasthan, in particular, experiences dust-raising winds more frequently than the other sub-divisions.'(Srinivasan **et al** 1973, p40). At stations east of Rajasthan **Andhi** dust storms are probably more important (Joseph, pers. comm. 1986) as the induced lows associated with western disturbances seldom move further east than the Punjab.

From observations at Delhi airport Joseph **et al** (1980) conclude that 'most of the dust storms that occur at Delhi are of the **Andhi** type' (Joseph **et al** 1980, p431) and in one of the earliest studies of dust storms at Agra Sreenivasaiah and Sur (1939) also suggest that **Andhi** is the most important type of dust storm occurring at that station.

It has been noted above that the frequency of thunderstorms is often underestimated over large areas, especially at stations manned by part-time observers, and also because, being localised systems, they may occur in areas where no observer is present to record them. The first of these problems has been overcome for India by Rao **et al** (1971) who base their analysis of thunderstorms on records from stations that cater to aviation and thus have continuous observations, and are particularly alert to adverse weather conditions such as thunderstorms or peals of thunder that are observed and reported for warning purposes.

The map of annual thunderstorm frequency produced by Rao **et al** (1971) is shown in figure 6.12 in combination with a schematic representation of the findings of the dust tracking exercises: the most common source area of pressure gradient dust storms and the trajectories of transported dust which are more or less synonymous with **Loo** events. Although the large majority of **Andhis** occur during the pre-monsoon hot season (April-June) the annual frequency map of thunderstorms is broadly similar to the monthly frequency distribution for these three months, in that minimum activity is shown over Rajasthan and north Gujarat increasing northwards to the Himalayan foothills and southeastwards down the the Gangetic plain. Since this frequency distribution includes all thunderstorms, both **Andhis** and 'wet' systems, it does not, of course, prove conclusively that **Andhis** become increasingly important as dust-raising systems north and east of Rajasthan. In conjunction with observations in the literature referred to above, however, and the views of the forecasters at the Indian Meteorological Department the general pattern broadly confirms this suspicion. The relationship is not simple, however, in that with increasing southeasterly distance down the Gangetic plain dust storm frequency declines (see fig 6.4) whereas thunderstorm frequency continues to increase to a maximum over northern Bangladesh and Assam. This lack of correlation is due, of course, to higher rainfall totals towards this area.

Although the above conclusion, that pressure gradient dust storms become less important and **Andhis** become more important north and east of Rajasthan, seems reasonable, the situation should not be oversimplified. **Andhis** occur in Rajasthan (Vaidyanathan 1969) and in Gujarat (Upadhyaya 1954) and **Loo** or thick dust haze conditions may be experienced in these states due to pressure gradient dust-raising to the west, on the Indus plains. Indeed, Vaidyanathan (1969) documents an instance of severe dust haze (visibility <500m, wind speed <6 knots) that was apparently due to an unrecorded **Andhi** occurring in the Rajasthan desert.

The complexity of the dust storm system has been illustrated by the apparent lack of correlation between dust storm frequency and rainfall, both for northwestern India and for many other of the world's dry lands (Goudie 1983). From this study it seems that a significant proportion of dust events in north-west India are due to dust raised and advected to a station from afar (**Loo**). In most cases, however, such events are not recorded as dust storms, which does not explain the lack of correlation in this area. In the case of **Andhi** the characteristic advancing dust wall is a dynamic entity that probably consists of material both raised and brought from afar and material that is continually being raised along its path, so that this convective system combines both the ideas of a dust storm and a thick dust haze. If the land over which such a storm passes has a constant erodibility, and the downdraft maintains constant velocity and turbulence characteristics, therefore, it follows that the **Andhi** would increase in intensity as it continues along its path until the wind's carrying capacity is reached. Case studies using a dense network of continual visibility observations would evaluate such a suggestion.

In Rajasthan, some degree of correlation between annual dust storm frequency and monsoonal rainfall of the preceeding year is shown by Krishnan (1977). At Delhi the incidence of dusty weather in May and June was found to be closely related to the frequency of western disturbances affecting the weather of Delhi State by Roy (1954) as would be expected from the above analysis of the synoptic meteorology of dust-raising events. In Roy's study 'dusty weather' refers to both numbers of **Andhis** occurring and also the number of hours of visibility reduced to <1000m, more representative of **Loo** events as **Andhis** typically involve severe visibility reduction of less than one hour. Nevertheless, the incidence of western disturbances and dusty weather were not related by a simple rule of ratio, as antecedent rainfall among other factors, is also an important influence.

6.5.5 CONCLUSION

Dust storms occur throughout south-west Asia with the major exception of the Indian Peninsula. Two areas of particularly high frequency have been identified in the region. The highest dust storm frequencies occur in an area at the convergence of the borders of Iran, Afghanistan and Pakistan that comprises the Seistan Basin, Registan and north-west Baluchistan. Here the **Sad-ou-bist bad** (Wind of 120 days) blows from the north in the rear of the Indian summer depression, and is particularly strong in May, June and July. The stream of air appears to originate in Soviet Turkmenistan and attains hurricane force due to channelling through mountain gaps and katabatic effects into desert basins which act as ready sources of fine grained material washed down from surrounding desert surfaces by periodic rainfall events.

The other high frequency area is on the plains of Afghan Turkestan where dust-raising winds are generated by frontal lows penetrating from Soviet central Asia and topographically enhanced flows. The material raised in this area is from the widespread Quaternary deposits in the region and the alluvial deposits of the Amu Darya.

Generally, dust storms in south-west Asia are characteristic of the dry season months of spring and summer.

Diurnal maximum in activity is for the most part during the afternoon hours when the lower troposphere is least stable.

Other important dust storm systems in south-west Asia include the summer northeasterly flow down the Arabian Gulf associated with the prevailing **Shamal** raising alluvial material from lower Mesopotamia, the pressure gradient winds caused by western disturbances and their associated lows over Pakistan and north-west India, raising dust from the sandy soils of the Thar Desert, and the mesoscale thunderstorm downdraft storms on the alluvial plains of the Indus and Ganges rivers.

Dust from the region is transported over great distances, to the Asian states of the USSR, southwards over the Arabian Sea and possibly eastwards over south-east Asia.

Dust storm frequency during the peak season on the alluvium of the Ganges was found to be poorly related to rainfall, mean wind speed and a climatic wind erosion factor developed by Chepil **et al** (1962). Further investigation of the dust storm system in this area was conducted by tracking four individual synoptic scale dust storm events caused by pressure gradient winds over the Thar Desert. Material raised in this way is advected in light winds as a dust haze (**Loo**) southeastwards down the Ganges Plain, northeastwards to be washed out over the Himalayan foothills and some material follows a recurved transport path, flowing northwestwards over the northern Thar to the upper Indus Plains. Thunderstorm downdraft dust storms (**Andhi**) are less important in the Thar Desert but occur with increasing frequency down the Ganges Plain.

CHAPTER SEVEN : EUROPE & THE SOVIET UNION

7.1 EUROPE

Dust storms are not a major feature of the climate of Europe, largely because the climate encourages good vegetative cover of wind susceptible soils. The most important European area of dust storm occurrence is in the southern USSR which is covered in sections 7.2.1 and 7.2.2. In this section comment is made on those areas in Europe where dust-raising occasionally occurs, in many instances due to the actions of human populations disturbing the natural vegetational balance through poor agricultural practices.

In Britain wind erodible soils are largely concentrated in eastern areas: east Yorkshire (Radley & Simms 1967), Lincolnshire (Robinson 1968), Norfolk and Suffolk (Pollard & Millar 1968) and the East Midlands. The wind susceptible soils are the peats of the Fens and the glacial sands and gravels and Pleistocene coversands of the other eastern areas, while Fullen (1985) reports wind action on freely drained brown sands and brown earths in east Shropshire.

Wind erosion events may be as severe as a dust storm, Spence (1955) describes an incident in the Fens in which visibility was reduced to about 300m with winds gusting to over 45 knots. The local name for a dust event in the Fens is a **Blow** and Spence notes that severe **Blows** occur about once every five years, although Pollard and Millar (1968) suggest that the incidence of **Blows** in the Fens seems to have increased in recent years.

The strong winds responsible for dust storms in Britain are most commonly associated with frontal passage (eg Robinson 1968, Fullen 1985) and, as Morgan (1980) points out, most workers agree that the cause of the problem in the United Kingdom lies in farming practices. These include the removal of hedgerows, the expansion of arable farming on to land not suitable for that purpose and the cultivation on wind erodible soils of inappropriate crops such as sugar beet which develops a cover late in spring. In many cases it is only particular fields or parts of fields that are prone to dust storm generation. Certainly fields that have been recently ploughed are most susceptible to wind erosion (Fullen 1985), and Spence (1957) notes that serious **Blows** in the Fens often occur in the day following a rainy night. Seemingly the earth clods tend to break down to smaller particles when drying after rain.

Similar wind erosion and dust storm problems may occur in other European countries. For example, Rasmusson (1962) has studied wind erosion of arable land on fluvioglacial deposits in the Vomb Valley, Scania, Sweden and Møller (1986) highlights the wind erosion problems on glacial and fluvioglacial deposits in Denmark. Other human actions that may generate dust-raising in localised areas include construction work, as studied by Goosens (1985) on a torn-up stretch of road in Belgium.

Dust events in the Mediterranean countries of Europe are largely in the form of haze or falls of material from North African sources (see section 3.3.4), but the dry areas of Spain and southern Italy may experience local dust-raising **Sirocco** winds.

Dust storms are more common features of the cold arid lands of northern Europe where off-glacier winds raise fine grained material from glacial outwash plains. Events in these remote areas go very largely unrecorded by meteorological observers. Hobbs (1931) has noted such dust-raising in Greenland and Heidam (1984) notes that local sources may be important contributors to Arctic aerosols in summer months. A number of papers have appeared on dust storms in the ice deserts of central Iceland: Ashwell and Hammell (1958); Hammell and Ashwell (1958); Ashwell (1966 & 1972).

Hammell and Ashwell (1958) identify both a foehn type storm and a katabatic flow in central Iceland, the latter occurring during the late afternoon as stone and sand desert surfaces are strongly heated creating strong convection currents, with initial dust devils giving way to the downslope winds from ice caps causing widespread dust storms. Such storms in central Iceland are known as **Mistur** (Vivian, pers. comm. 1986) and Ashwell (1966) notes that the severely eroded lands of central and lowland areas were probably denuded by overgrazing and woodcutting from the first years of settlement about 900AD.

7.2 THE SOVIET UNION

Although dust storms are common over wide areas of the USSR, the Soviet meteorological authorities have not made any data available on their occurrence. This section, therefore, is based upon material published in Russian. This body of literature is fairly extensive, and several important books (eg Nalivkin 1983) and papers (eg Zakharov 1966, Sapozhnikova 1973, Klimenko & Moskaleva 1979) dealing with frequency, distribution and associated controls on dust storm generation have been translated into English. In addition to these I have had a number of other papers translated in full (Kharitonova 1969, Kes 1983) and I have translated relevant sections of some other papers (Kravchenko 1961, Dolgilevich & Sazhin 1973). The abstracts of other Russian papers, presented in English in Meteorological and Geophysical Abstracts, are also referred to where appropriate.

Much of the Russian work has been inspired by the widespread problems posed to agriculture by dust storms in the Soviet Union. In the same way that the Dust Bowl of the 1930s generated a major thrust in dust storm and wind erosion research in the USA, the problems caused by the ploughing up of 'virgin lands' on the dry steppes of Kazakhstan and western Siberia during the 1950s increased academic interest in the topic. Much of this work has been devoted to so-called 'black storms' (**Chernye Buran** or **Kara Buran**) formed on black earth and chestnut coloured soils characteristic of the southern European part of the USSR and the virgin lands of Kazakhstan. 'Buran' is the name commonly used for a high wind storm on the steppes and deserts of central Asia. It is derived from the Turkish language, and is also used in Chinese central Asia (see chapter 8).

The black storm is one of four types of dust storm identified in a classification developed by Zhukov (1964 quoted in Kes 1983) based on the colour and composition of the dust. Zhukov also distinguishes 'yellow storms', generated on sand, sandy loam and loam soils and occurring throughout Soviet Central Asia; 'red storms', also characteristic of the desert provinces but occurring on surfaces reddened by oxidation of iron compounds; and 'white storms', dust/salt storms formed by erosion of saline soils, saline deposits and other salt-bearing porous rock. White storms are not as widespread as the other three types, but have become the major storm type around areas such as the Aral Sea .

Storms may belong to more than one category when dust is raised over wide areas. Nalivkin (1983) points out that from the virgin lands of Kazakhstan to the deserts of Kazakhstan and Central Asia the types of storm follow a gradual transition, from predominantly black in colour over the virgin lands, to yellow-red storms over the deserts of southern Kazakhstan to dominantly yellow storms prevailing in Central Asia.

From the general works on dust storms in the USSR a number of main areas of activity can be identified. These are shown in figure 7.1 indicating the distribution of dust storms in the Soviet Union based on data over a 25-year period, for the most part from 1936 to the early 1960s. Table 7.1 shows the seasonality of dust storm activity in the main regions. These areas are covered in detail in the sections below. They are:

1. The Ukraine.
2. Southeastern European USSR.
3. Kazakhstan.
4. Western Siberia.
5. The Central Asian states of Turkmenistan, Uzbekistan, Tadjikstan and Kirghizia.

7.2.1 UKRAINE

Three maxima in dust storm frequency are identifiable in the southern parts of the Ukrainian SSR (Klimenko & Moskaleva 1979). The locations of these areas are indicated in figure 7.2.

1. The Kherson-Kakhovka district, situated on the floodplain of the Dnepr (Dneiper) River where it flows into the Black Sea. Here the number of days with dust storms reaches a maximum at Nizhnie Serogozy (17.3 a year).

2. The Voroshilovgrad district (Donbass), in the valley of a tributary of the Sev Donets River.

3. The southwestern part of the Odessa region, centred

in the area of Sarata, on the floodplain of the Kogilnik River as it flows into the Black Sea.

Dust storms occur in the southern Ukraine at all times of the year, but are especially frequent and destructive in the spring when 52% of all storms occur, mostly associated with easterly winds (Shikula 1981). During this time of year loose, unvegetated agricultural fields are major sources of dust storm sediments, the fields having been recently ploughed for cultivation. The strong easterly winds that cause dust storms over the southern Ukraine seem to be commonly associated with steep pressure gradients caused by the eastward movement of cyclones from the north Mediterranean into the area south of the Black Sea in combination with a high pressure cell centred over central European USSR. Dust storms may also occasionally be caused by thunderstorm downdrafts according to Shikula (1981).

Particularly destructive storms (called **Chernye Buria** in southern Ukraine) have eroded hundreds of millions of hectares in 1891, 1892, 1903, 1928, 1946, 1960 and 1969. Dust from the 1892 event was transported westwards as far as Berlin and north to Sweden and Finland (Nordenskjold 1894, Zakharov 1965). In 1928, 15 million tons of black soil were removed from an area of one million km^2 and deposited in the Ukraine, Rumania and Poland (Babichenko 1965). Material from the storms of January and February 1969 was reported to have travelled to southeastern Sweden, where dustfall in red snow was estimated at 0.5 to 1.0 g/m^2 (Lundqvist & Bengtsson

1970), Holland and England (Shikula 1981).

The storms of March and April 1960 are described as having been the worst for 20 to 30 years in terms of intensity, duration and extent (Kulikov 1961). There were three storm periods that lasted for 20 days in all, between March 18 and April 18, over an area of one million km^2. The storms were blown by easterly winds caused by a cyclone taking up position over central Turkey, having travelled over the Balkans and Appenines, at the same time as a high pressure system arrived over central European USSR from the Barents Sea. This pressure distribution and the area of dust storm occurrence and dust transport are shown in figure 7.3. The storms during the April period (3rd to 9th) are estimated to have removed between 500 and 600 m^3/ha of soil from some fields (Kulikov 1961).

Some work has been carried out on prediction of dust storms according to certain soil and meteorological conditions in the southern Ukraine. Shikula (1981) and Volevakha and Mel'nik (1982) showed some degree of predictive success considering soil moisture deficit with wind velocities of observed storms, so that for a given wind velocity a dust storm is likely to occur when soil moisture deficit reaches a certain critical value and vice versa.

7.2.2 SOUTHEASTERN EUROPEAN USSR

The region of the lower Volga and northern Caucasus is considered to be one of the major dust-producing areas in the world by Grigoryev and Lipatov (1974), although dust storms are not as frequent here as in the deserts of Soviet Central Asia (see below). In addition to the delta and floodplain of the lower Volga, Klimenko and Moskaleva (1979) identify the Kuma Manych depression and the Sal'sk steppe as being major dust sources, both of which experience more than 20 dust storm days a year on average. On the lower Volga dust and salt storms occur during spring and autumn when salt lakes are dry, and in the Kuma Manych and Sal'sk activity occurs in spring and summer and is closely associated with dry weather, particularly **Sukhovei** winds (Bova 1957).

Grigoryev and Lipatov (1974) monitored a large dust storm over the Kalmyk Steppes (just west of the lower Volga) extending from the Tsimlyanskaya Reservoir almost to the Caspian Sea on Soviet AES Meteor-4 satellite imagery. The storm covered an area 450km by 250km in the dry steppe zone that is widely covered by sparsely vegetated loessoid deposits. The dust-raising winds in this region are largely southerly or southeasterly , in this case generated by an increase in barometric pressure gradients on the north-west periphery of a high pressure ridge that extended to the regions of the Volga River and Central Asia. This increase in pressure gradients was caused by the movement of a cyclone away into northern Europe and the spread of the low pressure trough to the south of the cyclone. At Volgograd superadiabatic temperature gradients were observed upto 2km in altitude, above which was an isothermal layer, indicating that the dust layer extended up to 2km. Dust deposition during major dust storms in the Volga Delta may be up to $115 g/m^2$ (Bruevich & Gudkov 1954).

Dust has been tracked from western Kazakhstan and the region of Astrakhan northwestwards to Volgograd and Saratov, even reaching as far as Gorky. The accumulation of salts brought by dust storms causes a number of problems in southeastern European USSR, the effects being particularly noticeable on soils, vegetation and other surfaces. Salts brought from the north-east coast of the Caspian Sea, for example, cause damage on electricity transmission lines between Astrakhan and Kazan as they settle on the insulators.

During the early spring months in the Rostov region, dust and snow storms occur. Nalivkin (1983) describes an incident in March 1939 in which coloured snow and rain fell on Rostov-on-Don, the dust having been raised locally by the cold front of a cyclone entering the region from across the Black Sea. Such dust/snow storms are common in south European USSR. Protsenko (1965) documents a series of severe snow storms that occurred in the region of Stavropol, just south of the Sal'sk steppe, caused by a deep cyclone moving southeastwards over Rostov and Volgograd to Kalmyk and the northern Caucasus in January 1964. East of Stavropol, where there was no snow cover, the cyclone caused dust storms, and grey snow fell 200km south-east near the Caspian Sea coast with a dust load of $14g/m^3$. Dust storms in the Stavropol region occur on an average 10 to 12 days a year, most frequently in April and May.

On the steppe zone of the Northern Caucasus dust storms blow on the ploughed Chernozem soils especially from January to April when fields are bare ready for planting, and soil aggregates have been broken down by freeze/thaw and wetting and drying in autumn and winter. These effects may also be exaggerated by poor agricultural practices (Tregubov **et al** 1977). Nalivkin (1983) shows a zone of major wind erosion around Baku on the shore of the Caspian associated with westerly winds. Dust has been reported to reach the Lower Caucasus from Turkmenistan, across the Caspian Sea, and in exceptional circumstances from North Africa in upper level winds, such as in March 1962 when material from southern Tunisia was transported over Syria, southern Greece and Turkey and beyond the Caucasus as far as the north Caspian (Nalivkin 1983, p158).

7.2.3 KAZAKHSTAN

Dust storms are widespread on the dry steppes of Kazakhstan, a vast area of southern central USSR (fig 7.4). East of the Urals an area extending northwards from the north coast of the Caspian experiences 20 or more dust storm days a year, with a maximum at Dzhambeiti in the Ural'sk region of 45.9 days and the Mangyshlak Peninsula with an average of 44 days a year. Areas with over 20 days a year stretch south and east of the Urals from the Turgay Plateau (Turgay, 20.9 dust storm days a year), the Tselinograd/Pavlodar/Karaganda region, between the Irtysh River and the Syr Darya (Dzhaltyr, 26.5 days), to the sands of Sary-Ishikotrau, between Alma Alta and Lake Balkhash (Bakanas, on the Lli River, records 47.7 days a year).

In a study of dust storms on the Mangyshlak Peninsula, on the east coast of the Caspian, Kharitonova (1969) found that they were generated locally, on average 44 days a year. Activity is largely from March to October, when wind speeds reached 6-7m/s and over. The dust storms were generated on dusty sands and light loams. Their duration lasted from several minutes to 18 hours, with an average duration of 3-4 hours. Average dust contents per season indicate that highest dust concentrations occur in summer and amounts to 40 microgrammes/m^3. Kes (1983) mentions a white storm on the Mangyshlak Peninsula in 1910 in which 500,000 sheep and goats and 30,000 camels perished.

The principal factors contributing to dust storm occurrence on the steppes of Kazakhstan are the presence of large areas of dark chestnut and chernozem carbonate soils in many areas ploughed up without adequate protection against wind erosion; light sandy loams that are easily subject to wind erosion; sparse tree cover in the north and absence of trees in the south; and synoptic systems that produce dry weather and strong winds during the warm season (Zhirkov 1964). Dust storms on the steppes begin in April, after the snow cover has melted and soil has dried out to some extent, and continue until October. In the desert regions of the south, where dust storm frequencies are generally higher and the percentage of territory with high frequency (>20 days) is greater, they occur all year round, but with a definite maximum from May to September. Nalivkin (1983) suggests that the main cause of dust-raising winds in Kazakhstan is the passage of cold fronts, and Zhirkov (1964) shows that although winds from all directions can cause storms in the steppes, the most common directions are from the south-west and west, corresponding to the prevailing winds in the area. In the desert regions south of Lake Balkhash local winds from the mountains to the south and east also produce dust storms (see below).

Dust storms in the Tselinograd/Pavlodar region have been studied by Chakvetadze (1962) who notes that cultivation of virgin lands in the 1950s led to more than one million hectares of various agricultural crops being damaged by wind erosion in the period 1955-60 in Pavlodar Oblast alone. Pavlodar is in the dry steppe zone, with 200 to 300 mm rainfall a year. Local physical conditions make some areas more liable to wind erosion than others, so that relatively high dust storm frequencies occur on the low right bank of the Irtysh River due to an almost complete absence of trees and the presence of large areas of dark chestnut sandy loam soils particularly prone to wind erosion. On the left bank of the Irtysh, by contrast, carbonate soils are much less subject to wind erosion and dust storm frequency is far lower. Zhirkov (1964) shows that dust storms are more common in this area during dry years. At Tselinograd in 1955 there were 49 storms when warm season precipitation was 28% of the long-term average, while in 1947 when precipitation amounted to 120% of the long-term average there were just two storms, although the wind regime was substantially similar to 1955. Dust storms are also more likely in years with unusually dry winter and spring seasons.

The duration of storms in Kazakhstan is in most cases quite short, lasting upto one hour. Numerous storms continue for 10 to 12 hours but rarely over 15 hours. Wind speeds are generally 4 to 10 m/s but many storms attain speeds of 11 to 20 m/s and on rare occasions more than 20 m/s (Nalivkin 1983). Zhirkov (1964) notes a direct relationship between the frequency and duration of storms: the more frequent they are at any given station the longer they last. In most cases, during the course of dust storms air temperatures rise and relative humidity falls, and dust storms are also closely linked to **Sukhovei** winds. At Tselinograd, for example, 55% of the dust storms that occurred during 1936-60 blew on **Sukhovei** days or on the following days, with the remaining 45% on ordinary dry days.

There are two local winds in the region of Lake Balkhash that are important dust storm generators (Nalivkin 1983). The **Balkhash Bora** is a cold dry northeasterly that blows from the crest of the Chengiz Mountains to the shores of Lake Balkhash. This wind is developed when anticyclones are located north and north-east of the crest of Chengiz, and the flow is probably increased by gravitational forces. The **Balkhash Bora** is strongest in winter when it often blows with hurricane strength, giving rise to fierce snow storms. In spring and summer, although the flow is weaker, it still gives rise to dust storms of considerable strength. South-east of Lake Balkhash the **Ibe** blows from Yebinor Lake in western China to the Balkhash-Alakul depression through the Dzungarian (Jungar) Gate, a valley corridor 200km long and 10 to 20km wide. Due to the funnelling effects of the Dzungarian Gate the **Ibe** reaches speeds of >50m/s, and a speed of 80m/s has been recorded (Lydolph 1977). The winds cause severe blizzards in winter and dust storms in summer on the debris-covered plains below the Dzungarian Gate. The terrible violence of the **Ibe** is noted in the journals of the Russian explorer Colonel Prejevalsky (1879) and the American Dr. Schuyler (1876) during their respective travels in Central

Asia.

In recent years a new source of dust and salt storms has emerged in Kazakhstan on the fringes of the Aral Sea. The sea level began to be reduced in 1961 and had fallen seven metres by 1980, exposing a strip of former lake bed 50-60km wide on the eastern side, by 1984 the level had fallen another 3m with a corresponding increase in exposed area (Kondratyev **et al** 1985). Grigoryev and Kondratyev (quoted in Kes 1983) have used satellite imagery to monitor large-scale dust events on the drying sea floor that were unknown before 1975. In that year dust storms were raised on an area of about $3000km^2$ on the northeastern shore, with a further dropping of sea levels that area had increased to $4800km^2$ by 1980. A storm in May 1975 blew with a dust cloud covering $14,000km^2$ with an estimated mass of 300,000 tons, and a storm generated by the same wind speed in May 1979 covered an area of $45,000km^2$ with a mass of about one million tons.

These dust/salt storms are generated by two particular synoptic systems:

1. The formation of a gale zone due to a steep pressure gradient set up between anticyclones having developed over the southern Urals, western Siberia or west and north Kazakhstan and cyclones settling over the west of Central Asia. These gales may blow for two to five days and typically occur in March to April and September, raising dust over a length of shoreline 170+/-88km.

2. Invasion of cold air masses behind cold fronts that

may blow for a period between 10 hours and several days. Such cold air intrusions arise at the end of May and in June and occasionally in August, and the storms typically raise material over a length of shoreline 270+/-140km.

The salts are mainly chlorides and sulphates, the major portion of which is blown upto 500km from its source to the south and south-west, causing problems of soil salinization and decreasing agricultural productivity especially in the oases of the Amu Darya and Syr Darya deltas (Kondratyev **et al** 1985).

7.2.4 SIBERIA

Dust storms occur on the steppes of western Siberia and southwestern Central Siberia in Tuvinia (fig 7.4). On the steppes the average number of days with dust storms varies from 0.2 to 4 in the south of Tyumen Oblast, and from 0.3 to 12 in the Omsk region. In the 1950s the Omsk region was opened up to agriculture, and the ploughing of the virgin lands had a dramatic effect on dust storm frequencies which increased by an average factor of 2.5 (table 7.2), and according to data from Pokrov-Irtyshsk by a factor of more than 5.

In the Novosibirsk region and Altay Territory the average number of days with dust storms ranges up to 20 to 28 (Aleisk, in Altay, records 28.2 days a year [Klimenko & Moskaleva 1979]). In the Altay region dust storms occur from

April to October, with a maximum in May and June. Their duration varies from half an hour to 24 hours, and these commonly occur between 0800 and 1500 local time (Kukis 1968). Dustfalls have been measured in the glacier ice of the Abramov glacier in the Altay Mountains by Nozdriukhin (1970) who found that maximum falls occur in June (2.5 tons/km^2), July (6.7 tons/km^2) and August (10.4 tons/km^2). This material was derived from the lowland regions, neighbouring dry valleys and the slopes surrounding glaciers. Virtually no dustfalls were recorded during the winter months.

Generally in western Siberia dust storms are most frequently associated with southwesterly winds, and Dolgilevich and Sazhin (1973) found a very high positive correlation (r=0.93) between duration at a station and the mean annual number of dust storm days.

In the south of Krasnoyarsk Territory and in Tuvinia dust storms are observed most often in May and June, although they also occur in winter when snow cover is light or absent. Annual dust storm frequency ranges upto 10 to 12 days (Klimenko & Moskaleva 1979).

7.2.5 TURKMENISTAN, UZBEKISTAN, TADJIKSTAN & KIRGHIZIA

Some of the highest dust storm frequencies are recorded in the dry desert areas of Soviet Central Asia, in the states of Turkmenistan, Uzbekistan and Tadjikstan (fig 7.4). These areas are overlain by loose alluvial sands, sierozems and, particularly in Tadjikstan, loess coverings (Goudie **et al** 1984b) and large parts of the region receive less than 200mm of rainfall a year. Klimenko and Moskaleva (1979) quote Repetek (Turkmenistan) as recording 65.5 dust storm days a year, and Kurguzul, located on a sandy spit dividing the Kara-Bogaz-Gol Gulf from the Caspian Sea, records 81 days. Sapozhnikova (1973) reports that the highest annual dust storm day frequency in Soviet Central Asia is at Takhiatash (Uzbekistan), located on the floodplain of the Amu Darya just south of the delta as it flows into the Aral Sea. Here 108 days with dust storms is the average annual frequency. Sapozhnikova notes that higher dust storm frequencies are experienced at more stations in the very hot dry zone of these three states than the dry zone (table 7.3). In Tadjikstan highest dust storm frequencies occur in the south of the state (about 20 days a year). Storms occur all year round in these areas, but with a maximum in the spring and summer months. Kes (1983) reports that in Turkmenistan the average annual number of hours with dust storms varies from 120 to 390 hours. The record duration of 721 hours in a year was recorded at Repetek; in July 1942 alone there were 176 hours of storms, averaging six hours each day.

Many of the stronger winds that raise dust on the lowlands of Soviet Central Asia are associated with vigorous intrusions of colder air from the west and north-west, behind rapidly moving cold fronts (Lydolph 1977). In south Turkmenistan storms are generally associated with the prevailing northwesterly and northerly winds except close to mountain ranges where more localised winds become important (see below). Thus, the greater part of material entrained is transported southeastwards towards Afghanistan. In north Turkmenistan, however, most storms are easterly, carrying dust to the Caspian Sea and the Lower Caucasus. These winds are associated with anticyclones (Nalivkin 1983). The majority of dust storms in Turkmenistan are local, according to Nalivkin, with dust being raised to a few hundred metres and occurring over small areas. Average duration ranges from a few minutes to a few hours, with the longest lasting storms continuing for two or three days. Kes (1983) shows that the hours with maximum dust-raising in Turkmenistan are 1100 to 1600 local time, and most storms occur with wind speeds of 7 to 14 m/s (table 7.4).

A typical post cold frontal wind associated with intensive cold outbreaks from the north or north-west is the **Afghan Wind** or **Afghanets.** This strong dry wind causes severe dust storms in the southeastern part of Central Asia, particularly on the upper reaches of the Amu Darya around Termez. This wind blows as a southwesterly, from the plains of Afghan Turkestan, or as a westerly across the Kara Kum. At times the **Afghan Wind** may penetrate up the Amu Darya into the Tien Shan mountains, and it also affects the Syr Darya area, particularly the Hungary Steppe and the Fergana Basin, where it is mentioned by Lansdell (1885) during his travels distributing religious literature. Locally the **Afghan Wind** dust storm is known as a **Kara Buran** (see section 7.1 above), and Nalivkin (1983) gives a graphic account of one storm in Termez in 1927:

> All of a sudden it became dark at noon and the lights had to be switched on. An enormous dark brown cloud covered the entire sky. The wind became stronger and stronger and all of a sudden it became hot due to the baked dust. The wind blew like a wild animal and a frightening roar was heard all around. Everyone hid as best he could. The windows were covered as much as possible but still the hot dust penetrated everywhere. There was no escaping it. Due to the movement of the burning dust, all the surrounding objects, buildings and air were charged electrically. (Nalivkin 1983, p137).

At Termez the **Afghan Wind** dust storms attain speeds of 41m/s, and dust may be raised to 3500m, taking ten days to settle after the wind has ceased.

Foehn-type winds, common throughout the mountainous areas of Soviet Central Asia, may attain unusually high velocities in some locations due to constrictions in airflow through mountain passes or around mountain edges. One such wind is the **Garmsil** ('hot wind' or ' hot storm') which blows from the Kopet Dag mountains in north-east Iran northwards over southern Turkmenistan, often raising large quantities of dust, changing to a typical hot, dry dust storm (Nalivkin 1983). This wind often blows from the warm sector of a cyclonic storm moving in from the south-west (Lydolph 1977). These systems commonly form over northern Arabia tracking through the region from Baghdad to Tehran, before crossing the Kopet Dag.

The **Garmsil** is characterised by high wind speeds, low relative humidities, high temperatures and great quantities of dust which dries and sears vegetation. Leaves are twisted, become covered in brown spots and fruits dry up and remain undeveloped. Nalivkin documents one case where an intense **Garmsil**, lasting a few hours, resulted in a 20-30% drop in yield of agricultural crops in south Turkmenistan. These winds blow at any time of the year but are generally most prevalent in summer, when their effects are most damaging.

A description of a **Garmsil** that crossed Kopet Dag on 13 March, 1953 and continued over Ashkhabad into the Kara Kum is given by Astapovich (1955). The area enveloped in the dust cloud was not less than $8400km^2$ and the total weight of material that comprised the cloud was more than 100,000 tons. This cloud passed over the interior deserts of Iran before crossing into the Soviet Union.

Dust-raising foehn winds also blow down the Chirchik Valley towards Tashkent at the southwestern end of the valley. Such winds may blow for one to four days, causing serious dust storms in the Chirchik Valley (Lydolph 1977).

North-east of Tadjikstan, in the mountainous state of

Kirghizia, Klimenko and Moskaleva (1979) note that dust storm activity is very localised, with high average annual numbers of days (14 to 16) occurring in the western part of the Issyl-Kul depression and in the lowland part of the Chuya Valley. Storms occur on 12 to 13 days a year in the Kochkor Valley and the Novorossiika district, and on 10 days a year at Darant-Kurgan.

7.3 CONCLUSION

The most active areas of dust storm occurrence in Europe are situated in the southern parts of the USSR, in the Ukraine and the lower Volga and Caucasus. Easterly pressure gradient winds commonly raise material over large areas of agricultural land in the Ukraine, agricultural fields also provide dust storm sediments in the Caucasus while to the north-west of the Caspian Sea the alluvial floodplains and delta of the lower Volga are an important dust source. In this area of southeastern European USSR pressure gradient winds are the main synoptic system generating dust storms, and across all the European areas of the USSR these winds are caused by cyclones tracking into the area from the west. The main dust storm season in these regions is the spring.

In Soviet Asia dust storms occur across large areas of the steppes of Kazakhstan and western Siberia, and in the arid and semi-arid regions of Soviet Central Asia. The wide expanse of flat steppe soils are particularly susceptible to dust storm generation with the passage of cold fronts and the ploughing of virgin lands in the 1950s resulted in large-scale increases in dust-raising. In more mountainous areas, topographic influences induce locally enhanced wind speeds, causing severe storms in desert depressions and on alluvial spreads. Recent dust/salt storm sources have been exposed since the 1960s on the fringes of the Aral Sea as water levels have dropped.

Highest dust storm frequencies are recorded in the states of Soviet Central Asia. Dust winds in these regions are associated with frontal passages and topographic influences, raising material from desert sources, the alluvial sediments of floodplains and deltas and loess coverings.

The effect of human action on the dust storm systems of the USSR is most pronounced in agricultural areas, where the problems of aeolian erosion have inspired a substantial body of research into dust-raising in the USSR. Silvestrov (1971) suggests that:

> Historically, extensive erosion and deflation in the USSR reflected the faulty use of agricultural lands prior to the October Revolution, but such use has not been completely eliminated today. Agriculture is both the chief cause and the principal victim of erosion and deflation, and the damage to its economy is both great and varied. (Silvestrov 1971, p161).

Long-range transport of material from the USSR occurs from the Ukraine and the Caucasus to eastern and northern Europe, and European USSR receives material from North Africa in exceptional circumstances while dust from Middle Eastern regions is transported over mountain barriers to Soviet Central Asia.

CHAPTER EIGHT : CHINA

8.1 INTRODUCTION

Some of the earliest written records of dust storm activity anywhere in the world are recorded in the ancient Chinese literature. They refer to dustfalls in northern China, and are variously known as "dust rain", "dust fog" or "yellow fog", usually occurring in the spring months. The earliest known record of "dust rain" was in 1150 BC, and is found in an historical book: **Zhu Shu Ji Nian** (Chronicles Recorded on Bamboo Slips, quoted in Liu **et al** 1981).

At the turn of this century numerous travellers in Chinese Central Asia have made vivid descriptions of dust storms (Prejevalsky 1879, Deasy 1901, Hedin 1903, Stein 1904). Sven Hedin, the Swedish explorer, notes the local names for dust storms in the Taklimakan and Gobi deserts: **Kara Buran** (black storm), a Turkish name also used in the Soviet Central Asian deserts today, and **Sarik Buran** (yellow storm). Tregear (1980) suggests that the **Kara Buran** is commonly rare and shortlived, lasting less than 24 hours, and adds that the **Kyzyl Buran** (red storm) is more common in these areas, lasting two or three days. During such storms the people of Xinjiang say "it is dusting" as opposed to "it is raining", the rain drops being cemented with dust.

Hedin made wind speed measurements during a ferocious reddish yellow storm in the Taklimakan that reduced visibility to zero. "Close to the earth" the wind blew at 18 m/s, and at six feet above the ground it was 26 m/s. In an account of a **Kara Buran** also in the eastern Tarim Basin he describes towering pillars of dust coalescing to form an advancing wall (plate 8.1), in very much the same way as Freeman (1952) describes the initial stages of a **Haboob** in Sudan (see section 3.3.5A). Hedin and Prejevalsky separately note that dust storms in the Taklimakan are typically blown by northeasterlies, and Stein refers to an ever present dust haze over the Tarim Basin with a definite roof at 1000 feet.

8.2 FREQUENCY, DISTRIBUTION & SEASONALITY

The distribution of stations used in the following analysis is shown in figure 8.1, with the major deserts and other topographical features. Figure 8.2 shows the frequency distribution of dust storms in China today and table 8.1 gives an indication of the seasonality of activity. The frequency distribution includes 30-year averages (1951-1980) for 20 stations provided by the Meteorological Administration at Beijing and incorporates data shown in Goudie (1983, fig 8) which shows annual averages taken over a three-year period. Although the areal extent of stations is by no means comprehensive, a number of high frequencies are recorded at stations in China, despite the classification of the Taklimakan and Gobi deserts as low energy wind environments in terms of sand movement (Fryberger 1979). Areas without data may also experience dust storms, notably the arid parts of the Tibetan Plateau.

The high frequency stations shown on figure 8.2 can be related to the main dust source areas in China as suggested by the above-mentioned literature, my own research and a number of papers that track large-scale dust storms across the country and dust transport over the Pacific Ocean (Ing 1972, Liu **et al** 1981, Xu **et al** 1979, Shaw 1980, Iwasaka **et al** 1983). These areas will now be covered in detail.

8.2.1 TARIM BASIN : TAKLIMAKAN DESERT

Situated in the central part of the Tarim Basin the Taklimakan is one of the world's largest sand deserts, and the whole area receives less than 100mm of precipitation a year. Kes and Fedorovich (1976) refer to it as one of the dustiest places in the world, recording 100-174 dust storm days a year, although no visibility limit is mentioned. Hotien (or Khotan) records an average 32.9 dust storm days a year, with a mean annual rainfall of 35mm. The dustiest months are from March to June, when 67% of storms occur (table 8.1). The diurnal variation in dust storm frequency (fig 8.3) shows that activity occurs throughout the 24-hour period - many of the turn of the century travellers describe storms lasting for several days - with a definite maximum around midday during the most active months.

The dust is derived both from the sand desert, from alluvial sediments and fans of rivers flowing into the basin from the surrounding mountains, from playas and perhaps also from loess deposits on the northern and southern periphery of the basin. The peripheral aeolian sands of the Tarim Basin are generally fine and include large amounts of silt (Breed **et al** 1979), and sand in the vicinity of Hotien is particularly fine grained (Petrov 1976, shows 81.5% of material to be in the 0.10mm to 0.01mm size fraction, 18.5% in the <0.01mm fraction). Large-scale dust-raising is caused by frontal passages moving westward and local storms may be produced by katabatic wind flow.

The annual variation in dust storm frequency at Hotien from 1954 to 1980 is shown in figure 8.4. The interannual variation is not great (CV=24%). A gradual decline in frequency over this period is discernable, but no increasing trend in rainfall is apparent, and correlations between dust storm days and annual rainfall were found to be poor, linear correlation coefficient = -0.03.

North-east of the Taklimakan, Hami records 33.0 dust storm days a year. The very arid Hami Depression to the east may act as a local dust source that adds to material transported from the Dzungarian Desert. North of the Tien Shan mountains, on the southern edge of the Dzungarian, Urumchi records 6.3 days a year. These northern slopes of the Tien Shan mountains are moister than the southern slopes due to the influence of air from the Arctic Ocean.

8.2.2 GANSU CORRIDOR

The Gansu Corridor (sometimes referred to as the Kansu or Hexi Corridor) is a narrow belt of land extending along the northern piedmont of the Qilian Shan mountains and bordered to the north by the Longshou Shan mountains. The orientation of the corridor, WNW-ESE, is such that outbreaks of cold polar air, often associated with the passage of southeasterly moving frontal systems, are funnelled through the corridor causing large-scale dust storms. These "black storms" which occur principally in the spring, cause widespread damage to property in the area. Xu **et al** (1979) defines the black storm in the Gansu Corridor as a gale with wind speed >25m/s, together with a visibility <50m. Dust storms caused by thunderstorm downdrafts may also occur in the corridor (Wells, pers. comm. 1985).

The corridor is dry, receiving less than 300mm rainfall a year, and the surficial sediments comprise sandy loess in a belt along the south of the corridor, and alluvial sediments to the north brought from the Longshou ranges. The station of Minqin is situated at the eastern end of the corridor and averages 37.3 dust storm days a year, with an annual rainfall of 116mm. Dust storms largely occur during the spring months, from March to July 63% of all storms are recorded, while rainfall is largely concentrated later in the year, during the summer months. The annual variation in frequency from 1954 to 1980 (Fig 8.4) shows no particular trend, and the interannual variation is not large (CV=27%). There seems to be a vague relationship between dust storm frequency and the rainfall of the previous year, and this relationship was investigated in more detail but was found to be statistically insignificant (fig 8.5), a test of annual dust storm frequency against average annual rainfall over the previous three years also showed little relationship ($r=-0.03$).

A particularly powerful black storm that swept along the Gansu Corridor on 22nd April 1977 is described by Xu **et al** (1979). The storm occurred behind an advancing front of cold air that originated to the north-west of the Tien Shan mountains and moved eastwards along the corridor at a speed of 22m/s. Successive towns along the corridor were engulfed by the advancing wall of dust that was 1000m high at Zhangye (Cheung Yik). As the storm arrived average wind speeds rose from less than 5m/s to gale force Beaufort 10 (25 to 28m/s), with gusts reaching Bft12 (>33m/s). The duration of winds of Bft8 or higher (>17m/s) ranged between two and five hours. With the arrival of the dust wall and concurrent dramatic rise in wind speed, a sharp increase in pressure and a rapid drop in temperature occurred (+2.8mb and -6.8^{o}C at Zhangye), a pattern of events reported in cold frontal dust storms all over the world. The black storm extended over 400km along the Gansu Corridor, caused extensive damage to farm land, and claimed a large number of casualties, both human and animal.

8.2.3 GOBI DESERT

The Gobi Desert is largely comprised of sandy areas such as the Tengger (Ala Shan) and Mu Us (Ordos) deserts on its southern periphery, and sand/gravel plain or **Gobi,** which extends across the border into the Mongolian People's Republic. This area is a major source of dust storms. Paotou and Hohhot, situated on the southern margins of the Gobi, record 19.0 and 8.4 days with dust storms a year respectively, spring time again being the most active season (table 8.1). As with most storms in northern China, the passage of cold fronts is the most important mechanism of dust-raising in the Gobi, although Space Shuttle photographs show that thunderstorm downdraft storms also occur in the region (Wells, pers. comm. 1985).

8.2.4 LOESS PLATEAU

The Loess Plateau of Inner Mongolia is situated immediately south of the Gobi, Tengger and Mu Us deserts, and is an important source of modern dust storm sediments. The area is fairly intensely cultivated with annual rainfall for the most part between 250mm and 750mm, although the loessic soils are very fine grained and easily eroded. Winds associated with frontal systems that sweep across the plains from the west and north west in spring are the most important synoptic situation for large-scale dust transport. Dust storm frequencies are not as high as in the drier desert regions, Zhengzhou on the south-east periphery of the loess plateau, averages 7.2 days a year. The season of maximum activity is spring, as in the other areas of China, although winter activity, starting in January, is also important (table 8.1). Rainfall over the plateau has a marked summer maximum.

8.2.5 OTHER CHINESE AREAS

High dust storm frequencies are recorded in western Szechuan and Changutu Province; Changtu and Kantzu stations have 12.8 and 35.0 days with dust storms each year respectively, occurring largely in winter and spring (December to May). Dust-raising from point sources on the 4000m Tibetan Plateau is generated by the upper westerly airflow (Wells, pers. comm. 1985) particularly in winter when vegetation cover is sparse. The average of four years of meteorological observations at Lhasa show 2.3 dust storm days a year, but large areas of this region are lacking in available data.

Generally dust storms decrease in frequency south and east of the major source areas outlined above. As seen above, the time of maximum frequency is the spring, due to the combined effects of low rainfall, the ploughing of cultivated soils for planting and the circulation over northern China that is marked by strong winds associated with cold fronts and movements of cold air masses from mid-latitudes (Ing

1972). Breed **et al** (1979) show that spring is the time of maximum Drift Potentials in northern China.

8.2.6 MONGOLIA

No data or information on dust storms was obtained from the People's Republic of Mongolia, but the following general comments can be made. The Gobi Desert extends across the southern regions of Mongolia and is a significant source of dust storm sediments as mentioned above. In the north of the country problems of aeolian soil erosion and dust storms are currently being experienced where 'virgin' steppe lands are being plowed up for wheat growing, without the employment of appropriate soil protection measures. Sanders (1982 and pers. comm. 1983) gives the example of the Atar ('Virgin Land') state farm, west of Ulan Bator, where 7000 hectares of soil have been eroded since its inception in 1977.

8.2.7 JAPAN

Blowing dust, known as **Fuhjin** , and to a lesser degree dust devils, occur fairly regularly on the Kanto Plains around Tokyo from January to early March, especially after a dry winter. The **Fuhjin** generally occurs behind an eastward moving cold front and visibility may be reduced to 400m with dust reaching an altitude of 900m (Fortner & Ihara 1964).

8.3 VARIATIONS THROUGH TIME

Dust storm frequency for Hotien and Minqin (fig 8.4) show no particular trends over a 27-year period for these two stations that represent two of the major dust sources in northern China. It is interesting to note that the variability at these stations, as expressed by the coefficient of variation, is low relative to other parts of the world.

Over a period of centuries Stoddart (1978) shows the variation in the number of "earth rain" events over north China, with distinct peaks in the thirteenth and seventeenth centuries. These peaks are confirmed in an analysis of dustfalls in China from AD 300 to the present by Zhang (1985). Other peaks during this period are from 1060-1090, 1470-1560 and 1820-1890, and Zhang has correlated these phases of high dustfall frequency with cold climatic phases. He also shows that dustfalls are most frequent from February to May, with 26% of all falls occurring in April. The most common areas of reported falls correspond closely to the areas of loess deposits in China.

8.4 LONG-RANGE TRANSPORT

Several lines of evidence attest to the long-range transport of dust from northern China and Mongolia. Dust throughout China obscures the horizon and causes red sunsets ('red soils in the sunset') as far south as Hong Kong (Bell **et al** 1970). Serious consequences may be attributable to the long-range transport of radioactive material by dust events in the southern parts of Xinjiang Autonomous Region (stretching from the Taklimakan to the Gansu Corridor). Dust from the nuclear testing ground at Lop Nor, where China has been testing since 1964, is thought to be responsible for numerous unexplained cases of early death from cancer, deformed newborn children and lambs (Becker 1986).

Dustfalls, referred to above, have been studied in Beijing by Liu **et al**(1981) and Yang **et al** (1981), who showed that material brought in the upper westerly jet stream from the Gansu Corridor and western Inner Mongolia resembles the typical Pleistocene loess of China in composition and texture. This particular dustfall (in April 1980) can, therefore, be classified as a modern loess. They estimate an annual dustfall on Beijing of about 0.1mm/yr.

Nearer the main dust sources, Derbyshire (1983) reports modern dust deposition rates of "several mm/yr" at Lanzhou on the Loess Plateau where Holocene loess deposits are upto five metres thick. The particle shape, size and fabric of loess samples from this area are consistent with derivation by deflation of silts from wadis, fans, playas and desert plains to the north and west of the Loess Plateau (Derbyshire 1983, 1984).

Dustfalls are also common phenomena in Japan where they are known as **Kosa. Kosa** is yellow or brown airborne dust which is transported from the Asian continent and causes local colouration of precipitation such as red snow, black rain and yellow fog (Tsunogai **et al** 1972). These soil particles are considered to play an important role as ice nuclei in the formation of snow crystals in the season of the north-west monsoon (Ishizaka 1972). Observations at Nagasaki in southern Japan over a 62-year period show an average of 5.3 days a year with **Kosa** events, with an annual range of 0 to 18 days. During this period 85% of all events occurred between March and June (Arao **et al** 1979, quoted in Uematsu **et al** 1983). Dust concentrations over Japan during a **Kosa** event may reach 60 microgrammes/m^3 (Kadowaki 1979) and Tsunogai **et al** (1972) estimate that **Kosa** deposition is equivalent to an annual rate of sedimentation of 2 to 20g/m^2 in the pelagic sediments of the north Pacific. Arao and Ishizaka (1986) estimate the mass flux of **Kosa** dust over Japan in latitudes 30^{o}-41^{o}N to be about 4.1-5.3 x 10^6 tonnes/yr. The mass size distribution of material during April 1979 events had a single mode at 4 microns.

Although **Kosa** events have been conclusively shown to be derived from the Chinese and Mongolian Asian dust sources, by satellite tracking of individual events (eg Ing 1972), dust transported in the upper air may pass over Japan without recognition of a **Kosa** event on the ground (Kobayashi **et al** 1985). Lidar (laser radar) measurements at Nagoya in April 1979 showed two distinct dust layers: one at 0.5 to 2.5km altitude and another at 4 to 8km (Iwasaka **et al** 1983). Air mass trajectory analysis showed that the two layers originated in distinctly different source areas: the upper layer from the Taklimakan desert and the lower layer from the Gobi Desert and the Huang Ho (Yellow River) Basin.

Evidence for the regular transport of material over the Pacific occurs in the areal distribution of quartz and illite in north Pacific sediments. The greatest concentrations are found at mid-latitudes (about 30°N to 40°N) beneath the prevailing westerlies (Rex & Goldberg 1958, Griffin **et al** 1968, Windom 1975). The quartz distribution is concentrated in these latitudes in an east-west band that stretches from Japan to California (Heath **et al**, quoted in Rahn **et al** 1981), with a well-defined southern boundary and a more diffuse northern margin. Indications of two lobes to the north may correspond to frequent transport paths of dust in the atmosphere (Rahn **et al** 1981). Further support for an aeolian origin of deep-sea sediments in the north Pacific is provided by Ferguson (1970) who shows that their clay-mineral assemblages are similar to those of atmospheric dusts over the north Pacific. On a geological timescale Rea **et al** (1985) have shown that dust flux in Pacific cores varies inversely with ice volume since the Cenozoic, indicating that interglacials were periods of greater dust production than glacials in the dust source areas, contrasting the findings of workers in the North Atlantic (see section 2.4.4).

Rex **et al** (1969) have shown that the particle size of quartz in Pacific deep-sea sediments, as indicated by oxygen isotope ratios, decreases with increasing distance from the Asian coast, and the quartz and mica fractions in Hawaiian soils are unlikely to be of local origin (Dymond **et al** 1974).

In North America, ice nuclei collected in the atmosphere over College, Alaska and Blue Glacier, Washington are apparently transported from the Chinese and Mongolian deserts (Isono **et al** 1971). Trajectory analysis at the 500mb level suggests the transport time for this material was four days to Blue Glacier and 10 days to College. Darby **et al** (1974) also suggest the dust on the Arctic pack ice north of Barrow, Alaska is transported from Asia. Rahn **et al** (1981) have shown that a series of Arctic haze bands over Barrow consisted of crustal dust with a mass-median radius of 2 microns, indicating a transport path in excess of 5000km. In fact, actual Asia-Alaska trajectories are roughly 12,000 to 15,000km (Rahn **et al** 1981).

Other individual large-scale atmospheric mineral dust events have been observed over the Pacific at Hawaii (Shaw 1980, Darzi & Winchester 1982) and at sea at $30^{\circ}N$, $160^{\circ}E$ (Tsunogai & Kondo 1982).

Measurements of aerosol concentrations at islands confirm the seasonal variation in output of dust from the region. Monitoring at Enewetak Atoll (11°20'N, 162°20'E) showed relatively large dust concentrations (2.3 microgrammes/m^3) at the start of an experiment in April 1979, decreasing steadily to 0.02 microgrammes/m^3 during the next five months (Duce **et al** 1980). Duce **et al** used atmospheric Al concentrations as an indicator of continental dust produced by soil and crustal weathering processes. Some decreasing trend was also found from April to August in three lipid class compounds monitored by Gagosian **et al** (1981). The decreases in the two indicators were not exactly correlated, but this is not surprising as particles containing these lipids (plant epicutular waxes, soil or loess) probably have different terrestrial sources from dust alumino-silicates. Similarly, ^{210}Pb activity of particles collected during the Enewetak programme dropped markedly during the five-month experiment (^{210}Pb derives ultimately from a gaseous land source (^{222}Rn) but is formed and attaches to particles in the atmosphere).

Other seasonal variations have been observed by Parrington **et al** (1983) in upper level winds at Hawaii over a four-year period. More recently, Uematsu **et al** (1985) recorded particulate Al and total suspended particulate matter (TSP) in surface waters of the northwestern north Pacific showing mean particulate concentration of Al in early July 1978 (5.5 microgrammes/kg) to be almost eight times higher than in mid-August.

In January 1981 a network of dust monitoring stations in the north Pacific became operational as part of the Sea/Air Exchange (SEAREX) Program. The SEAREX Asian Dust Study (SADS) consists of seven stations. Observations from the SADS network during its first 15 months showed a seasonal transport pattern at most sites, with high atmospheric Al concentrations from February to June and low concentrations from July to January (Uematsu **et al** 1983). Annual output of dust transported to the central north Pacific was estimated to be 6×10^6 to 12×10^6 tonnes. A latitudinal gradient in the mean annual atmospheric dust concentration was found, with the greatest concentrations occurring at the northernmost station (0.89 microgrammes/m^3 at Shemya, $54^\circ 44'$N, $174^\circ 06'$E) with a smooth transition to the southernmost station (0.05 microgrammes/m^3 at Fanning, $03^\circ 55'$N, $159^\circ 20'$W). A similar latitudinal gradient has been found for mean ^{210}Pb concentrations (Turekian & Graustein 1984).

Although seasonal variations in dust concentrations at Pacific stations appear to reflect a seasonal variability in source strength that is confirmed in this work, some of the variation over the Pacific is due to the seasonal frequency of air mass trajectories from different directions. Some of the SEAREX investigations are assessing this aspect of long-range transport from Asian sources. Seasonal isentropic trajectory analyses for the station at Midway suggest that the Asian dust source strength decreases by a factor of 1.5 to 2.0 from the dusty to the clean season (Merrill & Bleck 1985). These analyses also show that stations in the eastern

Pacific may be affected by material from the Americas. While no South American sources have been definitely identified, Merrill (pers. comm. 1986) suspects that widespread burning may contribute to some of the ion (nitrate and excess sulphate) variability.

8.5 CONCLUSION

Four main areas of dust storm activity are identified in the large expanse of arid and semi-arid territory in northern China. Dust from the Taklimakan Desert in the Tarim Basin is derived from fine desert sands, alluvial sediments, fans and playas. In the Gansu Corridor wind speeds are locally increased due to funnelling between the Longshou Shan and Qilian Shan mountain ranges, strong winds entrain fine grained material in alluvial deposits brought down from the mountains and sandy loess deposits along the southern part of the corridor. Dust storms are also common in the Gobi Desert and the Loess Plateau areas.

Low pressure fronts that cross these regions in the spring are the most important synoptic scale meteorological system generating dust storms in northern China. Spring is the time of maximum activity due to the prevalence of these systems, and the human impact can be seen during the spring ploughing season when the less arid areas, such as the Loess Plateau, are made more susceptible to dust-raising. Thunderstorm downdraft dust storms occur in the Gobi Desert, and katabatic effects on wind flow may be important in the Tarim Basin.

Considerable long-range transport occurs from these source areas. Most of this transport occurs in the upper westerlies and is carried southeastwards to the Loess Plateau forming modern loess deposits, eastwards over Japan to the Pacific as far as Hawaii and some material follows a curved transport path to Alaska. Dust from Asia makes a significant contribution to north Pacific sedimentation, the seasonal variation in continental material reaching Pacific stations reflects both the seasonality of dust-raising in northern China/Mongolia and also seasonal variations in air mass trajectories reaching Pacific monitoring stations.

CHAPTER NINE : AUSTRALIA

9.1 INTRODUCTION

Data were made available by the Bureau of Meteorology, Melbourne, in the form of monthly frequencies of dust storm days for mainland Australia and Tasmania. The lengths of records vary, but a sufficient number of stations were operational by 1957 to enable a reasonable distribution map to be compiled showing mean annual dust storm day frequencies for the period 1957-82. This map is shown in figure 9.1, with annual isohyets and the area of maximum dust storm activity as defined by Loewe (1943) who produced the earliest comprehensive analysis of dust storms in Australia using observations made during the period 1938-42. His area of maximum activity is not, however, quantified in any way.

This map was published with some comments on the distribution illustrated by the data (Middleton 1984). The areas of high frequency in the Simpson Desert and the Mallee are understood by reference to surficial sediments and synoptic meteorology, but the high frequency area in northwestern Western Australia is not so easily explicable. Further investigation of the dust storm system in this region revealed that this high frequency area was not a true reflection of dust storm activity, where a dust storm is defined as an event reducing visibility to <1000m. After consultation with the Bureau of Meteorology Head Office in Melbourne, it was discovered that all dust events, whether causing visibility to be reduced to below 1000m or not, were included in the compilation of dust storm day data until 1961 (Buckley, pers. comm. 1986). This meant that localised dust devils or minor blowing dust episodes were included in the records up until 1961. From 1962 onwards only those occasions when the visibility was obscured to below 1000m were used to compile numbers of dust storm days. Thus, figure 9.2 was prepared to show the distribution of dust storms using the data for the period 1962-82 (21 years).

9.2 FREQUENCY, DISTRIBUTION & SEASONALITY

Figure 9.2 shows that dust storms have been reported from all parts of Western Australia except an area in the extreme south-west, and from all over South Australia except for some coastal stations in the extreme south-east. In Northern Territory and Queensland dust storms are very largely absent north of about $15^{o}S$, but occur in all other areas. Most parts of New South Wales experience some degree of dust-raising activity but the state of Victoria is largely only affected in the northern regions. In Tasmania dust storms have been reported in all areas except the west coast. As Loewe (1943) observed, the greater part of Australia is affected by dust storms, but a considerable part of this area experiences too few per annum to represent any regular serious hazard to human activities.

On figure 9.2, areas of average frequency greater than four days with storms a year are shaded. The greatest frequencies occur in the most arid region, centred on the Simpson Desert and running from Alice Springs (with 5.9 dust storm days a year, the highest in the country) in Northern Territory to the Sturt Desert and Strzelecki Desert in the north-east corner of South Australia. This high-frequency area lies largely within the 200mm annual average isohyet (fig 9.2). A secondary area of high activity, centred on Mildura (4.8 dust storm days a year), is in a region of irrigated agriculture between the 200mm and 400mm isohyets.

These two areas of highest frequency are broadly similar to those shown on the original map (fig 9.1) but the area of high activity in northwestern Western Australia shown on figure 9.1 is not confirmed by the true average dust storm frequency distribution. This difference illustrates a number of points, both on the deficiency of the data and the nature of the dust-raising in these areas. These points are made with reference to figures 9.3, 9.4 and 9.5 showing the long-term dust day frequencies for Onslow, Alice Springs and Mildura respectively.

As noted above, the data upto and including 1961 are for all dust events, irrespective of visibility limits, and those data from 1962 are for dust storms (visibility <1000m) only. The inflated 1957-82 average annual dust day frequency for Onslow is clearly a reflection of a peak in activity in 1959 and 1960 when 40 and 37 dust event days respectively were recorded. This period was also one of markedly increased dust-raising activity in many regions of Australia, and is depicted in figures 9.4 and 9.5 at Alice Springs and Mildura. The annual rainfall for Alice Springs (fig 9.4) indicates that this was a period of low rainfall at this station. Indeed, 1961 was a year of serious rainfall deficiency over 46% of the national land area, while the northeastern part of Western Australia (including Onslow and Carnarvon) experienced drought in 1959 (Gibbs & Maher 1967). Thus, although the inclusion of data for the less severe dust events in figure 9.1 obviously lead to inflated average annual values at many stations in Australia (as compared to fig 9.2) this inflation was exaggerated since these years (1957-61) included periods of below average rainfall. During these periods of pluviometric shortage the frequency of all types of dust events can be expected to have been above average. While in the two areas of high activity, in the Simpson Desert and around Mildura, this increase seems to have been in proportion to their long-term average level of dust-raising activity, in northwestern Western Australia the increase was more marked than in other relatively unimportant dust-raising areas.

Buckley (pers. comm. 1986) suggests that the large numbers of dust event days at Onslow and Carnarvon prior to 1962 is due to the prevalence of dust devils at these locations. Dust devils occur on hot, dry days throughout

Western Australia, but Buckley notes that they are 'generally only reported by those stations staffed by Bureau of Meteorology personell- eg Onslow up till 1975, Carnarvon, Kalgoorlie, Forrest, Meekatharra etc.' (Buckley, pers. comm. from the Australian Bureau of Meteorology, 1986). This indicates that the numbers of dust event days reported from stations manned by part-time observers may be under-representative of actual dust-raising activity, especially in the less severe categories.

The two main areas of highest dust storm activity indicated in figure 9.2 contrast with that described by Loewe, who found that dust storms were most frequent not in the most arid parts of the continent but in the slightly moister regions of marginal agriculture in parts of the Wimmera, the Mallee and Riverina. There are a number of possible explanations for this difference in distribution.

Firstly, as has been indicated above, Loewe's data may include events other than dust storms with visibility below 1000m, producing locally inflated frequencies of relatively minor dust events. Thus Loewe may be describing a different set of dust processes. In the light of this suggestion it is important to note that much of eastern Australia in New South Wales and South Australia experienced a period of drought from 1939 to 1945 with generally more frequent strong winds (Ward 1984). This drought covers four years of Loewe's five-year data period. Whether or not Loewe's definition of a dust storm includes a <1000m visibility limit, and the information from the Bureau of Meteorology suggests that it did not, this period of below average rainfall may still have produced a local increase in 1000m dust storms. If there has been a shift in locus of most intense dust-raising activity, however, this might also be explained by a change in the erodibility of soil types (perhaps due to improved conservation practices in Wimmera, Mallee and Riverina).

Related to these explanations is the fact that Loewe's data period is just five years, as compared to the 21-year averages used to construct figure 9.2, and the number of operational meteorological stations during the period 1962-82 is much greater than during Loewe's data period (1938-42). Further, Loewe himself questions the difference in recorded activity between the driest and more marginal areas, suggesting that it may be more apparent than real as dust storms in marginal areas may be more strictly recorded because they are less common in such regions.

Dust storms in Australia are markedly seasonal in their occurrence, and appear to be a feature of late spring and summer months: September/October to February/March (Table 9.1). No seasonality is indicated in this table for stations outside the main areas of activity as annual frequencies are so small, but Loewe indicates that this seasonality holds true for all Australian regions, including the northern region of prevailing summer rains, where Loewe suggests that:

> ....the instability of stratification produced by the

increased heating of the ground and the stronger evaporation act more powerfully than the stronger cohesion of the surface deposits during the rainy season. (Loewe 1943, p8).

Even at Alice Springs and Oodnadatta, in the driest part of the country, average dust storm frequency is highest during the 'rainy' months (fig 9.6). At Alice Springs dust storm frequency, rainfall and average wind speed are all at maxima during the late spring/early summer months, as rainfall is brought by the meriodional frontal passages and associated thunderstorms that also initiate dust-raising. At Oodnadatta a broadly similar pattern emerges, although at this station monthly rainfall totals are very low and dust frequency seems to be more closely related to wind, as represented by Drift Potential.

9.2.1 DIFFERENCES IN DUST STORMS RECORDED INSIDE & OUTSIDE URBAN AREAS

A notable aspect of data from cities with more than one meteorological station is the differences in frequencies of dust storms recorded. Table 9.2 shows comparative average annual frequencies for pairs of stations at Sydney, Adelaide and Brisbane. In these cities the stations located on the outskirts of the urban areas (the airports at Sydney and Adelaide and the Area Meteorological Office at Brisbane) show greater frequencies than stations located within the urban area. This phenomenon is explicable in terms of the outer stations being closer to potential dust sources while potentially wind erodible areas within cities have greater protection by the urban surfaces. A similar trend for stations located inside and outside urban areas in Soviet Kazakhstan and Turkmenistan is noted by Sapozhnikova (1973).

9.3 METEOROLOGICAL SYSTEMS GENERATING DUST EVENTS

The synoptic systems that generate dust-raising winds have been well summarised by Sprigg (1982). He identifies three types of dust-bearing wind systems that are clearly related to the average summer season isobaric chart.

The classic southern Australian dust storm is an advancing turbulent dust wall caused by an eastward moving frontal system. Some of the dust may be raised high enough to enter the upper level westerly airflow which may create dust haze conditions along the New South Wales coast and lead to long-range dust transport over the Pacific as far as New Zealand (see section 9.9 below).

The second type of dust-raising wind is associated with the hot northerly wind of late winter and spring that preceeds these frontal systems. Dust raised by these winds is carried southwards usually as a less obviously wall-like structure. Material raised in the hot interior may reach the southern coast in these hot winds. Count Strzelecki recounted how on one occasion arriving from New Zealand he was prevented from entering Port Jackson for two days by the violence of the hot wind. The temperature 60 miles offshore was 90°F (32°C) and the ship's sails 'were covered with a quantity of impalpable dust' (De Strzelecki 1845, p166). Such dust may also be carried great distances if it reaches the upper westerly air stream.

The third type of dust storm is caused by winds from the south, blowing along the concentration of isobars in advance of approaching anticyclones (moving west-east), the passage of which are characteristic of the circulation over Australia south of about 30°S. These winds are primarily southwesterlies in the south that back to southerlies and then to SSElies in the Simpson Desert. In the southern interior, especially in Victoria, these hot, dusty winds are known as **Brickfielders**, a term that was brought by gold prospectors from the Sydney area where strong southerlies would raise much dust from brick fields around Sydney. The terms **Cobar Shower** and **Darling Shower** were used ironically in New South Wales for such winds (Gentilli 1971). Further north dust raised by these winds merges into the south-east Trade winds and then the zonal easterlies in the region of Tennant Creek. The dust-raising winds and entrained material may carry on towards north-west Australia. Dust haze is not uncommon over some considerable distance off the north-west coast in the belt of the tropical easterlies (Gentilli 1971, see also McDonald's 1938 haze maps). These three generic categories of dust-raising synoptic systems are depicted in figure 9.7.

In addition to these synoptic scale influences, some mesoscale dust-raising systems should be noted. Thus, in the Alice Springs district, where the frontal passages that are so prominent in the southern coastal districts are experienced to a lesser degree, dust storms are often caused by downdraft winds from thunderstorms that frequently accompany frontal passages in this region (Parkes **et al** 1985). Thunderstorm downdraft dust storms are also common in the hot interior of South Australia during summer. Lindsay (1933, 1936) suggests that they are usually associated with severe thunderstorms (line squalls). These systems typically move west-east and often mark the end of a heatwave. One example of the characteristic dust wall structure hit Anama Station in February 1930 extending over a front of 190km and travelling for nearly 320km. Thunderstorms are also frequent in the north of the continent during summer due to intense heating of the land surface creating atmospheric instability.

In the north dust may be raised during the dying stages of tropical cyclones, as when Hurricane Cecile crossed the coastline of Northern Territory and tracked inland towards the dry interior in March 1984 (Middleton **et al** 1986). The approach of a tropical cyclone may also cause dust storms in Western Australia in areas not yet affected by the precursor rain bands. Such dust storms occur north of 26^{o}S and west of 122^{o}E from November to April (Buckley, pers. comm. 1986).

Dust devils are common throughout the arid and semi-arid regions of Australia, Gentilli (1971, p107) reviews a number of vivid accounts of these phenomena and notes that the terms **Willy Willy** and **Cockeye Bob** are used fairly indiscriminately for tropical cyclones, tornadoes and small whirlwinds.

Since some form of relationship between aridity and high dust storm frequency is suggested by figure 9.2, further investigation was carried out using Spearman's rank correlation coefficient. For 30 of the stations shown in figure 9.2 dust storm frequency was correlated against mean annual rainfall and the annual number of 'rainy' days (defined as >0.25mm). The relationship in both cases was weak, however, with correlation coefficients -0.46 and -0.53 respectively.

These findings suggest that factors other than simple rainfall amounts are important in determining dust storm activity in Australia - a conclusion reached for a number of other countries in this work and also by Goudie (1983). Nevertheless, it is interesting to note that the map of rainfall variability, as expressed by the coefficient of variability, shown by Gentilli (1971, p158, fig32B) depicts two areas of highest variability. One small region is centred on Onslow (Western Australia) with 60% variability, and a larger area is roughly centred on the Simpson Desert with a maximum of 50 % variability.

9.4 SOIL

Some broad-scale observations can be made on the relationship between dust storm frequency and soil type by referring to the map of Australian soils produced by Northcote **et al** (1975).

The main area of activity comprises the red siliceous sand dunes of the Simpson and Sturt Deserts, crusty red duplex or texture-contrast soils, massive non-calcareous earths, and self-mulching clays. The region of high frequency centred on Mildura is made up of depositional, self-mulching cracking clays and calcareous earths on aeolian dunes.

By contrast the large tracts of Western Australia where less than one dust storm is recorded per year are made up of loams with red and brown hardpans, pedal and non-pedal hard yellow duplex soils, earthy sands and sands of the Great Sandy and Gibson Deserts (although these dune areas have very few data points). The area within the one-dust-storm isokon in south-east Western Australia (the Nullarbor Plain) largely comprises powdery calcereous loams, siliceous sands and calcareous and non-calcareous earths.

Although these general soil characteristics can be highlighted with regard to differing dust storm frequencies, direct relationships between soil type and dust production can be made only at much finer scales of analysis. For example, it may be that dust production is concentrated on a soil type of limited areal extent that is not shown on a large-scale map, and human impact through land management may be responsible for major differences in wind erodibility on the same soil type. Further analysis is presented below for the two main dust storm regions.

9.5 SIMPSON DESERT / LAKE EYRE BASIN

Further investigation of the dust storm system in Australia has been undertaken by McTainsh (1985a), who notes that the area of maximum occurrence in Australia is broadly coincident with the Lake Eyre Basin, and as an area of sediment supply it closely resembles the Lake Chad Basin in North Africa. Material is entrained from alluvial spreads brought by the southward flowing Eyre, Diamantina and Cooper rivers by strong winds caused by the synoptic and mesoscale systems outlined above.

Dust may be raised over wide areas or at specific point sources as evidenced by Space Shuttle imagery (plate 9.1). Twidale (1972) considers these contemporary riverine sediments to be important sources for Simpson Desert sands. He shows so-called 'debris mounds' on the alluvial plains that feed into the parallel linear dunes that are common throughout the Simpson. It is possible that these same debris mounds act as point sources for dust plumes as shown in plate 9.1. Material is also raised from playas and salinas of the Eyre/Poepples area (the Kallakoopah pans) of the southern Simpson Desert (Sprigg 1982) and McTainsh suggests that the saltating dune sands of the Simpson may act as a dust entrainment mechanism (McTainsh 1985a).

9.6 MALLEE

Land management may have major impacts on the wind erodibility of soils, such as in the area of high dust storm activity in north-west Victoria and south-west New South Wales, where dust sediments are derived from Mallee topsoils which have been affected by the expansion of agriculture into these marginal semi-arid lands (Heathcote 1983). During drought periods these areas may be particularly susceptible to wind erosion and dust storm generation. Drought in 1977 triggered the loss of large amounts of solonized brown soils to wind action in the Mallee area (Thompson **et al** 1983), and the summer of 1982/83, when southeastern Australia experienced one of its worst droughts on record, was characterised by dramatic large-scale frontal dust storms (Lourensz & Abe 1983, Garratt 1984).

Cultivated soils may not be the only dust sources in this area, for as Dare-Edwards (1984) points out the Murray Basin region has many relict lakes, dry swamps, palaeo-river channels and old groundwater discharge points, any of which may act as local sources. It is possible also that the Mallee dunefield, that includes high proportions of clay pellets, may produce reworked fine sediments liberated by attrition.

Plate 9.2 shows the advancing dust wall of one of the frontal storms that struck southeastern Australia during the summer of 1982/83. This photograph was taken in Melbourne at about 1500 LST on 8th February 1983, from the roof of the 27 storey building which houses the Head Office of the Australian Bureau of Meteorology. The dust was raised by north to northwesterly winds blowing in a pre-frontal trough (cf. Sprigg's second category above) as a cold front moved eastwards across south-east Australia (Lourensz & Abe 1983). The winds blew with an average speed of 33km/h (18knots) with gusts upto 80km/h (43knots) and raised material from drought-stricken Mallee topsoils, the area of open grazing and wheat lands which had become a virtual desert during the prolonged drought period. The raised dust extended for some 500km from Mildura down to the coast near Melbourne and was up to 100km wide. At source the dust was reported at heights of 2800m with one report of 3650m in the Mildura area, while as the dust wall approached Melbourne it was 320m high.

In Melbourne half an hour before the wall arrived the temperature was 43.5^{o}C, the highest recorded February temperature in the city since records began. At the leading edge of the storm the temperature dropped instantaneously to 29^{o}C (McGuckin 1984). The storm took about 90 minutes to pass through the city, reducing visibility to below 100m. The three Melbourne airports were closed for 35 minutes at the height of the storm and road and rail transport in and around the city was brought to a standstill. Power lines were blown down, telephone exchanges jammed with emergency calls, and many of the city's inhabitants suffered irritation to the eyes and throat (The Sun 1983). Dustfall deposition rates in the Melbourne suburbs during the storm were estimated to be 106kg/ha (Lourensz & Abe 1983).

9.7 WESTERN NEW SOUTH WALES

The dust storm system in western New South Wales has been studied by Thompson (1982) who related seasonal incidence, direction and duration of storms at Broken Hill to wind flow, moisture status and pasture conditions. The majority of dust storms occur in spring and summer and blow from westerly directions, with 50% of all storms from the north-west, except in the summer when southerly events are dominant. Spring and summer are not the seasons of strongest winds, however, Jones (1962) shows that winter is the season of greatest number of hours with high velocity winds at Broken Hill. This apparent inconsistency between dust storm activity and wind flow is explained by Jones, who suggests that spring and summer winds are generally more turbulent than winter winds as the warmer seasons are characterised by higher temperature lapse rates. A second important parameter is that of pasture cover, which affects both the seasonality and year-to-year variation in dust-raising activity (Thompson 1982).

Wind is a primary cause of erosion in the semi-arid and arid areas of western New South Wales; extensive overgrazing in the late 1800s and early 1900s, together with frequent droughts, resulted in the rapid deterioration of pastures, which changed from perennial to annual species in many areas during this period. At Broken Hill vegetation clearance and construction around the town in conjunction with mining activities added to the erosion problems. Summer dust storms became so bad in the 1930s that the mining companies provided funds to the town council to plant protective tree belts around the town from 1936 (Heathcote 1983). By the 1970s a green belt was established, but as Thompson (1982) points out, the continuing occurrence of dust storms reflects the arid climate, when even in good years vegetation rarely covers more than 60% of the surface of susceptible soils around Broken Hill.

9.8 VARIATIONS THROUGH TIME

The increase in dust storm activity during dry years in Australia is noted by Loewe (1943) for the country as a whole and by Thompson (1982) at Broken Hill. This observation is partly confirmed by long-term dust event frequency data from a number of stations from all parts of the country (eg figs 9.3, 9.4, 9.5).

The drought of 1977 is reflected by minor peaks in dust storms at Alice Springs and most of the stations to the south, but not to the north.

It is unfortunate that the hypothesis put forward above suggesting that Loewe's high frequency area is biased due to drought over eastern Australia during his data period cannot be confirmed from these data. No stations situated in Loewe's high frequency area have records dating back to 1942 or before, and the quality of the data does not allow any reasonable observations to be made on this hypothesis. Heathcote (1983), however, suggests that the 1940s was a period of high dust storm activity, indicating that dust from the wind eroded wheatfields of southern Australia coloured the sunsets of the eastern capitals and even reached the snowfields of New Zealand.

The validity of relating annual rainfall totals to concurrent annual dust storm totals also holds true in particularly wet years. Gibbs (1975) notes that 1973 and 1974 were years of extremely heavy rainfall over much of the country, and these years show minimal dust storm frequencies at all stations. Bigg and Turvey (1978) studied the sources of atmospheric particles over Australia, sampling from aircraft in October 1974, November 1975 (years of high rainfall and low dust storm activity) and May 1977 (although a drought year, during the season of generally low dust storm activity as indicated in table 9.1). Largely as a result of the sampling periods, therefore, they concluded that 'the land surface itself is almost ineffective in generating particles' (Bigg & Turvey 1978, p1649).

The effect of antecedent rainfall on wet deposition dustfalls has been studied over a period of six summers by

Hutton (1980) at Merbein, just north of Mildura, Victoria. Calcium carbonate suspended in the atmosphere is derived from the fine calcite particles present in the soils of the area, and the amount deposited per hectare in rain during the summer was found to be inversely related to the rainfall recorded during the two previous springs and winters. Hutton observes that the total annual weight of carbonate deposited appears to be independent of the amount of rain. However, when the annual totals of dust event days are examined in conjunction with Hutton's figures (table 9.3) the situation becomes clearer. In 1958 for example, when annual rainfall was at a maximum for the six-year period, deposited calcium was below average for the period, which seems to be related to the small number of dust event days (4). In 1960, by contrast, although annual rainfall was over 100mm lower, the number of dust event days was much greater (29) and consequently the deposition rate was at a maximum for the study period.

9.9 LONG-RANGE TRANSPORT

There is a considerable body of evidence indicating that dust is transported over long distances from Australian dust sources, both currently and in the recent geological past. Long-range transport to New Zealand is a well-established phenomenon. For example, in November 1902, after a particularly dry winter in Australia, 'brown rain' was observed in Tasmania (Dixon & Dove 1903) and 'red rain' in New Zealand (Marshall 1903). The composition and mineralogy of particles in this dustfall in New Zealand was similar to that of widespread dustfalls in 1928 over South Island and half of North Island, and these dusts were similar to material from Gippsland and the Nullarbor Plain in South Australia. The dustfall of 1928 was between 3 and 30grammes/m^2 (Marshall and Kidson 1929). A further fall was reported in 1929 (Kidson & Gregory 1930), and Windom (1969) measured dust in snowfalls over the southern Alps of New Zealand in 1966, estimating a deposition rate over the south-west Pacific of 0.8mm/1000yr.

The most comprehensive analysis of Australian dust reaching New Zealand was for material that fell in rain on South Island in November 1982. An Australian origin was established in this case by trajectory analysis, consideration of antecedent Australian weather conditions and mineralogical analysis (Collyer **et al** 1984). A review of aeolian transport across the Tasman Sea is provided by Glasby (1971) who notes that pollens and spores are also carried from southern Australia to New Zealand (Newman 1948, Moar 1969).

Australian dust may be transported as far as Antarctica (Shaw 1979) which is supported by simulations in a General Circulation Model by Joussaume **et al** (1984), and Durst (1935) suggests that material from north-west Australia may reach Java and Singapore. Material carried northwestwards from this part of Australia occurs by means of a strong circulation in the zone of convergence between a summer (November - March) heat low located over north-west Australia and anticyclones that may be resident over the Great Australian Bight for several days. Such winds blow from the south-east across northwestern Australia, and may transport material out over the Indian Ocean (Heirtzler 1974). McDonald's (1938) haze frequency maps confirm these long-range transport paths.

Modern dust deposition rates in snow collected in the Australian Alps may be as high as 1700kg/ha for individual events (Walker & Costin 1971) and there is considerable evidence for dust deposition during the Quaternary. The soils of large parts of Victoria, New South Wales and southern Queensland may have received a significant aeolian input during arid phases of the Quaternary. There seems to be a general agreement that there is no loess in Australia, the term 'dust mantle' is preferred (Wasson 1982), otherwise known as 'parna' in the Riverine Plain (Butler & Hutton 1956). Bowler (1976) has developed a model of dust flow during the arid phases of the Quaternary, identifying two major areas of long-range transport, one off the north-west coast and another off the south-east coast. These flows fit in well with the current distribution of dust storm activity shown in figure 9.2 (McTainsh 1985a), while the present-day area of the 'roaring forties' across south-east Australia, entraining material from the Mallee area, may represent a shift from an area of 'roaring thirties' that would have represented a greater level of activity in the Lake Eyre Basin (Sprigg 1982). Bowler's model of two major dust flow systems also fits in well with the ocean sediment record as shown by Kolla and Biscaye (1977) who indicate an area of high quartz content over the Indian Ocean off north-west Australia for example, and Griffin **et al** (1968) and Kolla and Biscaye (1973) who identified high kaolinite concentrations in the same area. Similar evidence is presented for the Tasman Sea (Kidson & Gregory 1930) and the south-west Pacific (Thiede 1979).

9.10 CONCLUSION

This chapter has illustrated a number of important problems with dust storm data. A first analysis of dust storm frequency and distribution used data supplied by the Australian Bureau of Meteorology, but more detailed investigation of the high frequency areas indicated by these data revealed that the data set was not homogeneous. A complete revision of the analysis was carried out using dust storm day data that incorporated a visibility limit of 1000m. The new analysis revealed that inclusion in the original analysis of five years' data that incorporated no visibility limit had inflated average annual values throughout Australia, with particular inflation in an area of north-west Western Australia. Data from pairs of stations at individual cities indicate that stations on the fringes of urban areas record higher frequencies than those located inside cities.

Two areas of high dust storm frequency have been identified in Australia. The highest level of activity occurs in an area centred on the Simpson Desert and broadly coincident with the Lake Eyre Basin. Dust sized debris in this region is derived from the alluvial spreads of rivers flowing southward into the basin and desert depressions. The role of saltating dune sands may be an important dust entrainment mechanism. The second area of high activity is in the semi-arid zone of irrigated agriculture centred on Mildura. Here dust is available from agricultural fields when vegetation cover is absent and from a range of palaeofeatures such as old lake beds, former dune fields and palaeo-river channels.

Dust storms are characteristic of late spring and summer when dust is raised by winds associated with passing frontal systems or along steep pressure gradients associated with areas of high pressure, both types of system moving eastwards across southern Australia. Other dust-generating meteorological systems include thunderstorm downdraft storms in many areas and tropical cylones as they track inland over the northern parts of the country.

Investigation has been made of the relationship of dust storms to meteorological parameters. For 30 Australian stations negative but insignificant relationships were found between average annual numbers of dust storm days and annual rainfall and average annual numbers of rainy days. Closer analysis at two stations shows that dust-raising occurs during months when wind speeds and rainfall are at a maximum, due to the fact that frontal systems and local thunderstorms are largely responsible for dust storms at Alice Springs and Oodnadatta and these systems also bring rainfall to those areas and cause high wind speeds. At Broken Hill, dust storms are not characteristic of the windiest months (winter) but of those months when winds are more turbulent due to higher solar heating rates (spring & summer). Pasture cover is also an important factor affecting dust-raising at Broken Hill, and periods of low rainfall or drought generally result in increased dust-raising through its effects on vegetation cover and soil moisture properties.

There are two major paths for long-range transport of dust from Australia which are confirmed by dustfalls and the evidence of deep-sea sediments. These paths are southeastwards from southern Australia towards New Zealand and over the south Pacific and northwestwards from north-west Australia towards south-east Asia.

CHAPTER TEN : NORTH AMERICA

10.1 USA

It has been estimated that of five billion tons of soil lost in the USA annually one billion tons is due to the effects of wind erosion (Pimental **et al** 1976). The presence of soil-eroded atmospheric particulates from desert, farm fields, unpaved roads and construction areas prevents 33 major cities in 14 states from meeting legislated air quality standards according to The Council on Environmental Quality (Kimberlin **et al** 1977). Partly because of the importance of aeolian erosion, and partly because of the dramatic impact on the national consciousness of the Dust Bowl of the 1930s on the Great Plains, dust storms have been studied in the USA more intensely than in any other country. These studies include analyses of dust storm frequency and distribution, generating synoptic systems, properties of erodible soils, the human impact on dust storms, wind erosion control strategies and predictive modelling. Thus, most of this section is made up of a review of these works, with some new material on certain areas, particularly the Great Basin Valleys Air Basin of southeastern California.

Two papers deal comprehensively with the frequency and distribution of dust storms in the USA: Orgill and Sehmel (1976) and Changery (1983). Both studies use summarised hourly visibility data from National Weather Service and military stations. Orgill and Sehmel's study includes the whole of the contiguous USA and defines a dust storm as a dust event reducing visibility to <7 miles (11.3km) using data for the period 1940 to 1970. Changery, however, concentrates on the western half of the country, using data for a number of visibility limits (<7 miles, <3 miles, <1 mile, and <5/8 mile [about 1000m]) in the period 1948 to 1977.

Orgill and Sehmel identified the highest dust frequency as being in the southern Great Plains, with secondary dust frequency maxima in the western states, the northern Great Plains, the southern coastal Pacific and inland valleys and some areas of the south-east. Changery confirms the Great Plains as the area of maximum occurrence, with the other high frequency areas being located in southeastern New Mexico along the Pecos River valley into south-west Texas; southward from the White Sands, New Mexico region into extreme west Texas; two areas in Utah; western Nevada; southeastern California into south-west Arizona; the San Joaquin Valley, California; and other small areas in southern Idaho and much of eastern Washington and northeastern Oregon (Fig 10.1). These areas will be investigated in detail below, organised into the following regions.

1. The Great Plains and southern New Mexico.
2. Arizona and Southern California .
3. Nevada, Utah and other regions.

10.1.1 THE GREAT PLAINS & SOUTHERN NEW MEXICO

The Great Plains is the centre of maximum dust storm activity in the United States. It is an area prone to recurrent drought and accompanying windstorms, where human actions have at times had major impacts on the wind erosion system. The Dust Bowl of the Great Plains in the 1930s is perhaps the best-known and most oft-quoted example of large-scale dust storm activity anywhere in the world. The core of the Dust Bowl area comprised the western third of Kansas, south-east Colorado, the Oklahoma Panhandle, the northern two-thirds of the Texas Panhandle and north-east New Mexico, although most of the Great Plains experienced dust bowl conditions at some time during the 1930s. Indeed, some of the most severe conditions were found as far north as Wyoming, Nebraska and the Dakotas.

The most severe dust storms occurred in the Dust Bowl between 1933 and 1938, with activity being at a maximum during the spring of these years. The concentration of dust storms on the Great Plains in March 1936 is shown in figure 10.2, the anomalous scale of events is indicated when this figure is compared to the long term average annual frequency shown in figure 10.1. At Amarillo, Texas, for example, at the height of the Dust Bowl, one month had 23 days with at least ten hours of dust, and one in five storms had zero visibilty (Choun 1936).

The 1930s was a period of drought on the Great Plains, and although dust storms are frequent in the area during dry years the scale and extent of the 1930s events were unprecedented. The reasons for this most dramatic of ecological catastrophes have been widely discussed, and blame has largely been laid at the feet of the pioneering farmers and 'sod busters' who ploughed up the plains for cultivation.

Cultivation of the Great Plains started in the late 1870s, and the natural sod-forming grasslands were slowly transformed into wheat fields. Waves of settlers arrived in the area during high-rainfall years, and a surge of settlement from 1914 to 1930, in conjunction with the increasing use of mechanised agricultural techniques, catalysed by high wheat prices, led to unprecedented large-scale wind erosion when drought hit the Plains in 1931. Worster (1979) emphasises the role of social and economic forces in the mismanagement of the Plains, the product of a culture set on dominating and exploiting the land and its natural resources. The conversion of large areas of grassland to crops, without regard for the suitability of the soil or the climate for such farming, certainly amplified the effects of the drought (Lockeretz 1978). Although this view of the agricultural mismanagement as a primary cause of the Dust Bowl is generally accepted, it has been argued that the result was an essentially natural phenomenon:

No more brazen falsehood was ever perpetrated upon a gullible public than the allegation that the dust storms of the 1930s were **caused** by 'the plow that broke the Plains.' (Malin 1956, p356).

Indeed, although open ploughed fields were undoubtably a very important source of dust storm sediments, the large areas of alluvial plains in northern Texas, Oklahoma, Kansas and Colorado (eg the Cimarron River and the Canadian River) may also have represented significant dust sources. These areas were almost uncultivated, but had lost much of their protective vegetation cover during the drought years (Wells, pers. comm. 1985). The proportions of agricultural and non-agricultural land affected has never been estimated, but in 1937 the US Soil Conservation Service estimated that 43% of a 6.5 million hectare area in the heart of the Dust Bowl had been seriously damaged by wind erosion.

Although in the public mind the Dust Bowl is associated with the advancing dust wall of the classic thunderstorm downdraft described by Krumm (1954) and Elser (1959), it is more likely in fact that the most important synoptic influence on dust storms originating in the Dust Bowl were eastward moving frontal depressions. Borchert (1971) shows that the drought of the 1930s was accompanied by a stronger westerly circulation and above average number of frontal passages, but with little moist air. Thunderstorm downdraft dust storms are relatively short-lived and of a smaller scale than the storms associated with frontal passage, which may last for several days and may raise material from a number of states (Riley 1931). The continental scale of frontal storms is shown by Parkinson (1936) in his study of Great Plains storms between 1933 and 1935. He notes that many storms were caused by unstable modified Polar Pacific air, and cleared away from a particular area by the arrival of stable Polar Continental air masses. Several of the more severe frontal storms carried dust from the Great Plains to the Atlantic coast and the Gulf of Mexico. One event, that raised material in Montana and Wyoming on May 9th, 1934, carried an estimated 350 million tons of soil eastward, depositing dust in Boston and New York in the morning of May 11th, and on ships' decks 500km out in the Atlantic during the next day or so (Worster 1979).

The transport of the Great Plains to the east coast brought the problems of the area home to the rest of America. For the inhabitants of the Great Plains, suffering like the rest of the country in the Depression years, the Dust Bowl produced a range of additional problems. At its worst in Kansas, for a period of six continuous weeks in March and April 1935 it was unusual to see a clear sky from dawn to sundown (Worster 1979). Apart from the constant aggravation of the dust storms, the dirt everywhere, and dust pervading into every aspect of day to day life, there was extensive damage to property. The fine siliceous dust caused epidemics of respiratory infections among the human population, while cattle suffocated in the severe storms and fish died in rivers whose surfaces were coated with dust. Many thought the end of the world had come, and for those who had lost all of their agricultural livelihood the answer was to leave. The 'Okies' and 'Exodusters', outmigrants from the Great Plains, contributed to one of the largest migrations in US history.

There is an approximate twenty-year drought cycle in the Great Plains. Major droughts have occurred in the 1890s, 1910s, 1930s, 1950s and 1970s, and as mentioned above these droughts are periods of exaggerated dust storm activity. Although the dust storms of the 1950s were not as spectacular as those of the 1930s, more land was actually damaged by wind erosion in the Great Plains, and soil loss in the 1970s was on a scale comparable to that of the 1930s (Lockeretz 1978). The Dust Bowl inspired major advances in wind erosion research, most noteworthy of which are the works of Chepil and his co-workers carried out at the US Department of Agriculture experimental station in Garden City, Kansas (see also chapter 2).

A major thrust in wind erosion and dust storm research in the Great Plains has been towards deriving a predictive index to enable dust storm forecasting. The Universal Equation for measuring wind erosion, developed by workers at the US Agricultural Research Service (Woodruff & Siddoway 1965) was derived in part from field studies in the Great Plains. Chepil **et al** (1963) developed the Wind Erosion Climatic Factor, C (see section 6.5.2 for further details) and found a good correlation with dust storm frequency in the Great Plains (see fig 10.3 for Dodge City, Kansas). Similarly, Fryrear's (1981a) Big Spring Index, devised to estimate dust storm frequency 12 months in advance, is based upon open pan evaporation, rainfall and wind movement. The technique estimates the yearly number of dust storms (of visibility <7km) in the southern Great Plains within four days 56% of the time and within eight days 82% of the time.

Despite the relative success of dust storm prediction, and the knowledge of wind erosion prevention techniques that Lockeretz (1978) points out was available in the 1930s, the US Soil Conservation Service annual reports of land damage leave little doubt that the factors responsible for wind erosion are not being controlled with present technology (Fryrear 1981a). Long-term decreasing trends in yields of dryland crops in the Texas Panhandle may be due to a combination of soil erosion (both aeolian and fluvial), annual cropping practices and increasing hazards such as insects and disease (Fryrear 1981b) but certainly improvements in crop varieties and cultivation practices have not kept pace with decreased crop production.

The weather conditions associated with dust storms in 1950 at Lubbock, Texas, are studied by Warn and Cox (1951). Localised storms caused by thunderstorm downdrafts, known as **Dusters**, were important, but the most severe storms in that area were associated with cold frontal passages. Localised downdraft storms, from high level thunderstorms, can be very severe as Elser (1959) describes an event at El Paso on 28th June, 1959 that reduced visibility to zero for four hours. The most important dust storm sediments from the Lubbock area in the early 1950s were associated with agricultural practices (Laprade 1957). These sources included cultivated or open fields, overgrazed pastures and ancient dunes reactivated by devegetation due to overgrazing, burning or cattle trampling. Other sources included bare floors of the numerous playas in the area and small quantities from graded roads and ditches.

In the 1970s a new phase of inappropriate cultivation practices was highlighted. Perhaps the worst single dust storm occurred after two years of drought, in the Portales Valley area of eastern New Mexico, as a low pressure frontal system moved eastwards across the Great Plains on 23rd February, 1977. In the Portales Valley dryland wheat farming has moved onto marginal land as a result of economic factors and high technology land use practices. In addition, certain government policies reduce the disincentives to cultivate marginal land, so that The Wheat Disaster Assistance Programme compensates farmers for loss of crops to wind erosion. Thus, cultivation of wheat on marginal lands, some of which were formerly dune fields, was encouraged by high prices that followed export sales of great quantities of wheat in 1975. These crops were irrigated by center pivot irrigation which requires the removal of linear wind breaks made up of trees planted since the Dust Bowl era to help prevent wind erosion (McCauley **et al** 1981). The dust palls from Portales Valley and an area in eastern Colorado/western Kansas were tracked on GOES-1 (Gurka 1977) and NOAA-5 (Vermillion 1977) satellite imagery southeastwards, obscuring 400,000km^2 of ground surface in the south central USA, crossing Georgia and South Carolina (Purvis 1977) and out over the mid-Atlantic Ocean.

The most important synoptic conditions associated with dust storms over the Great Plains have been described above, and a study of 35 severe storms between 1968 and 1977 by Henz and Woiceshyn (1980) has produced a definitive classification of the synoptic systems according to scale and duration of the event (table 10.1). Identification of a number of common predictors, such as upper level wind speeds, the location of upper and low level jets and cyclogenesis location and deepening rate, produced a fairly reliable man-machine mix forecasting technique. The technique was used to successfully predict six out of seven severe dust storms during the 1976/77 winter.

10.1.2. ARIZONA & SOUTHERN CALIFORNIA

The climatology and temporal and spatial variations in Arizona dust storms have been comprehensively studied by Brazel and Hsu (1981) and Nickling and Brazel (1984). Phoenix records the highest number of dust storms a year (2.7) and storms here tend to be the most severe, although Yuma, on the

Californian state border (1.6 storms a year) records more <11.3km dust events annually (9.4 to 6.6 at Phoenix).

The most intense and frequent dust storms in Arizona occur in summer, and are caused by strong downdraft winds from intense thunderstorm activity (Idso **et al** 1972, Idso 1973). Several less intense storms , but of longer duration, also occur in most parts of the state in late winter and early summer with a peak in April. These storms are usually associated with cyclonic storm activity, including cold frontal passages and upper level cut off lows which are common throughout the state at this time of year (Nickling & Brazel 1984). Tropical disturbances (storms & hurricanes) are the least frequent dust-generating systems, but are typically very intense, long-lasting events (Brazel & Nickling 1986).

The effects of human activity in destabilizing desert surfaces and rendering them more susceptible to wind erosion were illustrated during World War II when General Patton set up a desert training camp at Parker, Arizona. Surface disturbances due to the building of temporary camps and roads and the movements of tanks caused dust storms to increase in frequency by several times, and frequencies were as great during this period as a few years later when the area was suffering drought conditions (Clements **et al** 1963).

Dust storms represent a major hazard to road traffic in Arizona, the abrupt visibility reduction caused 32 multiple vehicle accidents between 1968 and 1975 on Interstate 10 alone (Hyers & Marcus 1981). The very close relationship between 'dust-related' accidents and thunderstorms is shown by Brazel and Hsu (1981), the large majority of accidents occurring in the late afternoon during summer. The thunderstorm systems that generate dust storms originate in southern and eastern Arizona and northern Mexico during summer. The seriousness of the problem has inspired the development of a Dust Storm Alert System on Interstate routes 8 and 10 in central southern Arizona. The system involves 40 remotely-controlled signs with changeable messages to alert motorists to blowing dust hazards and special dust alert messages broadcast on local radio (Burritt & Hyers 1981). Analysis of land use and surface sediments along Interstate 10 reveals a strong locational association between sparsely vegetated abandoned farmland and dust-related accidents. The lack of vegetation on abandoned farmland makes it a particularly effective dust source, and the problem is exacerbated by such land's accessibility to human activities such as off-road vehicle use and horse riding which disrupt the soil crust, making finer sediments available to wind (Hyers & Marcus 1981).

The contribution to atmospheric dust loadings of unpaved road traffic in Arizona has been investigated by Turner **et al** (quoted in Hall 1981). The amount of dust entering the atmosphere was found to be directly related to the silt content of gravel roads. In Pima County (Tucson region) unpaved roads were estimated to produce about $11\text{x}10^5$ kg/day of atmospheric dust which produces an atmospheric dust load of 25 microgrammes/m^3, reducing visibility to about a third of its pure value.

Among the other dust hazards in Arizona is the transport of potential disease-producing spores in blowing dust. One of the most important dust-borne diseases in Arizona is Valley Fever, which accounts for more than 27 human deaths each year in the state (Leathers 1981). The increases in the incidence of valley fever after severe dust storms testifies to the importance of this mechanism in spore dispersal. Dust transport in Arizona is believed to be of geomorphological significance as the source of $CaCO_3$ in caliche, and an important source of iron-manganese for desert varnish. An annual dust deposition rate of 54g/m^2 has been calculated for Tempe by Péwé **et al** (1981).

In California dust storms are largely confined to the southern regions. Highest frequencies occur in the Mojave Desert (5 dust storms a year) and the Colorado Desert (4 a year) with lower annual frequencies recorded in the San Joaquin Basin and the valleys of the Great Basins. Clements **et al** (1963) suggest that the most common synoptic system causing dust storms in southern California occurs in late winter and spring when the Pacific high is near to its southernmost position and a deep low pressure cell is centred in the Great Basin, in the general vicinity of Tonopah, Nevada. The resulting pressure gradient from west to east results in strong winds that usually blow in a southwesterly or southeasterly direction. Less common are the **Santa Ana** winds, that blow from the north or north-east as a hot dry katabatic airflow across the southern deserts and coastal plains, warmed by adiabatic or mechanical heating as air descends from the Western Cordillera and is compressed (Sergius **et al** 1962). The **Santa Ana** is generated by a reverse pressure distribution from that described above, with a high pressure cell centred over Nevada and a low present off the Californian coast. This situation is most common in the winter months.

These systems may be exaggerated in their dust-raising effects locally by channelling due to topographic features, such as in the San Gorgonio Pass north of the Colorado Desert where most winds are the result of marine air blowing inland to areas of lower barometric pressure. Under certain conditions an inversion layer forms over the the San Gorgonio Pass producing a 'capping' effect which further enhances the channelling.

Santa Ana dust plumes from the western Mojave Desert carry material to the Californian Channel Islands and have been tracked using Landsat imagery (Muhs 1983). Present day dustfall rates are of the order of 28 to 31 $g/m^2/yr$, comparable to other rates from the south-west USA (see table 10.2) and the presence of mica in soil profiles suggests that dustfalls may have been contributing to soils on the islands over the long-term. The effects of aeolian inputs to sedimentation off the southern Californian and northern Mexican coasts are discussed below (see section 11.1).

Natural dust sources in the Mojave Desert include dry playa floors and alluvial fans, but the role of human activity in disturbing natural desert surfaces, thus rendering them more susceptible to dust storm generation, has been highlighted in recent years by a number of authors (Bowden **et al** 1974, Wilshire 1980, Nakata **et al** 1981). Landsat and ERTS-1 satellite imagery of a **Santa Ana** wind on 1st January 1973 showed a series of six dust plumes emanating from the western Mojave that subsequent field investigation proved were areas of serious human impact. Among the causes of surface destabilization were the building of roads and utility corridors, seasonally bare agricultural surfaces, stripping of vegetation prior to urban development, overgrazing, stream channel modification and the use of off-road vehicles. Nakata **et al** (1981) warn that the extensive and growing destabilization of Californian desert surfaces provides for ever-increasing dust yields in future storms.

The surface destabilization effects of large-scale cultivation in the Imperial Valley, Colorado Desert have been highlighted by Clements **et al** (1963). Interviews with several long-term residents of the area suggested that upto 30 severe dust storms occurred annually in the first few years of tillage. Today, with the adoption of many wind erosion control practices and permanent crops on large acreages, the incidence is much less, on the order of four to six a year.

Inappropriate agricultural techniques contribute in part to dust storm generation in the San Joaquin Valley. A dramatic dust storm in December 1977 severely eroded surfaces that had been devegetated by the effects of drought in 1975/76 combined with overgrazing, a general lack of wind breaks on agricultural land, broad areas of land recently ploughed in preparation for planting and the stripping of natural vegetation from areas prior to cultivation. Other devegetation and surface destabilization had been caused by clearing land for urban development around Bakersfield, denudation of land around the oil fields north of Bakersfield and the use of off-road vehicles.

The storm resulted from a pressure gradient that steepened between a high pressure cell over Idaho and a low pressure system that intensified as it moved eastwards from its position over the Pacific 1400km west of California. The resulting easterly and southeasterly airflow across the southern end of the San Joaquin Valley was locally enhanced by valley channelling, producing winds of 83m/s, and mobilised more than 25 million tonnes of soil from grazing lands alone within a 24-hour period. The storm caused moderate to severe damage to structures, crops, orchards, vehicles, wildlife and soils in an area of about 2000km^2 in the southern part of the valley. The depositional plume, that reached over 500km to Sacramento, caused respiratory problems for the populace of the whole valley and carried valley fever spores northwards, provoking a dramatic increase in the incidence of the disease.

Among the other problems attributable to airborne dust in California are the interference with relay contacts in telephone transmission lines, contamination of drinking water, and the seepage into storage areas for military equipment and machinery (Clements **et al** 1963).

10.1.2A OWENS LAKE & MONO LAKE

Further illustrations of the way in which human actions can influence dust storm generation are found in two locations in the Great Basin Valleys Air Basin, a long narrow region of 36,000 km^2 east of the Sierra Nevada in east central California (Moore 1975). The region consists of a number of valleys oriented NNW-SSE with elevations rising from -86m in Death Valley in the south-east to near 2000m in Bridgeport Valley to the north (fig 10.4). The climate of the area is arid, with Bishop for example averaging 145mm rainfall a year. The lowlying parts of several of the valleys are occupied by playas. Two specific locations in the air basin have become significant sources of dust storm generation in recent decades. In the Owens Valley, Owens Lake has been dry since 1930, and further north, in the Mono Valley, Mono Lake's level has dropped by 14 vertical metres between 1941 and 1981. The dry lake beds are currently susceptible to wind erosion, and blowing dust and dust storms are common.

Owens Lake existed during the Pleistocene and probably dried completely during interglacials, but the latest desiccation, initiated by climatic change as glaciers receded and levels dropped, has been accelerated by human action. In the second half of the 19th century irrigated agriculture in the valley began lowering lake levels, but at the turn of the century levels were still deep enough for steam boats to float cargo across the waters. During the first 20 years of this century the construction of a 360km water export system by the Los Angeles Department of Water began a great acceleration of water table lowering in the Owens Valley. In 1912 the lake level was 7.6m, by 1930 the lake was completely desiccated, ordinary farming became impossible, and the 220km^2 dry lake bed became a source of dust except during intermittent wet year flooding. From 1930 to 1969 the Los Angeles system was extended northwards to tap streams flowing into the saline Mono Lake, lowering the level as mentioned above, the dry margins of the present lake (an area >8000 hectares in 1981) becoming dust sources.

Dust is raised from the Owens Lake when wind speeds are as low as 15 to 20 knots, but the major dust storms are generated by stronger winds, particularly westerly **Chinooks** and strong northerly flow, although southerlies cause less severe dust events. Minor storms occur occasionally in October or November, but the big dust storms are generated in January, February and March. Dust storms raised by northerly flow, essentially controlled by the valley's orientation, have the maximum impact on the valley settlements. Reinking

et al (1975) outline the synoptic scale circulations that drive the northerlies. Preceeding a dust storm, upper airflow (at 500mb) is westerly across the Sierra. With intensifying advection of Polar air a trough begins to develop over the western US aloft, and surface pressure falls with cyclogenesis in the Great Basin or less often as the southwestern extremity of a cold front advancing through dry air. When the low aloft deepens, massive fluxes of dust advect down Owens Valley and regional winds at all levels shift to the NNW and strengthen. The winds are enhanced by local **Chinooks** and Venturi flow effects between the steep walls of the valley. Dust storm onset occurs shortly after cyclogenesis at the surface to the east, or after the passage of a dry front, the onset being directly and consistently correlated with the initial sharp rise in pressure. Dust is turbulently mixed by the strong surface wind shear to depths upto 12,000m , although the dust is trapped in major valleys by sinking and stabilizing cold air aloft.

The dust plumes from Owens Lake are visible on satellite imagery 240km from source, covering at least 9000 km^2. A particularly severe event in March 1970 reduced visibility to 400m for five hours at China Lake 100km to the SSE of the lake (Reinking **et al** 1975).

The dust clears as winds at high levels weaken and back toward the west or veer north-east as high pressure builds into the region from the west. A return to the normally very clear visibility of a hundred miles or so, may take several days. Reinking **et al** suggest that the consistency of weather features leading to dust storms indicates that these episodes should be as predictable as the troughs, winds and cold air advection aloft.

10.1.2B NATURE & SOURCE OF DUST

Observations of dust storms in the Owens Lake area show that dust is sometimes raised from barren land created by excessive groundwater pumping (DeDecker, pers. comm. 1985), but is very largely derived from the dry lake bed, which consists of a saline crust overlying alkaline muds. On top of this saline crust a powdery deposit often forms, composed of sodium chloride, sodium sulphate, sodium carbonate and a host of other mixed minerals. This deposit is formed by capillary efflorescence when the water table is high, in years with heavy rainfall or when the Los Angeles Department of Water and Power pumps excess water into the lake. In the summer, this crust is a hard durable collection of small salt crystals in their monohydrate or anhydrous form mixed with sand and clay. These crystals are quite strong and hold the clays together.

In winter the crust grows faster, even though evaporation is less. As temperature drops, the hydrated versions of the sulphates and carbonates form, but as sodium chloride is not hydrated until well below freezing point its presence causes osmotic migration of the water of hydration from the other alkalis to the NaCl. This leads to the formation in a short time of the anhydrous form of sulphates and carbonates, while the growth of the hydrated forms causes a great volume change that results in disruption of the clay surface, making the clays available to wind erosion. This same process takes place at Mono Lake and in Saline Valley to produce wind erodible 'puffy solonchaks'. It is interesting to note, therefore, that due to the importance of a high water table in crustal formation dust storms are far worse in years when rainfall is heavy (St. Amand, pers. comm. 1985).

10.1.2C ENVIRONMENTAL PROBLEMS

A variety of environmental problems are associated with the dust storms on Owens Lake. Visibility reduction is a hazard to highway traffic and regional aviation operations at the Naval Weapons Center at China Lake some 100km to the SSE which is in the path of the dense dust plumes caused by northerly airflow as mentioned above. The dust is also a nuisance to tourists who come to enjoy the natural beauty of the unspoilt landscape and the 'normally' very clear atmosphere, and it has potential impact on solar power facilities proposed for the area. Botanists fear that the dust may be having an adverse effect on pine trees on the nearby Inyo Mountains (DeDecker, pers. comm. 1985). Particularly important are the effects on human health. Patients suffering from emphysema, asthma and chronic bronchitis are subject to increased morbidity (Reinking **et al** 1975), and the general population complains of coughing, sneezing and irritation to the eyes. Psychological problems emerge as people become apprehensive due to breathing difficulties. The concentration of atmospheric sulphate during dust events often violates State of California standards (Barone **et al** 1981). Dust events are common enough on the Owens Lake shoreline for residents to have adopted a local name for them, they are referred to as **Keeler Fog.**

The dust problems in the Mono Valley are essentially similar, although some may be more serious. The deterioration of the local ecology and natural beauty associated both with atmospheric dust and other problems resulting from the drop in lake levels are perhaps more critical in this more remote region where the local economy is almost entirely based on outdoor recreation. Mono Lake is more saline than Owens Lake and Mono Lake dust has ten times the concentration of sulphates, representing a more serious threat to human health.

The dust storms at Owens and Mono Lakes are an interesting example of the human role as an agent of dust storm generation. The environmental problems caused in the region are varied and widespread, and the situation is still a politically sensitive topic that stimulates intense passions in the inhabitants of the locally affected valleys. While plans for the stabilization of the Owens Lake surface are under consideration a strong environmental movement has been set up to save Mono Lake from further desiccation (Mono Lake Committee 1985).

10.1.3 NEVADA, UTAH & OTHER AREAS

High dust storm frequencies occur in a localised area in western Nevada, east and north of Reno where dust storms and devils are generated by downslope winds from alluvial sediments (eg Hallett 1969). In central Nevada research by Young and Evans (1986) has shown that dust storm deposition rates are very high on the lee side of playas in the Grass Valley ($2930g/m^2/yr$) and that more widespread but lower rates (20 to $30g/m^2/yr$) may have important influences on surface soil horizons in desert ecosystems. Subaerial deposition rates were highest during winter when dust storms are common on the playa surfaces due to low rainfall and high wind speeds associated with the alternate movement of high and low pressure cells. Deflation from the surface increased if the surface was alternately wetted and dried, due to salt crystallisation on the soil surface and mud curls produced by drying polygonal cracks. The playa is also fed with fine particles by ephemeral streams and by overland flow bringing fine material from unvegetated braided channels.

Two areas in Utah experience dust related visibility effects. Much of southeastern Utah east of the Wasatch Mountains is affected, particularly in the lower visibility categories (Changery 1983), and a second area in western Utah extends from the Great Salt Lake Desert southward along the west slopes of the Wasatch Mountains. Further north, in southern Idaho, one or two dust storms occur annually in the Snake River region and areas of eastern Washington on the Colombia and Snake Rivers.

In drought years other regions in the USA may become the source of dust storm activity. In Illinois Changnon (1983) shows that high dust storm activity was associated with continuing severe droughts in the period 1933-37 (when a statewide point average of 2.7 dust storm days a year were recorded) and 1976-77 (1.4 a year). More recently, however, in 1981 the statewide point average was 2.8 a year, which was not during a severe drought. These storms, in east and central Illinois, were due in part to below average seasonal rainfall, but these effects were exaggerated by early spring ploughing of the fine prairie soils and the gradual destruction of hedgerows over the previous 20 or 30 years.

10.2 CANADA & ALASKA

The Canadian Prairie provinces of Alberta, Saskatchewan and Manitoba are situated along the northern limits of the Great Plains of the USA and are primarily flat lands devoted to agriculture and subject to recurring drought. Annual soil loss due to wind and water erosion on the Prairies is estimated at 277 million tons, of which 58% is ascribed to aeolian action (Wheaton & Chakravarti 1986).

As in the Great Plains, the 1930s was a period of prolonged severe drought in the Canadian Prairies which exposed inappropriate agricultural practices and effected a 'dust bowl', although not on the scale of the US experience. Turner (1955) highlights the particular susceptibility of bare summer fallow lands, where the lifting effect of the wind is greatest due to convectional eddies generated by greatest absorption of the sun's rays by the black soils.

The most severe drought since the 1930s in the Prairies occurred from July 1976 to the end of April 1977, and saw less than half the normal precipitation fall in most areas. In many places the topsoil was blown off the land, and satellite imagery tracked huge dust clouds blowing across the northern US and Canada (Ladochy & Annett 1982). Atmospheric dust loads in the Prairie provinces were studied during this period using Total Suspended Particulate (TSP) data from 26 stations (in 11 cities) by Ladochy and Annett who confirmed that drought periods have higher atmospheric dust levels than non-drought periods, although differences were small. At Winnipeg, in the centre of drought severity in 1976/77, the TSP level was only 13% above 'normal'.

Analysis of dust storms on the prairies of Saskatchewan, using data for the period 1977-83, shows that the southernmost areas are worst affected by dust storms, a maximum annual frequency occurring at Regina (5.0 days). The seasonal maximum is in the spring months and three meteorological situations can be identified as causing dust-raising: localised convectional systems; frontal passages, and pressure gradient winds (Wheaton & Chakravarti 1986).

In 1984 drought on the Canadian Prairies saw conditions reminiscient of the 1930s. Dust storms with winds gusting upto 36m/s (70knots) closed highways and indirectly caused several fatal traffic accidents. Some fields had to be reseeded several times in consequence (Sweeney 1985). In 1985, after one of the driest springs in 100 years, June rainfall was less than half normal totals in southern Alberta and south-west Saskatchewan. One particularly violent dust storm on 8th and 9th June caused widespread crop, tree and property damage (Farmer 1985).

The occurrence of dust storms in polar regions, particularly in north-west Canada and Alaska, has been studied or commented upon by a number of authors. Tarr and Martin (1913) outlined the conditions favourable for dust transport by wind in the Copper River Basin, Alaska that are applicable in similar situations all across this region: abundant fine sediments in melt waters from glaciers; anastamosing and meandering channels producing wide floodplains; seasonal shrinkage leaving large areas of sparsely vegetated floodplain open to wind action from off-glacier, down-valley winds, and dry climate with often highly seasonal precipitation. Dust storms in comparable situations have been reported on the floodplains of the Matanuska Valley, Alaska (Tuck 1938), the Delta River, Alaska

(Péwé 1951), the Knick River, Alaska (Birkeland & Larson 1978) and the Slims River Valley, Yukon Territory, Canada (Nickling 1978 & 1983).

Nickling's work in the Slims River Valley is the most comprehensive of those concerned with dust-raising in Polar latitudes. Dust storms are generated by strong off-glacier winds and are most frequent from May to July when the Slims River is at a relatively low stage, precipitation is small (less that 100mm in total during these months) and temperatures are sufficiently high to dry the surface of the delta (Nickling 1978). The rate of sediment transport is directly affected by the surface moisture content and the presence of soluble salts at or near the ground surface, both factors tending to stabilize the surface by holding individual grains in place (see chapter 2). Interestingly, the most intensive dust storms were found to occur on days following heavy or extended rainfall, in contrast to the more commonly held belief that rainfall prevents dust entrainment due to the cohesion caused by wetting of surface sediments. In this case, however, the wetting effect is outweighed by the rainfall leaching of soluble salts down from the sediment surface, reducing or removing one of the principal cohesive forces opposing the entrainment of sediment by wind. Subsequent evaporation from the top few millimetres of sediment makes it available for aeolian entrainment, usually within 24 hours of the rain stopping.

10.3 CONCLUSION

Wind erosion of agricultural soils is a major problem in North America. The Great Plains of the USA and the Prairie provinces of Canada are the most active regions, and although dust storm frequencies are low compared to other world regions (a maximum of just six events annually on the Great Plains) the estimated annual soil loss due to wind from USA and Canada is about 1.2 billion tons. This figure exceeds estimates of total dust production from the Sahara, but methods of estimation may make comparison of such figures dubious. There can be no doubt, however, that dust storms are a significant hazard to agricultural and other activities in North America, and dustfall deposition rates from the south-west USA are of the same order of magnitude as those from other world regions (see tables 10.2 & 2.3).

These problems have generated a large volume of research into wind erosion and dust storm occurrence. A number of common meteorological systems associated with dust storms have been recognised. The passage of cold fronts is the most important large-scale system. Other synoptic scale systems include the effects of surface cyclones, including the intense cyclonic gyre generated by a more-or-less stationary low (an example is shown in plate 10.1). Katabatic and foehn-type enhancement of wind speeds has been identified in a number of more local situations (eg **Santa Ana** and **Chinook** winds), while mesoscale systems include the thunderstorm downdraft storm common at particularly high levels in the troposphere over the Great Plains.

Overused agricultural land is certainly a major source of dust storm sediments in North America, and these sources are particularly active during drought periods which appear to follow a 20-year cycle in the Great Plains. Many of the lessons learnt from the Dust Bowl years of the 1930s have been disregarded in subsequent times as new agricultural techniques have been adopted, often resulting in renewed severe soil loss during poor rainfall years. Other dust sources include devegetated palaeodunes in the semi-arid south-west USA, while in the more arid regions sources include pans, alluvial fans, alluvial plains and glacial outwash sediments in the cold arid areas of Alaska and northern Canada.

There are numerous examples of human activities affecting the North American dust storm systems, besides inappropriate use of agricultural lands. Increasing water demands by urban areas in southern California have led to the desiccation of lake beds, creating significant new dust sources at Owens Lake and Mono Lake. Vegetation stripping and surface destabilization by mining and construction activities and recreational use of semi-arid ecosystems in the south-west USA have also been shown to create local sources of dust storm activity.

Satellite monitoring of long-range transport from the USA indicates that dust is transported to the North Pacific and North Atlantic Oceans by major events. It has been suggested that material reaching the upper air may be transported as far as Europe (Jackson **et al** 1973) and simulations by Joussaume **et al** (1984) suggest that North American material may reach Greenland.

CHAPTER ELEVEN : LATIN AMERICA

11.1 MEXICO

Meteorological stations in Mexico do not record dust storms, but data were collected from the meteorological records kept by observers at airports in that country. In total, average dust storm frequencies were calculated for 27 airports. The length of records varied, but all stations used had records for at least five years. The longest continuous record was for Mexico City International Airport (1952-83). The data were gleaned from daily meteorological observations in which visibility is recorded at half-hourly intervals. The observation codes used for phenomena relating to dust in the atmosphere are as follows:

D: Dust in suspension.
BD: Blowing dust.
TBD: Thunderstorm and blowing dust.
DH: Dust and haze.

Visibilities associated with these dust codes ranged from zero visibility to 11km, but because of the problems involved in distinguishing between visibility reduction solely due to dust and that due to pollution in the atmosphere, only visibilities less than 5000m were used in this study. This problem is most emphasised in Mexico City where pollution from motor vehicles and industry is particularly bad, and present throughout the year.

In a number of papers Jauregui (1960, 1969, 1971, 1973) has investigated dust storms in Mexico City, the capital, and the salient points of these works are incorporated into the sections on Mexico City below. Otherwise there is just one paper dealing with the subject in the rest of the country. Idso (1976b) describes the **Chubasco,** violent storms that reach hurricane strength and often occur along the Pacific coast of Mexico and Central America. These storms, that originate from severe thunderstorms forming over the coastal mountains during the early afternoon, frequently spawn dust storms very similar to the North African **Haboob.** They are most common during the coastal rainy season (July to September).

The deserts of northern Mexico seem to be the source of mineral dust over the northern hemisphere low latitude eastern Pacific, as seen in the haze distribution maps of figure 3.1. The region is most active in June, July and August, and to a lesser extent in March, April and May. The only extensive aerosol study in this region was conducted by Prospero and Bonatti (1969), who showed that the mineral dust was comprised mainly of plagioclase which suggests andesite sources, consistent with the geology of the proposed source region: dry western and southern Mexico, and possibly Central America. The prevailing winds along the western coast of

Mexico are northwesterly all year, which would transport the dust to the cruise area, and analysis of the weather records showed 30 dust-raising events to have occurred in Mexico and Central America during the two week cruise in late February/early March.

An analysis of aeolian sedimentation in the Pacific off the coast of northern Mexico (Bonatti & Arrhenius 1965) suggests that the silty clay material with a silt fraction of clastic minerals (mainly quartz) is largely derived from a period of higher frequency of easterly moving dust storms from the Sonoran Desert and Baja California, dated by radio carbon methods at more than 40,000 years BP.

11.1.1 DUST STORM DISTRIBUTION

The distribution of dust storms in Mexico (fig 11.1) shows that these low visibility events are not a very common phenomenon, even in the arid north of the country. Just three airports record more than one dust storm day in an average year. Cuidad Jaurez, on the border with New Mexico and Texas, records 4.7 days a year and is situated at the southern edge of an area of high dust storm frequency that extends from the White Sands, New Mexico region into extreme west Texas as shown in figure 10.1 (Changery 1983). Torreon records an average of 11.8 days a year (the highest frequency in Mexico) and Mexico City 4.5 days a year, although there has been a marked decrease in frequency at both these stations during the period of available records (see below). Nuevo Laredo, in the arid north on the Texas border, records 0.7 days and San Luis Potosi 0.8. Perhaps surprisingly, dust storms are rare along the dry northwestern coast of Mexico and in Baja California. South and east of Mexico City rainfall totals are much higher and dust storms do not occur in this area.

It should be emphasised at this point that the number of stations shown on figure 11.1 leaves large parts of the country without data, and thus some areas in which dust storms occur may not be illustrated on this map. This problem may be particularly relevant in the arid and semi-arid regions of northern Mexico. One such area that is not apparent on figure 11.1 is a semi-arid highland basin about 100km east of Mexico City: La Cuenca de Oriental. The 300km^2 basin is intensively farmed using groundwater irrigation, but the very fine, powdery volcanic soils are particularly susceptible to aeolian erosion, especially during the dry season months of January to April. These are the months when maize is planted and the fields are bare, and salty marsh lands in the basin are dry. During this time the air in the basin is constantly hazy with material from the bare fields, dry marshlands and river beds (plate 11.1), and more severe dust storms occur when stronger winds blow.

The distribution of blowing dust days (visibility <5000m) shown in figure 11.2 confirms the high frequency locations discernable from the dust storm distribution. It is interesting to note, however, that although Torreon averages two and a half times the number of dust storm days at Mexico City, it records slightly fewer blowing dust days than the capital. Thus, at Torreon 36% of blowing dust days involve a severe visibility reduction to below 1000m, whereas just 13% of Mexico City's dusty days are severe events. At Cuidad Juarez the percentage of severe events is 18%.

Some minor dust blowing occurs in the north-west of Mexico and on Baja California, and two or three days a year with blowing dust occur on the east coast of the Gulf of Mexico. On this east coast the dust-raising is always associated with northerly winds and occurs from November to April, with a maximum in February/March. This activity is due to the winds of the **Norte** (see below).

On the Isthmus of Tehuantepec the airport at Cuidad Ixtepec averages 0.7 days of blowing dust a year, despite a rainfall of 1000mm. This activity occurs in April and May when agricultural fields are burnt in this area of southern Mexico, making smoke haze the most prevalent source of visibility obstruction, although after burning the dry fields also produce some terrigenous aerosols. It seems likely that the high incidence of dust haze in March, April and May over the Pacific south of the isthmus shown by McDonald (1938) (fig 3.1) is at least partly due to this seasonal burning.

11.1.2 MEXICO CITY

Mexico City was founded in the fourteenth century, on an island in Lake Texcoco. The Spanish conquered the city in 1522, and the settlement expanded under the conquistadors' rule; this expansion required better management and drainage of the many lakes that made up the floor of the Valley of Mexico. Dust storms, or **Tolvaneras** as they are locally known, blown up from the fine sediments of the dry lake beds, probably first made an appearance in the Valley of Mexico from the mid-seventeenth century (Jauregui 1960), although the earliest documentation of their occurrence dates from 1876 (Garay, quoted in Jauregui 1960).

The climate of Mexico City can be broadly divided into the dry season (October to April), when anticyclonic weather types prevail, and the wet season when cloudiness and precipitation are associated with the easterly Trade current (Jauregui 1973). Precipitation amounts vary within the urban area; Tacubaya, the headquarters of the National Meteorological Service, is situated on hilly terrain to the west of the city centre, where there is an orographic effect on rainfall. Cloudy skies and rain are more frequent over the hills to the west and south, whereas the central plains near Lake Texcoco (north-east of the present city and adjacent to the International Airport) are semi-arid (BS Köppen).

Major fluctuations in temperature occur during the winter caused by cold polar outbreaks associated with cold fronts that pass over the Gulf of Mexico (**Nortes**). During the second half of the dry season the dry winds from these polar continental air masses desiccate the surfaces and lower the water tables of the former lake beds, providing very dry topsoil conditions, and this is the season of maximum dust storm activity (January to June).

Dust is raised under these conditions by intense downdrafts during the afternoon from so-called "high-level thunderstorms" (Krumm 1954) that sweep over the plains. The worst storms in the Valley of Mexico are associated with these dry thunderstorms at the end of the dry season, as the westerly winter circulation gives way to the moist easterly current (Jauregui 1973). Another important influence on the development of dust storms in Mexico City, and more important still in the north-west of Mexico (see section 11.1.3 below) is the effect of the jet stream associated with southeasterly moving troughs of low pressure from the western seaboard of the USA and Canada. These jets typically occur at 500-300 mb, with a core at about 400 mb, or 7000m and are commonly developed during the months of November to February (Armendareiz, pers. comm. 1985). The development of thunderstorms associated with low pressure over California/Nevada is also a common synoptic situation for dust storms in the Valley of Mexico.

11.1.2A MEXICO CITY : DUST STORM FREQUENCY VARIATIONS

A study by Jauregui (1960) of dust storms at Tacubaya over the period 1923-58 shows a marked interannual variability in frequency, but with no trend over the period. Jauregui points out in another paper (1969) that dust storm intensities are generally lower at Tacubaya than at the airport, but as Jauregui's data (1960) for Tacubaya have no apparent visibility limit it is impossible to compare his frequencies with those taken from the airport. Nevertheless, he found no trend in dust storm frequency from 1923 to 1958, whereas a marked downward trend in frequency is apparent for all visibility classes used in the data from the airport over the period 1952-83 (fig 11.3). This decrease in frequency can be explained with reference to three major factors.

1. **Rainfall.** A gradual increase in rainfall in Mexico City over the data period (fig 11.3) is shown for precipitation figures from Tacubaya. A similar steady increase has been noted for the station at San Juan Aragon, about 7km north north-west of the airport and on the same side of the urban area, annual rainfall totals here being of the order of 200mm less than those at Tacubaya (Jauregui & Klaus 1982). Jauregui (1960) found little correlation between one or two year's rainfall and the following year's dust storm frequency, so that a year or two of low rainfall did not necessarily result in a high number of dust storms the

following year. For the data period analysed at the airport correlations with antecedent annual rainfall were also poor. Comparing annual dust event frequency (visibility <5000m) with the previous year's rainfall produced a Spearman's rank correlation coefficient of -0.14 (56% probability of a relationship) (see fig 11.4), and a correlation with the average annual rainfall over the preceeding three years was also poor: r = -0.22 with 23% probability (fig 11.5). Nevertheless, a gradual increase in rainfall can reasonably be expected to have had some effect on a concomitant decline in dust storm frequency over the same period when considered with the other factors outlined below.

2. Urban Development. The growth and spread of the urban area around the airport may also have had some impact on dust storm activity. Although perhaps in the short term urban sprawl may act to further destabilize dust sources during phases of construction, the encroachment of the urban area around the airport that has occurred since the 1950s will have acted to protect susceptible soil surfaces after the initial stages.

3. Proyecto Lago Texcoco. The identification of the dried bed of Lake Texcoco as a major source of dust storms resulted in the initiation of a project in 1972 that aimed to stabilize this area. Lake Texcoco, at low levels during dry years, has been a source of dust storms over Mexico City for at least one hundred years (Jauregui 1960), but drainage for industry, agriculture and human use resulted in the lake's complete desiccation in the early 1950s. During this decade the lake, situated immediately north-east of the present airport, became a major source of dust storms over the airport itself and the city. The project has planted pastures of **zacate salado** (**Distichlis spicata**), irrigated with recycled urban waste water (**agua negra**) and constructed a number of reservoirs on the former lake bed. In 1971 the area accounted for 40% of Mexico City's dust storms (SARH 1985); that percentage was reduced to zero by 1984 (see also below).

It is interesting to note that the average annual number of dust storm days occurring at Mexico City shown in the present study is very much less than that quoted by Jauregui (1973) and subsequently referred to in a number of more general texts on dust storms (eg Goudie 1978, Coude-Gaussen 1984a). The value used here is 4.5 days a year, Jauregui's (1973) value is 68, which is quoted from a previous paper (Jauregui 1960), which, as noted above, uses no visibility limit in the dust storm data. Further, Jauregui (1973) uses the value 68 as average number of dust storms whereas in the original paper (1960) he shows an average 68 dust storm days. This example serves as a useful illustration of the problems and dangers of definition and data useage.

11.1.2B MEXICO CITY : DUST STORMS BY DIRECTION & SPEED

The reduction in the frequency of dust days shown in figure 11.3 is further illustrated by tables 11.1 and 11.2 and figure 11.6. The tables show the number of occurrences of dust storms for two periods (1952-61 & 1975-83) for two visibility classes; the pie charts show percentages of dust storms with direction.

The average number of dust events has decreased between the two periods:

At <5000m from 47.7 a year to 16.4, a reduction of 66%.

At <1000m from 8.6 a year to 0.8, a reduction of 91%.

The most striking aspect of these tables is that in the first period the north-east sector, representing dust storms from Lake Texcoco, shows the largest percentage of dust storm occurrence (36% of events reducing visibility to <1000m, 24% of those with visibility <5000m), and in the second period this sector shows no dust events at either visibility class. This can be very largely attributed to the success of Proyecto Lago Texcoco.

The number of events for all sectors has declined, at both visibility classes, except in the northern sector from which the average annual number of dust events has remained constant for the two periods at eight (at <5000m). Thus the percentage of occurrences from the north has increased to 49% at <5000m, and this is currently the major direction from which dust events occur at the airport.

The tables also show how threshold velocities for dust-raising have increased over the two periods:

At <1000m in period 1, lowest wind speed for dust is Bft2.

At <1000m in period 2, lowest wind speed for dust is Bft4.

At <5000m in period 1, lowest wind speed for dust is Bft1.

At <5000m in period 2, lowest wind speed for dust is Bft3.

Similarly,

The maximum number of dust events at <1000m in period 1 is at Bft5.

The maximum number of dust events at <1000m in period 2 is at Bft6.

The maximum number of dust events at <5000m in period 1 is at Bft5.

The maximum number of dust events at <5000m in period 2 is at Bft5.

This increase in threshold velocities indicates a more stable wind erodible surface which is attributable to higher soil moisture conditions (from higher rainfall totals) and/or increases in protective covering such as vegetation or urban concrete.

11.1.2C DIURNAL VARIATION OF DUST EVENTS

Figures 11.7 and 11.8 show the diurnal frequency percent by month of dust events for the same two periods as used in the preceeding section. Two visibility classes are used, <5000m and <1600m (about 1mile). The <1000m class was not used here as the number of events of this severity during the second period was minimal.

These figures confirm the decline in dust frequency illustrated above, but these data do so at a higher resolution as frequencies have been calculated at three-hourly intervals. Dust-raising is very largely concentrated in the late afernoon and early evening, with activity around noon also being important during the months of January to April. Jauregui (1969) has shown that the time of greatest average wind speed for all months of the year at the airport is 1600-2000 local time.

11.1.2D MEXICO CITY : DUST EVENT DURATION

A comparison of the durations of dust events between the two periods 1952-61 and 1975-83 at three visibility levels is shown in table 11.3. As has been shown in previous sections, the absolute numbers of events at all visibility levels are markedly fewer during the second period, even though the second period is for 9 years and the first 10 years. At all visibility levels the large majority of events last for an hour or less, during both periods.

All events reducing visibility to <1000m during the second period lasted for an hour or less, whilst in the first period 4% of these events lasted for 1-2 hours. For <1600m visibility events the percentage lasting an hour or less, 1-2 hours and 2-3 hours is fairly similar in both periods, although one event in the first period lasted for 3-4 hours. At <5000m there is a noticeable difference in the percentages of events for 1 hour or less between the two periods: period two recording 13% less than period one. For events longer than one hour, there were slightly more events by percentage recorded during period two than during period one at 1-2 hours, 2-3 hours and 3-4 hours, and the longest event during either period was for 7-8 hours and occurred in period two. This suggests that although the number of dust events during the second period has been severely curtailed relative to the first (from an average of 46.7 events a year to an average 17.2) their average duration is slightly longer.

11.1.2E DUST OVER MEXICO CITY

It has been shown that dust storms affecting Mexico City blow mainly from the north, north-east and east. As a dust wall advances material is deposited over the city by gravity or is washed out by precipitation as 'mud rain'. In some cases the dusty air does not reach the southern sector of the city (Jauregui 1973). The mean monthly dustfall for the year

1959, as measured by Bravo and Baez (1960), demonstrated that the north and east fringes of the city are most affected by dustfalls. The mean monthly deposition rate was at a maximum over the north-east (>50 ton/km^2/30days) and minimum over the south-west (<10 ton/km^2/30days), the mean monthly dust deposition for the total urban area in 1959 was calculated to be 26 ton/km^2/30days. More recently, Marquez (1969, quoted in Jauregui 1973) conducted a similar study in the year from August 1967 to August 1968 and found an average monthly deposition over the whole city of 20.9 ton/km^2/30days. Although the number of dust days with visibility <5000m was similar in these two 12 month periods (36 in 1959, 33 in Aug67/Aug68) the number of more severe dust days (visibility <1000m) was more markedly reduced between the two periods (13 to 7).

11.1.3 TORREON

Torreon records the highest average annual number of dust storm days in Mexico (11.8), although the annual number of <5000m dusty days is slightly fewer than at Mexico City (32.5 at Torreon, 35.2 at Mexico City). The dustiest months are February to August, when 92.3% of <1000m dusty days are recorded and 84.1% of <5000m days. This period starts when mean monthly rainfall is at its lowest, but overlaps the rainy season months of May to October (fig 11.9). Torreon's climate is arid, BW according to Köppen's classification, with 248 mm of rainfall a year, and average potential evapotranspiration exceeds average precipitation in every month, representing soil moisture deficiency throughout the year.

11.1.3A TORREON : DUST STORMS BY DIRECTION & SPEED

Dust storms are experienced from all compass directions; least frequent are those from the west (2% of <1000m events; 4% of <5000m events). The greatest percentage of dust events are associated with winds blowing from the north-east and east (39% of <1000m events; 50% of <5000m events). Dusty days associated with calm winds represent 5% of all events at both visibility levels (see fig 11.10).

Data in table 11.4 represent the number of cases of dust occurring in the two visibility classes used. From the table it can be seen, as might be expected, that the lightest winds (Beaufort Scale 1) are not strong enough to raise dust. However, dust events are associated with calms occasionally, which can be understood as representing material raised at some distance from the airport and advected to the observation station. The number of dust events increases with increasing wind speed to a maximum number of events associated with Bft 4 in the case of <5000m events and Bft 5 for <1000m. This tendency is in keeping with the knowledge that a greater wind speed is required to raise a greater amount of material from the same source, thus producing a lower visibility. At higher Bft numbers, the frequency of events decreases from these maxima. For Bft numbers upto and including 5, at both visibility classes, the most frequent direction of dust-raising winds is from the north-east and east, but at speeds greater than Bft 5 winds from the south, south-west and north-west are more frequent bearers of dust. This may be for a number of reasons. Firstly, that winds stronger than Bft 5 are more frequent from the south, south-west and north-west. A further explanation may lie in the nature of the dust sources in these directions. Neither <1000m or <5000m dust events are recorded from the south or south-west until winds blow at Bft 4 or greater, whereas for all other directions dust is raised by winds of Bft 2 or more (except the south-east: Bft 3 or more). From the north-west dust is not entrained until winds reach Bft 5 at <1000m. This perhaps suggests that the dust sources from the south, south-west and north-west are more stable than those in other directions, either due to protective covering such as soil crusts, or due to the erodibility of the soil itself.

11.1.3B TORREON : DIURNAL VARIATION & FREQUENCY CHANGES THROUGH TIME

Figure 11.11 shows the average diurnal frequency per cent occurrence of dust storms for two periods: 1949-60 (from Hinds & Hoidale 1977) and 1962-69 (this study). The pattern and magnitude of frequency show some marked changes between the two periods. The frequency of dust occurrence has decreased, so that dust events are almost absent from September to January in the more recent period. Further, the diurnal distribution of frequencies has contracted, so that for example, in the months of February, March and April during the first period dust events could be experienced at almost any hour of the day or night, while in the more recent period they were largely concentrated during the late morning and afternoon hours.

The change in the frequency of dust occurrence commented on above is also confirmed by a substantial decline in the average annual number of <1000m dust events recorded during the two periods. During the earlier period an average of 30 compares with an average of 12 during the latter period. If a similar comparison is made between the average number of dust storm days in the two periods, however, the decline is much reduced: an average 12.9 days in the first period and an average 11.8 days in the second period. This difference serves as a reminder of the problems inherent in the use of dust storm days as opposed to an analysis that takes the duration of events into account.

11.1.4 CHIHUAHUA & MONCLOVA

Data on monthly blowing dust diurnal frequency for the northern stations of Chihuahua and Monclova are also presented in Hinds and Hoidale (1977) and shown in figure

11.12. Both stations display a blowing dust maximum in the period from February to May, at Chihuahua diurnal maximum is from 0900 local summer time (LST), while at Monclova dust is more evenly blown throughout the 24-hour period with an afternoon maximum.

11.2 CENTRAL & SOUTH AMERICA

This section is by necessity limited, as data and work on dust-raising in this area are very sparse. Only the Argentine Meteorological Authority provided data on dust storms in reply to numerous requests to national observing services. In Chile, Brazil and Colombia dust storms are not recorded by meteorological observers, and other countries that experience dust-raising activity, such as Peru and Paraguay, produce very erratic meteorological data.

11.2.1 CENTRAL AMERICA

Although Central America has no arid areas, many days with dust haze occur during the dry season of December to April (Portig 1976). McDonald's (1938) haze frequency maps show 15-20% of ship observations report dust haze at sea during the spring months in the area south of the Gulf of Tehuantepec. **Nortes** (Northers) are winds associated with cold fronts that move southeastwards over the Caribbean (see section 11.1.2) that blow as warm, dry winds on the Pacific side of the Central American mountains, due to a foehn-type effect. These winds stir up dust in areas where dust sources are available, reducing visibility considerably. The dust haze displays a sharp upper surface, probably the Trade wind inversion, through which the mountains penetrate into clear air with unlimited visibility. The high concentrations of aerosols measured by Prospero and Bonatti (1969) over the Pacific south of Central America are dealt with in detail in the section on Mexico above. Some of the material they observed may also have been derived from the Central American region. More recently, Maenhaut **et al** (1983) have observed a substantial crustal component of large particle Si and Fe (geometric mean concentration = 0.44 microgrammes/m^3) in aerosols sampled during a trip from Panama to the Galapagos Islands in April/May. This they attributed to long-range dust transport "probably originating in the vicinity of Central America" (Maenhaut **et al** 1983, p5362).

Dust haze in the region may contain a component from West Africa, although much of this material is probably washed out in transit over the mountain ranges along almost the entire length of Central America. Prospero and Bonatti (1969) speculate that some West African material may be transported to the atmosphere of the eastern Pacific across Panama, which is the only significant gap in the mountain barrier, where some flux of water vapour in the atmosphere occurs from the Caribbean (Weyl 1968).

11.2.2 ECUADOR, PERU, CHILE & BOLIVIA

The arid areas of northern Chile and Peru might be expected to represent a major source of dust in South America. Johnson (1976) reports that dust devils, caused by strong insolation heating of the land, may often be seen moving across some of the more southern highland basins in Peru, Bolivia and Ecuador. Dust storms can appear, usually in the early afternoon, during the dry season. They are particularly noticeable in the Altiplano, where strong gusts are a hazard to drivers and prevent sailing on Lake Titicaca after noon. Dust storms may also be caused by thunderstorm downdrafts, the typical advancing wall of dust being reddish brown in the limestone areas.

In the Atacama Desert, elongate gypsum dust lobes of windblown material are characteristic of the salt-encrusted playas known as **salars** (Stoertz & Ericksen 1974). These authors also describe a dust and sand storm that blew for two days in June, reducing visibility to a few feet, with winds exceeding 27m/s at times. East of the Atacama Desert, on the Andean altiplano of south-west Bolivia, Salar de Coispasa and Salar de Uyuni (the latter the world's largest salt flat) can reasonably be suggested to be sources of large-scale dust/salt storms (see also section on Puna de Atacama in Argentina, section 11.3.1, below).

The diurnal percentage frequencies of blowing dust (visibility <11km) for two of the very arid Peruvian coastal settlements of Talara, near the Ecuador border, and Lima are presented in figure 11.13. Both stations record blowing dust largely between April and October; at Talara activity is concentrated in the late morning and afternoon, whereas at Lima dust blowing starts earlier in the morning and continues into the early evening. Along the Peruvian coast from 6° to 18° S, wind speed maxima remain normally below 12 m/s with the exception of a 200 km stretch of sandy desert south of Pisco (14°S). The general southerlies along this stretch of the coast are locally intensified, where a coastal escarpment influences airflow, and between 10 or 11 am and early afternoon may reach gale force (Lettau & Costa 1978). The local name for these exceptionally strong daytime winds is **Paracas**. Sand and dust driven by the **Paracas**, which can blow with gusts exceeding 20 m/s, reduces visibility considerably and makes driving hazardous on the roads in the Pisco region.

Prospero and Bonatti (1969) have sampled aerosols derived from the arid coasts of Peru and Chile over the Pacific. The mineralogy of the aerosols was largely made up of quartz, with appreciable quantities of micas, chlorites and smectites also present. The smectites are believed to be from Ecuador or northern Peru, where weathering processes favour their formation. Dust loadings, sampled in the spring, were low, averaging 0.14 microgrammes/m^3 and the synoptic meteorology of this part of the Pacific is not conducive to the transport of material far out to sea, as the prevailing winds flow parallel to the coast. The haze frequency maps (fig 3.1) show this tendency with distance from the land, haze being relatively rare a few hundred kilometres from the coast. The haze frequency maps show seasonal maxima in December, January and February, in contrast to the the diurnal blowing dust frequency figures for the coast of Peru (fig 11.13), and also in June, July and August in accordance with figure 11.13.

The distribution of quartz in surface sea sediments seaward of the Peru-Chile trench resembles the pattern of the south-east Trades, which supports the idea that a large proportion of quartz in deep-sea sediments is carried from the continent by wind, especially high altitude winds (Molina-Cruz 1977, Molina-Cruz & Price 1977). Recently, Schneder **et al** (1983) have investigated the particle size distributions of n-alkanes and ^{210}Pb as indicators of terrestrial aerosols off the coast of Peru.

11.2.3 COLOMBIA & VENEZUELA

Some minor dust-raising activity occurs in the dry marine region of the Caribbean lowlands of northern Colombia: Calamar and Soplaviento (Bermudez, pers. comm. 1985). In Venezuela dust-raising occurs along the dry Caribbean coastal strip where annual rainfall is in some places below 500 mm. Figure 11.14 shows the diurnal percentage frequency by month of blowing dust (visibility <11 km) at Caracas. Activity only occurs during the driest months of January to May during the entire 24-hour period, with less activity recorded during the midday hours.

11.2.4 BRAZIL & PARAGUAY

Haze is very common in almost all regions of Brazil (the Portuguese phrase is **Nevoa Seca** - dry haze and/or dust). Although the nature of the particles has not been studied they must be both mineral dust and products of combustion. Ratisbona (1976) suggests that haze occurring in the summer in the semi-arid north-east and the southwestern part of the country (the Planicie Riograndeuse) consists mainly of dust particles originating in these dry parts of the country. Dust from Brazil may be transported in the south-east Trades dominant during the summer and autumn as far as Cayenne, French Guiana, where Prospero **et al** (1981) have monitored aerosols. At Cayenne the largest influxes of dust are transported from West Africa during February-April/May when monthly arithmetic mean concentrations were 25 microgrammes/m^3 and 23 microgrammes/m^3 in March of 1978 and 1979 respectively. The West African dust consisted largely of mica and illite, with significant amounts of kaolinite, quartz, chlorite and plagioclase, whereas the material assumed to be of South American origin was characterised by higher concentrations of iron oxide minerals such as geothite and aluminium oxide minerals such as gibbsite. This mineralogy is consistent with the the highly leached lateritic soils common in the north-east of South America. The most likely source is the semi-arid region of north-east Brazil, an area that was experiencing drought conditions during the Cayenne aerosol sampling programme, although it is only speculative to suggest that deflation in this area was increased as a consequence of the drought.

McDonald's (1938) haze frequency maps show haze at sea off the south-east coast of Brazil in all seasons. The only study of aerosols in this part of the southern Atlantic was conducted in February/March by Parkin **et al** (1972) who found that atmospheric dust in the area was principally derived from the Namib Desert of southwestern Africa. However, some haematite found in samples off Brazil is thought to have been carried by northerly winds from soils in the south-east of the country.

The only information relating to dust storms in Paraguay was found in Grubb (1911) who describes a "great black cloud" in the Paraguayan Chaco that arrived from the south, immediately preceeding the rain of an advancing thunderstorm.

11.3 ARGENTINA

Data were available from the Argentine National Meteorological Service in the form of monthly totals of dust storm days for 170 stations over the period 1968-78. It should be noted, however, that there is no visibility limit used when recording dust events at Argentine meteorological stations. Thus, the mean annual dust event frequency distribution shown in figure 11.15 is not directly comparable to those for other countries included in this work. Nevertheless, the map gives an indication of the main areas of dust-raising activity in Argentina. In the account of these main areas which follows, reference will be made to the distribution of loess in Argentina (after Teruggi 1957) and the broad division of natural farming regions that is made by the Instituto de Suelos y Agrotecnia based on water availability (fig 11.16).

11.3.1 DUST EVENT DISTRIBUTION

Dust storms have been reported from almost all parts of Argentina, although with only minor occurrences in the north-east and the region of Entre Rios east of the Parana River where rainfall is fairly high (over 750mm a year), in large parts of Patagonia (although the density of stations is low here), and some stations high in the Andes. The high frequency areas can be divided broadly in two: the Andes and foothills and the semi-arid loess region.

11.3.1A ANDES & FOOTHILLS

An area of high dust storm frequency in the extreme north-east of the country occurs on the Puna de Atacama, a plateau region some 3000-4000 m in altitude. The Puna de Atacama is characterised by extreme aridity and year round strong winds from the west/north-west, which entrain material from a series of **salars** and other dry lake beds. An illustration of this activity is shown in the Space Shuttle image in plate 11.2. The image was taken in September, 1983, during the season of maximum activity in this area of Argentina : June to September (see fig 11.17 for La Quiaca). The image shows dust plumes of two distinct colours; the red plumes are being blown from rust-coloured silty sediments that are probably fine muds of former lake basins, and the white plumes, from the **salars,** are salty dusts of evaporite minerals such as gypsum. Not all the salt lakes shown are acting as dust sources which may be due partly to the nature of the surface, such as the presence of too much moisture or a protective salt crust. Conversely, the image suggests that local topography is an important influence, both in funnelling winds, and in generating turbulent flow in the lee of mountain ridges.

Many of the valleys of the Altiplano have their own wind systems that locally entrain fluvial sediments brought down from the Andean chain. At La Quiaca from November to March winds are dominantly from the north-east due to the advection of rain-bearing tropical air masses. From May to September, however, dust storms are generated by the predominant mountain winds (from the south due to local topography) together with the westerlies of the general circulation (Prohaska 1976).

Foehn-type winds are also observed along the whole length of the of the Andes, where they are known as **Zonda,** after a locality in the valley of the Rio San Juan. Figure 11.17 shows that **Zonda** is most active from September to February at San Juan. Temperature rises associated with the **Zonda** can exceed $30^{\circ}C$, and relative humidity may fall to 5-10%. Dust raised by a powerful **Zonda** blowing across the very sparsely vegetated cordillera may take several days to settle.

11.3.1B SEMI-ARID LOESSIC REGION

This region can be subdivided conveniently into the northern **Chaqueña** (Chaco) and the southern **Pampeana** (Pampas), both agricultural regions. In the Chaco, dust storms are associated with strong winds from the north and south. These winds are generated by a quasi-stationary low which is situated over western Paraguay and northern Argentina in winter (causing frequent southerly winds in the Chaco) and over the piedmont area of the Pampean sierra (La Rioja -

Catamarca) in summer, causing northerly and northeasterly winds in the Chaco (Prohaska 1976). The wind erosion problem in the Chaco has been aggravated this century by human activities. Prego (1961) reports that uncontrolled forest clearance and overgrazing in years with above average rainfall have created desertification problems and an increase in dust storm activity on the fine-grained loessic soils. Santiago del Estero is notorious for its **Tormentas de Tierra** (Earth Storms), and the largest annual number of dust storm days in the area is recorded at Monte Quemado (18.3).

The Pampas is generally more intensely cultivated than the Chaco. It supports a larger population and has worse wind erosion problems. Sand dunes are widespread in this semi-arid region, fences piled with soil and sand on the leeward side are a common sight, and there are many closed depressions (pans) with lunette dunes on their lee sides which bare further testimony to the operation of deflation processes. Dust storm clouds seen from afar are referred to as **Volcanes** (literally meaning 'volcanoes'). Dust storms occur largely from July to December (see fig 11.17 for Rio Tercero), and maximum annual frequencies are greater than 30 at Rio Quarto and General Pico. These storms often travel from north to south with the winds prevailing over the region, although dust storms are also generated by the southeasterly coastal wind **Pampero** (known as **Pampero Sucio** when it raises dust), and from thunderstorm downdrafts and squall lines. In the central province of Cordoba tornadoes are particularly frequent (Altinger de Schwarzkopf 1982).

A severe dust storm that travelled northeastwards across the region around Coronel Suarez in January is described by Woelcken (1951). The dust storm was related to the passage of a cold front with the characteristics of a squall line, producing a dust wall very similar to a classic thunderstorm downdraft (although just 500m in height). At Coronel Suarez the arrival of the dust front reduced visibility to between one and 30 metres for 20 minutes and was accompanied by the classic pattern of variation in temperature and pressure as described in North Africa by Freeman (1952) and Morales (1979b). Pressure rose by 2mm of mercury and temperature dropped dramatically from 34°C to 21°C, the cold air mass brought a light rainfall which increased the relative humidity from 35% to 75%.

Dust is also raised by vehicles on the long, straight dirt roads that cross the Pampas linking the main roads and railways (Wallington 1964). Towards dusk, as convectional forces weaken, the clouds of dust behind motor vehicles remain undispersed with a ceiling at between two and six metres and visibility is drastically reduced at these low levels for many minutes.

Despite the high wind speeds of the prevailing westerlies and the arid climate of Patagonia, there are few dust storms on the grass and shrub steppe of this meseta landscape. The exception to this general observation occurs in a coastal area around Camarones and Comodoro Rivadavia which average 4.3 and 3.9 dust storm days a year respectively from September to April (fig 11.17). In the sub-polar steppe of Tierra del Fuego the station at Rio Grande averages 2.5 a year (fig 11.17).

11.3.2 SEASONALITY & METEOROLOGICAL SYSTEMS

Analysis of the seasonality of dust storms for 32 of the 170 stations used in figure 11.15 shows that with increasing degrees of latitude the month of peak activity occurs later in the year: from August/September at 20-25^{o}S to December/January/February at 40-45^{o}S. Similarly, the season commences earlier in the north of the country (June at 20-25^{o}S, continuing until October) and gets later with increasing degrees of latitude : the season at 40-45^{o}S starting in October/November and continuing into February/March. Generally, dust-raising occurs most frequently during the afternoon hours at all stations in Argentina (Alaimo, pers. comm. 1984).

From the above analysis four main meteorological situations can be identified as being of primary importance in the generation of dust storms in Argentina:

1. Strong winds from the west/north-west on the Puna de Atacama.
2. **Zonda** foehn-type winds at the foot of the Andean cordillera.

3. Strong winds from the north or south on the Pampeana-Chaquena plain.
4. Thunderstorms/squall lines/frontal passage across the Pampas.

11.4 CONCLUSION

Comprehensive dust storm data were only available from Mexico and Argentina. From Mexico these data were collected from daily airport observations taken at half-hourly intervals, while the Argentine data were for monthly frequencies of 'dust storm days' but without any visibility limit. Lack of data, therefore, prevented an assessment of the dust storm system in the Atacama Desert.

In Mexico the dust storm systems at Mexico City and Torreon (the most affected stations) have been investigated in detail, showing the relationship to wind speed and direction. At both stations dust storms are largely concentrated in the first six months of the year (the dry season) and are generated by downdrafts from high level thunderstorms that are associated with cold northerly airflow (**Nortes**). Diurnal maxima are largely during the afternoon hours in Mexico.

At both stations dust storm frequencies have declined over the data period. At Mexico City this decrease is related to an increasing trend in rainfall totals, expansion of the urban area and a project set up to stabilize Lake Texcoco adjacent to the airport, that had been a major source of dust storm sediments since its desiccation in the 1950s by drainage. A poor relationship was found between dust storm activity and antecedent rainfall at Mexico City however.

In Argentina two major areas of dust-raising are identified. On the Puna de Atacama dust sediments are raised from dry **salars** and lake beds by the strong westerly airflow. In the semi-arid loessic region dust storms are caused by frontal passages and thunderstorm downdrafts. Overuse of these agricultural regions has caused some increase in dust storm activity. With increasing degrees of latitude the dust-raising season occurs later in the year: from June to October at 20°-25°S, to October to February at 40°-45°S, with a diurnal maximum in the afternoon hours.

CHAPTER TWELVE : SUMMARIES & CONCLUSIONS

12.1 DEFINITIONS & DATA

To date, no comprehensive definitions of dust events have appeared in the literature, a reflection of the early stages of dust storm studies. Thus, it has been necessary to make a set of definitions of these phenomena for this investigation, as outlined in section 2.2. The definitions used in this study are made in accordance with international standards of meteorological observations, and they can be used in future investigations in order to achieve some standardisation in the work on dust storms and related phenomena. These definitions make the important distinction between a dust storm, a severe dust event observed at source, and dust haze, a dust transport event.

The use of meteorologically observed data involves problems of observer error, infrequent recording, scarcity of meteorological stations in remote areas and unavailability of data for various reasons. Employment of the dust storm day as a global measure of dust storm activity also adds the problem of poor temporal resolution to the inaccurate indication of a storm's volumetric extent, but these difficulties were considered acceptable given the deliberately global perspective of the present thesis. More detailed local investigation has been carried out where data were available and also with the aid of remote sensing imagery, but there will be some degree of inaccuracy in the surveys of some areas due to the quality of data used.

Meteorological data are widely available, but care needs to be exercised in their use. One of the definitional problems highlighted in this study has been that some dust events may incorporate aspects of both a dust storm and a thick dust haze; in the case of a thunderstorm downdraft storm, material carried in the characteristic dust wall may be being raised at the point of observation and also comprise dust raised from afar and carried forward in the turbulent dust lobe. Such problems have important implications for the identification of dust source areas.

The main body of data used in this thesis is from terrestrial observations, with additional information provided by remote sensing imagery and the literature on remote sensing, aerosol monitoring and deep-sea coring programmes.

12.2 GLOBAL DISTRIBUTION, FREQUENCY & SOURCES OF DUST STORMS

There have been a number of attempts to indicate the global distribution of dust storm activity (eg Grigoryev & Kondratyev 1980, Péwé 1981, Goudie 1983, Coudé-Gaussen 1984a). In figure 12.1 I have attempted to develop these world maps further by identifying those sources that experience more than 15 dust storm days a year. This map also shows the major trajectories of dust transport from these and other source regions. Figure 12.1 is complemented by table 12.1 which names the major source areas and indicates the mean number of dust storm days at stations that represent the scale of activities in these regions, using the longest available meteorological records.

The global distribution of dust storm activity reflects the controls on the dust storm system: the presence of fine grained sediments; their availability for entrainment, as reflected in the degree of surface protection, and the occurrence of a strong wind to blow the material into the air. On the global scale, the areas of greatest dust storm frequency are those with an arid or semi-arid climate and are located in the broad band of such regions that stretches from West Africa to northern China. The reason for this general correspondence with dry lands is largely derived from the sparse vegetation that is characteristic of such climates, meaning less protection for potentially wind-erodible surfaces. At the more local scale the role of appropriate geomorphological settings with sufficient quantities of dust sized debris becomes important, and the prevalence of weather systems capable of generating dust-raising winds.

12.3 CHANGES IN DUST STORM FREQUENCY

Although the major world areas of dust-raising activity have been outlined, it is important to note that the frequency and intensity of dust output from these regions may vary quite considerably from one year to the next, quite apart from seasonal variations. On the geological timescale this variation may be identified in the ocean sediment record, polar ice caps and the world's peri-desert loess deposits. Since the beginning of routine and standardised meteorological observations, shorter term variations, on the timescale of decades, become apparent. The spatial extent of such variations ranges from the stabilization or destabilization of a local lake bed for example, to a region the size of the Sahelian fringe of the southern Sahara.

A number of examples of changes in dust storm frequency have been noted, and these changes are often related to variations in meteorological parameters (particularly rainfall) and the activities of human populations. Thus, widespread increases in dust storms have been noted over large areas such as the south-west USA in the 1930s and 1950s, the steppes of the USSR in the 1950s and most recently across the Sahel since the late 1960s.

12.4 CONTROLS ON THE DUST STORM SYSTEM

It would appear from figure 12.1 that the most important broad-scale control on dust storms is climatic. It has also been noted that a second important control on dust storm occurrence is the availability of large quantities of fine grained material for entrainment. From the investigation into the world's dust storm systems a number of common dust-producing geomorphological units can be identified, and these are shown in table 12.2.

It has also been possible to identify a number of typical meteorological systems which give rise to dust-raising winds, and these are shown in table 12.3. Some of the major world regions where these systems operate are indicated in figure 12.2.

Having outlined a number of common dust-producing geomorphological environments and meteorological systems, it will be useful to summarise the role of other controls on the dust storm system and how those controls vary from time to time and in different locations.

The effect of rainfall on the wind erosion system is difficult to assess. It has been noted that dust storm activity is more characteristic of semi-arid regions rather than hyper-arid areas, which is, perhaps, a reflection of the importance of water in producing large quantities of fine debris. There is, however, no simple relationship between annual dust storm frequency and annual rainfall totals. Nor is there any good correlation with the annual number of rainy days, arguably a more accurate indicator of the long-term moisture status of a ground surface. There is, however, clear evidence that in many world regions dust storms are characteristic of dry seasons (eg Afghanistan, north-west India, Mauritania) when soil moisture is low and perhaps vegetation cover is sparse. This seasonality is not universal, however, as in other areas dust storm activity is intense during the 'wet season' (eg Sudan, Arizona) as local atmospheric systems common in the wet season are effective dust-raising agents and moisture in the soil surface layers is quickly evaporated in high seasonal temperatures.

On the day-to-day timescale the intensity and duration of rainfall is important, so that long duration low intensity rain, or frequent rainfall of small amounts will keep soil moisture levels high enough to prevent the entrainment of particles. Less frequent but more intense rainfall, however, may leave sufficient time between rainfall events to allow evaporation from surface layers, leaving dry particles available for entrainment. In areas where salts are important binding agents of surface particles, intense rainfall may effectively leach soluble salts, thus reducing or removing one of the principal cohesive forces preventing aeolian entrainment and allowing dust to be raised on days after heavy rain (eg Nickling 1978). Similarly, Spence (1957) notes that serious **Blows** in the Fens of England are often preceeded by rainy nights, so that drying earth clods tend to break down to smaller particles susceptible to wind erosion.

Clearer relationships have been found between rainfall and dust-raising activity over longer periods in areas subject to drought conditions that may continue for several years. In such cases, low rainfall not only affects soil moisture but as time goes on vegetation cover is reduced, and after a few years of drought this has further affects on soil structure due to the smaller amounts of organic matter reaching the soil. Decreasing rainfall over a period of ten years or more has been found to have a fairly close relationship to increases in dust storm days in the Mauritanian Sahel for example.

The role of antecedent rainfall on dust storms was recognised by Sidwell (1938) who noted that high dust storm activity followed years of poor rainfall. Although in many areas some degree of relationship of this nature is discernable (eg at Minqin in northern China), this is not always the case. The effects of antecedent rainfall depend upon the nature of the dust source. At Owens Lake, southern California, for example, dust storms are most frequent following years when rainfall has been higher than average (St. Amand, pers. comm. 1985). In that case, high rainfall results in high water tables, enabling the better development of a 'puffy' solonchak, readily erodible by wind.

For some meteorological stations months of high dust storm frequency are coincident with months of high average wind speeds and Drift Potentials, which incorporates a threshold velocity for particle movement (eg Dhahran, Saudi Arabia). Correlations between average annual dust storm frequency and average annual wind speed have not been good, however, (eg north-west India) but this lack of correlation is not surprising given the coarseness of the wind speed measures. Indeed, the role of wind is better assessed with reference to weather systems responsible for generating dust storms, especially as the direction, turbulence, frequency and duration of winds are all important parameters for which sufficient data were not available at the scale of investigation of the present study.

The relationship of dust storms to the wind erosion climatic factor, C, developed by Chepil **et al** (1962), was tested in north-west India and found to be poor, but a clearer understanding of the dust storm system in this area has been reached by tracking individual events and with an appraisal of the generating meteorological systems.

In many cases human activities are another important factor in the already highly variable and complex wind erosion system. The direct effects of human activities are largely through interference with vegetation, ground surfaces and surface water. Numerous examples have been noted, at a variety of scales, where human actions have increased dust storm occurrence. Widespread destruction of vegetation and surface destabilization occurs as a result of inappropriate agricultural techniques; over large areas examples include the Dust Bowl of the USA in the 1930s, the Virgin Land schemes of the Soviet steppes in the 1950s, and the Sahel area in the 1970s and early 1980s. Other examples of increased dust-raising due to agricultural development have been reported from the Chaco region of northern Argentina (Prego 1961), in a number of Californian valleys (Clements **et al** 1963), on the steppes of northern Mongolia (Sanders, pers. comm. 1982) and in the cold arid areas of Iceland (Ashwell 1966).

Examples of increases in dust storms due to surface destabilization include the tank battles of the North African campaign in World War II (Oliver 1945), recreational and off-road vehicle use in southern California (Wilshire 1980), and construction activities in the Gulf region (Aspin, pers. comm. 1983).

Areas of silts, muds and clays on the margins of lakes, revealed as water levels are reduced by human water use, have become sources of dust storms in a number of regions. On a small scale the draining of Lake Texcoco, to the north-east of Mexico City, provided a ready source of dust storm sediments over the Mexican capital, while an increasing water use by the city of Los Angeles has resulted in lowered water levels at Owens and Mono Lakes in southern California, resulting in new regional dust sources. On a much larger scale, the falling levels of the Aral Sea are partly attributable to increasing water use for irrigation, revealing hundreds of square kilometres of salty lacustrine deposits susceptible to large-scale dust storm generation.

The example of Lake Texcoco illustrates how revegetation and reservoir construction on the former lake bed have been successful in stabilizing this dust source.

12.5 THE WORLD'S DUST-BEARING WINDS

It has been seen that dust storms are a common feature of many world regions, and the frequency of dust-bearing winds in these areas has spawned a great number of local names which are shown in table 12.4. This table indicates the etymological derivation of the name where known, the area affected, the season and direction of the wind and the meteorological systems involved.

From this table and the classification according to meteorological systems presented in table 12.3, a number of distinct types of dust event can be identified according to their structure.

1. DUST DEVIL.
2. DUST WALL - travelling system in which dust is raised at the front of a seperate air mass (eg thunderstorm downdraft or frontal).
3. DUST PLUME - point source in high winds.
4. LARGE-SCALE DUST-RAISING - eg pressure gradient or cyclonic gyre.
5. DUST HAZE - eg **Brume Sèche** or **Loo.**

12.6 CONCLUSION

Aeolian deflation, as represented by dust events, is a process that occurs in many world environments but is particularly characteristic of arid and semi-arid lands. Dust storms have a great many environmental implications for which it is necessary to derive a methodology to quantify their frequency, distribution and areal extent. Standard meteorological data have been used to give an indication of regional frequencies and seasonalities and favourable environments for deflation, while consideration of regional meteorology has enabled an assessment of the generation of dust-raising winds. Orbital reconnaissance techniques have allowed specific dust sources to be pinpointed, they have provided a check on the reliability of surface station reports and above all they have provided a true perspective on the scale of dust transport.

Terrestrially observed dust storm data indicate that the most intensive areas of activity are located in the broad band of dry lands that stretches from West Africa to northern China, while less intense areas are located in Australia and North and South America. Care should be exercised in the use of such data, however, and although high frequency areas can be identified it is noted that significant changes in dust storm frequency occur on all timescales.

These changes in dust storm activity reflect the controls on the wind erosion system. On the geological timescale analysis of deep-sea and ice cores indicates variations related to large-scale climatic change. On a timescale of decades, meteorological data illustrate fluctuations caused by drought and the actions of human populations. The seasonal variation in dust storms is due to meteorological controls such as rainfall (and its effects on soil moisture and vegetation), wind speeds and the occurrence of dust-generating weather systems, while diurnal variation in dust-raising is largely dependant upon the effects of solar heating that induces instability and turbulent airflow in the troposphere.

This thesis has been global in its perspective. Details of the dust storm systems of many world regions remain to be discovered, including more specific identification of source areas, analysis of meteorological systems generating dust storms, and estimates of volume, density and mass budgets of dust raised. The present thesis has been designed to provide a methodological and regional framework for these future investigations.

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