PHYSICS OF DESERTIFICATION

EDITED BY FAROUK EL-BAZ, M.H.A. HASSAN

MARTINUS NIJHOFF PUBLISHERS

Physics of desertification

Edited by

FAROUK EL-BAZ

*Itek Optical Systems
Lexington MA
USA*

M.H.A. HASSAN

*University of Khartoum
Khartoum, Sudan*

1986 **MARTINUS NIJHOFF PUBLISHERS**
a member of the KLUWER ACADEMIC PUBLISHERS GROUP
DORDRECHT / BOSTON / LANCASTER

TABLE OF CONTENTS

List of Contributors	XI
Report on Workshop on Physics of Desertification M. H. A. Hassan	1
Desertification of Arid Lands H. E. Dregne	4
Wind Versus Water Erosion in the Context of Desertification Michael M. Verstraete and Jean-Pascal van Ypersele	35
The United Nations Organization and the Issue of Desertification Michel M. Verstraete	42
The Impact of Desertification on Land Productivity in Sudan M. O. H. El-Karouri	52
Desertification Problems in a Humid Region Jens Tyge Møller	59
The Formation and Motion of Dunes and Sand Seas Farouk El-Baz	70
Two-Dimensional Analysis of Dune Profile and the Effects of Grain Size on Sand Dune Morphology Haim Tsoar	94

LIST OF CONTRIBUTORS

Professor R. Ahmad
Faculty of Science
University of Karachi
Karachi 32, Pakistan

Dr. I. A. Eltayeb
School of Mathematical
Sciences
University of Khartoum
Khartoum, Sudan

Dr. Hans Karl Barth
Universitaet Bremen
FBI-Schewertukt Geographie
Post Fach 330440
D-2800 Bremen 33
Federal Republic of Germany

Dr. Jørgen Fredsøe
Institute of Hydrodynamics
and Hydraulic Engineering
Technical University
of Denmark
KD-2800 Lyngby
Denmark

Professor Harold E. Dregne
International Center for Arid
and Semi-Arid Land Studies
Texas Tech University
P. O. Box 4260
Lubbock, Texas 79409 USA

Dr. Dale A. Gillette
Cires Program
University of Colorado
Boulder, Colorado 80307 USA

Dr. Farouk El-Baz
Itek Optical Systems
10 Maguire Road
Lexington, Massachusetts 02173
USA

Dr. Ronald Greeley
Department of Geology
Arizona State University
Tempe Arizona 85281 USA

Dr. M. O. H. El-Karouri
Agricultural Research Council
P. O. Box 6096
Khartoum, Sudan

Dr. M. H. A. Hassan
School of Mathematical
Sciences
University of Khartoum
Khartoum, Sudan

tists of various academic backgrounds including Geomorphologists, Ecologists, Geographers, Soil Scientists, Engineers, Physicists, and Mathematicians. The general objective of the workshop was to review and improve our present knowledge and basic understanding of the physical processes in deserts and arid lands, which are responsible for the desertification mechanism. Special attention was given to sand dunes, soil erosion, and dust storms that are affecting the livelihood of millions of people in those regions.

THE PROGRAM

The workshop lasted for three weeks. The first two weeks were devoted to a series of lecture courses that reviewed in depth some of the recent research advances in the various fields. Each lecture course was presented by an international authority.

The first week provided a broad description of the physical problems in deserts and arid lands and proved to be extremely useful and highly informative to the participants, particularly to those who knew relatively little about desertification.

Although the lectures presented during the second week greatly improved our understanding of sand and dust movements, a firm theoretical background is still far off as compared with the physical problems described in the first week. It should be mentioned, however, that the modelling approach is still in its infancy, and that a substantial effort must still be made before a clear understanding of problems of the deserts are firmly based on mathematical modelling.

During the third week two working groups dealing with desert dust and desert dunes were formed. As a basis for discussions, a number of specialized papers were presented.

GENERAL REMARKS AND CONCLUSIONS

The workshop on Physics of Desertification differed widely from all previous workshops and conferences on desertification: it provided the first opportunity even for a multidisciplinary international group of theoretical and applied scientists to exchange and coordinate their views, transcending all traditional professional boundaries. The lectures presented reviewed our up-to-date basic knowledge of the physical processes in deserts and arid regions, and the ensuing discussions identified several relevant problems where more research is needed. The required research should be interdisciplinary and must be guided by physical and mathematical modelling. Among the principal areas where basic knowledge is insufficient are:

1. Comprehensive studies using wind tunnel experiments and theoretical analyses to understand the three types of sand movement near the surface: saltation, surface creep, and suspension.
2. Detailed studies to understand the formation and movement of various types of sediment ripples and sand dunes (longitudinal, transverse, barchan, and star dunes). The studies should be carried out using wind tunnel and field experiments together with modelling techniques.
3. Studies using laboratory wind tunnels and modern instrumentation together with theoretical analysis to determine accurately the appropriate meteorological and soil factors responsible for dust production, transport, and deposition.
4. Further research and development of the wind erosion equation to determine the primary factors responsible for wind erosion.

ACKNOWLEDGEMENT

On the behalf of the participants of the workshop on physics of desertification, I would like to express our very sincere thanks and gratitude to the International Center for Theoretical Physics and its director, Professor Abdus Salam, for generously supporting the workshop. Without this support it would have been impossible to hold the workshop. We also thank Professor Salam and the Center's staff for their very kind hospitality.

I would also like to extend a note of thanks to Professor Bertocchi (Head of Scientific Training Programs) and Mrs. Sue Benich (the Workshop's secretary) for the care and attention that they had given to the local organization of the workshop. Their endless help, patience, and understanding are deeply appreciated.

Last, but not least, I wish to thank all those who participated in the workshop. The success of the workshop owes much to their enthusiasm and constructive discussions.

DESERTIFICATION OF ARID LANDS

By

H. E. Dregne

ABSTRACT

Desertification is a land degradation problem of major importance in the arid regions of the world. Deterioration in soil and plant cover have adversely affected nearly 50 percent of the land areas as the result of human mismanagement of cultivated and range lands. North America and Spain have the largest percentage of their arid lands affected. Overgrazing and woodcutting are responsible for most of the desertification of rangelands, cultivation practices inducing accelerated water and wind erosion are most responsible in the rain-fed croplands, and improper water management leading to salinization is the cause of the deterioration of irrigated lands. In addition to vegetation deterioration, erosion, and salinization, desertification effects can be seen in loss of soil fertility, soil compaction, and soil crusting. Urbanization, mining, and recreation are having adverse effects on the land of the same kind as is seen on range, dry farming, and irrigated lands. Combatting desertification can be done successfully using techniques already known if financial resources are available and the political will to act is present.

INTRODUCTION

Desertification of the arid lands of the world has been proceeding--sometimes rapidly, sometimes slowly--for more than a thousand years. It has caused untold misery among those

most directly affected, yet environmental destruction continues. Until recently, few if any lessons seemed to have been learned from the past, in part because the problem was an insidious one that went unrecognized in its early stages or was seen as a local one affecting only a small population, and in part because new land was always available to start over again. As long as remedial action could be deferred by moving on to new frontiers, land conservation had little appeal. It was not until the 20th century--when easy land expansion came to an end--that governments and people finally realized that continued careless degradation of natural resources threatened their future.

INTERNATIONAL DIRECTIONS

The decade of the 1950's witnessed the first worldwide effort to call attention to the problems and potentials of arid regions. It started when the United Nations Educational, Scientific, and Cultural Organization (UNESCO) launched its Major Project on Scientific Research on Arid Lands in 1951. That project led to publication of a newsletter, the provision for funds for establishing and strengthening arid land research institutes, organization of conferences and symposia, and publication of a series of research reviews and special reports on a wide range of topics. The Major Project was terminated in 1962 and the arid land program was merged with the broader UNESCO natural resource program.

The impetus generated by the UNESCO project led to expanded interest in, and support of, arid lands studies throughout the world. By 1970, knowledgeable scientists were well aware of the magnitude of the land destruction that had taken place in the past, and that was becoming even more serious as population pressures increased.

At about the same time, one event served to focus world attention on desertification: the 1969 to 1973 drought in the African Sahel. Recognition of the severity of the drought affecting six countries on the southern border of the Sahara (Mauritania, Senegal, Mali, Upper Volta, Niger, and Chad) was slow to develop. Droughts, after all, were not unusual in the Sahelian countries; an equally bad or even worse one had struck the same region during the years from 1911 to 1914, and several other droughts had occurred before and after that time. It was only when the situation had become catastrophic that relief measures were undertaken on a large scale. Among the aftereffects of the human toll and the millions of livestock that died due to the drought, was the call by the United Nations General Assembly for the convening of an International Conference on Desertification. The conference was held in Nairobi, Kenya, in August and September of 1977, attended by representatives of nearly 100 nations and many

international organizations; governmental and nongovernmental. In addition to providing a forum by which the world's attention was drawn to the land destruction that was occurring, the conference addressed the problem of combatting desertification of arid lands. A plan of action was drawn to the land destruction that was developed and approved by the delegates (United Nations, 1978). Responsibility for following up and coordinating the implementation of the Plan of Action to Combat Desertification was entrusted to the United Nations Environment Program.

DEFINITION

Desertification is a term that has been in use since at least 1949 when Aubreville, a perceptive and well-informed botanist and ecologist, published a book on "Climate, Forêts, et Desertification de l'Afrique Tropicale" (Aubreville, 1949). Aubreville thought of desertification as the changing of productive land into a desert as the result of ruination of land by man-induced soil erosion. He associated it with the humid and subhumid tropics where he worked. The causes of land destruction were tree cutting, indiscriminate use of fire, and cultivation, which exposed the soil to water and wind erosion. Desertification was not the result of the Sahara spreading outward but of localized activity that could begin anywhere. Aubreville was quite clear in his conclusion that desertification in tropical Africa was due to man's activity, and that there had been no significant climatic change during the past thousand or more years. Most of that destructive activity has occurred within recent historic time by the action of agricultural populations.

Despite the fact that a world conference has been held on the subject, there is no generally accepted definition of desertification. For most people, the word probably evokes a mental picture of a barren forbidding landscape resembling Death Valley or the Sahara. Fortunately, that grim picture does not apply to most of the land that has undergone desertification.

A definition of desertification should recognize that it is a land degradation process that involves a continuum of change, from slight to very severe degradation of the plant and soil resource, and is due to man's activities. The definition used here is the following:

Desertification is the impoverishment of terrestrial ecosystems under the impact of man. It is the process of deterioration in these ecosystems that can be measured by reduced productivity of desirable plants, undesirable alterations in the biomass and the diversity of the micro

and macro fauna and flora, accelerated soil deterioration, and increased hazards for human occupancy.

Figure 1 delimits the arid regions with which this monograph is concerned. The map is redrawn from a UNESCO map of the arid regions (UNESCO, 1977).

DESERT SPREAD AND DROUGHTS

A common misapprehension about desertification is that it spreads from a desert core, like a ripple on a pond. The truth is that land degradation can and does occur far from any climatic desert; the presence or absence of a nearby desert has no direct relation to desertification. Desertification usually begins as a spot on the land-scape where land abuse has become excessive. From that spot, which might be around a watering point or in a cultivated field, land degradation spreads outward if the abuse continues. Ultimately the spots may merge into a homogeneous area, but that is unusual on a large scale.

A second misconception is that droughts are responsible for desertification. Droughts do increase the likelihood that the rate of degradation will increase on nonirrigated land if the carrying capacity is exceeded. However, well-managed land will recover from droughts with minimal adverse effects when the rains return. The deadly combination is land abuse during good periods and its continuation during periods of deficient rainfall. The famous satellite photograph taken during the 1970's drought of a green rectangle (the Ekrafane Ranch) surrounded by brown denuded land in eastern Niger demonstrated dramatically the value of good range management during favorable and unfavorable years.

HISTORICAL DEVELOPMENT

Land degradation is by no means a new problem, despite the attention focused upon it in recent years. In some quarters, there is a tendency to blame desertification upon land pressures generated by the population explosion of the middle 20th century. While the rapidly expanding population has greatly exacerbated the situation, desertification is not a new phenomenon.

Historical evidence shows that serious and extensive land deterioration occurring several centuries ago in the arid regions had three epicenters: the Mediterranean Sea, The Mesopotamian Valley, and the loessial plateau of China. There were other places where destructive changes in soil and plant cover had occurred but they were small in extent or are not well known.

Figure 1
World Distribution of Arid Regions. Redrawn from 1977 UNESCO map.

Last One Hundred Years

Desertification in Africa and elsewhere began long before the 1969-1973 drought struck the Sahel. Stebbing (1937a, 1937b) was the most persistent of those sounding the alarm over the rapid deforestation of the Sahelian and Sudanian vegetative zones. He looked upon desertification in West Africa as forest degradation that leads to erosion and, as a last stage, to barren sand or rock. That process is hastened, he contended, by blown sand from the Sahara being deposited on the deforested land. It was this latter activity that led him to refer to the "encroaching Sahara," a term he later regretted using (Stebbing, 1938). He disclaimed any attempt to give the idea that the Sahara was a vast sand field advancing in great waves like the incoming tide of a sea. The latter concept, however, has proved so attractive to numerous writers on desertification that it now represents a common view on the subject (Cloudsley-Thompson, 1974). There apparently is something fascinating about the idea of an expanding desert threatening mankind. Encroachment of moving sand dunes on desert oases and transportation routes is an aspect of desertification that is of small areal extent but is locally important and highly visible.

Warnings similar to those of Stebbing for the Sahel were made by other scientists in southern Africa, North America, South America, Asia, and Australia during the 1920's and 1930's. Research has been undertaken in many countries to develop techniques of grazing management and soil and water conservation that would halt and reverse desertification. As a result, there is now a good understanding of the basic principles of land conservation. Field application of those principles has been slow, unfortunately, and land degradation continues to undermine efforts to improve human well-being.

OCCURRENCE OF DESERTIFICATION

Desertification affects nearly all of the arid regions, to varying degrees, except for the extremely arid climatic deserts such as the Sahara, Atacama, and Taklimakan. An indication of the extent and intensity of the desertification that has occurred in the past is presented in Figures 2 through 7.

The classification system used in the preparation of continental desertification maps is based on four classes of desertification: slight, moderate, severe, and very severe. The criteria for each class are as follows:

Slight:	Little or no degradation of the soil and plant cover has occurred.
Moderate:	(1) 26 to 50 percent of plant community consists of climax species, or (2) 25 to 75

Figure 2
Desertification in Africa.

percent of original topsoil lost, or (3) soil salinity has reduced crop yields 10 to 50 percent.

Severe: (1) 10 to 25 percent of plant community consists of climax species, or (2) erosion has removed all or practically all of the topsoil, or (3) salinity controllable by drainage and leaching has reduced crop yield by more than 50 percent.

Very Severe: (1) Less than 10 percent of plant community consists of climax species, or (2) land has many sand dunes or deep gullies, or (3) salt crusts have developed on very slow permeable irrigated soils.

The "very severe" category represents the extreme condition that many people associate with desertification. It is land so badly degraded that its utility by man or animals is virtually zero and the degradation is economically irreversible, for most purposes. While there are many small areas of land that fit into this category, there are few areas large enough to be shown on the continental maps. Practically all of the world's desertification can, at this point, be reversed.

Delineations on the maps, because of the small scale employed, usually are combinations of different desertification classes. For example, an area shown as moderately desertified may have inclusions of slightly desertified and severely desertified land. The guidelines established to determine the map classification for any delineated area are given in Table 1. Table 2 gives data on the land area of the arid regions of the world in the four desertification classes. Most of the land used for agriculture in the arid regions is at least moderately desertified. The 52.1-percent figure for land in the category of slight desertification consists mainly of the naturally barren climatic deserts (hyperarid regions) where man's presence has been minimal.

AFRICA

The severe Sahelian drought that extended from 1969 through 1973 focused the world's attention on the human aspects of land degradation and led to the convening of the 1977 United Nations Conference on Desertification. Drought, however, is not the cause of desertification; man is. The drought served only to place additional stress on the biological resources of the Sahel. If resource management has been good, little, if any, permanent damage is done by droughts. However, if resource management has been unwise, a drought accentuates the adverse impact of that management and accel-

Table 1
Desertification Classification Criteria

Map Classification	Percent of Area in Various Desertification Categories
Slight desertification	>50% of area in slight category <20% in severe category <10% in very severe category
Moderate desertification	<50% in slight category <30% in severe and very severe category
Severe desertification	>30% of area in severe category 0-30% of area in very severe category
Very severe desertification	>30% of area in very severe category

Table 2
Desertification of Arid Lands of the World

Area Affected		
Desertification Class	Square Kilometers	Percent of Arid Lands
Slight	24,520,000	52.1
Moderate	13,770,000	29.2
Severe	8,700,000	18.5
Very severe	73,000	0.2
Total	47,063,000	100.0

Table 3
Desertification of Arid Lands of Africa

Area Affected		
Desertification Class	Square Kilometers	Percent of Arid Lands
Slight	12,430,000	71.7
Moderate	1,870,000	10.8
Severe	3,030,000	17.5
Total	17,330,000	100.0

Table 4
Desertification of Arid Lands of Asia

Area Affected		
Desertification Class	Square Kilometers	Percent of Arid Lands
Slight	7,980,000	50.9
Moderate	4,480,000	28.6
Severe	3,210,000	20.5
Total	15,670,000	100.0

erates land degradation (Weaver and Albertson, 1940). The latter is what has occurred widely in the Sahel and elsewhere.

All of the usual forms of desertification are present in the arid regions of the African continent and are manifested as serious local or regional problems (Figure 2). Overgrazing has reduced range productivity virtually everywhere outside the tsetse fly regions, in north, west, east, and south Africa. Wind and water erosion have devastated landscapes in the cultivated regions and in much of the rangelands (Rapp, 1974). Shortened fallow periods in the shifting cultivation system south of the Sahara have led to severe depletion of plant nutrients. Salinization and waterlogging of irrigated land is worst in the Nile Valley and North Africa but also occurs elsewhere. Mining has left surface scars wherever it is practiced. The environmental degradation continues and shows no sign of slowing down.

Desertification Characteristics

A number of factors have increased land degradation and the vulnerability of the African arid regions to desertification. Most of them have had similar effects in Asia and Latin America. They can be grouped in three categories: (1) increased human and animal population, (2) improved health services, and (3) injudicious use of technology.

Due to the increased sedentary population, pressures on cultivated land led to a shortening of the fallow period in the shifting cultivation cycle and the extension of cropping into the more precarious drier regions. Crop harvests became less reliable and more variable as the desert edge was approached. Concurrently, nomadic pastoralists were deprived of some of their best grazing lands as the cultivators moved in (Delwaulle, 1977). At the same time the rangeland area was contracting, populations of pastoralists and their livestock were increasing and the provision of improved veterinary services and the lack of a viable marketing system helped assure that animal numbers would grow rapidly (Widstrand, 1975). The result was inevitable: overgrazing and accelerated desertification.

Overgrazing inadvertently was made worse, particularly in the Sahel, by the drilling of additional wells that provided drinking water for livestock throughout the year. Without the rest period that intermittent water supplies previously assured, forage conditions deteriorated around the wells where water was no longer a limiting factor in livestock survival. Local authorities did not or could not impose a control system that would allow forage plants to recover from heavy grazing.

A crude estimate of the amount of crop yield reduction that could be attributed to salinization and waterlogging in

the Nile Valley indicates it to be at least 17 percent on 80 to 90 percent of the old irrigated land. In the newly irrigated lands east and west of the delta, the figure is likely to be a minimum of 25 percent yield reduction. Both problems are serious ones that can become even more serious in the future if remedial measures are not taken. The situation in Algeria and Tunisia is similar to that in Egypt with respect to yield reductions.

Solutions to desertification problems in Africa are known--and in general--can be implemented readily if resources are available to do so. An exploding population in the developing countries, however, means that land pressures will continue to build. Reducing livestock numbers in the grazing lands until forage productivity can be improved is very difficult, at best, and introducing range use controls on communal land is not easy. There is little likelihood that marginal dryfarm lands will be returned to pastoral use unless production on the better lands is increased enough to offset the loss of cultivated land. While the latter can be done, progress in that direction is very slow.

Destruction of woody vegetation has been hastened by the ever-increasing need for firewood to meet the demands of the larger population. The destruction is especially noticeable around the rapidly growing urban centers, where the circle of deforested lands gets larger every year (Delwaulle, 1973).

While desertification was a long-standing problem even in the absence of droughts, the gradually increasing vulnerability of the land made the impact of the inevitable droughts worse than ever (Dahl and Hjort, 1979). The factors responsible for that vulnerability are still operating, desertification continues, and future droughts will have ever-greater damaging effects.

Salt-affected soils are common in Tunisia and Algeria where both surface and well irrigation waters frequently are saline. Salinization is important in other irrigated areas but on a small scale.

Magnitude of Desertification

About 18 percent of the arid region of Africa is severely desertified, and most of that represented by grazing lands and rain-fed cropping lands on the south side of the Sahara (Table 3). The other large area that is severely affected is the mountain slopes and the plains of North Africa. Moderate to high salinity affects about 30 percent of the irrigated land in Egypt (Aboukhaled et al., 1975).

Wind erosion is dominant in the drier regions and water erosion on the wetter sloping lands. Ethiopia, Kenya, and the Maghreb countries of Algeria, Morocco, and Tunisia have been subjected to especially serious water erosion, whereas wind

erosion has been most damaging in sub-Saharan West Africa. While good data on the effect of land degradation on crop and livestock yields are not available, it seems likely that soil fertility losses, alone, have reduced dryland crop yields by 25 to 50 percent in the severely desertified areas. Animal productivity may well have declined by at least 50 percent nearly everywhere that domestic livestock are raised. In many areas south of the Sahara, rangeland forage production probably is less than 25 percent of the potential.

ASIA

Desertification in the arid regions of Asia is characterized by overgrazing of the rangelands of the Middle East and Central Asia, water erosion of cultivated lands from eastern China to the Mediterranean Sea, and salinization and waterlogging on a large scale in Iraq, Pakistan, China, and the Soviet Union. Mining (including oil and gas production) has caused severe land damage wherever it is carried on.

Overgrazing, soil erosion, and salt damage to irrigated land are long-standing problems in the Middle East and Central Asia, as is water erosion on the rain-fed cultivated lands of India, Pakistan, and the loessial plateau of China. Waterlogging and salinization are problems centuries old in the lower plain of the Yellow River in China, but are of relatively recent origin in the Indus Basin of Pakistan and India. Ameliorative measures to reduce land degradation have been undertaken in every country affected, with varying degrees of success. The problem is an immense one in terms of the area involved and the amount of degradation that has occurred (Figure 3).

Desertification Characteristics

Man first placed his imprint on the landscapes of Asia thousands of years ago. Historical records make it clear that the destructive impact of man was already apparent two or three thousand years ago in China and the Middle East. It was then that the first period of accelerated, man-made erosion began (Dregne, in press). Woodcutting and cultivation of sloping lands were responsible for most of the land degradation that occurred in the uplands of China and around the Fertile Crescent. Sedimentation of canals, waterlogging, and salinization combined to raise havoc in the Tigris and Euphrates valleys of Mesopotamia (Jacobsen and Adams, 1958).

Grazing pressure seems to have become a much greater problem in the past two or three decades than it was before. There has been an explosion of human and livestock population recently from India (Office of Environmental Planning and Coordination, 1977) to the Middle East (Pearse, 1971).

Figure 3
Desertification in Australia.

Figure 4
Desertification in Asia.

Cultivation has been extended into pastoral lands, wells have been constructed to make pastures usable throughout the year, and improved transportation facilities have made formerly remote areas accessible to livestock. As a consequence, cropping has become a high-risk land use in the climatically marginal areas that formerly were the best grazing lands. At the same time, overgrazing of the remaining range lands is accelerated by the increased concentration of livestock on drier lands. Destruction of vegetative cover on sandy lands has led to widespread wind erosion, depletion of soil fertility, and the undertaking of expensive erosion control projects in China, Iran, Saudi Arabia, and elsewhere.

Magnitude of Desertification

Asia has large areas where desertification is so severe that it borders on being classed as very severe desertification, but about half of the land is only slightly desertified (Table 4). Water erosion is spectacular in the loessial plateau in the middle reaches of the Yellow River. The eroded silt is sometimes so thick in the river that at Shansien, east of Xian, the sediment load has been measured at 46 percent. The river annually carries 1.3 billion tons of sediment, enough to cover 1,800 square kilometers of land 30 centimeters deep (de Crespigny, 1971). Virtually all of that represents loess eroded from 300,000 square kilometers of highlands. Efforts are being made to reduce erosion but the problem is so huge that progress is slow. Only the great thickness of the aeolian deposits has allowed cultivation to be continued on the eroding slopes.

Water erosion also approaches the spectacular on the slopes of the dryland farming areas of western India and northern Pakistan. Fortunately, the soils there are more resistant to displacement than in loessial material. Nevertheless, the damage--in the form of reduced yields--is serious nearly everywhere, and it is getting worse.

Salinization and waterlogging are as bad--and as difficult to control--in the irrigated region of southern Iraq as they are anywhere in the world. Deep fine-textured soils and almost flat topography make it hard to provide drainage that is adequate to maintain a deep water table and prevent salt accumulation. Techniques of soil and water management that are suitable for conditions in Iraq have been developed and tested successfully, but the technical problems are less difficult to resolve than the social ones (Dougrameji and Clor, 1977). Land reclamation is much easier on the more permeable soils of the Indus Valley and Central Asia.

Range productivity over most of the Middle East and western Pakistan is among the lowest in the arid world. A small beginning has been made in bringing the pastoral lands

up to their potential productivity. Much remains to be done, however, because range management is only a young science in Asia. Recent emphasis on sand dune stabilization has led to a number of successful efforts to find effective methods and to identify adapted plants. Accomplishments have been outstanding, under very difficult circumstances.

A rapidly growing population and limited land resources mean that combatting desertification will be difficult for the developing nations of Asia and very difficult for the poor countries. Land pressures must somehow be relieved, and that can only be accomplished by raising productivity and yields per unit area. A start can be made with presently available knowledge, but research and training will be needed to broaden the knowledge base and to provide guidance to the pastoralists and cultivators.

AUSTRALIA

Land degradation in Australia is, from the standpoint of area affected, a problem of overgrazing of range lands. Wind and water erosion and salinization of irrigated and nonirrigated land are serious locally, primarily in small areas across the southern part of the continent, and their economic impact probably exceeds that due to overgrazing. Land degradation is usual around farming communities and popular tourist areas such as Ayers Rock.

Overgrazing began 50 to 100 years ago with settlement of the interior arid lands at a time when there was no information on the carrying capacity of range land, and little knowledge of climatic averages and extremes. During the first quarter of the 20th century, problems of wind and water erosion, salinization and waterlogging of irrigated land, and saline seepage of dryland areas arose as cultivation spread into the drier areas. Remedial action was undertaken on a concerted basis in the 1930s and 1940s when states enacted soil conservation legislation and established units to correct land abuses and prevent further deterioration. At present, it probably is safe to say that overgrazing is much less of a problem than it was before 1940, that wind and water erosion still occur but to a lesser degree than before, that salinization of irrigated land in the Murray River catchment (watershed) is increasing, and that saline seepage of semiarid and subhumid lands is expanding steadily.

Desertification Characteristics

About 25 percent of Australia's arid regions are unoccupied (Condon, 1978) due to the unpalatability of spinifex grasses or to the remoteness of the land from population centers. Most of the remaining 75 percent is moderately deser-

tified (Figure 4). Severe desertification is most extensive in the saltbush-bluebush (*Atriplex-Maireana*) vegetation type occurring in New South Wales and South Australia, where overgrazing has caused degeneration of the plant cover. Degradation of the vegetation due to overgrazing has also been severe on the fine-textured lowland soils (Vertisols) of eastern Australia and in the flood plains and surrounding slopes of coastal river valleys, particularly the Gascoyne, Ord, and Victoria catchments in Western Australia and the Northern Territory. Gully erosion has been severe on the Ord River catchment and significant on the other catchments, with greatly increased runoff where the trampling of livestock has developed water-repellent surface crusts. The presence of large numbers of feral donkeys in the Victoria River watershed compounds the problem of grazing control there.

Areas delineated on the map as severely desertified are not uniformly degraded. The typical spotty character of desertification shows up in river catchments as badly overgrazed and gullied land on flood plains and on the gentle slopes adjoining them, with moderately desertified areas in the less accessible parts of those shrub steppe zones.

Distribution of saline seeps on drylands, and salinized and waterlogged irrigated lands (Mabbitt, 1978), is also spotty. Seepage of salt-laden water on slopes has plagued cultivated lands in the semiarid and subhumid regions of Western Australia, South Australia, New South Wales, and Victoria for decades. The resulting saline soils on lower slopes and in waterlogged depressions ordinarily cover a few or several hectares and are surrounded by nonsaline soils. In Western Australia where the problem is most acute, seepage of saline groundwater into streams has also had a significant adverse effect on water quality.

Waterlogging in the Murray River and its tributaries is characterized by groundwater mounds affecting tens of thousands of hectares of irrigated land (Pels, 1978). The high groundwater levels lead to secondary salinization of soils as the salts originally present in the subsoils are carried to the surface through evaporation from the shallow water tables. Since waterlogging and salinization are functions of the irrigation water management and the presence or absence of a drainage system, salty land is interspersed with nonsalty land.

Wind and water erosion have been severe in the past but are less so now, by and large. Wind erosion is an old problem in the grazing and cultivated lands of New South Wales (Wagner, 1978) and the wheat lands of the Murray Mallee (Matheson, 1978). One of the special problems in the semiarid Australia is the appearance of "scalds," when sandy surface soils are blown or washed away and the slowly permeable subsoil of texture-contrast (duplex) soils is exposed.

Revegetating scalds is difficult unless steps are taken to impound water on the surface. Mining towns such as Broken Hill have experienced accelerated wind erosion and dust storms as the land around the towns was denuded of vegetation.

Magnitude of Desertification

A 1969 report on land degradation in Australia (Newman and Condon, 1969) included an estimate that about 14 percent of the arid rangelands were severely degenerated in pastoral value, 23 percent were moderately degenerated, and 63 percent showed minor, little, or no degradation. Condon (1978) notes that much of the serious erosion and degeneration of rangelands that was evident until the late 1940s was a consequence of heavy use during the late 1800s. Since the 1940s there has been considerable improvement in range conditions, partly due to abnormally favorable rainfall over the last thirty years. The magnitude of desertification in Australia is shown in Table 5.

About 1 percent of the cultivated land in Western Australia is affected by saline seepage, with lesser percentages affected in the other states. Northcote and Skene (1972) estimate that 85,000 hectares of irrigated land in Australia are salt-affected, most of it in the Murray Basin. They also state that at least 197,000 hectares of dryland are affected by secondary salinization. Waterlogging and salinization go hand-in-hand in the Murray region.

Techniques to control desertification are well-known in Australia, the problem is recognized as a serious one, and state agencies are available to assist in carrying out reclamation and prevention programs. Grazing pressure probably has been reduced by half since the late 1800s and early 1900s and wind and water erosion in the more populous southern part of the continent is much less of a problem now. Soil salinity and waterlogging are more difficult to control because of the cost arising from the special conditions in the Murray irrigated area. One method used to dispose of saline drainage water without increasing the salinity of the Murray River water downstream is to divert it into evaporation basins. The principle is good but the cost is high and there are uncertainties about future contamination of groundwater by effluents from the basins.

NORTH AMERICA

Overgrazing has left a lasting imprint on the arid lands of North America, and little has been accomplished in attempts to restore the land to its original productivity. Wind and

Table 5
Desertification of Arid Lands of Australia

Area Affected		
Desertification Class	Square Kilometers	Percent of Arid Lands
Slight	2,330,000	36.6
Moderate	3,510,000	55.2
Severe	520,000	8.2
Total	6,360,000	100.0

Table 6
Desertification of Arid Lands of North America

Area Affected		
Desertification Class	Square Kilometers	Percent of Arid Lands
Slight	440,000	9.8
Moderate	2,720,000	61.6
Severe	1,200,000	27.1
Very severe	67,000	1.5
Total	6,360,000	100.0

Table 7
Desertification of Arid Lands of South America

Area Affected		
Desertification Class	Square Kilometers	Percent of Arid Lands
Slight	1,340,000	43.6
Moderate	1,050,000	34.1
Severe	680,000	22.1
Very severe	6,000	0.2
Total	3,076,000	100.0

Table 8
Desertification of Arid Lands of Spain

Area Affected		
Desertification Class	Square Kilometers	Percent of Arid Lands
Slight	0	0
Moderate	140,000	70.7
Severe	60,000	29.3
Total	200,000	100.0

water erosion are extensive and salinization and waterlogging--of varying degrees of severity--are common in nearly all of the irrigated valleys. At present, desertification probably has been stabilized, with some improvement in plant cover of rangelands and in control of erosion and waterlogging over the past 30 years. Salinization of irrigated land is still occurring and the dryland saline seep problem is getting worse.

The overgrazing that destroyed or severely altered the original grass cover of the rangelands of Mexico and the United States began in Mexico after the Spanish conquest and spread into the U.S. Southwest. By the early part of the 19th century, overgrazing was already a fact on both sides of the international boundary. With the explosive expansion of cattle numbers in the Southwest when the railroads arrived, range carrying capacities were greatly exceeded, and that situation continued well into the 20th century. The increased gully (arroyo) formation that occurred in overgrazed rangelands during the latter part of the 19th century has been associated with range deterioration, erosion, salinization, and waterlogging first received attention from research organizations in the late 1800s and early 1900s. Many of the basic principles for controlling desertification were established at that time. However, application of the principles did not become widespread until the 1940s and 1950s. Even though solutions are available now, much remains to be done to put them into practice.

Desertification Characteristics

There are about 450 million hectares in the arid regions of Canada, the United States, and Mexico. Approximately two-thirds of that total is moderately desertified and less than one-third severely desertified, with a considerable area of slightly desertified land and four small delineations of very severely desertified land (Figure 5).

In general, rangelands are moderately desertified in the north and severely desertified in the south. The dividing line approximates the ecotone between sagebrush (cool climate) and creosote bush (hot climate). Desertification is the result of a combination of overgrazing and water erosion. The four delineations of very severely desertified land on the map are in rangelands. Three of those four areas in New Mexico, Texas, and Chihuahua suffered first from overgrazing, then from wind erosion. The result is small hummocks of mesquite-topped sand dunes separated by completely barren blow-outs. The fourth--and largest--of the very severely desertified areas is on the Navajo Indian Reservation of northern New Mexico and Arizona. Overgrazing by sheep has removed most of the perennial grasses

Figure 5
Desertification in North America.

exposed the easily eroded, shale-derived soils to extensive sheet and gully erosion. It is unlikely that any of those areas will recover naturally in the next hundred or more years, even if grazing is completely prohibited.

Wind erosion on dry-farmed lands is the second most extensive desertification process. The ravages of the drought-inspired Dust Bowl of the 1930s were but an extreme example of the damage wind erosion has caused in the western Great Plains, the Pacific Northwest and elsewhere. Wind erosion of cultivated land, whether dry farmed or irrigated, has been a problem ever since the original sod cover was broken in the dry regions. It is an ever-present threat, worsening during droughts and lessening during wet periods, but always occurring somewhere every year. While sandy soils are the most susceptible, even fine textured soils will blow when conditions are right, and along with the dust goes organic matter and nutrient elements. The introduction of sprinkler irrigation systems, especially the center pivot systems, has enabled previously suitable rolling sandy lands to be cropped successfully. If and when those soils are abandoned, for reasons of economy or shortage of water, the United States will face an even greater wind erosion threat than it has had to cope with in the past.

Accelerated water erosion has been especially serious on overgrazed rangelands and on cropland in the Palouse region of the Pacific Northwest and in the highlands of Mexico. Normal geologic erosion has been severe in the Badlands of South Dakota and in the shale materials of Wyoming, Colorado, Utah, and New Mexico but man has managed to make matters worse through overgrazing and cultivation.

While salinity and waterlogging problems are found in nearly every irrigated valley, the only areas large enough to be shown are in the San Joaquin Valley of California, the contiguous Imperial and Mexicali valleys, and the lower valley of the Rio Grande. Land degradation is more severe in the Mexicali Valley than in the adjoining Imperial Valley because (1) the salinity of the irrigation water is worse where Mexico diverts Colorado River water than where the Imperial Valley does, and (2) the Mexicali Valley does not have the intensive drainage network that the Imperial Valley has. Waterlogging is more of a problem in the northern states of the United States, whereas the combination of waterlogging and salinization is more serious in the south. Serious salinity problems are associated with marine shales in the numerous small irrigated valleys in the upper Colorado River watershed.

The special problem of saline seeps in nonirrigated croplands is widespread in the states of Montana, South Dakota, and North Dakota and in the prairie provinces of Manitoba, Saskatchewan, and Alberta (Vander Pluym, 1978). The affected areas are too scattered to be delineated on the map.

Soil compaction and soil crusting are problems associated with medium-and fine-textured soils. Man-induced compacted subsoil layers are especially common in fine-textured irrigated soils, and in medium-to moderately coarse-textured dryland soils. Surface crusting occurs on irrigated, dryland, and rangeland soils.

Magnitude of Desertification

Nearly 90 percent of North American arid lands are moderately and severely desertified (Table 6). The status of the U.S. rangelands does not appear to have been much improved in recent years, and less than 20 percent is producing anywhere near its potential. It is likely that the same condition prevails in Mexico.

Data on the status of wind and water erosion in the arid regions are not available. A 1979 report of the U.S. Soil Conservation Service said that 58 percent of the nation's cropland needed conservation treatment of an unspecified character. That figure was down from 64 percent in 1967. The major conservation treatment undoubtedly would be water erosion control.

An indication of the magnitude of the wind erosion problem can be obtained from annual surveys conducted by the U.S. Soil Conservation Service for the Great Plains. Since 1935, surveys have been made of the amount of land damaged by wind erosion in susceptible parts of the Great Plains. Annual totals have varied from about 400,000 hectares to a maximum of 6,000,000 hectares. The lowest amount of land damage occurred during the wet 1940s and 1960s; the highest was in 1955 during a protracted drought. In any one year, the percentage of damaged cropland varied from 0.6 to 9 percent of the total cropland surveyed. There is no discernible trend upward or downward in the area damaged; the dryness of the particular year seems to determine how bad wind erosion will be.

Wind erosion in the prairie provinces of Canada--and the associated loss in soil fertility--has long been recognized as a serious problem (Soil Research Laboratory Staff, 1949). Soil and crop conditions are similar on both sides of the international boundary in the northern Great Plains, and so are erosional problems. Wind erosion is much less of a problem than water erosion in the arid regions of Mexico, where sandy soils are less prevalent in cropped areas than is the case in Canada and the United States.

Salinization is estimated to affect adversely about 25 percent of the irrigated land in the United States and a somewhat greater percentage in Mexico. Dryland saline seepage is a Canada-United States problem, with about 0.8 million hectares severely affected and perhaps another 2.7 million slightly or moderately affected. Waterlogging may affect about

the same area as salinization but the two problems are not always found together.

Techniques for controlling desertification in North America are well-known. Recognition of the seriousness of the problem varies from year to year, usually depending on the amount of notoriety droughts or floods receive. In the United States, where soil conservation has been given considerable financial support at the federal level, only about one-third of the nation's cropland has been adequately treated against soil erosion. Much less than one-third of the depleted rangeland has been improved. The situation may be better in Canada but probably is worse in Mexico.

The reasons for the relatively poor performance in combatting desertification are numerous. High among them is the emphasis on short-term versus long-term benefits (Boggess et al., 1979), inability to finance improvements, lack of clear-cut proof that control is profitable in the immediate future, and the absence of a public sense of urgency toward controlling land degradation (General Accounting Office, 1977).

SOUTH AMERICA

Overgrazing, cultivation of unsuitable land, land degradation by woodcutting--and the associated water erosion--are long-standing problems in the arid regions of South America. Wind erosion is a major threat in the semiarid pampas in Argentina, where moving sand dunes have ruined much land. Salinity and waterlogging have adversely affected irrigated regions in western Argentina, particularly along the Rio Salado, and in the numerous narrow irrigated valleys crossing the coastal plain of Peru. The very large salinas and salarea (salt lakes) found in western Argentina, northern Chile, and the altiplano of Bolivia and Chile are of natural origin rather than being due to man's activities. There is little evidence of land improvement, and conditions are either worsening, or at least not improving, nearly everywhere.

Desertification Characteristics

Heavy woodcutting for mine timbers and fuel accompanied the settling of western South America by the Spaniards in the 16th and 17th centuries. Along with that went localized overgrazing of livestock and the cultivation of the nearby slopes, with the inevitable accelerated erosion. It was not until the 19th and 20th centuries, however, that land degradation on the slopes of the Andes Mountains and the coastal ranges become widespread. Population increases and land tenure systems in effect in most countries of South America have com-

bined to put further pressure on the arid regions in recent years (Figure 6).

Desertification has been held to a moderate level in semiarid northeastern Brazil by the erratic character of the rainfall there. Land destruction probably would have been much worse if the rainfall were more reliable and if severe droughts did not occur so frequently. The original caatinga vegetation is xerophytic and is adapted to the long dry periods that occur each year (Banco do Nordeste do Brasil, 1964).

The coast of Peru is crossed by a large number of short rivers flowing from the Andes to the Pacific Ocean. Many of the irrigated valleys are affected to some degree by salinization and waterlogging. The valleys constitute only a small part of the coast desert of Peru and even less of the Chilean desert. Most of the desert has experienced little development or desertification.

Farther south, in the semiarid coastal mountains of Chile, land degradation due to overgrazing and cultivation of sloping lands has been severe around population centers. Indiscriminate woodcutting has also been an important negative factor in the development of the region.

In Argentina, which has more arid land than any other South American country, overgrazing has led to the degradation of range vegetation from the high plateaus in the north to the cold Patagonian desert in the south. Wind erosion plagues both range and cultivated lands, especially in the southern half of the nation.

Magnitude of Desertification

Overgrazing of rangelands has been the most destructive aspect of desertification in all the arid regions of South America. There has been a progressive elimination of palatable species by cattle, sheep, and goats over many decades. Along with this has gone the cutting of trees for timber, firewood, and charcoal. In the Gran Chaco of Argentina it is difficult to find good stands of the valuable red and white quebracho trees that once grew everywhere.

Water erosion has followed overgrazing through the continent, and wind erosion has become an acute problem in parts of Argentina (Prego et al., 1971). The most serious wind damage has been done in the dry-farmed areas of central Argentina where cultivation has led to the formation of mobile dunes. In 1963 an estimated 16,000,000 hectares of land were affected to varying degrees by wind erosion, and the situation has become worse since then. Soil fertility has declined and runoff has increased.

Salinization and waterlogging affect about 20 percent of the irrigated land in the coastal valley of Peru (Comite

Figure 6
Desertification in South America.

Peruano de Zonas Aridas, 1963) and large sections of the semiarid Gran Chaco, in and around the province of Santiago del Estero. The same problem, to a lesser degree, appears in the irrigated regions in arid western Argentina and in northeastern Brazil.

Although only about 22 percent of the arid regions of South America are severely or very severely affected by desertification (Table 7), the problem is critical because the best lands have been degraded and the degradation is continuing. Virtually no improvement of grazing lands has occurred anywhere, nor have soil salinization and water-logging been controlled. Some progress has been made in stabilizing sand dunes in the dry farming areas of Argentina but much remains to be done. Water erosion in the Andes and the semiarid western coastal ranges continues to present problems and may be getting worse. A fast-growing population virtually assures greater desertification in the years ahead unless governments become determined to cope with the problem.

SPAIN

All of the arid regions of Spain have been moderately to severely desertified for decades, if not for centuries (Figure 7). Most of the damage has been done by heavy grazing and woodcutting, but wind and water erosion on cultivated land has also been extensive. Salinization and waterlogging of irrigated land is a major problem in parts of the Ebro River watershed in northeastern Spain. Destruction of the native vegetation and the subsequent water erosion of thin soils on slopes has had a devastating effect on the plant environment and on the potential productivity of the land. Mobile sand dunes are found along the Mediterranean coast.

Desertification Characteristics

During the past several centuries, heavy grazing by sheep and goats has led to the destruction of much of the herbaceous and woody vegetation on the noncultivated land (Albareda, 1955). Water erosion has been severe on the overgrazed slopes as a result of the loss of vegetative cover and the torrential character of the rains. Cutting of wood for fuel and construction and the extension of dryland farming into the pasture lands has accompanied overgrazing. Plant cover has changed to a more xeric type and surface runoff has increased.

A monoculture of grain in the cultivated regions has depleted the native fertility of the soil and has been responsible for increasing the susceptibility of the land to wind and water erosion. Extended droughts from time to time have

Figure 7
Desertification in Spain.

served to accelerate desertification. Water erosion is severe nearly everywhere on sloping land.

Salinization and waterlogging do not affect a high percentage of the total cultivated land in Spain but important and large areas of affected soils do occur in irrigated valleys. The major salt-affected areas in the northeast are in the Ebro River watershed in the vicinity of Zaragoza and Herida. Seepage water from irrigation on the higher land has caused waterlogging and salinization of lower-lying areas (Martinez, 1978). Gypsum is a common constituent of the soils.

The other major salt-affected areas are in southwest Spain near the coast. The soils are composed of fine-textured sediments that were subjected regularly to flooding by seawater in the past. Surface and internal drainage of the irrigated land is poor and water tables generally are close to the surface. Pumping is required to lower the water tables (Ayers et al., 1960).

Magnitude of Desertification

Approximately 50 percent of Spain is arid. In the arid regions about 70 percent is moderately desertified and 30 percent is severely desertified (Table 8).

Virtually all of the rangeland has suffered severe land degradation. Range productivity is probably stabilized at a low level now, with little improvement anywhere. Erosion continues on the extensive dryfarm lands except in a few places where soil and water conservation measures have been instituted. Soil fertility remains low.

Salinization and waterlogging affect about 240,000 hectares of irrigated land to various degrees (Ayers et al., 1960). Reclamation has been undertaken in several areas. Because of the need for more agricultural production and the gradual worsening of the salt problem, irrigation has received special attention in recent decades. Much remains to be done.

Soil and water conservation techniques for dryland farming are known but their application to the field is limited. Range management is not a well-supported science in Spain. Considerable progress has been made in the reclamation of saline soils and procedures for doing so are quite well-known (Martinez, 1978).

REFERENCES

- Aboukhaled, A., Arar, A., Balba, A.M., Bishay, B.G., Kadry, L.T., Rijtema, P.E., and Taher, A. (1975) Research on Crop Water Use, Salt Affected Soils and Drainage in the Arab

- Republic of Egypt. Near East Regional Office, FAO, Cairo, 92 p.
- Albareda, J.M. (1955) Influence des changements de la vegetation dans les sols arides. In: Plant Ecology, Arid Zone Research V, UNESCO, Paris, p. 84-88.
- Aubreville, A. (1949) Climats, Forest, et Desertification de l'Afrique Tropicale. Societe de Editions Geographiques, Maritime et Coloniales, Paris, 255 p.
- Ayers, A.D., Vasquez, A., de la Rublia, J., Blasco, F., and Samplon S. (1960) Saline and Sodic Soils of Spain. Soil Science, v. 90, p. 133-138.
- Banco do Nordeste do Brasil (1964) O Nordeste e as Lavouras Xerofilas. Banco do Nordeste do Brasil, S.A., Departamento de Estudos Economicos do Nordeste, Fortaleza, Ceara, 238 p.
- Boggess, W., McGrann, J., Boehlje, M., and Heady, E.O. (1979) Farm Level Impacts of Alternative Soil Loss Control Practices. Jour. Soil and Water Conservation, v. 34, p. 177-183.
- Cloudsley-Thompson, J.L. (1974) The Expanding Sahara. Environmental Conservation, 1: 5-13.
- Comite Peruano de Zonas Aridas (1963) Informe Nacional sobre las Zonas Aridas. Republica del Peru, Ministerio de Agricultura, Lima, 105 p.
- Condon, R.W. (1978) Land Tenure and Desertification in Australia's Arid Lands. Search, v. 9, p. 261-264.
- Cooke, R.H. and Reeves, R.W. (1976) Arroyos and Environmental Change in the American Southwest. Clarendon Press, Oxford, England, 213 p.
- de Crespigny, R.R.C. (1971) China. St. Martin's Press, New York, N.Y., 235 p.
- Dahl, G. and Hjort, A. (1979) Pastoral Change and the Role of Drought. Swedish Agency for Research Cooperation with Developing Countries Report R2: 1979, Stockholm, Sweden, 50 p.
- Delwaulle, J.C. (1973) Desertification de l'Afrique au sud du Sahara. Bois et Forets des Tropiques, v. 149, p. 3-20.
- Delwaulle, J.C. (1977) La situation forestiere dans le Sahel. Bois et Forets des Tropiques, v. 173, p. 3-22.
- Dougrameji, J.S. and Clor, M.A. (1977) Case Study on Desertification. Greater Mussayeb Project. Iraq. United Nations Conference on Desertification A/CONF., 74/10, 102 p.
- Dregne, H.E. (in press) Historical Perspective of Accelerated Erosion and Effort on World Cultivation. American Society of Agronomy, Madison, Wisconsin.
- General Accounting Office (1977) To Protect Tomorrow's Food Supply Soil Conservation Needs Priority Attention. Report to the Congress, CED-77-30, Washington, D.C., 59 p.
- Jacobsen, T. and Adams, R.M. (1958) Salt and silt in Ancient Mesopotamian agriculture. Science, v. 128, p. 1251-1258.

- Mabbutt, J.A. (1978) Desertification in Australia. Water Research Foundation of Australia Report No. 54, Kingsford, N.S.W., Australia 132 p.
- Matrinez Beltran, J. (1978) Drainage and Reclamation of Salt-Affected Soils in the Bardenas area, Spain. International Institute for Land Reclamation and Improvement Publication No. 24, Wageningen, The Netherlands, 321 p.
- Matheson, W.E. (1978) Soil Loss Made South Australia Come Down to Earth. Journal of Soil Conservation Service of N.S.W., v. 34, p. 88-100.
- Newman, J.C. and Condon, R.W. (1969) Land Use and Present Condition. In: R.O. Slayter and R.A. Perry (Editors), Arid Lands of Australia, Australian National University Press, Canberra, Australia, p. 105-132.
- Northcote, K.H. and Skene, J.K.M. (1972) Australian Soils with Saline and Sodic Properties. C.S.I.R.O. Soil Publication No. 27, Melbourne, Australia, 62 p.
- Office of Environmental Planning and Coordination (1977) Country Report: India. Department of Science and Technology, Government of India, New Delhi, 71 p.
- Pearse, C.K. (1971) Grazing in the Middle East: Past, Present, and Future. Journal of Range Management, v. 24, p. 13-16.
- Pels, S. (1978) Waterlogging and Salinization in Irrigated Semi-arid Regions of NSW. Search, v. 9, p. 273-276.
- Prego, A.J., Ruggiero, R.A., Alberti, F.R., and Prohaska, F.J. (1971) Stabilization of Sand Dunes in the Semiarid Argentine Pampas. In: William J. McGinnies, Bram J. Goldman, and Patricia Paylore (Editors), Food, Fiber and the Arid Lands, University of Arizona Press, Tucson, p. 369-392.
- Rapp, Anders (1974) A review of Desertification in Africa--Water, Report No. 1, Stockholm, Sweden, 77 p.
- Soil Research Laboratory Staff (1949) Soil Moisture, Wind Erosion and Fertility of Some Canadian Prairie Soils. Department of Agriculture Publication 819, Ottawa, Canada, 78 p.
- Stebbing, E.P. (1937a) The Encroaching Sahara: The Threat to the West African Colonies. The Geographical Journal, 86: 506-519.
- Stebbing, E.P. (1937b) The Threat of the Sahara. Journal of the Royal African Society. Extra Supplement to vol. 36, 35 p., May 25, 1937.
- Stebbing, E.P. (1938) The Advance of the Desert. The Geographical Journal 91: 356-359.
- UNESCO (1977) World Distribution of Arid Regions. Map Scale: 1/25,000,000, UNESCO, Paris.
- United Nations (1978) United Nations Conference on Desertification. Round-up, Plan of Action, and Resolutions, United Nations, New York, 43 p.

- Vander Pluym, H.S.A. (1978) Extent, Causes and Control of Dryland Saline Seepage in the Northern Great Plains Region of North America. In: H.S.A. Vander Pluym (Editor), Dryland-Saline-Seep Control, Agriculture Center, Lethbridge, Alberta, v. 1, p. 48-58.
- Wagner, R. (1978) Soil Conservation in New South Wales, 1938-1978. Journal Soil Conservation Service of N.S.W., v. 34, p. 124-132.
- Weaver, J.E., and Albertson, F.W. (1940) Deterioration of Midwestern Ranges. Ecology, v. 21, p. 216-236.
- Widstrand, C.G. (1975) The Rationale of Nomad Economy. Ambio, v. 4, p. 146-153.

WIND VERSUS WATER EROSION IN THE CONTEXT OF DESERTIFICATION

Michael M. Verstraete and Jean-Pascal van Ypersele

ABSTRACT

While the Workshop on the Physics of Desertification focused on the process of wind erosion, it is important to realize that desertification is a much broader issue, and that land degradation itself is not limited to wind erosion. This paper attempts to compare wind to water erosion from the point of view of their respective characteristics and impacts. The relative importance of water and wind erosion processes must be assessed on a local basis, but both forms of erosion are very damaging, and threaten the long-term economic viability of many arid and semi-arid regions.

INTRODUCTION

The Workshop on Desertification, organized in November 1980 by the International Centre for Theoretical Physics (Trieste, Italy) as part of a larger course on the Physics of Flow in the Oceans, Atmosphere, and Deserts, focused on the problems of wind erosion and associated phenomena like sand dune movement and dust transport. If these processes may have a severe impact on the agropastoral systems of numerous countries, it remains that they represent only particular cases of erosion. Furthermore, the concept of desertification is itself much broader than erosion alone.

In this short paper, we compare wind and water erosion from the viewpoint of their characteristics and their impact on the ecosystems and economies, in order to provide some

Figure 1

Relationships of water erosion (continuous lines) and wind erosion (broken lines) with increasing mean annual precipitation. The curves for water erosion indicate the relationships with mean annual precipitation for (A) areas of natural vegetation cover and (B) bare ground. The curves for wind erosion indicates the relationships with mean annual precipitation for (C) areas of natural vegetation cover and (D) bare ground. These curves are based on what would be expected from the relationship of wind erosion to vegetation cover and to moist soil (Marshall, 1973).

broader perspective on the role of wind erosion in the context of desertification.

The subject of wind erosion is adequately covered in the other papers of this volume, and the subject of water erosion would take itself at least as much space for even a cursory treatment. It is therefore obvious that this paper can only present a broad outline of the subject.

SOIL EROSION: WHERE AND WHEN

Perhaps the first question to ask is, What is soil erosion? Briefly, it is soil degradation in the form of top-soil removal and, usually, transport to unwanted locations. This is deleterious to plant production and has a potentially large impact on primary production. In order to gain a global perspective on the problem, the next question to investigate is Where does the phenomena occur? Our assessment is based on a study by Marshall (1973), and it is worth spending a few minutes on Figure 1, taken from this publication.

It is important to stress that, under normally vegetated conditions (curve A), water erosion presents a maximum for semiarid conditions, where the lack of vegetation cover leads to low infiltration rate, and decreases to virtually zero for very dry and very wet regimes for obvious reasons: under very dry conditions the vegetation cover is sufficient to hold the soil and facilitate infiltration. In the case of wind erosion (curve C), the vegetation cover increases monotonically with mean annual precipitation and help to protect the soil. Curves B and D show the typical soil losses, for water and wind erosion respectively, when the vegetation cover is completely destroyed, for example, through overgrazing, wood collection, or improper harvesting methods. These curves do not represent the maximum erosion possible because the surface can still be disturbed mechanically (e.g., using heavy agricultural machines, or by overtrampling), in which case the total erosion losses may still be higher.

Clearly, most ecological situations are time-dependent and a given ecosystem will become more sensitive to wind erosion during dry and windy periods, or to water erosion during wet periods, especially when rains come as thundershowers. Furthermore, the climatic conditions that favor wind erosion may also increase the risk of water erosion. For example, the impact of heavy rains at the end of a dry spell may be more dramatic than at the end of the rainy season.

WIND VERSUS WATER

Table 1 is presented here as a quick way to compare wind and water erosion in terms of their characteristics and impacts. Each entry in this table should require a detailed

Table 1

	Wind Erosion	Water Erosion
Where?	Deserts and arid regions; improperly managed semi-arid regions. <i>Humid regions with much or partly arid</i>	Semi-arid and sub-humid regions, <i>everywhere</i> especially when improperly managed
When?	Dry periods (dry season, drought...)	Wet periods (wet season, wet spells, thunderstorms...)
Condition	Friction wind speed above threshold value, which depends on size distribution of surface soil aggregates, soil moisture, vegetation.	Precipitation rate higher than infiltration rate
Transport	Wind displaces particles selectively	All sizes particles are roughly displaced <i>and a good</i> over similar distances: 10^2 - 10^5 m <i>large - less small - greater</i>
What & How Far?	According to diameter (mass), through different processes: - Sand (mainly creeping & saltation): 10 - 10^4 m - Dust (mainly suspension): 10^4 - 10^7 m	Nutrients can be washed away by <u>undersurface flow</u> <i>not in erosion (surface)</i> <i>dry</i>
How Fast?	- Sandstorm: 10^3 - 10^4 s - Duststorm: 10^4 - 10^7 s	Thunderstorm: 10^2 - 10^3 s
How Much?	10^9 ton yr^{-1} for USA (48 contiguous states only) (from Pimentel, 1976) 6 - 20 10^7 tonnes yr^{-1} of Saharan dust (Junge, 1977)	4 10^9 t yr^{-1} for USA (same remark and source)
Main Impacts	- Loss of small particles, organic materials, hence soil fertility - Radiative energy transfers and regional climate - Dust-induced health hazards - Sandblasting effects on plants	- Loss of soil and nutrients - Gullies, sheet erosion - Obstacles to agricultural mechanization - Silting of dams
Who is Affected?	Principally: pastoralists, nomads Secondarily: cultivators <i>All people and animals, etc. (air pollution)</i>	Principally: cultivators Secondarily: pastoralists, nomads <i>people in flood plains</i>

treatment, but we limit ourselves here to some particular points.

Soil particles are moved by the surface wind flow when the friction speed exceeds some threshold value that is dependent on numerous local factors, including surface roughness, soil moisture, etc. In the case of water erosion, the soil is displaced mainly by water flowing freely at the surface, a situation that occurs essentially when the precipitation rate is greater than the infiltration rate, also a function of various local parameters. Roose (1970) observed in Senegal that essential plant nutrients, in the form of colloids and minerals in ionic solution, can be removed by subsurface water flow.

It must be kept in mind that the soil fraction moved by either form of erosion is very different in each case: if the wind moves most particles over large areas, the heavier ones (sand) are displaced over rather short distances (often as dunes), while the finer material may stay in the atmosphere for days and weeks, and be transported from one continent to another. For example, Saharan dust is regularly collected in Central America and even Florida (depending on seasons and wind patterns). By opposition, water runoff displaces all soil particles but the heaviest ones, and may create gullies and other forms of geomorphological degradation. In most cases, the soil displaced by water erosion is lost to the ocean floor when it does not fill dams and reservoirs. However, in a few instances, the material is collected in specific locations: for example, erosion may be a means to move agricultural lands from impractical slopes to better located valleys.

The differences in the nature of selectivity for wind and water erosion have to be remembered when one wants to assess the respective importance of wind and water erosion: for example, wind-eroded material may be more precious than an equivalent quantity of water-eroded material because of its higher density in organic material, but on the other hand, if the water-eroded material is taken in the form of gullies covering a fraction of a given surface, these can make it impossible to use agricultural machinery over the whole surface.

THE SOIL AS A QUASI-NONRENEWABLE RESOURCE

Desertification is often considered to be a long-term ecological degradation, and both forms of erosion are normally treated as desertification processes. It may therefore seem contradictory to show how fast they are: indeed sand or dust storms are synoptic disturbances whose time scales are measured in hours and days, and the heavy thunderstorms responsible for the most severe water erosion losses are also

rather short-lived systems. The point however is that those events have long-lasting consequences, because of a very strong positive feedback: the loss of soil, which is often originated by a reduction of the previous vegetation cover, further reduces the ability of the vegetation to recover from such destruction, and thereby generates further loss of soil. Soil erosion is the major mechanism whereby desertification becomes irreversible for all practical purposes. It should be clear, however, that all soils undergo some level of erosion, given by the curves A and C of Figure 1. It is the additional, accelerated erosion loss due to the removal of vegetation that is the main cause of concern because it is not counterbalanced by soil formation, a very slow process itself dependent on vegetation cover.

All land uses may generate increased erosion. But for a given final result (agricultural output), some of them are most efficient at conserving the soil. The soil itself must be considered as a precious nonrenewable resource, because the soil formation is such a slow process when compared to the soil erosion. Furthermore, some agricultural practices may in fact diminish the expected erosion losses, for example: by reducing the slope of the fields with terraces; protecting the land with crop residues or other surface treatments after harvesting; all methods devoted to increasing the infiltration capacity of the soil, reducing the velocity of runoff and wind, or maintaining the soil structure.

CONCLUSIONS

The dependence of all societies on agricultural (primary) production is such that the control of soil erosion should receive a very high priority in all national budgets. From the above, it must be clear that both forms of erosion have a significant impact on the natural resources of large areas in the world.

Desertification is not an academic problem but concerns millions of people living in arid, semiarid and subhumid regions, and the relative importance of each form of erosion should be estimated according to their respective impact on the ecosystem and the population considered, keeping in mind the possible interactions between different processes, and the external constraints.

As the population density, as well as the biomass is significantly higher in semiarid and subhumid zones than in arid ones, it should be recognized that the risk of water erosion, a function of population density itself, concerns more people globally than wind erosion. However, since the control of both forms of erosion is not only a response to local ecological conditions, but is also a function of local social, economical, and political situation, it is not possible to

make recommendations of general value as to the priority that should be given to one or another form of erosion.

We hope that this short paper will be useful to clarify the main characteristics, roles, and impacts of wind and water erosion, and will stimulate further awareness of the worldwide issue of soil erosion, which is but one of the numerous soil degradation processes related to desertification.

Acknowledgements

Financial support to one of the authors (JpvY) has been provided by the Fonds National de la Recherche Scientifique (Belgium) to attend the Workshop on Desertification.

REFERENCES

- Junge, C. (1977) The Importance of Mineral Dust as an Atmospheric Constituent, Preprint to the Workshop on Saharan Dust, Gothenburg, Sweden, April 1977, pp. 25-28.
- Marshall, J.K. (1973) Drought Land Use and Soil Erosion. In: The Environmental Economic and Social Significance of Drought, J.V. Lovett, (ed.) Angus and Robertson, pp. 55-77.
- Piometal D. (1976) Land Degradation: Effects on Food and Energy Resources, Science 194, pp. 149-155.
- Roose, E.J. (1970) Importance relative de l'erosion, du drainage oblique et vertical dans la predogenese actuelle d'un sol ferrallitique de moyenne Cote d'Ivoire, Cah. ORSTOM, ser. Pedol. 8, pp. 469-482.

THE UNITED NATIONS ORGANIZATION
AND THE ISSUE OF DESERTIFICATION

By

Michel M. Verstraete

ABSTRACT

The wide interest in processes of desertification and possible remedial actions is rather recent: it dates back to the severe drought that struck the Sudano-Sahelian zone of West Africa about a decade ago. The United Nations system had a definitive role in catalyzing this interest and organizing a worldwide response to this relatively new issue. This paper reviews the role played by various agencies and organizations of the United Nations in the planning, holding, and follow-up of the international conference on desertification in Nairobi, Kenya. Research and training opportunities within the UN system related to desertification are also reviewed, and a series of useful addresses are listed in an annex.

DISCUSSION

The importance, nature, and extent of desertification, an issue that emerged as a world problem during the last decade, became well-known largely as a result of various activities of the United Nations (UN), in particular the UN Conference on Desertification (UNCOD).

The objectives of this short paper are: (1) to outline how the UN first became involved with the issue of desertification and then a leader of the world's efforts to combat it, and (2)

to summarize, for the benefit of the participants of this workshop, some useful information about the research and the training on desertification that has been organized by various Agencies of the UN system.

While accelerated land degradation has probably been known to man since the time he started agriculture, the worldwide concern about desertification dates back only fifteen years or so. In fact, it is often said that the problem experienced during the well-publicized Sudano-Sahelian drought (1968-1974) served to attract and focus the world's attention, not only on relief measures for the concerned populations, but on the environmental degradation processes that seemed to have greatly enhanced the impact of the drought on the social, political, and economical structures of the countries concerned.

In 1973, the United Nations Sahelian Office (UNSO) was established to act as the central coordinating mechanism of the United Nations for the implementation of medium and long-term recovery and rehabilitation programs for the drought-stricken Sahelian countries. But soon the nature and extent of the problem became evident, and it was realized that drought relief measures alone would not be sufficient in the long run.

In 1974, the General Assembly (GA) adopted a set of resolutions relative to the preparation and convening of the World Conference on Desertification (UNCOD) already mentioned. The objectives of this conference were threefold: (1) to raise worldwide awareness about desertification, (2) to bring together all the scientific and technical knowledge available at that time on the problems and their solutions, and (3) to launch and implement, in affected countries, the necessary plans and programs to combat desertification, with the ultimate goal of stopping further degradation by the year 2000.

Some 500 delegates from 94 countries and from a number of international and nongovernmental organizations participated in the Conference, which was held in Nairobi from August 29 to September 9, 1977.

The boldest step taken by the Conference was certainly the adoption of the UN Plan of Action to Combat Desertification (PACD). This Plan of Action consists of 28 recommendations for action at the regional, national, and international levels to avoid desertification through better planning, to alleviate its effects through improved management and the introduction of appropriate technologies, and to ensure long-term solutions through education and integrated development.

Another document that played an important role and was the subject of much discussion during the Conference was the World Map of Desertification, at the scale 1/25,000,000, prepared on behalf of the UN Environment Programme (UNEP) by the Food and Agriculture Organization (FAO), in cooperation with the UN

Educational, Scientific and Cultural Organization (UNESCO) and the World Meteorological Organization (WMO). Both the Plan of Action and the Map were distributed to all participants in the Workshop.

In addition to these, the following three series of documents, which were also prepared as background or discussion material for the Conference, should be mentioned here because of the scope and the depth of scientific information they contain: (1) a series of four component reviews dealing respectively with the climatic, ecological, sociological, and technological aspects of desertification; (2) a series of six commissioned case studies on the degradation processes occurring in carefully selected locations, and (3) a series of national case studies presented by the governments of most of the affected countries. These three sets of documents form a very comprehensive compilation on the problems of desertification in the world.

From all of these studies and documents, a rather bleak perspective on desertification emerged. It is not possible to summarize here even the main scientific findings of the Conference, but three of the estimates made at that time will be recalled here in order to give an idea of how the situation was perceived:

1. About one-third of all land surfaces are desertic, arid, or semiarid, and most of the regions under the last two categories are prone to desertification
2. More than 600 million people live in these areas, of which some 78 million are already directly threatened by desertification
3. Close to 60,000 km² of land are lost annually due to desertification in the world.

Some of these estimates have been challenged since then, but the prevalent opinion was that the problem was very serious and needed both immediate and sustained attention.

Not surprisingly, the Plan of Action dealt explicitly with the institutional and financial requirements needed to deal adequately with such a global issue. These provisions are now reviewed in detail because they specify how and where developing countries may obtain additional support to combat desertification.

Clearly, somebody should look after the implementation of the PACD. However, most governments and organizations were opposed to the creation of yet another institution to do this. Moreover, many of the existing specialized agencies of the UN organizations were already dealing with specific aspects of the issue. After long and difficult discussions, the Conference recommended UNEP as the primary organization responsible for the follow-up and coordination of the imple-

mentation of the PACD. The UN General Assembly later endorsed the Plan of Action and thereby confirmed UNEP in this role. As a result, UNEP set up a Desertification Unit (now Desertification Branch) within its own structure to carry out the various tasks scheduled in paragraph 103 of the PACD.

Meanwhile, the number of countries assisted by UNSO had increased from six in 1973 to eight in 1975. Three years later, in December 1978, the General Assembly designated UNSO (which now stands for the United Nations Sudano-Sahelian Office) as the arm of the United Nations to be responsible for assisting, on behalf of the United Nations Environment Program, the efforts of the fifteen countries of the Sudano-Sahelian region situated south of the Sahara and north of the equator to implement the Plan of Action to Combat Desertification.

This additional mandate was further extended to eighteen countries in 1980; these are Cape Verde, Chad, Djibouti, Ethiopia, Gambia, Guinea, Guinea Bissau, Kenya, Mali, Mauritania, Niger, Nigeria, Senegal, Somalia, Sudan, Uganda, United Republic of Cameroon, and Upper Volta.

Both the Desertification Branch and UNSO are geared toward helping the developing countries of their respective zones of responsibility to avoid, alleviate, or combat the effects of desertification. Although these bodies send exploratory missions to the concerned countries to assess the situation and needs, and to raise awareness on these problems, it is generally up to the governments to request assistance as and when required.

Typically, such assistance would permit the government to design and to run a national system for the surveillance of the nature and status of the country's ecological resources, and, in particular, to assess and monitor any trends or developments in desertification. A sophisticated and well-established methodology for doing so has been developed by UNEP's Global Environmental Monitoring System (GEMS), based on three levels of data acquisition: ground surveys, aerial reconnaissance flights, and satellite imagery.

It could also assist in the drafting of a comprehensive national PACD, to be integrated in the national development plan. Or it might assist in designing a particular project in an affected region, whether it is oriented toward production, education, information, or research.

This list is by no means exhaustive but it is certainly representative of the kind of services provided by the Desertification Branch and by UNSO to the desertification-prone countries.

The other essential aspect of the combat of desertification is, of course, the mobilization of the financial resources to carry out all the recommendations for action whenever and wherever necessary. At the time of Conference, it was esti-

mated that an amount not less than 400 million US dollars should be spent annually for 20 years in affected developing countries, in addition to the amounts already scheduled for their general development, in order to stop further loss of land through desertification by the year 2000. This does not include the cost of any attempt at recovering areas already affected or desertified.

The PACD accordingly recommended the establishment or strengthening of subregional cooperation, as well as bilateral and multilateral assistance. As this was judged insufficient, the implementation of three specific proposals was also recommended to address this question of financing. These have been undertaken as follows:

1. The General Assembly established a Special Account for the implementation of the PACD. This account was meant to receive voluntary and/or automatic contributions from governments, international taxation, donations, multilateral financing institutions and interest-free loans. So far, a few governments have already contributed to this account.
2. A study to identify additional ways and means of financing the combat against desertification has also been made. It reviewed the various proposals made within and outside the UN system over the last 20 years to raise money for development, and assessed the feasibility and practicality of these in the light of the present political and economic situation. This study was presented to the General Assembly at its 35th Session in 1980, and is followed-up by concrete investigations on those proposals deemed practicable.
3. Finally, the Executive Director of UNEP is convening, at approximately 18-month intervals, a Consultative Group on Desertification Control, also known as DESCON, to bring together the likely sources of financing and the developing countries suffering from desertification. A small number of projects are carefully selected before each meeting of DESCON, which acts therefore as a forum where the concerned parties have an opportunity to raise or show interest for specific antidesertification activities. The Desertification Branch is the secretariat of DESCON.

So far, only some of the new mechanisms set up within the UN system for the specific purpose of combating desertification have been described. It should be recalled here again that many of the Specialized Agencies of the UN organization have considerable experience and numerous activities in some aspects of desertification. In order to improve the availability of this experience and to coordinate these activities, an

Inter-Agency Working Group on Desertification (IAWGD) has been set up. One of its realizations is the preparation and updating of a complete directory of all projects and activities of these Agencies in the field of desertification, organized by subject and by country. The Desertification Branch is also the secretariat of IAWGD.

What precedes can only be a broad outline of the operative mechanisms and institutions available within the UN to deal with the problems of desertification. More information on specific programs or for selected countries should be obtained by writing directly to the appropriate Agency or Organization. A list of useful addresses is enclosed in the annex.

The last part of this paper will concern itself with research and training opportunities in desertification offered by various UN Agencies.

RESEARCH OPPORTUNITIES

Original research, fundamental as well as applied, is a limited and rather exceptional activity for most of the agencies of the UN Organization (UNO). But two UN bodies are specifically oriented toward such objectives, even though desertification might not be a priority area, and these are the United Nations University (UNU) and the International Centre for Theoretical Physics (ICTP). Both these organizations have various schemes to support outstanding individual researchers from developing countries. The UNU also supports a limited number of research projects in developed countries.

UNESCO is of course very much interested in promoting research in developing countries, although research per se is not one of its main objectives. Of much interest in the present case is UNESCO's Man and Biosphere (MAB) Program, and, in particular, its third component: MAB-3, dealing explicitly with the problems of arid ecosystems. The Integrated Project in Arid Lands (IPAL) is a research project presently implemented in Kenya and Tunisia under MAB-3.

Most other Agencies of the UN organizations may have occasional research and activities on desertification in the framework of specific projects, but these are rather marginal. Also, limited support for research projects could be obtained in principle in the framework of a national PACD or through DESCON. In both cases, however, it is up to the concerned government to assign a high priority to these projects and to submit them to UNEP for financing.

Finally, particular attention should be given to the results of the recent UN Conference on Science and Technology for Development (UNCSTD). A fund for promoting Research and Development in developing countries has been set up and is administered by the UN Development Program (UNDP). It is too

early however to give precise indications on the opportunities offered.

Before leaving this subject, it should be added that the UN Plan of Action stresses the need for strengthening national and regional research facilities to improve our understanding of the processes of desertification and the efficiency of the methods used to combat it. If no specific step forward has been made so far to implement this recommendation, it is not unlikely that UNEP might, at some point in the future, support the creation or strengthening of some regional research center on desertification.

TRAINING OPPORTUNITIES

Coming now to the training opportunities, the situation is relatively different. It has long been recognized that one of the major drawbacks of development in developing countries was the lack of trained manpower at most levels of economic activity. Accordingly, training receives a much higher priority than research in the activities of the UN, as well as many other development aid programs.

Most of the Agencies of the UN organize training courses, seminars, workshops, and other educational activities in their fields of interest. Some of them, like WMO, have even set up a worldwide network of training centers to meet specific requirements. A number of such activities may be marginally related to desertification, and the various Agencies should be approached to obtain additional information on such training activities.

In addition, UNEP is organizing, in cooperation with the respective governments, two yearly training courses directly and principally aimed at desertification, in the USSR and China. English is the working language. Lecturers are world authorities on the subjects treated, which range from sand dune fixation to rangeland management. These courses are four to six weeks long and usually include field trips. It is up to the governments of the desertification-prone countries to select potential candidates for these courses and to submit their names, through the office of the Resident Representative of UNDP, to UNEP, which will make the final selection. The cost of participation to these courses is borne by UNEP according to the terms of the project document, and includes travel, subsistence, and training materials.

CONCLUSIONS

To conclude, while desertification is perceived as a threatening issue for the global food supply and the stability of human societies, especially in developing countries, the fact that it develops slowly and cumulatively over periods of

time that are long with respect to the usual political and economic time scales also complicates the issue. In a way, it would be easier to deal with a sudden catastrophe rather than with a recurrent problem.

It is therefore essential to raise the awareness of all people, directly affected or not, from developing as well as developed countries, to initiate and sustain a large-scale and long-term effort to eradicate the global problem. This workshop was a small but significant step forward in that direction, especially as it introduced many participants to the subject of desertification for the first time.

This paper will hopefully be useful to them in providing a historical perspective and in pointing out the needs for action and the intellectual and financial resources available from the United Nations system.

GENERAL REFERENCES

1. Secretariat of the United Nations Conference on Desertification (1977) Desertification, Its Causes and Consequences; Pergamon Press; p. 448.
2. United Nations (1978) United Nations Conference on Desertification, Round-up, Plan of Action and Resolutions; United Nations; New York.
3. Michael H. Glantz, Editor (1977) Desertification, Environmental Degradation In and Around Arid Lands; Westview Press, Boulder, Colorado.
4. Victor A. Kovda (1980) Land Aridization and Drought Control, Westview Press, Boulder, Colorado.
5. Robert W. Wright, Editor (1978) The Reclamation of Disturbed Arid Lands, University of New Mexico Press, Albuquerque, New Mexico.

Following is a list of suggestions and useful addresses to obtain more information on the various topics dealt with in this paper. As a matter of principle, it is faster and more efficient, in developing countries, to contact first the office of the Resident Representative of the UN Development Programme (UNDP). He is the official representative in each country, of the United Nations at large and of UNEP in particular. Some specialized agencies of the UN also have national or regional representatives (UNESCO and FAO for example). The permanent Representative of WMO in all countries is the Director of the National Meteorological Service. For all these, no address is shown here as they can easily be obtained from the national telephone directory.

As explained in the text, UNEP is responsible within the UN for the overall follow-up and coordination of the implementation of the Plan of Action to Combat Desertification. This responsibility is administratively shared by two bodies,

which may be contacted for all questions related to desertification in the countries indicated:

- Enquiries concerning desertification in Cape Verde, Mauritania, Senegal, Gambia, Guinea Bissau, Guinea, Mali, Upper Volta, Niger, Nigeria, United Republic of Cameroon, Chad, Sudan, Ethiopia, Djibouti, Somalia, Uganda and Kenya should be directed to:
The Director
United Nations Sudano-Sahelian Office (UNSO)
1 United Nations Plaza
New York, NY 10017
USA
- Enquiries about desertification in all other countries of the world should be addressed to:
The Chief
Desertification Branch
United Nations Environment Program
P.O. Box 30552
Nairobi, Kenya
- If information is sought on the specific topic of environmental monitoring, in particular by satellite remote sensing, contact
The Director
Global Environmental Monitoring System (GEMS)
United Nations Environment Program
P.O. Box 30552
Nairobi, Kenya
- Only when the information needed cannot be obtained locally should the headquarters of the various organizations be approached:
FAO: The Director General
Food and Agriculture Organization
Via delle Terme di Caracalla
00100 Rome, Italy
UNESCO: The Director General
United Nations Educational, Scientific and Cultural Organization
7, Place Fontenoy
75700 Paris, France
WMO: The Secretary General
World Meteorological Organization
P. O. Box 5
CH-1211 Geneva 20, Switzerland
ICTP: The Director
International Centre for Theoretical Physics
P. O. Box 586 Miramare
I-34100 Trieste, Italy

UNU: The Rector
United Nations University
Toho Seimei Building
15-1, Shibuya-ku
Tokyo 150, Japan

THE IMPACT OF DESERTIFICATION ON LAND PRODUCTIVITY IN SUDAN

By

M. O. H. El-Karouri

Desertification, which is a progressive environmental degradation terminating in desert-like conditions, has been given different definitions and names. The controversy arises from differences of opinion regarding the relative importance of its various causes. However, the United Nations' definition seems to be widely accepted. It defines desertification as "a process leading to reduced biological productivity with consequent reduction in plant biomass, in the lands, carrying capacity for livestock, in crop yields, and human well-being."

Desertification has been associated in the minds of many people with aridity and moisture deficiency. If so, a simpler definition could be: "A reduced productivity of agricultural land due to physical disturbance of the rhizosphere through sand deposition or removal of soil by wind action." The reduction in plant biomass, in the land carrying capacity for livestock and the increased hazards to man's well-being are manifestations of the deterioration of the ecosystem due to desertification. Other agents that adversely affect the productivity of the land, e.g., water erosion, compaction, and nutrient depletion are secondary causal agents of desertification and should not be equated with wind erosion, which is the primary causal agent. Likewise, different names have been used for desertification, such as desertization, desert creep, desert encroachment, etc. These and other names have been subject to considerable debate, especially with respect to popular names, e.g., desert encroachment and desert creep, both of which imply that the deserts are moving outwards by

themselves. Scientifically, desertification or desertization are more appropriate names, but the popular names have a greater impact, conveying a clearer message to policymakers.

DESERTIFICATION IN SUDAN

The Sudan falls within the zone where the risks of desertification are great. The area threatened by desertification hazards lies between 12° to 18° N, extending across the country from east to west. It includes a northern strip along the Nile up to the Egyptian border between 30° - 32° E covering a total area of 65,000,000 hectares.

Sudan's arid and semiarid ecology is mainly determined by the total amount and distribution of rainfall and to a more limited extent, by soil type, topography, and elevation. The annual rainfall within the area ranges from 75 mm in the north to about 400 mm in the south. The rainfall is controlled by the nature and behavior of the Inter Tropical Convergence Zone (I.T.C.Z.), which itself is controlled by the atmospheric pressure in the vicinity of the African Continent (Mahdi, 1975). However, due to the nature of the I.T.C.Z., marked variations in rainfall exist, especially over the northern part of the area. The coefficient of variation of the annual rainfall means varies between 100% and 150% for most of the area, thus indicating the great variation in rainfall within the area.

In the southern part, the annual evapotranspiration is about 1,900 mm but in the extreme north, it is over 2,500 mm. Thus, the whole region suffers from a large water deficit, the mean annual balance being about -1,800 mm in the south and -2,500 mm in the extreme north. The seasonality of the rainfall, its variability and unreliability as well as the great water deficit combine to augment the vulnerability of this region to the hazards of desertification.

Sudan is one of the countries that are seriously affected by desertification. Comparison of aerial photographs taken in 1975 with maps prepared in 1958 indicated that the desert boundary has shifted southward. According to Kassas (1971), the overall southward progression of the vegetational belt in Sudan has moved by 150 km in 20 years. The reduced production of dura, sesame, millet and gum arabic in Western Sudan, combined with overgrazing and the dominance of less palatable grasses, as well as the extinction of wildlife species in areas where they once had flourished, are all strong indications of the seriousness of the problem.

In global terms, desertification is generally attributed to three causes, namely: (1) long-term climatic changes, (2) weather fluctuations, and (3) man's destructive use of the land.

In Sudan, however, there is insufficient evidence to substantiate the role of climatic changes and weather fluctuations in causing desertification. Regarding climatic change, there is no evidence of any major climatic change in this region for the last 100 years. There was a special study of the Sahelian drought that concluded there is no evidence of a continuous lessening in rainfall in this region during the periods for which records are available. According to the Sudan Soil Conservation Committee (1944) "there is very strong evidence that in Sudan that the climate of today with its normal variations has undergone no basic changes for better or worse since the close of the final major wet phase of Pleistocene times in which a Neolithic civilization flourished at least in central Sudan. The soil deterioration which is still occurring may safely be attributed to the work of man and to his domestic animals rather than to any change in basic climate." Weather fluctuations are characteristic of arid and semiarid regions. This is particularly true with rainfall. For Sudan, averages taken over long periods of time do not indicate any significant change in the amount of rainfall (Table 1).

Weather fluctuations in Sudan are regarded as accelerators rather than the main cause of desertification. As early as 1944, the role of man's activity in destroying the land was observed by the Sudan Soil Conservation Committee, which reported that "Town and village peripheries generally are deteriorating rapidly due to over-cultivation, over-grazing and cutting trees for wood or a combination of these factors." Recently, the role of man in destroying the ecosystem has been reemphasized (DECARP, 1976) and desertification in Sudan is recognized as a man-made phenomenon and it needs man's efforts to control it. The most destructive human activities leading to desertification are the following.

Irrational Cultivation

In many parts of Sudan cropping has been extended to marginal lands and areas of higher climatic risks. This is particularly true in areas where pressure on the land is increased through population growth or by the introduction of mechanized farming in marginal lands with fragile ecosystems. In such areas, when the vegetational covering is removed by cultivation in dry years, the soil is exposed to erosional hazards. The fine clay and silt particles are carried away as dust and the sand drifts into dunes. According to Menaching and Ibrahim (1977) the expansion of millet cultivation beyond its climatic borders is the major cause for desertification in Northern Darfour. Clearance of trees and eradication of vegetation to secure sufficient moisture for the millet crop has

TABLE 1. Average Annual Rainfall

Site	Rainfall, mm	
	1931 - 1960	1941 - 1970
El-Fasher	287	286
Khartoum	164	167
Port Sudan	110	111

TABLE 2. Estimate of Dry Forage Lost by Burning Annually

Province	Total Forage Production, ton	Total Forage Lost by Burning ton	Percent Loss
Kordafan	112,252,000	33,676,000	30
Darfur	218,712,000	65,614,000	30
Kassala	31,567,000	6,313,000	20
N. Province	--	--	--
Blue Nile	35,072,000	5,261,000	15
Khartoum	2,592,000	205,000	10
Upper Nile	67,046,000	23,466,000	35
Sahr Elgazal	204,096,000	81,638,000	40
Equatoria	119,168,000	89,626,000	45
Total	870,506,000	307,892,000	

exposed the topsoil to wind and water erosion, resulting in the removal of the top fertile soil. In the central rain lands of Sudan, which has great potential to become the granary of the Middle East, mechanized farming was introduced on a large scale. To open new areas, the land was cleared of all vegetation. Due to the lack of proper land management, e.g., shelter belts, rotation, fertilizer, drainage, etc., and due to the inefficiency of the "suitcase" farmers lacking vocational training in farm management, the soil was "mined," and put at the mercy of the wind and water, paving the way for the southward extension of the desert.

In the late sixties, mechanized farming was extended to western Sudan where the soil is lighter in texture, making the risks of erosion higher. These areas were traditionally under shifting cultivation, a practice involving a bush fallow of about 15 years. Signs of declining land productivity has already been observed. Sorghum yields were reduced by about 50% in five years, due to the activity of the wind and water erosion, which is enhanced by the removal of the natural vegetation cover (Simpson and Simpson, 1978).

Deforestation

In the past *Acacia senegal* gardens were growing extensively in a well-defined rotational system. The cycle started by burning the *Acacia* trees and clearing the land before being planted with millet, sorghum, sesame, and other crops for 4 to 10 years. The land was then left fallow in order to allow a regeneration of fertility. During the fallow period, the *Acacia* trees started to invade the area, and after 8 years could be tapped for gum arabic for 6 to 10 years.

When the trees started to die, they were burned and the cycle starts again. Under the pressure of a growing population, the cultivation period is extended by several years and the soil becomes too impoverished to recover. Overgrazing during the fallow period prevents the regeneration of the *Acacia* seedlings, while the mature trees are used as fuelwood. With time, the area became void of *Acacia* trees and Sudan lost an important cash crop.

Besides the *Acacia* trees, other trees and shrubs are uprooted by nomads and villagers for fuel and for building. According to a recent study, (DECARP, 1976) more than 548 million trees and shrubs are uprooted annually by nomads.

In the absence of forests, wind erosion becomes rampant, the microclimate becomes invariably more arid, the dry season becomes more accentuated and there is a probable reduction in the total precipitation.

Overgrazing

Overgrazing has become very acute in Sudan in the past few decades. Previously, the nomads maintained the delicate balance between the number of their animals and the carrying capacity of the pasture. Recently, this balance has been disturbed due to the expansion in dry farming in traditional grazing lands, and an increase in animal population due to the recent control of epidemic diseases and the provision of more watering points in areas traditionally grazed only in years of abundant rainfall. In the Kardofan region, livestock populations have increased nearly sixfold since 1947. Due to the breaking down of the dynamic equilibrium that existed between livestock and the natural pastures, the area became devegetated and the erosional agents became more active. This resulted in reduction of the livestock carrying capacity of the natural pastures.

Besides the degradation of the natural vegetational cover, over-grazing resulted in lowering the quality of the pasture where most of the palatable species have been replaced by less desirable ones. According to Lebon (1965), some of the more palatable plants such as *Aristida plumosa* have been replaced by the less palatable *Cyperus conglomeratus* near water courses in Kardofan.

Fires

Burning vegetation is a common nomadic practice in Sudan in order to encourage growth of young grasses and hence to improve the plants palatability, also killing ticks and other pests. Sometimes, the burning of the vegetation is accidental and usually this is associated with the clearing of land for cultivation by fire, which can easily get out of control. According to DECARP, fires annually destroy about 35% of the forage resource (Table 2).

Sometimes the heat generated by the fires becomes so great that it destroys the seeds, thus reducing the regenerating capacity of even the most resistant species.

EFFECTS OF DESERTIFICATION

Sand drift and dunes formation

Recent surveys have revealed that most of the rain-fed land between 17° and 15°N was lost to desertification due to the movement of sand from the Libyan Desert (FAO mission report, 1976). Sand dunes are engulfing productive agricultural lands and human settlements along the Nile as well. The entire length of the Nile between Dongola and Kariema is threatened by sand drift and sand dunes encroachment.

Reduced Food Production

One of the most serious consequences of desertification is the decline in food production in the Kardofan region, production of dura (*Sorghum vulgare*), the main staple food in many parts of Sudan, has declined from 1.02 tons/hectare in 1961 to 0.46 tons/hectare in 1973. Maize and dukhan (millet) yields have declined from 0.8 ton/hectare and 1.3 tons/hectare to 0.36 and 0.29 ton/hectare respectively during the same period.

Reduced Gum Arabic Production

Gum arabic, which is extracted from *Acacia senegal*, was second only to cotton as a main source of foreign currency. Due to the extermination of huge members of the *Acacia* trees, as mentioned earlier, desertification has not only affected the country's foreign exchange balance, but it has also reduced the world's supply, since Sudan was its major producer.

Wildlife

Desertification has also affected wildlife habitats. Many of the once abundant wildlife species have vanished and those remaining are endangered. In Dinder National Park, the number of ungulate species is declining rapidly due to competition for grazing with livestock. Already the gazelle (*Gazella someringi*), which was abundant until 1966, has completely disappeared from this area. Also, the populations of tiang are among some of the species in that area have already declined. In the area of Wadi Howar, the addax has also not been seen for the last 15 years, nor has the oryx been seen since 1973. Both of these species had often been seen in that area before.

REFERENCES

- DECARP (1976) Sudan's Desert Encroachment Control and Rehabilitation Program.
- Kassas, M. (1971) *Biolog. Conserv.* 4, 1-12.
- Lebon, J.H.G. (1965) *Land Use in Sudan*. London.
- Mahdi, A.E. (1975) *The Rains of the Sudan*. Khartoum Univ. Press.
- Menching, H. and Ibrahim, F. (1977) *Applied Sci. and Devel.*, 10, Inst. of Sci. Coop., FRG.
- Simpson, L.G. and Simpson, M.C. (1978) *Alternative Strategies for Agricultural Development in the Central Rainlands of the Sudan*. Rural Develop. Studies, Univ. of Leeds.
- Sudan Soil Conservation Committee Report (1944).

DESERTIFICATION PROBLEMS IN A HUMID REGION

By

Jens Tyge Møller

ABSTRACT

Many parts of Denmark are characterized by very sandy soils, but even where the climate is rather humid, increased destruction of vegetation has led to sand drift, resulting in large areas marked with blown sand. Agricultural methods causing soils to be devoid of vegetation for most of the year have resulted in sand drifting during the spring. Aeolian transport and countermeasures against sand drift will be discussed here.

DISCUSSION

The Danish landscape is characterized by glacial and fluvioglacial deposits (Figure 1). The glacial deposits are sandy or clayish, spread throughout the country. However, clayish soils are mainly found in eastern Jutland and the southern parts of the islands. The fluvial deposits are to a great extent forming sandy plains, especially in western Jutland. South of Esbjerg, the western parts of the sand plains are covered with salt marsh. In northern Jutland, large areas of uplifted seafloor are found, mainly covered with sandy deposits. Along the North Sea coast, the uplifted seafloors are overlain by dunes. As a whole, the western regions in Denmark are strongly marked by blown sand. Only a few regions in Jutland, facing the North Sea have glacial deposits close to the sea.

Figure 1

The main features of distribution of surface sediments in Denmark. 1, uplifted seafloors, lakes, and salt marsh. 2, glacial deposits. 3, fluvial deposits. 4, dunes.

Almost every year sand drifting occurs in the spring. In very dry periods, and during long storms, the sand drift may increase with disastrous results, such as destruction of crops, and undesirable deposition of sand. However, the sand can be removed and the fields resown. The hidden danger in such a process is the removal of the fine sediments such as organic matter and clay, leaving the fields as well-sorted sandy plains. Very little research has been done on the fine materials, but evidently it is transported long distances. After a heavy storm, organic matter from Jutland was found on a lightship 100 km to the east.

Average temperatures characteristic of the Danish climate are about 0°C in January, and 16 to 17°C in July. The precipitation varies from 500 to 800 mm/year. Along the coasts facing the North Sea, the annual precipitation is approximately 700 mm. Despite great variations from one year to the next, the climate is by no means dry. However, several cases of desertification have been described in the history of Denmark. These incidences of sand drift covering large areas cannot be explained satisfactorily, but damages caused by too many grazing cattle, inappropriate cultivation of farming land, and cutting of forest for timber and fuel have been of great importance. The establishment of colonies of small-holders parcels in the 20th century was of great importance too, because marginal, sandy soils were reclaimed. Also, holiday traffic in the dunes has caused extensive damages, especially since World War II.

The importance of sand drift in Denmark has been reflected in the countermeasures taken during historical times. In 1539, an act provided for the protection of the vegetation against destruction that was caused by the use of the plants for fodder and roof thatching. In 1824, experiments started with plantings and wind fences. According to statutory instruments issued in 1792, the dune landscapes are managed by a special authority. In the case of uncontrolled sand drift along the coasts, the government is taking more serious measures to combat the desertification such as planting, fencing, and establishing wind fences.

Since sandy soils are widespread in Denmark, great parts of the farming land are sandy. Most of the crops have to be grown in the summer due to the rather cold winters, and so fields are not covered with vegetation during the winter or spring. Between the time of the melting of the snow to when the crops are covering the surface, the sandy soils are subject to aeolian erosion and transport, when most of the fine materials are completely removed. The sand moves across the fields and is deposited in fences, grasslands, and woods, causing the seeds and artificial fertilizers to be removed from the fields.

Countermeasures against aeolian erosion in farming regions are not taken by the government but rather by agricultural associations, and the society for moorland reclamation. The sand transport is mostly fought by means of wind fences, but some farmers try to maintain a permanent vegetation covering in the shape of crops, e.g. rye and wheat, which are able to survive the winter. The wind fences are mainly hedgerows. In recent years, great efforts have been taken to renew old hedgerows and to revive the tradition of planting hedgerows, both of which are being supported by the government. Still, many farmers are displeased with the planting of hedgerows because they occupy too much land. On the whole, one-third of Denmark is highly marked by aeolian erosion and by the countermeasures taken against this form of soil erosion.

Amongst the professional farmers, there is limited understanding of aeolian erosion or the risk of desertification in Denmark. Tourists are literally obstructing countermeasures taken against the sand drift. They remove fences and plantations of grass in the dunes. Owners of holiday cabins in the dune areas are even excavating dunes in shortsighted attempts to save their houses.

Theoretical research on aeolian transport and erosion was first introduced in Denmark by Jensen (1954) and Kuhlman (1956), both inspired by the work of R. A. Bagnold (1941). Even if the importance of sand drift control was recognized very early in Denmark, giving rise to practical countermeasures, only small attention was paid to the theoretical background. Since hedgerows obviously led to an increase of roughness of the landscape and provided protection against soil erosion, most of the research was originally directed to this subject. The results of this early research are used extensively in the protection of sandy regions.

However, a requisite condition for progress in the prevention of sand drift is a deeper knowledge of the physical background of the sand movement.

As a means of introduction, some substantial definitions are mentioned here. As a consequence of the shelter effect, the wind velocity varies with the distance from the solid surface, which in this case is the soil surface. Owing to velocity variations from zero on the surface, to the wind velocity at a height in which the wind is not affected by the shelter, the ratio of wind velocity of one height compared to another is of the greatest importance. Consequently, the wind profile is most advantageously described against a scale of logarithmic height (Figure 2).

The dotted lines indicate profiles above a fixed sand surface. If v_z is the wind velocity at any height z ,

$$v_z = 5.76V_* \log \frac{z}{k}, \quad V_* = \frac{\sqrt{\tau}}{\rho}$$

Figure 2

Schematic view of wind velocity distribution (Bagnold, 1941). The numbers and functions are fictive. The slope of gradient is determined as the abscissa divided by the ordinate. If the increase of ordinate is 10 times, $V_* =$ the increase of abscissa divided by 5.76.

Figure 3

Impact thresholds, wind velocity profiles indicating the lower limit of wind velocity being able to start the movement of quartz grains of different size diameters in millimeters (Kuhlman, 1960).

k is the roughness, approximately 1/30 of the surface irregularity. If the sand is moving

$$v_z = 5.76 V_*' \log \frac{z}{k} + V_t, \quad V_*' = \frac{\sqrt{\tau}}{\rho}$$

ρ is the density of air, the drag on a fixed surface, and a surface with moving sand, respectively, and V_t is the threshold velocity (Bagnold, 1941). It has to be stressed that V_* and V_*' are not proper velocities, but velocity gradients; the increase in velocity in relation to height. In the equations, V_*' , V_* , V_t , k and k' can easily be determined graphically, if v_z is measured in two points for each profile. Wind profiles above a rough surface clearly showed a trend towards intersection close to a point, in Figure 2, 0' and 0. The profile with the lowest value of V_* indicates the velocity profile at which the grain movement starts.

The bundle of profiles in Figure 2 is not to be mistaken for the impact thresholds of Figures 3 and 4, indicating typical impact thresholds for grains of quartz and organic matter. At decreasing grain diameters below 0.01 mm, increasing values of V_* are needed to start the grain movement, because cohesive forces grow more and more important. Figures 3 and 4 show some examples of impact thresholds for grains of quartz and organic matter. Due to the varying conditions of roughness, shelter, and wind velocity, it is very difficult to measure the impact threshold. Above all, it is necessary to realize that the normal wind statistics from a meteorological service cannot be used for estimates of the impact threshold. The wind velocity has to be measured directly in the area in question, because even the variation in grain size and grain movement may change the impact threshold.

Topographic variations will of course be of primary importance. On the other hand, it can be seen from the figures that a surface roughness due to coarse sediments such as pebbles and cobbles evenly spread on or near the surface sediments will very soon prevent any erosion of the sediments at all. The larger the roughness of the surface, the lower the wind speed near the surface (Kuhlman, 1960). An adequate roughness may prevent erosion, but not transit of sediments in transport.

Wind profiles indicating the minimum velocity required for maintenance of creep and saltation appear in Figure 5. Comparing this figure with the impact thresholds, (Figures 3 and 4), it can be seen that if a $V_*' = 35$ is required to start erosion of a 0.6 mm quartz grain, only a $V_*' = 30$ is necessary to maintain the movement.

Figure 6 shows some selected wind velocity profiles indicating the smallest values required for suspended transport. For grains below 0.1 mm suspension is the only form of

Figure 4

Impact thresholds, wind velocity profiles indicating the lower limit of wind velocity being able to start the movement of grains, organic matter, of different sizes indicated in millimeters (Kuhlman, 1960).

Figure 5

Wind profiles indicating the minimum velocity being able to maintain saltation and creep of quartz grains (Kuhlman, 1960). d, diameter. c, creep. s, saltation.

Figure 6
Wind velocity profiles indicating the minimum velocity at which quartz grains can be transported in suspension (Kuhlman, 1960).

Figure 7
Wind transport of quartz grains. Critical wind velocities near the soil surface (Kuhlman, 1960). A, lower velocity limit of suspended transport. B, impact threshold. C, lower velocity limit of saltation. D, lower velocity limit of surface creep.

transport. It is clearly seen that the suspension threshold increases proportionately with the diameters. Once in suspension, it is almost impossible to get a grain settled if the wind does not fade away or if the physical qualities change leading to the movement of larger grains. Normally the suspended matter is lost for the region.

Figure 7 illustrates the critical wind velocities near the surface for the four types of transportation. A, suspended load. B, impact threshold. C, lower limit for saltation. D, lower limit for surface creep. The close connection between the two types of transportation, saltation and surface creep, must draw attention to the sorting process. The sample should contain some grains able to saltate and some grains small enough to be moved by the impact of the saltating grains.

One of the less probable consequences is the termination of movement, due to too great of a scattering of the size distribution and to the lack of creeping grains. If the distribution is very narrow, the lack of surface creep may lead to a lack of ripples.

Sand drift cannot be prevented totally, but it is possible to significantly hinder the movement either by means of an increase in surface roughness, in the shape of coarser sediments, or else by means of wind fences. A badly sorted sediment influenced by the wind will change its sorting. As time goes on, the finer parts will disappear at a given, constant wind velocity, leaving the coarser grains behind as a natural pavement. The existence of the coarser grain increases the amount of roughness, and at last the erosion will stop, leaving the original sediment protected by a layer of coarse grains. This surface will remain stable until the advent of a stronger wind. At last, the surface will attain a stability that will allow vegetation to move in if permitted by climatological conditions. If the pattern of a natural pavement is disturbed by traffic or cultivation, the erosion will recommence. Consequently, this kind of stabilization has only a small effect in agricultural areas. However, if the wind has a predominant direction, surface structures perpendicular to the wind direction such as furrows left by agricultural implements, may create adequate roughness, protecting the fields against wind erosion.

Although wind fences are a more advanced remedy against sand drift, they are very expensive if the climate does not permit the growth of hedgerows. The Danish dune areas along the coasts often nourish sand drifts as well as incurring damages upon the dune landscape itself. Due to the moving sand and the salt in the sea water, trees cannot grow close to the sea. Tourists' wear and tear in the dunes are an important problem, because the traffic that they create impedes any countermeasures against sand drift. To prevent traffic and to increase the roughness at the same time, pine branches are

stuck in the sand. Further, a special grass (*Ammophila arenaria*) is planted. This grass is stimulated by the drifting sand and the air content of salt, but cannot be used for reclamation close to the sea due to the waves during storms. In some places, wind fences are constructed as wooden grills. To bind the sandy surface, rye (*Secale cereale*) is sown together with artificial fertilizer and seeds of a grass that easily grows in sand. The rye will germinate very fast, binding the surface and protecting the seedlings of the permanent grass until a permanent vegetation cover could form.

Some important results from the research on shelter effects should also be mentioned. A wind fence in the shape of a wall without openings will give rise to eddies on the lee side of the fence. Furthermore, the shelter effect will only cover a distance of approximately 20 times the height of the fence. Eddies will occur even if the openings cover up to 35% of the surface. The maximum shelter effect occurs at fences with 50% holes in the surface area. In this case, the shelter effect covers an area up to 30 times the height of the fence. However, the openings must be spread evenly on the surface of the shelter (Jensen, 1954).

Figures 8 and 9 illustrate the shelter effect. The only differences are the values of roughness corresponding to a surface with crops (Figure 9) and without crops (Figure 8). The percentage of openings in the shelter, the normal wind velocities, and the required shelter effect have to be balanced, if a system of effective wind fences is to be planned. The greatest shelter is needed when the surface is without crops and the occurrence of turbulence is low (Figure 9). If the crops have grown so much that the turbulence is affected, the need for shelter is less. The openings in a shelter, screens or hedgerows, must be spread uniformly. Openings close to the surface will give rise to a greatly increasing turbulence.

REFERENCES

- Bagnold, R. A. (1941) *The Physics of Blown Sand and Desert Dunes*. London, 265 p.
- Kuhlman, H. (1960) Microenvironments in a Danish Dune Area, Rabjerg Mile. *Medd. Dansk Geol. Foren.* 14, 253-258, København.
- Kuhlman, H. (1960) Den potentielle jordfygning på danske marker. *Geogr. Tidsskr.* 59, København.

Figure 8

The shelter effect, s , of a wooden screen with openings covering 48% of the screen area. The abscissa is the distance to the screen with the screen height, h , as unit. The wind velocity was measured at height z . $z_0 = 18 \times 10^{-3} h$, where z_0 is the roughness expressed as the height in which the wind velocity is zero (Jensen, 1954).

Figure 9

As in Figure 8, but in this case the roughness is less, $z_0 = 2.2 \times 10^{-4} h$. In consequence of less turbulence, the shelter has a greater effect (Jensen, 1954).

THE FORMATION AND MOTION OF DUNES AND SAND SEAS

By

Farouk El-Baz

SUMMARY

Particulate material in the desert originates from the disintegration of rock by both mechanical and chemical weathering. This disintegration exposes the particles to the agents of erosion, which in turn cause more loosening of particles from the rock.

Wind is the major agent of erosion and deposition in the desert. The wind hurls the finest particles into the atmosphere as dust. The largest particles remain behind and armor the surface, protecting it from further erosion by the wind. Medium-sized particles are winnowed away and deposited downwind as sand sheets and sand dunes. This paper summarizes dune shapes and accumulations, their rates of movement and methods of stabilization.

Sand accumulations occur in a large number of forms, which makes them hard to classify. However, generally speaking, sand-sized grains accumulate in the form of sand sheets with rippled, undulated surfaces or as free dunes and obstacle-related dunes. Free dunes, which are by far the most abundant, can be isolated simple forms, coalesced compound groups, or complex accumulations. The basic geometric shapes of dunes are linear (longitudinal and seif dunes), crescentic (barchan and parabolic dunes), and domical (including dome, pyramid, and star dunes). All types of dunes may be active, semifixed, or fixed by vegetation. Dune shapes vary considerably in different areas and often in the same locality; little is known about the causes of such variation.

Figure 1
Sketch of the physical and chemical processes that result in the formation of sand-sized particulate material from solid rock.

A few dunes may form a bundle, and a larger number is commonly referred to as a dune field, which is here arbitrarily limited to less than 2,000 km². Larger accumulations of sand dunes are known as sand seas because of the repetitive nature of dune forms, much like waves on the sea surface. Detailed knowledge of the texture and morphology of these sand seas was made possible through the interpretation of photographs and images taken from space. For example, data from Meteosat allow the recognition of general sand movement patterns in all of the Sahara. In addition, Landsat images and photographs taken on manned space missions allow the mapping of dune accumulations such as the Great Sand Sea in the Western Desert of Egypt or the Taklamakan Desert in northwestern China.

The motion of dunes is fueled by the wind and depends on its direction and velocity. The rate of advance also depends on the size of the dune. For example, crescentic dunes in the Kharga depression in the central part of the Western Desert of Egypt move at the varying rates of 20 to 100 meters per year; the smallest dunes are the fastest moving.

The motion of dunes may have devastating effects on human settlements in the desert. Because of this, numerous attempts have been made in various deserts to halt the advance of sand dunes. Stabilization methods include the utilization of vegetation by seeding the dunes and by planting trees on the wind or lee sides. Another method, which has been employed in Egypt, India, and China, uses hay fences to limit the saltation of sand grains. Still other methods include the spraying of the sand surface with crude oil, as has been successfully tried in Iran and Saudi Arabia, and with chemicals that bind the sand grains on the uppermost surfaces of the dunes.

INTRODUCTION

Dunes are accumulations of windblown, sand-sized particles in the form of mounds, ridges, or hills. They occur in arid, temperate, or glacial environments where the requirements of availability of particulate matter and strong winds are met. Once formed, a dune becomes an obstacle to windblown sand and the lodgment of additional grains causes the dune to grow.

Sand-sized particles of snow accumulate in dunes in the frozen tundra. However, the majority of sand-sized particles are generated from solid rock by weathering processes, both chemical and mechanical (Figure 1). With the passage of time the exposed bedrock of the Earth's crust wastes away under the onslaught of changing temperatures and pressures. Frost shatters seemingly permanent rock; salt crystals scale it; chemicals decay it. This weathering sets the stage for the agents of erosion. These agents--ice, water and wind--all play a role in generating particles for the formation of dunes. This paper will summarize, in a simplified way, our present knowledge

Figure 2

The world's major deserts, as defined by the 300-mm isohyet. These include the tropical deserts, temperate deserts, and tundra areas (after Allen, 1970).

Figure 3

World map of the principal fixed and active dune areas (after Cooke and Warren, 1973).

about the mode of formation and motion of sand dunes and sand seas.

The environments of wind action that usually results in the formation of dunes include (Allen, 1970):

1. Desert environments, where the wind picks up particles from the uplands, alluvial fans, and playa deposits.
2. Coastal zones, where the wind transports particles that were deposited or precipitated by the sea along beaches.
3. Glacial outwash regions, where particles are winnowed away by the wind from the dry courses of braided outwash streams.
4. Temperate areas devoid of vegetation, where deflation by the wind generate sand-sized particles.

The action of wind on the land is closely related to the amount of vegetation cover. Therefore, the wind plays more of a significant role as the amount of surface or near-surface water in the soil decreases. The land areas of the tropics and subtropics that are considered dry are those defined by the 300 mm isohyet (the line connecting places having equal depth of rainfall over a year). This line encloses all of the dry regions of the world including the hot deserts (Figure 2). However, the action of wind is not limited to these regions; in temperate areas with relatively high precipitation, the wind is an effective force where plant cover is removed or cannot gain hold.

Contrary to widespread belief, sand does not cover much of the hot desert surfaces. It is estimated that only one-fourth of the area of these deserts is covered by sand; the rest is bare rock and rock rubble. It is further estimated that one-seventh of the desert surfaces are covered by dunes; the rest of the sand cover is in the form of thin blankets, and less commonly thick deposits. Furthermore, although most of the sand fields are active, some are fixed by vegetation (Figure 3).

The largest accumulation of dune fields is by far in the North African Sahara (Figure 4). The southern boundary of this vast desert is known as the Sahel, from the Arabic word meaning the shoreline. Migrations of this boundary to the north and south (Figure 5) are triggered by climatic changes, and to some extent by human activities (Mainguet et al., 1979).

WIND EROSION AND PARTICLE TRANSPORT

Because running water no longer plays a significant role in the transportation and deposition of sediment in the desert, wind is the main force of erosion. The wind carries

Figure 4

Saharan dune fields or ergs tend to occur near the centers of basins of interior drainage beyond the termini of the present river systems, suggesting an alluvial origin for the sands. Main dune trends shown (after Peel, 1941, with tracks of sand movement from Mainguet, 1975; after Mabbutt, 1977).

Figure 5

Paleoclimatic migration of the limits of the arid Sahel. Upper and lower dashed lines mark the zone of migration and represent the 150-mm and 1,000-mm isohyets respectively. Upper and lower solid lines represent the limits of active and fixed dunes respectively (after Mainguet and Callot, 1978).

enormous quantities of debris, which are deposited where its path is disturbed by topographic variations.

Much of our understanding of desert landforms came from research that just preceded World War II, since part of the war was conducted in North Africa, particularly in Libya and Egypt. Basic research on dune classification and sand movement by wind has been carried out in the Western Desert of Egypt, where the free interplay of sand and wind has been allowed to continue for a vast period of time (Bagnold, 1933). Geomorphic descriptions followed by laboratory experiments led to Bagnold's (1941) classic treatise on the physics of sand movement by wind, based largely on observations made in this desert.

During the past two decades, research on the landforms of deserts and arid lands was intensified, in part due to the scientific interest in the effects of prolonged droughts in the African Sahel. As a result of this emphasis, we now have an impressive collection of books that treat the geomorphology of arid lands (Butzer and Hansen, 1968; Glennie, 1970; Cooke and Warren, 1973; Doehring, 1977; Mabbutt, 1977; and McKee, 1979).

As shown by such recent studies, prolonged wind erosion creates isolated hills (Figure 6a); cliff retreat does not have to result from water erosion. Also corrasion features, shaped solely by the wind can be 100 km in length (Mainguet and Callot, 1978). Yardangs, the smaller wind-sculpted features that resemble inverted boat hulls, are being described from numerous new localities (Figure 6b), indicating new interest in wind-formed features. However, the role of vorticity in developing lineations by wind erosion was only recently established by Whitney (1978). This role eliminates the necessity for shifts in the orientation of ventifacts relative to the surface winds as postulated by Sharp (1949) among others.

A striking characteristic of the desert environment is that its wind deposits are not mixed up and scattered at random. Rather, they are discretely zoned in response to the capacity of the wind to sort out and segregate particles by grain size. The finest particles (clay and silt particles, up to 0.05 mm in size) are winnowed out wherever they are exposed and whirled away into the atmosphere as dust, settling out of suspension beyond the zones of high wind energy (Breed et al., 1980). Saharan dust storms, for example, carry such fine dust particles concentrated at altitudes of 1.5 km and 3.7 km in the atmosphere. It was estimated that 25 to 37 million tons of dust are transported through the 60° west longitude each year, which is equivalent to the present rate of pelagic sedimentation for the entire north equatorial region of the Atlantic Ocean (Prospero and Carson, 1972).

Figure 6a

Aerial view of isolated hills south of the northern escarpment of the Farafra depression, Western Desert of Egypt.

Figure 6b

Aerial view of a yardang field northeast of the Kharga depression, Western Desert of Egypt. Nearly straight line in lower right is paved road.

Similarly, in temperate regions, the wind picks up particles of glacial silt and clay and deposits them as loess. A buff-colored deposit of loess up to 100-meters-thick scoured from the Mongolian Plateau covers most of northern China. This deposit is the largest loess accumulation on earth.

In contrast to the suspension in the atmosphere of the finest particles, fine-to-medium sand saltates readily in the ubiquitously brisk desert winds. By the saltation process, the wind segregates particles of 0.05 to 0.5 mm sizes from the clays, silts and gravels and encourages them into dunes. The movement of sand by saltation depends on the nature of the surface. The sand particles jump to higher levels if they saltate on gravel than on a sandy surface. One measurement showed that 90% of sand moved below 87 cm on a gravel surface, whereas the same amount moved below 31 cm on a sandy surface (Bradshaw et al., 1978, p. 217).

Particles too large (0.5 to 2 mm in diameter) to be lifted off the surface by the bombardment of saltating grains may gradually and erratically move or roll along the surface. With high winds, the whole surface covered by such grains appears to be creeping slowly along the wind direction.

This process of segregation of particles by the wind into varying sizes, usually results in the formation of vast flat plains that are veneered with a surface of granules or pebbles, commonly well-sorted and usually one-grain-thick. This lag forms an armor that is apparently in equilibrium with the strongest winds. The process of surface creep, by which the wind distributes the coarse materials seems imperceptible except under rare high-velocity wind conditions. Removal of the coarse lag results in immediate deflation of the suddenly unprotected, underlying silts and sands, which produces lasting scars (Breed et al., 1980).

CLASSIFICATION OF DUNES

Sand Deposits

As stated above, sand may be deposited in the form of dunes or flat blankets called sand sheets. These sheets may cover enormous areas, for example, the Selima Sand Sheet, first described by Bagnold (1933) in the southern part of the Western Desert of Egypt, is over 40,000 km². Much of this sheet is surfaced not by sand-sized particles but by coarser granules.

More commonly, however, sand sheets occur as long and narrow sand deposits sometimes without any slip faces, referred to here as streaks. In the Western Desert of Egypt many such streaks occur. In places, these streaks occur in straight lines in the lee of topographic obstacles to the pre-dominant northerly winds, and in others they create

Figure 7
Truncated granule ripples in the Wadi Wassa in the southern part of the Gilf Kebir plateau, southwestern Egypt.

streamlined patterns similar to those in the lee of craters and knobs on Mars (El-Baz and Maxwell, 1979).

Where patches of fine sand occur without the stabilizing larger grains, the surface becomes rippled with undulations that are transverse to the wind. The wavelength of sand ripples is commonly from 10 cm to 50 cm. The ripple index, which is the ratio between the wavelength and the amplitude, is generally about 20 (Sharp, 1963). Ripples may also grow in sands with an armor of coarser granules, up to 4 mm in diameter, with wavelengths of 1 or 2 meters. These are called granule ripples, which may be truncated by very strong winds and thus create a very bumpy surface (Figure 7).

Dune Accumulations

Accumulations of sand take numerous forms, which may be simple, compound, or complex as defined below:

- Simple dunes are those where the individuality of each dune is clear and separable.
- Compound dunes are those in which two or more of the same type coalesce or combine by overlapping each other.
- Complex dunes are those that result from the combination of two different types of shapes.

Dune Shapes

Dunes occur in a very large variety of shapes. Numerous terms have been employed to classify dune shapes. In many cases terms used by local desert dwellers are used to describe the different types. However, this often results in referring to the same type by various names in different localities. In other classification schemes, a genetic connotation is implied. For example, the term "transverse dune" indicates formation normal to the wind direction. To provide a consistent and purely morphologic classification (and at the risk of oversimplification) dune shapes are here divided into three types, linear, crescentic, and domical.

- Linear dunes are sand accumulations in the form of a long ridge. They are known as "seif" dunes in parts of the Arabian Peninsula because of their likeness to the sword, specifically the curved scimitar. In many other deserts these have been described as longitudinal dunes and in Egypt as Ghurud. Such ridges are usually formed where there is a large supply of sand and a persistent strong wind, such as in the Western Desert of Egypt (Figures 8 and 9). The winds that are necessary to form such dunes are not fully understood. However, it is

Figure 8

Mosaic of Landsat images of Egypt showing the distribution of dunes, dune bundles, and the Great Sand Sea (left center) in the Western Desert. Dark areas in the upper left and near center are depressions that enclose oases. Dark areas in the lower left are topographic prominences.

believed by some workers that linear dunes probably become aligned by winds from two directions. In this case, the dune axes would parallel the resultant of the two wind directions. This may account for the occasional formation of slip faces on either side of the dune. In plan view, such as in Earth-orbital photographs, linear dunes may appear as straight lines, in echelon tracks, thin arrowhead accumulations, slightly curved lines, or sinuous ridges.

- Crescentic dunes are the best-known dunes, with the crescent-shaped or barchan dunes having arms pointing downwind (Figure 10) and U-shaped or parabolic dunes with arms pointing upwind. They range from simple shapes to coalesced crescents to crescentic or barchanoid ridges (Figure 11). These dunes occur where there is less of a sand supply and more of an energetic wind than in the case of linear dunes. For example, in the Western Desert of Egypt the Ghard Abu Muharik extends in a S-SE direction from the Bahariya Oases to the Kharga depression 300 km away. As this dune reaches the edge of the Kharga depression, its sand descends the escarpment and is redistributed into fields of crescentic dunes and crescentic ridges. The transition in dune type is apparently due both to the effect of the abrupt change in slope on the oncoming sand-moving winds and to the accompanying increase in wind velocity. Vegetation is virtually absent except in oases and villages in the center of the depression, and thus does not affect the formation of these dunes (El-Baz et al., 1979).
- Domical dunes are mound-like sand accumulations in the form of a dome, pyramid, or star: (1) Dome dunes are gently sloping mounds of circular or elliptical shape that normally show no slip faces. Some dome dunes show a small slip face suggesting the start of a crescent-shaped dune. It is generally believed that dome dunes occur in a strong-wind regime. In support of this, dome dunes occur at the southern terminae of linear dunes in the Western Desert of Egypt; (2) Pyramid dunes are faceted and peaked mounds that result from wind that comes from two different directions. Many such forms occur in the Taklamakan Desert in northwestern China; and (3) Star dunes are those peaked shapes with multiple slip faces that result from winds from several directions. They usually have three or more arms that extend in a radial fashion. Star dunes usually grow vertically, rather than migrate laterally (McKee, 1979, p. 12).

Figure 9

A linear dune, approximately 2 km long in the lee of a hill at the southeastern tip of the Gilf Kebir plateau, southeastern Egypt. Slipfaces with "sand slides" lie at a steep, nearly 40° angle.

Figure 10a.

Aerial view of a bundle of crescent-shaped, compound dunes within the Kharga depression, Western Desert of Egypt.

Figure 10b.
Crescent-shaped (barchan) dunes: (A) plan view; (B) section showing how the dune slipface develops; (C) a typical belt; and (D) one suggestion as to the formation of linear (longitudinal, or seif) dunes (after Bagnold, (1941)).

Figure 11.

Crescent-shaped (barchan) and lee-dune patterns: (a) from near Bilma, Niger; (b, c, and e) the development of different oblique elements; (d) oblique elements aligned with lee dunes (from Cooke and Warren, 1973).

Dune Activity

Dunes may be free, as described above, or obstacle-related. All types of free dunes may be active, semifixed, or fixed by vegetation. The degree of dune fixation depends on the amount of vegetation cover.

Obstacle Deposits

Topographic rises in desert surfaces nearly always affect the wind pattern and invariably cause the deposition of sand in the form of dunes. A few large topographic prominences such as mountains cause major variations in the wind flow. These variations result in the formation of different types of dunes windward, leeward, and on either side of a topographic barrier (see, e.g., Figure 12).

Small-scale topographic rises are just as effective in causing the accumulation of sand into numerous forms including (see Mabbutt, 1977):

- Sand shadows, which are elongate deposits in the lee of an obstacle, such as a rock or hill.
- Sand drifts, which are elliptical mounds less tapered than sand shadows, resulting from the deposition in gaps between obstacles or from vertical separation of flow in the lee of steep slopes.
- Lee dunes, which result from the buildup and downwind extension of sand shadows and sand drifts.
- Windward dunes, which result from the deposition of sand upwind from an obstacle, such as a hill.
- Echo dunes, which form some distance upwind from a major relief feature such as a scarp.
- Scarp dunes, which form when sand-laden winds encounter a hill or an escarpment and climb up the slope (climbing dunes), or when the sand-carrying winds deposit the sand as they descend the slopes of hills and plateaus (falling dune).
- U-shaped dunes, known commonly as parabolic dunes because their shape resembles a parabola, which develop from partial stabilization by vegetation and/or moisture. While the arms of the U-shape remain anchored, the nose migrates downwind.
- Circular dunes, known also as blowout dunes, which are anchored by vegetation and may have slipfaces sloping in one or many directions depending on what parts of their rims are not stabilized and therefore free to move.

Figure 12.

Theoretical case based on the application of the Fachi-Bilma (Chad-Niger) setting to model the effects of a major obstacle (after Mainguet and Callot, 1978:

(1) Echo dune; (2) Region of transport windward of obstacle and start of outward flow; (3) Corrasion due to turbulent flow; (4) Central area of current interference; (5) Central area of linear edifices; (6) Central area of cross-current edifices; (7) First outer aureole of "silks"; and (8) Lopsided aureole of barchans.

Figure 13.

Bundles of linear dunes (or "clusters of silks") with closed SW ends and open NE sides. Note the asymmetry in the lower (downwind) parts (after Mainguet and Callot, 1978).

DUNE GROUPINGS

Sand dunes rarely occur as a single unit. More commonly groups of dunes occur together. Such groupings are here described as dune bundles, dune fields and sand seas.

Dune Bundles

These occur where a few dunes join together. In the case of linear dunes they usually form the shape of a "V". Such groupings were described as a bouquet by Mainguet and Callot (1978, p. 106) as illustrated in Figure 13. Many such groupings occur in the northern part of the Western Desert of Egypt (Figure 8) just south of the Qattara depression (Gifford et al., 1979).

Dune Fields

These are areas here arbitrarily limited to less than 2,000 km², where there is a large number of dunes. Most such fields can be studied in a single photograph from space. Examples include the Algodones field in southern California, the Turpan field in the Turpan depression of northwestern China, and the Vallecito field in southeastern San Juan, Argentina. The latter was photographed by the Apollo-Soyuz mission and was named after the nearest town (Warner and El-Baz, 1979). The field is approximately 1,300 km². In its southern part complex ridges display steep western flanks and gently sloping eastern flanks. These ridges are diagonally crossed by smaller linear dunes (Figure 14). In the central part, crescentic dunes with elongated horns abound; these horns may be affected by a seasonal cross wind known locally as "zonda" (Warner and El-Baz, 1979). The complex array of dune elements displayed by the Vallecito dunes cannot easily be explained. The pattern may be controlled by changes in climate, underlying and surrounding topography, wind regime, amount of sediment available, and grain-size characteristics.

Sand Seas

Larger accumulations of sand dunes are known as sand seas because of the repetitive nature of dune forms, much like waves on the sea surface. Detailed knowledge of the texture and morphology of these sand seas was made possible through the interpretation of photographs and images taken from space (McKee and Breed, 1974; McKee et al., 1977; Breed et al., 1979). For example, data from Meteosat allow the recognition of general sand movement patterns in all of the Sahara. In addition, Landsat images and photographs taken on manned space missions allow the mapping of dune accumulations such as the

Figure 14.
Sketch showing crescentic dune elements (bold lines) and
linear dune elements (fine lines) of the Vallecito dune field
in the Mote Desert of San Juan, Argentina (after Warner and
El-Baz, 1979).

Great Sand Sea in the Western Desert of Egypt (Figure 8) or the Taklamakan Desert in northwestern China (Breed et al., 1979).

MOVEMENT OF DUNES

Although soil erosion, salinization, and removal of the scant vegetation cover (for firewood and grazing) represent grave problems to inhabitants of the deserts, the most formidable problem is posed by the shifting sands. Such a problem exists in the Western Desert of Egypt, particularly in oases such as Kharga.

The Kharga depression is bounded on the north by a 200-m-high scarp capped by limestone of Eocene age, but it is open to the south and southwest. Oases and villages lie in its center, along a north-trending fault that provides access to groundwater. Downwind extensions of large compound linear dunes enter the depression from the limestone plateau to the north. The largest of these, Ghard Abu Muharik, enters the depression and breaks up into crescent-shaped dunes (Figure 10a). The simple crescent dunes (barchans) range in length from 30 to 650 m, in width from 25 to 540 m, and in height from 0.5 m to 25 m (Embabi, 1970-1971). Their rate of movement is roughly inversely proportional to their height. From field measurements, over one year, Embabi (1978) established that: (1) the rate of dune movement varied between 20 and 100 meters per year; (2) 94% of the variation in the rate of motion can be explained by variations in dune size, ground length of windward side, and mean slope of windward slope; and (3) the Kharga barchans move under the effective, 20 km/hr or more, northerly winds.

DUNE STABILIZATION

In dealing with shifting sand dunes, one can avoid them, try to halt their advance, or learn how to live with, and even utilize them. Avoidance of migrating dunes can be done by planning in advance. For example, if a dune belt is moving at the rate of 50 m per year, a settlement built at least 25 km farther downwind would exist without the danger of inundation for about 500 years. However, this is not always done and inhabitants of the desert have to worry about dune stabilization to protect their livelihoods.

Armoring of the sand dunes by pebbles and cobbles has been employed by numerous desert peoples, such as those of southern Peru (Lettau and Lettau, 1969). In fact, the Hopi Indians still use this method of stabilizing sand dunes on their reservation in northern Arizona, and thus are able to use the stabilized sand surfaces for agriculture.

The more-popular method of halting the advance of dunes is planting trees in their path. Particularly in the Sahara, oasis dwellers plant tamarisk trees for this purpose, although some farmers plant date palm trees instead. The trees in this case serve as a physical barrier downwind. They are effective until the dune becomes tall enough to engulf the trees. In Tunisia and Algeria, taller eucalyptus trees are being planted. The eucalyptus trees work as windbreakers rather than physical barriers. Their height (up to 100 m) disturbs the wind regime, forcing the wind to unload the sand or limit grain transport before it approaches the wall of trees.

In India's Rajasthan Desert, parallel rows of dry hay fences, only 30 cm high are used to stabilize shifting dunes. More effectively, in China, checkerboards of hay fences divide the sand surface into 1 m² areas. In this method the hay fences are not designed to stop the motion of sand, but allow it to move within the 1-m squares.

Other methods of halting the advance of dunes include spraying them with petroleum (as was successfully but expensively done in oil-rich countries like Iran and Saudi Arabia; Kerr and Nigra, 1952), or with chemicals such as Sand Seal, a nontoxic water emulsified blend of liquid polymers and silicate derivatives. Stabilizing dunes by seeding them with grass, as done in the American Southwest, would not work in some other deserts because of the scarcity of water and lack of humidity.

REFERENCES

- Allen, J.R.L. (1970) *Physical Processes of Sedimentation*. George Allen and Unwin, London.
- Bagnold, R.A. (1933) A Further Journey Through the Libyan Desert. *Geogr. Jour.*, v. 82, p. 103-129.
- Bagnold, R.A. (1941) *The Physics of Blown Sand and Desert Dunes*. Methuen, London.
- Bradshaw, M.J., Abbott, A.J. and Gelsthorpe, A.P. (1978) *The Earth's Changing Surface*. John Wiley and Sons, New York, New York.
- Breed, C.S., Fryberger, S.G., Andrews, S., McCauley, J., Lennartz, F., Gebel, D., and Horstman, K. (1979) Regional Studies of Sand Seas, Using Landsat (ERTS) Imagery. In: *A Study of Global Sand Seas*, U.S. Geol. Surv., Prof. Pap. 1052, p. 305-397.
- Breed, C.S., Embabi, N.S., El-Etr, H.A., and Grolier, M.J. (1980) Wind Deposits in the Western Desert. *Geogr. Jour.*, v. 146, p. 88-90.
- Butzer, K.W. and Hansen, C.L. (1968) *Desert and River in Nubia*. The Univ. of Wisconsin Press, Madison, Wisconsin.

- Cooke, R.V. and Warren, A. (1973) *Geomorphology in Deserts*. Univ. California Press, Berkeley, California.
- Doehring, D.O., ed. (1977) *Geomorphology in Arid Regions*. Donald O. Doehring, Fort Collins, Colorado.
- El-Baz, F. and Maxwell, T.A. (1979) Eolian Streaks in Southwestern Egypt and Similar Features on Mars. *Proc. Lunar Planet. Sci. Conf. 10th, Geochim. et Cosmochim. Acta Suppl.* 11, v. 3, p. 3017-3030.
- El-Baz, F., Breed, C.S., Grolier, M.J., and McCauley, J.F. (1979) Eolian Features in the Western Desert of Egypt and Some Applications to Mars. *J. Geophys. Res.*, v. 84, p. 8205-8221.
- Embabi, N.S. (1970-1971) Structures of Barchan Dunes of the Kharga Oasis Depression, the Western Desert of Egypt. *Soc. Geogr. Egypte Bull.*, 43-44, p. 7-71.
- Embabi, N.S. (1978) Barchan Dune Movement and Its Effect on Economic Development in the Kharga Depression. *Bull. Mid. East Res. Centre, Ain Shams University* (in Arabic).
- Gifford, A.W., Warner, D.M. and El-Baz, F. (1979) Orbital Observations of Sand Distribution in the Western Desert of Egypt. In: *Apollo-Soyuz Test Project Summary Science Report, V. II, NASA SP-412*, p. 219-236.
- Glennie, K.W. (1970) *Desert Sedimentary Environments*. Elsevier, New York, New York.
- Kerr, R.C. and Nigra, J.O. (1952) Eolian Sand Control. *Bull. Amer. Assoc. Petrol. Geol.*, v. 36, p. 1541-1573.
- Lettau, K. and Lettau, H. (1969) Bulk Transport of Sand by the Barchans of the Pampa La Joya in Southern Peru. *Z. Geomorph.*, v. 13, p. 182-195.
- Mabbutt, J.A. (1977) *Desert Landforms*. The MIT Press, Cambridge, Mass.
- Mainquet, M. (1975) Etude comparee des ergs, a l'echelle continentale (Sahara et deserts 'Australie'). *Bull. Ass. Geogr. fr.*, v. 52, p. 135-140.
- Mainquet, M. and Callot, Y. (1978) L'Erg de Fachi-Bilma, Chad-Niger. *Centre Nat. Res. Sci., Paris*.
- Mainquet, M., Cannon-Cossus, L. and Chemin, M.C. (1979) Degradation dans les regions centrales de la Republique du Niger. *Trav. Inst. Geogr. Reims*, n. 39-40, p. 61-73.
- McKee, E.D., ed. (1979) *A Study of Global Sand Seas*. *Geol. Surv. Prof. Pap.* 1052, U.S. Gov. Printing Office, Washington, D.C.
- McKee, E.D. and Breed, C.S. (1974) An Investigation of Major Sand Seas in Desert Areas Throughout the World. In: *Third Earth Resources Technology Satellite 1 Symposium. NASA SP-351*, p. 665-679.
- McKee, E.D., Breed, C.S. and Fryberger, S.G. (1977) Desert Sand Seas. In: *Skylab Explores the Earth, NASA SP-380*, p. 5-47.

- Peel, R.F. (1941) Denudational Landforms of the Central Libyan Desert. *J. Geomorph.*, v. 4, p. 3-23.
- Prospero, J.M. and Carson, T.N. (1972) Vertical and Aerial Distribution of Saharan Dust Over the Western Equatorial North Atlantic Ocean. *J. Geophys. Res.*, v. 77, p. 5255-5265.
- Sharp, R.P. (1949) Pleistocene Ventifacts East of the Big Horn Mountains, Wyoming. *J. Geol.*, v. 57, p. 175-195.
- Sharp, R.P. (1963) Wind Ripples. *J. Geol.*, v. 71, p. 617-636.
- Warner, D.M. and El-Baz, F. (1979) Monte Desert of San Juan, Argentina, as Photographed by ASTP. In: Apollo-Soyuz Test Project Summary Science Report, V. II, NASA SP-412, p. 301-318.
- Whitney, M.I. (1978) The Role of Vorticity in Developing Lineation by Wind Erosion. *Bull. Geol. Soc. America*, v. 89, p. 1-18.

TWO-DIMENSIONAL ANALYSIS OF DUNE PROFILE
AND THE EFFECT OF GRAIN SIZE ON SAND DUNE MORPHOLOGY

By

Haim Tsoar

ABSTRACT

Theoretical consideration shows that the steady-state profile of a dune subjected to an unidirectional wind is flat, low, and convex with the ratio of the dune height (h) to its upwind length (L) $h/L = 0.05$. This profile can be found in many barchans and transverse dunes. A steeper profile or a triangular-shaped profile cannot remain in a steady state under such conditions, as such a profile is only assured by bidirectional or multidirectional winds.

Grain size is another factor governing dune profile. A coarse-grained sand dune has a greater difference in threshold velocity between the crest area and its lower levels than a finer-grained sand dune of a similar profile. A coarse-grained sand dune will have much longer periods of erosion taking place only at or near the crest, thus lowering the dune profile. This explains why all dunes composed of bimodal coarse-mode grains such as whaleback and "zibar" are flat and low.

DISCUSSION

Wind regime and grain size are the factors that determine the morphology and dynamics of sand dunes in desert areas (Cooke and Warren, 1973). Various studies of desert sand dunes have been published in recent years but the effect of grain size on dune morphology and the interaction between this

Figure 1
 h/L (dune height to half its length) versus X/σ (mean size to its standard deviation (sorting)).

morphology and the modification of the wind speed on the dune have not been thoroughly studied.

As early as the end of the 19th century, sedimentologists noticed that the modal size of aeolian dune sand was between 0.125 and 0.250 mm (Udden, 1898). Latest analyses have verified this finding, but have also revealed a variety of sand size characteristics in various parts of sand dune deserts (Folk, 1971). Cooke and Warren (1973) claim that most aeolian sediments can be seen as part of a continuum between two distinct types; one has a very limited range of sizes with a modal of 0.177 mm positively skewed; the other is bimodal with a coarse mode of above 0.6 mm and fine mode of about 0.062 mm. They believe that both types are of common origin.

Descriptions of coarse dune sand causing a distinct dune type are prevalent in literature. Bagnold (1931) described a wide, low dune built of grains of all sizes. He called this type "whaleback" and noted that the mobility of vehicles on this type is good. A similar description of a low, flat dune called a zibar is given by Holm (1960). Warren (1972), who also described this dune type, showed that it is composed of bimodal sand, while the seif dune sands in the same area are unimodal. Wilson (1972, Figure 1) presented a graph that shows a correlation between aeolian bedforms (ripples, dunes, and draas) wavelengths and grain sizes, in which the wavelength increases with particle size. Thus, one might infer that bigger dunes have coarser sands. Howard et al. (1978), who developed an equilibrium model of barchan dunes, concluded that a decrease in grain size should cause a steeper and more blunt dune. They also calculated and analyzed the situation that leads to the profile of equilibrium of barchan dunes in three dimensions.

Dune Morphology and Grain Size

Desert dunes display a great variety of shapes, profiles (Breed and Grow, 1979) and grain sizes (Folk, 1971). It is clear that most of these dunes are in a steady state that depends on grain size, wind velocity, and direction. Field work conducted in Arizona and southern California has yielded data (Figure 1) regarding the effect of grain size on dune morphology. Figure 1 shows the effects of grain size (determined by the ratio of mean size $\bar{X}$) to sorting ($\sigma\phi$) on dune morphology (determined by the ratio of dune height h to half of its length L). It seems that low, flat dunes display a low mean grain size (in ϕ values) and a high ϕ value of sorting while the only combination that can create steep dunes is high mean grain size and low sorting. Flat dunes (low h/L) can be composed of all types of sand with high and low $X\phi/\sigma\phi$.

Figure 2
Results of measurements of the amplification factor (A/z) over a triangular profile model in the wind tunnel.

Figure 3
Results of calculated A_z ($= U_2/U_1$) and q_2/q_1 over a uniform windward dune slope of $4^\circ 45'$.

Other descriptions in the literature (Bagnold, 1931; Holm, 1969; Warren, 1972) show that the coarser the sand, the flatter and less prominent the dune.

This paper offers some two-dimensional theoretical explanations for the profile of equilibrium of dunes, and the effect of grain size on their morphology.

Equilibrium Profile of a Dune

A dune has a profile that changes the wind speed above it. The change in wind speed is essential for the increase in the rate of sand transport on the windward slope which maintains the dune in a steady state. The sand transport depends on the cubic wind velocity (Bagnold, 1954).

$$q = C \left(\frac{d}{D} \right)^{1/2} \frac{\rho}{g} U_*^3 \quad (1)$$

where: q = sand transport (M/LT)

U_* = friction velocity

ρ = air density

d = grain diameter

D = reference grain diameter (=0.025 cm)

C = constant (=1.8 for high dune sand)

The change in wind velocity over the dune measured at a particular height (z) above the ground level can be expressed as the amplification factor $A_z = \bar{U}_2/\bar{U}_1$ (Bowen and Lindley, 1977), defined as the mean wind velocity at a height z above the dune ($\bar{U}_2$) divided by the mean velocity of the undisturbed wind flow at the same height z above a flat surface ($\bar{U}_1$).

We studied the amplification factor both theoretically and empirically over several simulated dune profiles. The dune profile in our study can be divided into two types: one with a triangular profile (Figure 2) and the other with a convex blunt profile (Figure 6). These two profiles exist on longitudinal and barchan dunes respectively (Tsoar, 1974).

Measurement of wind velocities carried out on triangular models in a wind tunnel (Figure 2) show that the rate of the amplification factor is not uniform. There is a large increase in A_z close to the crest and a large decrease at the upwind base (Figure 2). Theoretical calculations of A_z for escarpments with this profile can be made by means of Jackson's (1975) equation:

$$A_z = 1 + \frac{h}{4\pi L} \ln \frac{(z/L)^2 + [1+(X/L)]^2}{(z/L)^2 + [1-(X/L)]^2} \quad (2)$$

Figure 4
Results of hypothetical calculation of A_z needed to keep a triangular shaped dune in a steady state.

Figure 5
Results of measurements of the amplification factor (A_z) over a convex model ($h/L = 2/9$).

where: z = the height where the measurement was taken

L = the horizontal distance from the top of the hill to the point at which the height is half the maximum height

h = height of escarpment

X = horizontal distance from the middle of the windward slope downwind.

Results of these calculations give a curve of A_z similar in trend to that of Figure 2 with an increase at the crest and a decrease at the base (Figure 3).

Equation 1 gives us the relation between wind velocity and sand transportation. In order for a dune to be maintained in a steady state, the rate of sand transport must increase so that in every unit area (along the windward slope), q will be capable of carrying all the sand eroded before this unit area, and to continue eroding more sand. On the uniform windward slope of Figures 2 and 3, an increase of q in arithmetical progression will give a steady state or an equal forward advance at every point of the windward slope. Calculations of q_2/q_1 (q_1 being the sand transport where the velocity is $\bar{U}_1$; and q_2 where the velocity is $\bar{U}_2$) show that the increase in A_z close to the crest brings about a great increase in the sand transport, which will result in much higher erosion there (Figure 3). In other words, a dune that has the shape shown in Figures 2 and 3 will never be in a steady state. Figure 4 shows what rate of increase of A_z is required over a triangular profile dune in order to yield a steady state in which dq/dx (rate of sand transport per unit area) is the same everywhere on the windward slope. The A_z curve of Figure 4 shows a lower increase in velocity toward the crest, unlike the curves of Figures 2 and 3.

When the windward slope gradient is changed, e.g. over a convex hill, the rate of sand transport per unit area changes accordingly. The relative between the rate of sand transport per unit area to the slope is represented by the equation:

$$\frac{dq}{dx} = \gamma c \tan \alpha \quad (\text{Bagnold, 1954}) \quad (3)$$

where: γ = specific weight of loosely packed sand in bulk

c = rate of advance of the dune

α = the inclination angle of the slope

A dune in a steady state should have the same rate of advance, c , everywhere on the dune.

Figure 6

Results of measurements of the amplification factor $A(z)$ over a convex model ($h/L = 1/9$).

Figure 7

Results of calculated A_z over a convex hill (I - $h = 15$ m $L = 100$ m II - $h = 5$ m $L = 100$ m).

Theoretical calculations and measurements of A_z over curved models show a convex curve of A_z . Figures 5 and 6 show the results of wind tunnel measurements. It is clear that the lower the model (low h/L) the more blunt the A_z curve. Low h/L results in low $\tan \alpha$, which should result in lower wind velocity, and lower rate of advance maintaining the model in a steady state (equation 3). Theoretical calculations show similar results (Figure 7). They are based on equation 3.23 b of Jackson and Hunt (1975), which was developed for a hill whose profile is defined as:

$$\left(\frac{X}{L}\right) = \left[\left(\frac{X}{L}\right)^2 - 1 \right]^2 - L \leq X \leq L \quad (4)$$

where: X = horizontal distance from the top of the hill downwind

L = horizontal distance from the top of the hill to the point in which the height is zero.

A_z = is calculated according to the equation:

$$A_z = 1 + 4 \frac{h}{L} \left[\frac{\ln(L/z_0)}{\ln(z/z_0)} \right]^2 \left\{ \frac{4}{3} - 2 \left(\frac{X}{L}\right)^2 - \frac{X}{L} \left[\left(\frac{X}{L}\right)^2 - 1 \right] \ln \left| \frac{X/L - 1}{X/L + 1} \right| \right\} \quad (5)$$

where:

$$z = \frac{z_0}{8} \left(\frac{L}{z_0} \right)^{0.9} \quad (6)$$

z_0 = roughness length.

Sand transport (equation 1) and rate of advance of different points along the windward slope were calculated for two hypothetical dunes, one with a uniform windward slope (as in Figure 2) where $h/L = 0.02$ and the other having the profile of equation 4 with $h = 5$ m, $L = 100$ m (Figure 7). The grain size in the two models is the same (0.0177 cm). Figure 8 shows $c/\bar{c}$

Figure 8

Results of the calculated ratio of rate advance ($\bar{c}$) to average rate of advance (c) versus X/L . (X is the horizontal distance from the crest line downwind. L is the horizontal distance from the top of the hill to the point in which the height is zero.

Case I - dune with uniform slope and $h/L = 0.02$.

Case II - dune with the profile of Figure 7 Case II.

Figure 9

A profile of a barchan dune subjected to unidirectional wind. Sinai Desert.

versus X/L for the two hypothetical dunes, c is the rate of advance;

$$\bar{c} = \frac{1}{n} \sum_{i=1}^{i=n} c_i;$$

We can discern that the hypothetical dune in Figure 7 is much closer to a steady state than the dune in Figure 2, since every point on the windward slope advances at almost the same rate. The dune profile of Figure 7 (equation 4) is very similar to a barchan dune profile (Figure 9), i.e., profile of a mound of sand subjected to a unidirectional wind. Dunes that have a triangular profile as in Figure 2 are never in a steady state unless they are subjected to more than one wind direction so that their profile changes accordingly and the lee-side eddies play an important role in the formation of the sharp profile (Tsoar, 1978).

Threshold Velocities of Different Grain Sizes on Different Levels of a Windward Dune Slope

The profile of equilibrium as determined in the previous chapter exists as long as the wind velocity is above the threshold on all levels of the dune, from the base to the crest. It is obvious from Figures 2, 5, 6, and 7 that the threshold velocity (as measured at a certain elevation on a flat undisturbed surface where $A_z = 1$) will be lowest at the crest and highest at the base of the dune. We can conclude from these facts that every hill of sand will be lowered by a certain rate when the wind velocity (measured where $A_z = 1$) is at or above the threshold at the crest, and below the threshold on other lower levels. If the difference in the threshold velocities between the crest and lower levels is small, the short period during which the crest is eroded will be compensated for by a much longer period in which the wind velocity is above the threshold for all the dune levels. Moreover, the wind magnitude during the time in which only the crest is in movement is relatively low, and according to equation 1, the amount of erosion (sand transport) of the crest will be low. On the other hand, in those cases where are great differences in the threshold velocities between the crest level and lower levels of the dune, the situation will be different. The period in which the wind velocity will be above the threshold of the crest and below the threshold of the lower levels of the dune is much longer than the period in which all the dune levels are in motion (wind velocity is at or above the threshold of the base of the dune). The crest

Figure 10
Values of threshold velocities (U_{10t}) measured at 10-m height, for different grain sizes on different levels of windward slope (X/L) of terrestrial convex dune having the profile of Figure 7 case II.

Figure 11
Same as Figure 10 but with a dune having a profile of Figure 7 case I.

section will be eroded in this case and the dune will not be able to maintain steep profile.

By using equation 5 we can calculate the threshold velocity measured at a height of 10 m above the ground (standard height of meteorological anemometers) on open terrain (where $A_z = 1$) for different grain sizes on the windward slope of the dune.

Figures 10 and 11 give the results of calculations for the dune profiles shown in Figure 7. In these profiles, which are close to steady state, the threshold velocities around the crest do not change much. The trend of the threshold velocity is the same in the low dune (Figure 10) and the high one (Figure 11). In the case of the high dune, the threshold velocities near the crest are low. Hence, the difference in threshold velocity between the crest area and the middle of the windward slope is much greater and this trend increases with grain size increase. With coarse sand and high dune (Figure 11) there is a high percentage of time when wind velocity is above the threshold of the crest area and below the threshold of lower levels of the dunes. High dunes with coarse sand will not be stable during the relatively long period in which movement only occurs around the crest. The relatively short time in which all the dune's levels are in movement will not suffice to rebuild the eroded crest. This situation will not occur with fine sand (such as 0.0177 cm).

In conclusion, it appears that coarse sands are not stable in steep dunes because the crest area is eroded for a long period (Figure 11). When the dune becomes lower and flatter, the erosional period of the crest is shorter (Figure 10).

Summary and Conclusions:

1. Every dune that is subjected to a unidirectional wind has a steady-state profile by which the morphology of the dune is preserved. According to the amplification factor (A_z) of the wind over a hill, the steady state profile should be convex (Figures 6 and 7).
2. A triangular-shape profile will not be in a steady state under a unidirectional wind. Such a profile is subjected to strong erosion close to the crest. This profile can be formed by multidirectional winds, mainly oblique winds that create strong reverse flows on the lee side (Tsoar, 1978).
3. In addition to the steady-state profile, there is great importance given to the distribution of the threshold velocity at the windward slope for different grain sizes. When the dune is composed of coarse grains the threshold velocity is higher and there is a greater difference between the threshold velocity at the crest area and at the lower levels of the dunes

(Figures 10 and 11). It is obvious that a coarse sand dune will be subjected to a much longer period where only erosion occurs near the crest, lowering the dune to a profile that is incapable of having a slip-face. Fine-grained dunes have a far shorter period of erosion of the crest area. This erosion is compensated for by a longer period in which all the dune levels are in motion.

REFERENCES

- Bagnold, R.A. (1931) Journeys in the Libyan Desert, 1929 and 1930. *Geogr. Jour.*, 78, 13-39, 524-533.
- Bagnold, R.A. (1954) *The Physics of Blown Sand and Desert Dunes*. Chapman and Hall, London, 265 p.
- Bowen, A.J. and Lindley, D. (1977) A Wind Tunnel Investigation of the Wind Speed and Turbulence Characteristics Close to the Ground Over Various Escarpment Shapes. *Boundary-Layer Met.*, 12, 259-271.
- Breed, C.S. and Grow, T. (1979) Morphology and Distribution of Dunes in Sand Seas Observed by Remote Sensing. In: *A Study of Global Sand Seas*, edited by E.D. McKee, U.S. Geol. Surv. Prof. Pap., 1052, 253-302.
- Cooke, R.U. and Warren, A. (1973) *Geomorphology in Deserts*. B.T. Batsford, London, 374 p.
- Folk, R.L. (1971) Longitudinal Dunes of the Northwestern Edge of the Simpson Desert, Northern Territory, Australia, 1. Geomorphology and Grain Size Relationship. *Sedimentology*, 16, 5-54.
- Holm, D.A. (1960) Desert Geomorphology in the Arabian Peninsula. *Science*, 132, 1369-1379.
- Howard, A.D., Morton, J.B., Gad-El-Hak, M., and Pierce, D.B. (1978) Sand Transport Model of Barchan Dune Equilibrium. *Sedimentology*, 25, 307-338.
- Jackson, P.S. (1975) A Theory for Flow Over Escarpments. *Proc. Fourth International Conf. on Wind Effects on Buildings and Structures*, Heathrow, U.K., September, 1975, Cambridge Univ. Press, 33-40.
- Jackson, P.S. and Hunt, J.C.R. (1975) Turbulent Wind Flow Over a Low Hill. *Quart. J. Roy. Met. Soc.*, 101, 929-955.
- Tsoar, H. (1974) Desert Dunes Morphology and Dynamics: El-Arish (Northern Sinai). *Z. Geomorphl. Suppl.*, 20, 41-61.
- Tsoar, H. (1978) *The Dynamics of Longitudinal Dunes*. Final Technical Report, Eur. Res. Office, U.S. Army, London, 171 p.
- Udden, J.A. (1898) *Mechanical Composition of Wind Deposits*. Augustana Library Publ., 69 p.

- Warren, A. (1972) Observation on Dunes and Bimodal Sands in the Tenere Desert, Sedimentology, 19, 37-44.
- Wilson, I.G. (1972) Universal Discontinuities in Bedforms Produced by the Wind J. Sed. Pet., 42, 667-669.

STABILIZATION AND AFFORESTATION OF SAND DUNES THROUGH BIOSALINE CULTURE TECHNIQUE

By

R. Ahmad

ABSTRACT

In view of the success met with in growing salt-tolerant plants in a sandy belt using highly saline water for irrigation, it is suggested that this technology be adopted for afforestation and stabilization of coastal sand dunes, where underground brackish water is often available. It is further proposed to adopt a natural ecological sequence in plantation. Biosaline cultivation coupled with man-made plant succession is expected to shorten the duration of time that is otherwise required by nature for reaching the stage of climax vegetation. Names of the plants that are suitable in this endeavor are mentioned in the text.

DISCUSSION

Dunes constitute an integral part of sandy deserts and are found in abundance in the areas classified as arid or semiarid on the basis of the Thornwaites Index. The arid regions comprise about one-third of the land surface of the Earth, of which roughly about one-third is occupied by coastal or sub-coastal sand deserts. It will be useful if sandy deserts are classified on the basis of their location into "coastal" or "inland" sandy deserts for the purposes of afforestation. Coastal regions have a definite advantage over the regions located thousands of miles away from sea. Their environment,

in terms of maximum/minimum temperatures, wind velocity and direction, relative humidity, annual precipitation, the presence of subsoil brackish water at comparatively low water tables and the frequent presence of perpetual fog and cold oceanic currents make them different from the inland sandy deserts.

The Sahara desert of Africa, the central deserts of the Arabian Peninsula and that of other countries of the Middle East; the Dashte Kavir and the Dashte Lut of Iran, the Thar Desert of Indo-Pakistan and the Takla-Makan Deserts of western China, all constitute the main inland hot, sandy deserts of Africa and Asia. The sandy regions of the Mediterranean coast, the Somali-Ghalbi Desert along the African shore of the Red Sea, the Gulf of Aden, the dry, sandy belt of the countries around the Persian Gulf, especially the Jafurah sand sea of Saudi Arabia, and the sandy belt of the Makran Coast of Pakistan, are all examples of coastal sandy deserts in Africa and Asia (Bhardwaj, 1961, McGinnies et al., 1968). The coastal desert of Namib at the western edges of Angola in southwest Africa presents a bit different environment, due to presence of cold offshore oceanic currents and perpetual fog.

The temperature range varies widely from one desert type to the other. In inland hot, sandy deserts, the average annual range of temperature is roughly between 10 to 35°C, whereas these values may become relatively mild at the coastal sandy belt. Similarly, the amount of rainfall varies greatly both in its seasonal and yearly distribution (Zohary, 1973). Leaving aside the extremes, one can say that the annual rainfall may vary from 10 to 100 mm in inland sandy deserts, though instances ranging from 0 to 500 mm may also be met with occasionally. Secondly, the frequency of rainfall is so unpredictable and the structure of the rainy season is so different that it is difficult to forecast for the next season. Semiarid coastal and subcoastal regions are in a more advantageous position due to the presence of comparatively high relative humidity and latent precipitation such as dew.

Nature has devised different techniques for stabilization of sand dunes, depending on the environment. The appearance of vegetation initiates microclimatic changes and as a result of plant succession, these changes become macroclimatic. With further passage of time they lead to a well-stabilized sand dune, thickly populated by climax vegetation. This is no doubt a slow and time-consuming process, but with the use of modern technology, man can initiate and accelerate this process. Advantage could be taken from the findings that plant tissue is capable of regulating absorption of minerals to a certain extent, and salinity-induced changes in the metabolism enable the salt-tolerant plants to bear osmotic and ionic shocks under a saline environment.

Nature has itself demonstrated the occurrence of pioneer vegetation followed by plant succession leading to stabilization of a sand dune (FAO, 1960). Plants having a foliage cover spread over the soil surface and those with adventitious roots are both significant sand binders. Citrullus colocynthis, Cucumis prophetarum, Cyperus arenarius, Ipomoea pes-caprae are some examples. Herbaceous plants capable of growing under arid conditions, form culms and act as binders of sandy strata. Plant communities of Haloxylon persicum, Heliotropium tuberculatum, Salsola baryosma, Suaeda fruticosa, Tribulus terrestris, Fagonia cretica, Alhagi maurorum, Zygophyllum simplex, Cressa cretica can all be observed growing on saline coastal sand. Among the shrubs Indigofera argentea, Aerva pseudodotomentosa, Leptadenia pyrotechnica, Atriplex canescens, Euphorbia caducifolia, Calotropis procera, Saccharum spontaneum, Gymnosporia senegalensis, Commiphora mukul, Capparis decidua, Calligonum polygonoides, Lycium shawii, and among the trees Acacia senegal, Salvadora oleoides, Tamarix troupii and Prosopis cineraria appear in various densities and stabilize coastal and subcoastal sand dunes. Keeping in view the environmental conditions of a barren coastal sandy desert, one can recommend introduction of certain plant types for that locality. Furthermore, plant species that have shown their emergence during the process of succession under natural conditions could be cultivated in a similar environment by using one of the techniques to be mentioned later in the text. This would amount to inducing a man-made succession of plant species through suitable knowledge of brackish water irrigation. To accelerate the stabilization of the sand dunes, the phases of introduction, succession, and the appearance of the climax vegetation could be made to be somewhat overlapping one another. Thus, the duration of primary stabilization could be shortened and connected much earlier with the phase of secondary stabilization.

Between the extremes of the minimal and the maximal values of desert conditions one may still find a period when the climate is favorable for plant growth. Nonavailability of water and unsuitable soil types are only two of the limiting factors under the above-mentioned conditions. The soil of the sandy deserts could prove to be a blessing though, if it were to allow most of the brackish water to percolate down without standing for a long time in the root zone of the plant. Hence, if brackish water of a reasonable salt concentration is present at a subsoil level, it could be used for irrigation purposes in sandy deserts.

Modern approaches of plant physiology have made it possible to use highly saline water for irrigation of sandy strata for afforestation and agriculture. It is important to calculate water salt balance as well as the amount of salts accumulated due to brackish water irrigation. The accumulated

salt should not exceed beyond a critical permissible limit. Kovda (1977) has proposed a simple equation for inflow and outflow of salts.

$$Y C + Q_1 C_1 = Q_2 C_2 + Q_{dr} C_{dr},$$

where

Y = amount of irrigation water inflow during a given time limit.

C = concentration of salts in irrigation water during a given time limit.

Q_1 Q_2 = natural flow rate of the incoming and outflowing ground water, respectively.

C_1 C_2 = concentration of salts of the incoming and outflowing groundwater, respectively.

Q_{dr} = flow rate of the outflowing drainage water.

C_{dr} = the weighed average actual salinity of groundwater in a drained field (salt concentration of the drainage water).

Effective drainage is the main factor in the regulation of salt balance in saline irrigation, which is of no problem on sandy deserts. The practice of excessive irrigation brings about oversaturation, which regulates the removal of salts with drainage water. Salinization of the soil will not occur if $YC + Q_1 C_1 = Q_2 C_2$. The presence of natural drainage (Q_{dr}) and the removal of groundwater salt (C_{dr}) by natural leaching on sandy strata help in creating a negative salt balance (desalinization). Timing for irrigation also plays an important part in avoiding the accumulation of salts in the substrate. Plants may preferably be irrigated just after sunrise or near sunset to avoid extensive evaporation of water after irrigation. Rains, if any, provide supplemental irrigation and wash down the salts accumulated in the root zone.

The maximum permissible limit of salinity in irrigation water keeps on changing in accordance with the permeability of the strata and the types of plants involved. Saline water having EC 2250-5000 micro mohs (salt concentration 0.15 to 0.3%) is being used in Morocco for growing cotton and maize in soils having good natural drainage (Durand, 1956). Experience of Tunisia, Egypt, India, and USSR in using water with a mineral content varying from 2 to 7 g/l is very encouraging for brackish water irrigation under favorable drainage conditions (Darra et al., 1964). Certain salt-tolerant plants

growing at the intertidal zone of a sandy desert (*Hordeum marinum* and *Beta maritima*) give a natural demonstration of irrigation with water of oceanic strength. Cultivation under extreme climatic conditions through saline irrigation is being practiced at Al-Hassa (Saudi Arabia) and useful results for the consumptive use of salty water have been obtained for the date palm, alfalfa, cabbage, barley and sorghum under desert conditions (Achtnich, 1975).

Boyko and Junke (1966) have done some pioneer work in this direction and used water having total dilutions of salts ranging from 2,000 to 6,000 mg/litre for flowing salt-tolerant plants on sandy gravel strata at the desert garden in Eilat. Circular frames of asphalt barrels were dug into sand after the removal of the bottom plate. Salt-tolerant plants were grown in these frames and irrigated with saline water.

Some modest technologies have been developed recently for agriculture and afforestation in arid regions. Building up catchment areas in depressions for rain water harvesting, runoff afforestation and agriculture, conservation of water by trickle or sprinkler irrigation, bituminous mulching of sand dunes, use of subsoil saline water for plant growth, and selection of stress-tolerant plants are some of the techniques recommended for afforestation of sandy deserts (Bishay and McGinnies, 1979, NAS, 1974). The experience of Iran's under dune development program is very encouraging in this respect (Niknam and Aharanjani, 1975). Direct broadcast seeding of both woody and grassy species by hand or aircraft has been found to be less effective than transplanting saplings and cuttings of stress-tolerant plants on sand dunes, followed by bituminous mulching at the surface, which has proved to be much more fruitful for stabilization and afforestation. Plants were irrigated up to a maximum of three times during the growing season during the first and subsequent years. Stabilizing sand dunes with heavy oil mulch in the Rajasthan deserts of India and that of barren stretches of the Sahara in Libya followed by plantation has been successful. Erection of wind and sand breakers made up of interwoven dried plant material on sandy strata and planting stress-tolerant plants in between these partitions has also been a successful technique world over. Exploitation of fossil water from deep soil sandstone aquifers in the Kufra Desert of Libya through a center-pivot sprinkler system has made it possible to produce crops such as wheat, barley, and alfalfa on land that never produced any plant life before, though it has not yet proved to be an economically sound proposition.

The author and his associates have developed a methodology for the use of saline water (or various dilutions of seawater) for afforestation and agriculture on a coastal sandy belt (Ahmad and Abdullah, 1979, 1980, 1981). A drum pot culture technique was practiced on the sand of the Makran coast in

Pakistan and seedlings of salt-tolerant plants were transplanted and irrigated with various dilutions of seawater mixed with chemical amendments. Pretreatment of seedlings with gradually increasing salinity in irrigation water prior to transplanting encouraged the development of salt tolerance, and saplings were in better positions to survive in their new surroundings. Concentrations of salts in subsoil water at various distances from the shore at the sandy belt of the Makran coast in Pakistan were determined by us to see if the salt content of the subsoil water is proportionally decreased by increasing the distance from the seashore. Results are presented in Figure 1.

It is evident that the salinity level of the subsoil water at 1,000 m away from the sea is reduced to about 1/10th the oceanic strength. However, our results indicate that the concentration of underground water away from the seacoast may vary greatly from one location to another, depending on the soil profile of the region and the amount of precipitation.

Chemical amendments in saline water further improve the situation. On one hand, they bring about the maximum possible balance in an ionic ratio, but on the other, they maintain a suitable pH of irrigation water and provide essential minerals for plant growth. It is a recognized practice to add gypsum (CaSO_4) for improvement of saline soils or brackish water. Similarly, the importance of K^+ amendment in improving the quality of irrigation water and the building up of salt tolerance is known in literature. Saline water has also been occasionally supplemented with phosphate and nitrate in order to provide beneficial NPK values for plant growth. The author and his associates have worked out a recipe of chemical amendments which, when mixed with brackish water, proved beneficial for plant growth on sandy strata. Our results have indicated that trickle or sprinkler irrigation with saline water causes more harm in accumulations of salt in the root zone than 'limited-flood' irrigation, which restricts the spread of saline water on the surface but provides a direction for its percolation towards a subsoil region. Overirrigation was practiced under the above-mentioned condition. We have been successful in growing maize and castor bean on coastal sand by irrigating them with the seawater starting from zero to one-fourth oceanic concentrations, and beet root, eucalyptus, muskeet, and tamarix with seawater starting from zero to one-half oceanic strength. Various dilutions of seawater were supplemented with chemical amendments according to our recipe. Dewfall in certain periods of the year provided pure water for foliar absorption and was quite helpful for plant growth. Hence, one can grow wild plants in their natural sequence of plant succession for stabilization of sand dunes using brackish water for irrigation.

Figure 1
Distance from seacoast.

Different stages of plant succession could be made to overlap for an early stabilization. Skipping the introduction of pioneer vegetation (herbs and undershrubs) and directly transplanting climax vegetation (shrubs and trees) without providing soil cover of prostrate plants (or without spraying heavy oil mulch) may not be very useful in this respect. However, substituting fodder grass or any other stress-tolerant plants of economic importance in the sequence of plant succession may provide some monetary benefit as well.

Recent advances in science have shown that an involvement with a multidisciplinary approach often helps in solving complicated problems. Disciplines like theoretical physics and mathematics have much to contribute towards the understanding of fundamentals involved in the shifting of sand and the formation of dunes under desert conditions. Climatology, geophysics, and desert hydrology involving remote sensing devices provide basic information to plant scientists for development a program of sand dune afforestation leading to their stabilization.

REFERENCES

- Achtnich, W. (1975) Cultivation of Plants Under Extreme Climatic Conditions Through Use of Salty Irrigation Water. Plant Research and Development 1:12-21. Bundesforschungsanstalt für Forst-und Holzwirtschaft. Hamburg.
- Ahmad, R. and Abdullah, Z. (1979) Saline Agriculture Under Desert Condition in "Advances in Desert and Arid Land Technology and Development". Vol. I 593-618. (Bishay, A. and W.G. McGinnies, eds.). Harwood Academic Publishers.
- Ahmad, R. and Abdullah, Z. (1980) Biomass Production Under Desert Conditions. Biotech. Bioeng. Symp. 10:121-124. (John Wiley and Sons, Inc.).
- Ahmad, R. and Abdullah, Z. (1981) Biomass Production of Food and Fiber Crops Under Desert Conditions Using Highly Saline Water. International Biosaline Symp. Mexico. Plenum Publishing Corp.
- Bharadwaj, O.P. (1961) The Arid Zone of India and Pakistan in "History of Land Use in Arid Region", Arid Zone Research XVII UNESCO (L. Budley Stamp, ed.).
- Bishay, Adli and McGinnies, W.G. (1979) Advances in Desert and Arid Land Technology and Development Vol. I. Harwood Academic Publishers, New York.
- Boyko, Hu-o and Junke, W. (1966) Salinity and Aridity, New Approaches to Old Problem. Dr. W. Junke Publishers. The Hague.

- Darra, G.L., Mehta, K.M. and Pareek, B.L. (1964) Quality of Irrigation Water in Rajasthan I. Indian Society of Soil Science. 12:121-130.
- Durand, J.H. (1956) The Quality of Irrigation Water. African Soils. 4:52-88.
- Food and Agriculture Council (1960). Soil Erosion and Its control in Arid and Semi Arid Zone. Proceedings of Karachi Symposium, Pakistan.
- Kovda, V.A. (1977) Arid Land irrigation and soil fertility; Problem of Salinity, Alkalinity Compaction. In "Arid Land Irrigation in Developing Countries, Environmental Problem and Effects". E.B. Worthington (ed.) Pergamon Press, Oxford.
- McGinnies, W.G., Boldman, B.J. and Paylore, P. (1968) Deserts of Worlds: and Appraisal of Research Into Their Physical and Biological Environments. University of Arizona Press, Tucson.
- National Academy of Science (1974) More Water for Arid Lands. Promising Technologies and Research Opportunities. Washington, D.C.
- Niknam, F. and Aharanjani, B. (1975) Dunes and Development in Iran. Pub. Imperial Government of Iran, Ministry of Agriculture and Natural Resources.
- Zohary, Michael (1973) Geo-Botanical Foundation of the Middle East. Vol. I & II Gustav Fischer Verlag, Stuttgart.

SOME ASPECTS OF SAND STABILIZATION IN EGYPT

By

S. S. Tag El Din

ABSTRACT

Sands and sand dunes occupy vast areas of Egypt. The majority of these areas are of the inland type; the remaining are of the coastal type. The inland dune chains cannot be stabilized by vegetation due to the lack of permanent irrigation in the extremely arid climate, nor by the application of chemicals, which are of prohibitively high cost considering the enormous size of the dune areas involved. Only small sand areas, mostly around shallow wells or in places where the water table comes close to the surface, are stabilized by the growth of natural vegetation.

The stabilization of coastal sand dunes, those where conditions allow stabilization by vegetation, could be carried out through certain steps enacted through a program taking local conditions into consideration. Techniques for the fixation and afforestation of the coastal dunes of Egypt are suggested.

INTRODUCTION

Sand dunes occupy vast areas in many countries of the world, posing a major problem to the development of these nations. They are unproductive lands that threaten to cover inhabited localities, main roads, fertile soils, and water resources. The control of these mobile dunes and their placement under production is of high priority if one considers the

fast growing rate of the global population and the need for newly reclaimed areas.

In Egypt, sand dune fixation is one of the major problems, as shifting dunes are widespread and very great damage has been caused during their long history.

Geographical distribution of dunes in Egypt (Figure 1).

Dunes are well known in the following districts:

A. The Western Desert:

- (1) Coastal sand dunes.
- (2) The Great Sand Sea.
- (3) The northwest--southeast longitudinal sand sea (south of the Qattara Depression).
- (4) Sand dunes and barchans between and within oases.
- (5) Nubian sand sheets and shadows.

B. The Delta and its Fringes:

- (6) Coastal dunes.
- (7) Turtleback or ancient dunes--like that of Quisna.
- (8) El Khanka dunes near Cairo.
- (9) El Salhiya--east of the delta.
- (10) El Tahreer--west of the delta.

C. The Eastern Desert:

- (11) Recent formation of migrating dunes in the region between Port Said and Suez.
- (12) Nubian sand sheets and shadows.

D. Sinai Peninsula:

- (13) Northern coastal dunes.
- (14) The western coastal zone (Gulf of Suez).
- (15) Inland dunes--like that of Maghara.

Inland dunes of the Western Desert

The Great Western Sand Sea, with a width of 200 to 250 km, begins about 50 km west of the Farafra oasis. There are a few smaller systems of sand dunes to the east of it affecting the oases of the New Valley. All of them have the same direction of movement from NNW to SSE in common. The dune field begins at the southeastern side of the Farafa Oasis and ends about 25 km north of the Dakhla Oasis, where the dunes disintegrate into parallel strips of migrating barchans. These dunal chains cross the oasis and reunite again into one large dune belt south of Dakhla (Hagedorn et al., 1977).

Even more conspicuous are the migrating dunes affecting the Kharga oasis. A number of parallel dune chains on the eastern side of Bahariya grow into a single strip only 4 to

Figure 1
Map of Egypt showing the location of sand dunes in Egypt (El-Baz, 1978).

6 km wide, called Ghard Abu Muharik. It crosses 250 km of level limestone desert, widening during the next 30 km to about 12 km, descends two escarpments and at the same time breaks up into about a dozen chains of dunes subdivided by narrow N-S corridors. Southward, most of the dunes unite again in a belt forming the western boundary of the Kharga oasis.

The migrating dunes cover only about 2.5% of the area in the Kharga Oasis, while it is about 12% for Dakhla. It is somewhat misleading because a relatively large part of the dunes of Kharga lie within the already cultivated area, whereas the majority of the sand dunes of Dakhla lie in two large areas outside the cultivated land. The area of dune land could be negligible, but it becomes important because of the migratory nature of these dunes. Individual barchans move at a velocity of 8 to 10, sometimes even 15 m per year, temporarily covering every object standing in their way.

The dune chains described cannot be stopped by vegetation, because it can only be done with permanent irrigation in the extremely arid climate, and because the application of chemicals would not be feasible due to their high costs and the enormous length of the dune areas north of the oases. Even if the dunes within the oases were fixed at high cost, there still remains an unlimited source of movable sand that is invading the stabilized areas and covering the stabilized dunes. The only economical way to handle the migrating dune chains is to get out of their way. This has already been followed during the expansion of the cultivated area. Agricultural land is located in the the corridors between the dune chains, and there is enough land in a safe position for future expansion of the irrigated area.

In addition to the migrating dunes there are small areas where sand dunes have become stabilized by the growth of natural vegetation--mostly around shallow wells or in places where the water table comes close to the surface.

The whole region is affected by the movement of loose sand. The wind speed close to the surface decreases in cultivated areas because their higher surface roughness causes blown sand to settle at a higher rate. The sand tends to develop into sheets of increasing thickness. The northern margins of the cultivated areas are the most threatened ones. The measured rates of annual sand accumulation reach a layer of 5 to 6 cm around any obstacle.

While protection against the migrating dunes is almost impossible, this type of moving sand can successfully be kept under control. Present practice is to construct mechanical defense lines of reed or Sorghum straw fences, crosswise or oblique to the direction of the prevailing wind. It was suggested that the fences should be about 1 meter high, at a distance of 5 to 7 m from each other, each fence consisting of two rows of reeds or Sorghum straw a few cm apart, the space

between them filled with grass or other vegetative material and both rows tied together with ropes or palm leaves. This type of fence would be impermeable to the sand, which consequently would settle before the obstacle. However, the barriers where the sand grains cannot go through are easily crushed by the lateral pressure of the depositing sand.

There are also shelter-belts of casuarine, 7 to 10 rows deep and furrow irrigated. These belts do not appear very promising, because the furrows are easily covered by the sand accumulating within the belts. Therefore, a system of shelter belts of casuarine, three rows deep, with the lower branches cut off to a height of about 1.5 m is recommended. This system lets the wind pass between the trees, still allowing an increase in its speed by compression of wind-flow, but with this practice irrigation canals will be filled by sand at a very slow rate, and trees will not be in danger of suffocation from accumulating sand.

The thick shelter-belts cause a rapid drop of wind speed immediately in front of and within the belt where, consequently, sand deposition takes place. The wind speed then rises quickly again behind the obstacle, even to the extent that whirlwinds develop. In contrast, velocity increase is lower behind the narrow forest belt. If a casuarine strip height reaches 10 m as estimated in New Valley within 10 years--wind velocity would be reduced to about 40% of the original value at a distance of 50 to 75 m. The reduction will be repeated if the next strip is planted at that distance. Therefore, the pilot project is for the construction of eight strips, both for additional security and for possible future agricultural use of the land between the southernmost strips. Three rows of mechanical shelter-belts will protect the area until the trees have reached a sufficient height.

There are further plans to stabilize individual dunes at the margin of a dune chain to the southeast, with the drainage water from the pilot project area, irrigating a shelter-belt both on the leeward and on the windward sides of an individual dune, enabling natural vegetation to grow on seepage water on the dune. Stabilization with the aid of drainage water would not only have the benefit of reducing the amount of blown sand, but it would also be useful in reducing the drainage problem in oases.

Coastal Sand Dunes

Coastal dunes are of marine origin and are found along the northern coast. The northwestern coastal dunes are maritime sand dunes and white in color. They are mobile, generally supporting no vegetation and only on scattered spots can some *Ammophila arenaria*, *Agropyron junceum*, or *Retama ratam* be found. They grow in a long but narrow belt along the

Mediterranean Sea, with height ranges from sea level up to 35 m above sea level. The texture of these coastal sands is generally coarse. The content of calcium carbonate is very high, ranging between 78% and 95%, it is strongly alkaline (pH 8.0 to 8.5), and very poor.

The total area of northwestern coastal maritime sand dunes is estimated to be about 54,000 hectares, of which about 65 to 75% appear to consist of cemented dunes and 25 to 35% of shifting dunes. Close estimation of cemented dunes is difficult because the crusted surface of a cemented dune is sometimes covered with a thin layer of sand. It was estimated that in the El Qasr pilot zone that shore dunes represents 9%, cemented dunes 54%, sloping area of dunes 24%, summit area of dunes 7% and depressions 6%. The shifting sand dunes are a great threat to various cultivated areas, and their reclamation is technically feasible and necessary, except where salt marshes lie behind them, in which case reclamation is not economically feasible.

Detailed studies and demonstration projects have been carried out in all locations of the northwestern coastal zone. Field investigations were made, sites were mapped and their ecological value was estimated, laboratory analyses were done and the costs and benefits of the proposed works were calculated.

The reclamation of maritime sand dunes can best be carried out through fixation by mechanical means as a first stage of protection, and afforestation as a second stage, for long-term protection. No agricultural activities on these poor, dry, shifting sand dunes, with less than 150 mm of annual rainfall, is possible in the first stage, except for figs on the low sand dunes in the eastern part of the region. Once the sand dunes are fixed and their soil conditions improve under the shade of the wood trees, other horticultural trees, or even forestry trees, can be planted.

Model for Dune Stabilization Program

No master plan for all stabilization projects can be given. Local conditions always have to be taken into account. Permanent stabilization is possible only by means of vegetation. Plants, however, will not grow without irrigation in hyper-arid areas such as the New Valley of Egypt, unless groundwater is very close to the surface.

Steps that are necessary for a stabilization program in an area where conditions allow stabilization by vegetation are as follows:

Initially a resource inventory has to be made in order to decide on the best and most economical steps to be taken. This inventory should give information on factors such as topography, types of dunes or moving sand, rates of movement,

hydrologic and climatic conditions, potential for plant growth in the dune area when sand movement has been stopped, availability of local material for fences, or the availability of labor in the area.

Once a plan has been made, the topography has to be changed to meet the needs of the subsequent program. Depressions will have to be filled up and steep slip faces will have to be changed to a lower slope, because otherwise vegetation will not grow there or fences will become difficult to construct.

Tracks have to be laid out for use by heavy equipment in the course of the work, and then the sand-collecting fences have to be set up. On the dunes this will most likely be done in a checkerboard pattern, unless there is only one wind direction during all seasons. Types of fences used and distances from each other will vary with local conditions. In most cases a fore-dune, or the installation of a three-fence system on the windward side of the project area will become necessary in order to stop further sand transport into the area. Protection of the flanks of the area should not be disregarded, in order to avoid damage from other than the principal wind directions. A system where rows of trees are used instead of fences is preferable.

Whereas fences are constructed to stop sand movement on the dunes vegetation will be used in the depressions between the dunes. At any rate, the interdunal areas have to be stabilized in order to cut off the supply to the dunes.

For large projects, a nursery will have to be set up, since as many as 200 trees may be needed per hectare. Seedlings have to be planted, together with fertilizer, and possibly some products that improve the soil structure should be applied to the planting holes. Irrigation will be necessary in most cases shortly after planting.

After the planting of the trees, seeds for the undergrowth of grasses and herbs will have to be brought out. This will speed-up the development of a dense vegetal cover, although experiments have proved that vegetation will also move into the area unaided once the sand movement has been brought to a halt.

Seeding can be combined with the application of a soil stabilizer. Test sprayings will be necessary in each case in order to determine the proper concentration of the emulsion. Water quality may influence the behavior of the stabilizers. Different soils may require the use of different dispersion agents. It should be observed that the areas treated with chemical soil binders must not be entered for at least the next six months. Therefore, all other work have to be done prior to spraying. A major item in all stabilization projects is the guarding of the project areas, both against accidental damage by people or animals and against destruction or theft.

Another factor that has to be observed for a new project is the proper season. Although there are variations because of climate, planting and seeding should be done between October and December in most dune stabilization projects. In this part of the year temperature and evaporation are relatively low, so plants will be able to develop a sufficient root system with minimal additional irrigation before the beginning of the hot season. The period from March to June should be avoided for seeding and planting, because severe winds are frequent in most of the regions under consideration during that season. This could mean mechanical damage to tender plants and high evapotranspiration rates.

In most projects the three months available for seeding and planting will be too short because of limitations set by equipment and manpower. Therefore, the project area has to be subdivided in units where work can be completed in one cool season. The size of the units that can be worked in one season may be increased, however, if seeding is already done in the hot season, as germination will not occur before the beginning of the cool season or before the application of irrigation water.

Techniques suggested for fixation and afforestation of the coastal dunes of northwestern Egypt:

First, a network of cross-fences should be prepared, the size of the squares formed depending on the steepness of the slope and the form of the dune. In some cases only linear fences set across the direction of the prevailing wind can be used. Dead material like reeds (Phragmites communis) or any other dry material available can be used. It is also advisable to use live material such as Sacharum Spontaneum, which grows very well on these maritime sand dunes. Fences should be sunk 15 cm and rise about 40 cm above the sand. With an average size of 7 x 10 m for each fence square, about 500 bundles of reeds are required per hectare, making the fence 3,000 m long. In places where intensive fencing is not economical, one or two rows can be located on the higher ridges of sand dunes or on the flat beach. When they are covered by moving sands, a new fence should be erected over the accumulated sand.

Once the sand dunes are fixed by mechanical devices, planting should be done with suitable drought-resistant tree species. Tall trees tolerant to strong winds should be chosen to secure a mixture of a soil ameliorator and quick protection. Mixing Acacia cyanophylla with Tamarix articulata in a proportion of 4:1 achieves these aims. In addition to these two very drought-tolerant species, the following can be tried; Zizyphus spina christi and Acacia cyclops on the higher part of slopes; Prosopis juliflora on medium sites, and Eucalyptus gomphocephala and Casuarina equisetifolias on depressions where the water table is accessible. Atriplex sp., Nitraria

retusa and Tamarix articulata, very salt-tolerant species, can be tried on the coast, at 150-200 meters from the sea.

Planting should be done at 60-120 cm depth, to maximize the moist bottom layers of sand. Big seedlings of about 1.2-1.5 m in height, grown in pots, should be used. Planting must be done very carefully, with only moist sand from the bottom of pits being used for filling the holes.

Other Dune Areas in Egypt

Coastal dunes occur along much of the delta coast. They are relatively stable and used as agricultural land, especially in the western region (Mariut area). It is irrigated by drawing water from the shallow layer of fresh groundwater resting on top of the salty ground water entering the ground from the sea. Overexploitation of this scant fresh water reserve could easily lead to increased intrusion of salt water, promoting the process of desertification.

Also, the Khanka dunes area near Cairo has been selected by the Desert Institute as a possible test site for stabilization methods best adjusted to the ecological conditions of the delta region.

The problem already exists in the eastern delta margin and has even been accentuated in response to the changes of Nile water discharge into the sea after construction of the Aswan High Dam. New dunes are developing from sands of recent beach deposits.

Increased economic activities in the region between Port Said and Suez are also leading to the mobilization of sands and to the formation of migrating dunes. Protective measures should be taken at an early stage of dune formation in order to avoid high costs in the future.

There are plans at the Desert Institute in Cairo to study the possibilities of sand stabilization by means of bacterial mucilage. This microbiological line of research might lead to new stabilization methods for areas where oiling is not advisable or ecological, as well as other grounds, but where there is also a lack of water for vegetal stabilization.

REFERENCES

- El-Baz, F. (1978) The Western Desert of Egypt--Its Problems and Potentials. Cairo Inter. Workshop--Appl. of Sci. and Tech. for Desert Development.
- Hagedorn, H., Giebner, K., Welse, O., Busche, D., and Grumert, G. (1977) Dune Stabilization. Geographisches Institut Universität Würzburg.

ECO-TECHNIQUE--MAGIC FORMULA FOR RECLAIMING DESERTIFIED
SEMIARID CROP LAND

By

Hans-Joachim Spaeth

ABSTRACT

In all semiarid regions summer fallowing is stimulating the process of desertification. So far, it is predominantly short-term economic considerations that keep this process alive.

Taking reclamation and conservation requirements as well as long-term ecological stability into consideration, destructive soil erosion must be avoided. This maxim, however, does not necessarily imply economical cutbacks. Rather, the application of selected and adapted eco-techniques coordinates both ecological stability and economical prosperity. Conservative minimum tillage, water harvesting, stubble mulching, application of crop varieties with a specific residue-share, and water use efficiency are such techniques. They interact with the interrelated factors "soil erosion," "soil moisture," and "yield" in such a way that a positive input into this ecological system yields a positive feedback and keeps the production potential at an ecologically and economically required level. The following is the agro-ecological dry boundary equation

$$F_{opt} = \frac{100 \times Re_{min}}{b \times d \times (100 - Ti)} - \frac{a + b \times c}{b \times d}$$

where F_{opt} = optimized moisture supply; Re_{min} = minimum residue rate for effective erosion control required at planting

time; T_1 = residue loss due to tillage between harvest and planting time in percent of pretillage residue rate; a = Y-intercept and b = slope of local grain/straw yield-regression equation; c = Y-intercept and d = slope of local available moisture/grain-yield regression equation. This formula provides a means by which any optional or required single eco-technique or combination of techniques can be qualified and quantified according to the local and/or regional ecological and economical setting.

INTRODUCTION

Based on the research results published in Spaeth (1980a and 1980b), which are also partly reflecting long-term research activity carried out by various branches of the United States Department of Agriculture and by the universities of the Great Plains States of North America, the following conclusions were reached:

- Vast areas of the semiarid zone are affected by a process that visibly destroys their agricultural potential.
- All observations indicate that soil erosion is initiated by inappropriate and excessive land use in general, and by exposure of topsoil in particular. This means overgrazing and clean tillage during the fallow season.
- So far, land use in semiarid regions has been economically aggressive, with unadapted or at least mismatched technologies. The ecological prime-principle of land use--soil conservation--has been neglected almost always and everywhere. This applies both for developed as well as for developing countries.

Such an inventory of waste and lost resources proves man's wrong and one-sided approach to dryland use. The bad news about droughts, dust storms, crop failures, and famine has been established with a cyclic recurrence in semiarid areas all around the globe. They suggest that we find a new approach to dryland use that is ecologically oriented, and aware of the need of the soil for protection and conservation, before it is too late and the remaining breadbaskets of the world are totally lost.

OBJECTIVES

The objectives of this paper are:

- To explain why summer fallow stimulates the process of desertification under present land-use systems

- To analyze the dynamic interactions between traditional summer fallow on one hand and the ecological model "soil moisture / soil erosion / crop yield" on the other hand
- To explain and quantify the purpose and the effect of selected eco-techniques on this model
- To develop a formula from the interactions of soil moisture, soil erosion, crop-yield and land-use practice, a formula that can serve as a tool in land-use planning and in reclaiming semiarid regions, and that is universally applicable and easily handled.

The central Great Plains of North America serve as a model problem area for this presentation. All findings and data are taken from Spaeth 1980a and 1980b, where a comprehensive list of references provides additional information.

ANALYSIS

Summer Fallow and Soil Erosion

During the spring season, crop and plant residue rates are at their lowest. Wind velocities are very high or even at their maximum. The clods in the upper soil horizon are disintegrated after the weathering process of the winter. These are the prerequisites for soil transportation, both by water and by wind.

Summer fallowing still implies repeated tillage in order to reduce evaporation and weed growth. As a consequence, the topsoil is loosened up and dries out, the remaining clods are destroyed, and the field surface is exposed and levelled by reducing the soil cover. This again further stimulates soil erosion during spring and summer. Table 1 shows the effect of wind erosion on the quality of the soil and provides a comparison with some of the ecologically relevant data gathered on eroded cropland and neighboring rangeland. The data represents conditions in western Kansas and eastern Colorado. P₁ stands for rangeland that has never been overgrazed for the past 25 years; the stocking rate has been 5 acres per head. P₂ and P₃ are taken phosphorus, and potassium (K) dominate in the noneroded range sites. Coarse sand, however, which is hard to carry away, dominates on both the eroded crop sites throughout the plow-horizon.

Interactions Between Summer Fallowing, Soil Moisture, Soil Erosion, and Crop Yield

Excessive erosion as well as the associated excessive deposition have an impact on the ecological quality of a given area. Eroded sites become more and more susceptible to droughts, as can be concluded from comparative interpretation

TABLE 1. Nutrient Content and Particle Size Combination on Cropland (P₂ and P₃) After More Than 25 Years of Wind Erosion Compared With Not Overgrazed Rangeland (P₁).

Distances: P₁ - P₂ = 20 m (65 ft), P₂ - P₃ = 80 m (260 ft)

X Scope: P₁ - P₂ = 0 - 1%, P₂ - P₃ = 0 - 1%

Profile Depth									
		0.002	0.05-	0.2-	2.0	0.M	Ca	P	K
	0.002	-0.05	0.2	2.0					
in cm	mm	Diameter of fine soil				%	ppm	ppm	ppm
	in %								
P ₁	0-15	32	32	22	11	3	4,0	4,600	27 360
	15-30	38	29	29	11	3	2,4	4,900	28 380
	30-45	41	28	21	8	2	1,7	5,600	14 460
	45-60	39	27	20	12	2	1,2	10,900	3 430
P ₂	0-15	38	30	14	11	7	2,9	5,200	21 340
	15-30	40	30	13	12	5	1,9	5,200	12 330
	30-45	39	30	20	9	2	1,3	7,500	9 390
	45-60	39	27	20	13	1	1,3	9,300	4 370
P ₃	0-15	38	31	16	10	5	2,1	4,800	21 320
	15-30	39	30	14	11	6	1,5	6,500	7 310
	30-45	37	30	19	11	3	1,3	7,300	4 390
	45-60	41	27	20	10	2	0,9	9,000	5 410

of moisture characteristic curves that are taken from such sites, as represented by Table 1. As time goes by, the content of fine soil particles is reduced in the plow horizon, and the organic matter content decreases. Both of these, however, are very important for storing moisture in the soil. Losing these two components, therefore, also means losing moisture storage capacity, and losing moisture, in the long run, ecologically means desertification.

Due to 25 years of erosional activity, loamy sand developed into a coarse sand. The moisture characteristic curves reflect shifted 1/3- and 15 atm-conditions. Consequently, the moisture budgets of sandy soils became more and more predominant in the plow-horizon on a regional scale. Plant-available moisture is reduced, and the production potential decreases as long as moisture is the key limiting factor.

The following two equations (1.1 and 1.2) describe how winter wheat yields relate to rates of available moisture; (available moisture defined as soil moisture given at planting time plus 100% of rainfall during vegetation period).

$$Y_{W-Kansas} = -15 + 2X \quad (r = 0.78) \quad (1.1)$$

$$Y_{E-Colorado} = -36 + 3.66X \quad (r = 0.83) \quad (1.2)$$

where Y = grain yield in bu/acre and X = available moisture in inches.

With regards to the necessity of summer fallowing, the Kansas-equation indicates that about 7.5 inches (= 190 mm) of available moisture are consumed for leaf and stem production before a single grain will be produced. From this marginal value, on a moisture rate of 1 inch (= 25 mm), about 2 bushels of wheat grain per acre will be produced.

Plant residue reduces surface runoff, wind impact, and evaporation rates, especially during the summer fallow season. High yield levels, therefore can help control erosion and influence the factor "soil moisture" by high rates of straw yield residue.

The relationship between winter wheat grain yield and straw yield in E-Colorado and W-Kansas is given in equation (2).

$$Y = -1561.5 + 158.8X \quad (r = 0.95) \quad (2)$$

where Y = straw yield in lb/acre and X = grain yield in bu/acre; valid for yield levels between 15 and 35 bu/acre.

Finally the influence of soil moisture content on wind erosion rates must be considered. Tests have been conducted on fine sands, sandy loams, silt loams, and silt clays in the wind tunnel at 30, 40, and 50 km/h measured at 15 cm above ground. For a complete test description see Spaeth 1979b,

p. 432. The general findings indicate that the mobility of the soil decreases with increasing content of hygroscopic moisture. It reaches zero in the vicinity of the permanent wilting point. This is because the formation of nonerrodible clods is a function of the soil moisture content; erodibility decreases with increasing content of nonerrodible clods over 0.84 mm in diameter and the formation of such nonerrodible clods is a function of the hygroscopic moisture content. This moisture content, in turn, grows with the fraction of soil particles smaller than 0.05 mm in diameter. This fraction, however, is also highly erodible, as shown above, and therefore holds the key position. The sandy soil in this test had the smallest content of these soil particles smaller than 0.05 mm in diameter. That is why its erodibility increased most rapidly and why it reached maximum values with growing moisture stress.

The ecological model in Figure 1 summarizes the dynamic interactions between traditional summer fallowing (which stand for "land use" in general) on one hand, and soil moisture, soil erosion, and crop yield on the other:

- Soil moisture regulates crop yield levels and erosion rates at any given site; reduced moisture rates reduce crop yields and increase erosion rates.
- Soil erosion in turn leads to a change in available moisture as well as in production potential; increased erosion reduces available moisture and crop production.
- The yield status finally bears the consequences of both the soil moisture and erosion.

The cycle of fundamental interactions in the erosion process, therefore, is closed. This erosion process is initiated and stimulated by the factor "land use"; once this process is in action it can only be stopped directly, by controlling soil erosion, or indirectly, by improving the moisture status in the topsoil and thus reducing erodibility.

Eco-Techniques

The key question of this paper--can eco-techniques really save the breadbasket of the world by stabilizing its budget of resources?--seems to be rather provocative in the face of worldwide desertification of semiarid regions. Nevertheless, the following text will explain in brief the purpose and the effect of some selected eco-techniques and will lead to an encouraging answer.

Eco-techniques in agriculture, by definition, are techniques that influence the local or regional ecological setting, a single ecological factor, or a whole ecological system in a positive way. In other words: The application of

Figure 1
Impact of land use practices (l μ m prac) on the system "soil
moisture/ soil erosion /crop yield." After Spaeth (1980b).

eco-techniques in dryland agriculture implies technological planning guided by ecological maxims such as:

- Conservation and reclamation of soil
- Avoidance of resource waste
- Application of a philosophy of growth, strictly based on the supply of renewable resources.

The objectives of any such techniques are:

1. To increase available moisture
2. To reduce and control erosion by water and wind
3. To maximize and stabilize yield levels.

All three objectives in turn cause positive chain reactions in the ecological system--soil moisture/soil erosion/crop yield.

Moisture storage can be effectively achieved by wind barriers of various materials; the unanimous results of a large volume of research show a reduction of wind speed, additional accumulations of snow, a relative increase in soil moisture, which in turn results in an additional yield increase. The shorter the barrier intervals, the greater the benefits. In areas without any snowfall, wind barriers help at least to control wind impact on field surfaces.

Soil moisture can also be optimized by installing terrace systems or water-spreading systems, as is currently under study in various parts of the world. The objective is to divert runoff from unproductive areas to fields for crop production. Another objective is to reduce runoff from such productive areas. Small changes of the land surface--such as the installation of grassed waterways, diversion ridges, level-pan areas, and terraces of various kinds--have demonstrated the potential of harvesting moisture for increasing yields dramatically under dryland conditions. Thereby, residue rates also increase and provide additional vital means for erosion control and evaporation reduction.

The benefits of a combination of micro-relief-alterations plus the application of plant residue in a fashion that is especially suitable for countries where farming is more or less a matter of manual labor, have been described by Fairbourne and Gardner (1974). Their findings show that especially microwatersheds with vertically installed linear mulches, but also to a lesser degree microwatersheds without these mulches, have proven to be very efficient and cheap alternatives for conventional flat-field surfaces when the optimization of the soil moisture budget is at stake.

It has been shown that an increase in plant residue as a ground cover tends to improve the soil moisture budget considerably. At Akron, Colorado, additional moisture gains lead to a yield increase of 5 bu/acre, at North Platte, Nebraska,

of 4 bu/acre, over the plots with the lowest residue rates. Compared to the regional average yield of 30 bu/acre these 5 and 4 bushels mean an improvement of 17 and 13% respectively.

Yet another phenomenon is shown by this data of Table 2. From mid-June to September a certain moisture loss occurs at North Platte, no matter how dense the ground cover may be. This is an example of average annual precipitation. At Akron, an example of above-average precipitation showed that only minor moisture gain occurred during this late-summer period. We conclude that we have to look for a means to reduce evaporation losses during the second half of the summer fallow period. Precision-timing of tillage activity after rainfall incidences is one approach to solving this problem. Related laboratory tests show that no tillage at all results in the highest evaporation losses. The earlier the timing of tillage activity after evaporation is initiated and the capillary flow is in action, the more substantial are the savings potentials. These laboratory tests, however, still have to be repeated under actual field conditions in order to determine the optimal timing for various types of field activity on various types of soil after rainfall, as well as the optimal depth of tillage (see Spaeth, 1976, p. 88).

In this context, strip farming and contouring should be mentioned. These are well-established techniques in some dryland areas. Nevertheless, they have hardly become a commonly adopted practice.

These tests and findings outline the capacity of selected eco-techniques, which could be applied individually or as a package in order to maintain or even increase the production level of semiarid croplands.

One example will summarize the benefits of an ecologically reasonable approach to dryland use. The example station is Ft. Collins, Colorado, with approximately 12 inches of average annual rainfall. So far, a dryland winter wheat-summer fallow-winter wheat-rotation has been established under the guidelines of the aspiration for "profit in land use". 25 to 30 bu/acre is the average yield per harvest. Blowing fields can be found almost every spring. Moisture harvesting capacity reaches 30% during the 14-month fallow season. This percentage can be boosted dramatically if some of the aforementioned methods were to be adopted. 60% moisture storage efficiency already may be within the reach of our current eco-techniques. According to Spaeth (1980a), about 2.9 inches (75 mm) of moisture are available in the rootzone (180 cm deep) for silt loam at the end of the 20th decade (= harvest time) at Ft. Collins. By applying the eco-techniques described above, moisture content in the rootzone will have increased to 4.87 inches (123.8 mm) by the 25th decade of the very same year (= planting time). Accounting for 80% of the rainfall that is falling during the vegetation period, a total of 14.2 inches

(360.6 mm) of moisture will be available for wheat production. Using this moisture in equation 1.2 above, a wheat yield of 15.36 bu/acre can be calculated. Note: Here only 80% of total rainfall that occurs during the vegetation period has been accounted for, whereas calculations of the U.S.D.A. research stations are based on 100. The 15.36 bushels per acre calculated above are the yield of a continuously cropped wheat-system, a yield that compares very favorably, even on economical grounds, with the 30 bu/acre of the wheat-fallow rotations of the present time, which also are associated with all the ecological disadvantages outlined in this paper. This example reflects the potential of eco-techniques, which ultimately could make ecologically advantageous, continuous wheat-systems even economically attractive in some of the areas presently summer fallowed. In areas with no options for continuous cropping, wheat fallow rotations and flexible cropping systems could be established on an improved production level due to improved moisture conditions.

The Agro-Ecological Dry Boundary--Formula

If soil reclamation and conservation are to be the key principles of future dryland use, then (regardless of all other aspects of land use) the complete absence of destructive soil erosion has to be the principal criterion of an ecological redefinition of the dry boundary of rain-fed agriculture, which has been called the "agro-ecological dry boundary" by Spaeth (1980a and 1980b). Consequently, an equation for determining the location of this dry boundary has to be deduced from the ecological system Soil Moisture / Soil Erosion / Crop Yield, which is basic to dryland farming.

The ecological system in turn is composed of the following chain of relations:

1. Soil type / soil moisture
2. Available moisture / wheat grain yield
3. Grain yield / straw yield
4. Minimum straw rate required for effective erosion control
5. Tillage type / residue rate remaining on the field surface at end of the fallow season.

The relationship between grain yield K (in bu/acre) and available moisture F (in inches) is given in equation (1.1) and (1.2). In general terms

$$K = c + d \cdot F \quad (1)$$

The relationship between straw yield Re' (in lb/acre) and grain yield K (in bu/acre) is given by equation (2) above. In general terms

$$Re' = a + b \cdot K \quad (2)$$

Ti (in %) is the factor by which tillage operations reduce the originally produced residue rate Re' during the fallow season. For example, a moldboard plow reduces 100% of Re' during a single run, a tandem disk by 50%, a chisel plow by 25%, and a rodweeder by 10% of the residue available on the field surface before the tillage operation.

Re is the residue rate remaining on the field surface in an upright position at the end of the fallow season.

Re' from equation (2) can be rewritten as equation (3)

$$Re \cdot 1 + \left(\frac{Ti}{100-Ti} \right) = a + b \cdot K \quad (3)$$

Spaeth (1980b, p. 227) describes this transformation in detail.

To combine (1) and (3), and solve for F leads to equation (4)

$$F = \frac{100 \cdot Re}{b \cdot d(100-Ti)} - \frac{a + b \cdot c}{b \cdot d} \quad (4)$$

Re will now be replaced by a minimum rate of standing stubble, Re_{min} , which is required for effective erosion control at planting time if no other erosion control is being practiced.

If the moisture supply of an area can be optimized, e.g., by using wind breaks or any other eco-technique, F from (4) will be replaced by F_{opt} .

The agro-ecological dry boundary is given at places where optimized moisture supplies are adequate for producing a required minimum rate of residue under a certain tillage system. Accordingly, the agro-ecological dry boundary is given where (5) is applicable.

$$F_{opt} = \frac{100 \cdot Re_{min}}{b \cdot d(100 Ti)} - \frac{a + b \cdot c}{b \cdot d} \quad (5)$$

Application of the Dry Boundary--Formula

One local example from eastern Colorado will demonstrate the use of this formula:

Let Ti be 30% (this figure is related to a winter wheat - fallow - rotation with minimum tillage corresponding to three operations with a sweep 7 cm or larger).

Let Re_{min} be 1,250 lb/acre (this rate is required on silt loam). a, b, c, and d have already been specified above.

$$F_{opt} = \frac{100 \cdot 1,250}{158.8 \cdot 3.66 (100 - 30)} - \frac{-1,561.5 + 158.8 (-36)}{158.8 \cdot 3.66}$$

$$= 15.59 \text{ inches } (= 396 \text{ mm})$$

If a site can produce this optimized moisture rate, it lies within the area recommended for dryfarming; if F_{opt} is smaller than 15.59 inches, the site lies beyond the ecologically justified dry boundary of non-irrigated agriculture and it should be reseeded to grass. It would have to stay in grass until a balance between F_{opt} and the right side of equation (5) can be achieved. This can be done by changing tillage practices and by reducing the Ti -factor, or by reducing Re_{min} rates by adding wind breaks or other additional erosion control structures. It can also be achieved by taking other crop varieties with greater straw proportion, or another moisture / grain-yield relationship.

This means that equation (5) should also be solved for the factor Ti so that all fallow operations will have to be adjusted to a desired or calculated Ti -value, or that equation (5) is solved for Re_{min} such as can be produced under local conditions. Planning problems of plant breeders finally will be answered by solving (5) for a, b, c, or d, respectively.

Due to the character of the formula, there can be no regional limitation to its applicability. The ecological model (Figure 1), which is basic to the formula, is valid in any agricultural system region. It is of no importance whether an area has a semiarid mediterranean type of climate with mainly winter rainfall, a semiarid continental climate with up to 80% of total annual precipitation occurring during the growing season, or a semiarid tropical climate with a rainy season of various duration the summer. If there is a coincidence of sufficient moisture supply and a tolerable temperature range, allowing a crop to burgeon and to mature in a semiarid area, then all planning problems in land use will be reduced to the question: Can enough residue be produced along with a crop as to prevent destructive soil erosion between harvest and planting time? If this cannot be achieved, any agricultural land use will ultimately lead to the problem of desertification.

In developing countries with a limited supply of climatic and agronomic data, the key relationships and the parameters basic to this formula have to be extrapolated from climatically similar regions. After one year of work, field tests

can provide sufficient data for a first approximation of actual local functions, and continued research in the area will continuously adjust these preliminary quantified key relationships to the actual local ecological setting.

REFERENCES

- Albrecht, W. A. (1974) Physical, Chemical, and Biochemical Changes in the Soil Community. In: Thomas, W. L., Jr. (ed.): Man's Role in Changing the Face of the Earth. Vol. 2, University of Chicago Press, p. 648-673.
- Burrows, W. C., Reynolds, R. E., Strickler, F. C., and Van Riper, G. E. (1970) International Conference on Mechanized Dryland Farming. Proc. of an International Conference on Mechanized Dryland Farming sponsored by Deere & Co. under the auspices of the Food and Agriculture Organization of the United Nations, Moline, Ill.
- Evans, C. E. and Lemon, E. R. (1957) Conserving Soil Moisture. USDA Yearbook 1957, p. 340-359.
- Fairbourn, M. L. and Gardner, H. R. (1974) Field use of Microwatersheds With Vertical Mulch, Agron. J. 66, p. 740-744.
- Falkner, F. R. (1938) Die Trockengrenze des Regenfeldbaus in Afrika. In: Petermanns Geogr. Mitt. H. 7/8, p. 518-523.
- Falkner, F. R. (1939) Beitrage zur Agrargeographie der afrikanischen Trockengebiete. Geographische Abhdl. 3, Reihe, H. 11, Stuttgart.
- Fenster, C. R. and McCalla, T. M. (1970) Tillage Practices in Western Nebraska With a Wheat-Fallow-Rotation. Nebr. Agr. Exp. Station Bul. 507.
- Hanway, D. C. (1970) Cultural Practices in Mechanized Dryland Farming. In: Burrows, W. C. et al. (eds.): International Conference on Mechanized Dryland Farming. Moline, Ill., p. 102-122.
- Heyne, E. G., Smith, F. W., Hobbs, J. A., Stickler, F. C., Anderson, L. E., and Wilkins, H. D. (1964) Growing Wheat in Kansas. Kansas Agr. Exp. Station Bul. 463.
- Johnson, W. G. (1964) Some Observations on the Contribution of an Inch of Seeding-Time Soil Moisture to Wheat Yields in the Great Plains. In: J. of Agronomy 56, p. 29-35.
- Johnson, W. C. and Davis, R. G. (1972) Research on Stubble Mulch Farming of Winter Wheat. USDA Cons. Research Report 16, p. 1-31.
- Kuska, J. B. and Mathews, O. R. (1956) Dryland Crop Rotation and Tillage Experiments at the Colby Kansas Branch Experiment Station. USDA Cir. 979.

- Mathews, O. R. (1954) Implements and Methods of Tillage to Control Soil Blowing on the Northern Great Plains. USDA Farmers' Bul. 1797.
- Penck, A. (1910) Versuch einer Klimaklassifikation auf physiographischer Grundlage. Sitz.-Ber. 1 Preuss. Akad. d. Wiss., Berlin, Pd XII.
- Spaeth, H.-J. (1976) Problems of Optimal Utilization of Dry Stepes with Cold Winters. Soil Erosion and Soil Moisture Balance in Dry Steppe Climates with Cold Winters - As Demonstrated by the Example of Central Anatolia. In: Natural Resources and Development, v. 3, p. 72-92.
- Spaeth, H. J. (1977) The Konya-Cumra-Project in Turkey. A Study of the Problems of Irrigated Agriculture in Dry Steppe Climates With Cold Winters. In: Applied Science and Development, v. 10, p. 122-145.
- Spaeth, H.-J. (1978) Water Balance and Water Management in the Beysehir-Konya-Basin, Southwest-Central Anatolia. In: Applied Science and Development, v. 12, p. 96-110.
- Spaeth, H.-J. (1979a) Agronomic Problems in Designing Water Erosion Control Structures in Semiarid regions--as Demonstrated by the Example of Libya. In: Applied Science and Development, v. 13, p. 27-64.
- Spaeth, H.-J. (1979b) Zur Funktion der Bodenfeuchte im Erosions - und landwirtschaftlichen Produktionsproze der zentralen Great Plains (USA). In: Verhandlungen der Gesellsch. f. Ökologie (Munster 1978) Band VII, p. 429-435.
- Spaeth, H.-J. (1980a) The Agro-Ecological Dry Boundary of Nonirrigated Agriculture in the Central Great Plains of North America. On the Redefinition and the Mobility of the Dry Boundary of Rain-fed Agriculture. Steiner Verlag, Wiesbaden, 160 p.
- Spaeth, H.-J. (1980b) Die agro-ökologische Trockengrenze. In: Erdkunde Bd. 34, p. 224-231.
- Thorntwaite, C. W. (1931) The Climate of the Earth. Geogr. Rev. Bd. 21, p. 633-655.
- Throckmorton, R. I. and Meyers, H. E. (1941) Summer Fallow in Kansas. Kansas Agr. Exp. Station Bul. 293.
- Woodruff, N. P. and Siddoway, F. H. (1965) Wind Erosion Equation. In: Proc. Soil Sc. Soc. Amer. 29, p. 603-608.
- Zingg, A. W. (1951) Evaluation of the Erodibility of Field Surfaces With a Portable Wind Tunnel. In: Soil. Sc. Soc. Amer. 15, p. 11-17.

MONITORING THE SAND-MOVING DESERT WINDS

By

Farouk El-Baz

ABSTRACT

In the desert, running water plays no significant role in the transportation and deposition of sediment; the desert environment is aeolian, being dominated by the wind. Thus, it is very important to understand the wind patterns of any desert region under investigation. However, in most deserts meteorological stations are placed in oases, which are located in depressions. Cliffs and scarps that surround these depressions exert a marked influence on wind direction and velocity. Additionally, many of the desert oases, and therefore the meteorological stations, are separated by great distances, sometimes exceeding 600 km. The result of these two factors is that the surface wind data are usually scant and often reflect local influences rather than the wind regime in the open desert.

Studied examples of the effects of local topography on the orientation of wind deposits in the Western Desert of Egypt include: the Kharga depression, where the major scarp is aligned close to the direction of sand transport; and the Farafra depression where a scarp is oriented nearly normal to the direction of the wind. In both cases, sand is deposited in dunes whose axes are parallel to the scarps rather than to the dunes in the surrounding open desert. Thus, it is important to consider local topographic effects when comparing wind directions and dune orientations, particularly in the absence of wind stations in the open desert above the depressions.

Space-age technology has paved the way for the utilization of automated stations that can gather meteorological data in remote desert regions. Data collected by such means can be transmitted to orbiting satellites, which then retransmit the data to ground receiving stations for processing, distribution, and analysis. There are three basic elements to such a scheme: (1) a data collection platform connected to the sensor recorders; (2) a radio transponder with receiving and transmitting capabilities on board a satellite; and (3) a data receiving station for retrieval, processing, and dissemination of the collected data.

At the present time, the U.S. Geological Survey is operating a program of remote monitoring of hydrologic data for assessment of water resources. Results of this program indicate both the cost-effectiveness of the method, and the advantages gained from real-time collection of the data. In one case, improved estimates of runoff were instrumental in avoiding flood damage and generating extra electrical energy during the runoff period. The same technology is being tested for meteorological data collection. Two experimental programs are being conducted in remote places that represent extremes of Earth weather; one in Antarctica and the other in the southwestern U.S. deserts.

The "Desert Winds Project" of the U.S. Geological Survey of Flagstaff, Arizona, aims at establishing automated weather stations capable of monitoring critical environmental factors that determine the capability of wind to erode and transport various types of materials in deserts. Stations now under study include sensors for measuring wind speed, wind direction, precipitation, air temperature, humidity, soil temperature, and barometric pressure. The stations are powered by solar panels and their data are stored in the collection platform prior to transmission to the GOES-1 satellite via an antenna. Results from these stations indicate that this method is exceedingly practical and should be utilized for the cost-effective and timely collection of meteorological data in remote desert regions.

INTRODUCTION

Running water does not play a significant role in the transportation and deposition of sediment in the extremely arid desert environment. Wind is the main force of erosion in the deserts. The wind carries enormous quantities of debris, which are deposited where its path is disturbed by topographic impediments.

On small scales, the wind often changes directions and forms vortices that cause air to flow in the reverse direction. Where furrows exist in the terrain, the wind is funneled through them with an increase in velocity, and thus in erosive

power. On larger scales the wind tends to intimately follow the contours of slopes and faces of scarps and cliffs.

In most countries meteorological stations are limited to locations in and around cities. In desert areas particularly, meteorological stations are widely spaced and are placed in oases, which are located in depressions, where surrounding cliffs exert a great deal of influence on the wind. These facts have to be taken into account when analyzing wind data from any desert region. Examples will be given here from studies of the general wind circulation and local wind regimes in the Western Desert of Egypt.

THE CASE OF THE WESTERN DESERT

Dune Orientation

The Western Desert of Egypt extends from the Libyan border on the west to the Nile River and from the Mediterranean Sea southward to the border with Sudan. The desert occupies 681,000 km², more than two-thirds of the total area of Egypt. The region consists of sandstones and limestones that dip gently to the north. These units have been eroded to an aeolian peneplain, veneered with windblown debris. The remarkable flat terrain is broken by numerous escarpments that bound depressions, by scattered inselbergs, and in the southwest corner, by the Gilf Kebir plateau and the isolated, circular mountains centered around Gebel Uweinat (El-Baz et al., 1979a).

Much of the basic research on dune classification and sand movement by wind has been carried out in this desert, where "the free interplay of sand and wind has been allowed to continue for a vast period of time, and (where) if anywhere, it should be possible in the future to discover the laws of sand movement and growth of dunes" (Bagnold, 1933, p. 121). Bagnold's (1941) classic treatise on the physics of sand movement by wind is based largely on observations made in the Western Desert. Sand flow patterns in the Sahara, including western Egypt, have been mapped by Wilson (1971) from the "sand storm resultants" and from barchan dune orientations as shown on topographic maps.

A Landsat mosaic of Egypt was made to study the regional patterns of large-scale aeolian features in the Western Desert (El-Baz et al., 1979b). Mapping of dune accumulations based on these images resulted in the sand distribution map shown in Figure 1. In addition to the numerous dune bundles in the desert, the map shows one of the largest accumulations of dune sand in North Africa, the Great Sand Sea. This sand sea occupies the west central part of the Western Desert beginning just south of the Siwa depression and continuing uninterrupted for 600 km to the Gilf Kebir plateau. The main sand mass con-

Figure 1
Distribution of sand dunes in the Western Desert of Egypt
(after El-Baz et al., 1979a).

sists of subparallel arrays of enormous compound linear dunes. These are broad "whaleback" dunes surmounted by sharp-crested linear or seif dunes (Bagnold, 1941). The whalebacks are relatively static, whereas the linear dunes, much like the rest of the dunes in the Western Desert are actively migrating southward.

The Wind Regime

Wind data on this desert (as most other desert regions) are available from the U.S. National Climatic Center (Asheville, North Carolina, U.S.A.) in the form of U.S. Air Force N-type summaries. The summaries are usually produced from observations made by local stations under the auspices of the World Meteorological Organization (WMO). From these data an annual summary of all winds was calculated for six stations in the six Western Desert oases (Figure 2). The wind directions did not exactly fit the dune orientations.

It was assumed that the wind stations in the six oases may not be enough to convey a reasonably complete picture of the wind regime in the whole desert (Wolfe and El-Baz, 1979). Data were summarized for 42 meteorological stations covering a much larger area (15° to 35° N and 15° to 41° E) including all of the stations in Egypt and parts of the surrounding countries (Figure 3).

Analysis of these wind data resulted in calculations of the direction and relative amount of sand migration or drift. The amount of sand drift depends on the average wind velocity (measured at 10 meters height in most stations), and the impact threshold wind velocity for a given period of time. The value reflecting the sand moving capacity of the wind for a given time is known as the sand drift potential, and is numerically expressed in vector units (McKee et al., 1977). The direction of the vector resultants of drift potentials from 16 compass directions is known as the resultant drift direction, and the magnitude of the same as the resultant drift potential. Circular histograms, or sand roses, representing the potential effects of winds show graphically the directional groupings or modes of the effective, sand-moving winds (Fryberger, 1979). Accordingly, a summary of resultant sand drift potential was calculated for all 42 stations (Figure 4). Although some of the patterns resemble the dune orientation patterns, there were still differences between the two.

In such regions where meteorological stations are widely spaced, the patterns of drift potential may be difficult to interpret. Sand drift potential streamlines were interpolated by further computations. This was done by overlapping a rectangular grid on the region and computing the direction and magnitude for each grid point (El-Baz and Wolfe, 1980). This

Figure 2
Annual summary of all winds in the six oases of the Western Desert of Egypt.

Figure 3
Location map of 42 wind stations in North Africa and the Middle East from which data were analyzed (after El-Baz and Wolfe, 1980).

was accomplished by assuming that the value at each grid point equals the average of the nearest value of a number of stations weighted by the inverse of their distance from that point. The generated streamlines give a general flow pattern (Figure 5) for comparison with data obtained from measurements of orientations of dunes (Figure 1). Comparison between the directions of generated streamlines and the dune orientations still showed differences. These differences are believed to be due to the local topographic effects on the winds as measured at the meteorological stations.

Effects of Topography

In the Western Desert of Egypt, there are numerous examples of the control of local topography on alignment of dunes and other wind deposits. The examples given here are in the depressions of Kharga and Farafra.

Kharga depression

The Kharga Oases occur within a depression, the only one in the Western Desert that is aligned in a north-south direction. The depression is open to the south and southeast. The width of the depression varies from 20 km in the north to 80 km in the south. The oases in the Kharga depression occur along a line that trends in a north-south direction (N10°W). This line is a major fault, which allowed the settlers to reach underground water that used the fault as a channel for upward movement. This fault is parallel to the eastern scarp of the Kharga depression, which varies between 100 and 300 m in elevation above the depression floor.

The sand atop the high plateau north of the Kharga depression accumulates in broad dunes of the linear type. The prevailing sand-carrying winds are towards the southeast (range from S32°E to S25°E) as indicated by the orientation of parallel lines of corrasion features and sand deposits (Figure 1). As these broad dunes reach the scarp, the sand creates patterns that are reminiscent of waterfalls. After the sand is carried south of the scarp, it forms small and relatively fast-moving crescentic or barchan dunes. Axes of the dunes are oriented in a N10°W, which is the same orientation of the scarp of the Kharga depression. This direction of orientation differs at least 15° from that of the corrasion features and sand deposits on the plateau above the depression. This difference is believed to be directly related to the control of wind direction by the scarp bounding the Kharga depression.

Farafra depression

The Farafra Oasis is located in a large triangular depression in the west-central part of the Western Desert of

Figure 4
Annual summary of resultant sand drift potential at the 42 stations in North Africa and the Middle East shown in Figure 3 (after El-Baz and Wolfe, 1980).

Figure 5
Streamlines of annual summary of resultant sand drift potential developed by trend surface analyses of data shown in Figure 4 (after El-Baz and Wolfe, 1980).

Egypt (Figure 1). It is bounded by scarps on the northern and eastern sides. The depression gradually widens until it meets the Dakhla depression 200 km to the south. El-Guss Abu-Said plateau west of the Farafra Oasis is oriented N45°W with its long axis measuring 63 km and its width about 28 km.

The Great Sand Sea west of the Farafra depression is composed of linear dunes on top of large, gently sloping, whale-back dunes with orientations that vary between N-S and N26°W; the average orientation is N13°W (Manent and El-Baz, 1980).

Between El-Guss Abu-Said plateau and the northern scarp of the depression there is a small dune field (A in Figure 6). Within this dune field the dunes are oriented N52°E. This orientation is vastly different from that of the Great Sand Sea dunes. It is similar to the resultant direction of orientation (N52°E) of the Farafra depression (N58°E) and of El-Guss Abu-Said plateau (N45°E). West of this area is a smaller dune field that is oriented in a N6°E direction (B in Figure 6). This field is surrounded by two scarps, one oriented N50°E and the other N38°W. The resultant orientation direction of the two scarps is N6°E, which is identical to the orientation direction of the dunes in the field (Manent and El-Baz, 1980).

These examples prove beyond doubt that local topography exerts a great influence on wind direction within depressions in the Western Desert of Egypt. Therefore, wind data measured at oases in similar deserts cannot be taken as representative of the general wind circulation.

AUTOMATED MONITORING

General Principles

Because of the effects on wind direction of local scarps and cliffs surrounding desert oases, there is a need to place meteorological stations in the inhospitable, inaccessible, and wind-blown terrain of the open desert.

Space-age technology has paved the way for the utilization of automated stations that can gather meteorological data from such remote areas. Data collected by such platforms can be transmitted to orbiting satellites, which then retransmit the data to ground receiving stations for processing, distribution, and analysis. A great number of satellites can receive and retransmit data collected on the ground. These include satellites of the Landsat type, Nimbus, GOES, Tiros, Meteosat, and the communications satellites Comsat, Intelsat, and French satellites of the Anik type.

At the present time, the U.S. Geological Survey is operating a program of remote monitoring of hydrologic data for assessment of water resources (Paulson, 1978). Results of this program indicate both the cost-effectiveness of the

Figure 6

Map of dune orientations in the Farafra depression in the central part of the Western Desert of Egypt. Dunes in fields A and B are oriented in a different direction from that of the Great Sand Sea dunes (after Manent and El-Baz, 1980).

method, and the advantages gained from real-time collection of the data. In one case, improved estimates of runoff were instrumental in avoiding flood damage and generating extra electrical energy during the runoff period.

Along the same lines, a meteorological data collection scheme may use Earth-orbiting satellites to relay data from a great number of stations to one or more receiving stations. There are three basic elements to such a scheme: (1) a field radio, usually called Data Collection Platform, that is connected to the sensor recorders; (2) a radio transponder with receiving station for retrieval, processing, and dissemination to investigators (Figure 7). Two programs are underway that utilize such a system in monitoring weather. These are for use in remote places that represent extremes of Earth weather; one in Antarctica and the other in the southwestern U.S. deserts.

The first one is the Automatic Weather Station (AWS), which uses a transportable unit consisting of a 3-meter triangular tower, environmental sensors, a data acquisition and radio transmitter unit, an omnidirectional antenna, and a power source. This AWS, at Asgard Station in Antarctica, was designed by members of the Radioscience Laboratory at Stanford University (T. Howard, pers. comm.). In the near future, the AWS will transmit sensor data to the ARGOS system aboard the TIROS-N satellite. The 400-Mhz radio antenna, temperature sensor, and wind speed/direction monitor sit on a horizontal spar atop the tower. The microprocessor-based data acquisition unit, radio transmitter, and pressure sensor are housed in an insulated steel enclosure mounted at about the mid-level of the tower. Power for the AWS is supplied by either a radioactive thermo-electric generator (RTG) or by lithium batteries. The RTG units can power the AWS for many years and the lithium battery supply, as presently used, can power the AWS for longer than two years.

The second case is that of the "Desert Winds Project" of the U.S. Geological Survey in Flagstaff, Arizona. In this case stations consist of an aluminum tripod with meteorological sensors as described in detail below.

The Desert Winds Project

Numerous data acquisition systems are being developed to acquire and store information in a number of analog signals for later evaluation. For example, Hochberg (1979) describes a low power system primarily developed for "wind prospecting" at remote locations. This system is designed to work with standard U.S. Weather Bureau (Type F-420-C) wind speed and wind direction transducers and two temperature sensors in its minimal configuration (Figure 8). The use of a standard transducer enables realistic comparisons to be made with existing weather records using the same standard. Among its advantages are its

Figure 7
Sensor components of an idealized station for automated meteorological monitoring of desert regions.

Figure 8
Typical fixed installation of a wind energy instrumentation system (after Hochberg, 1979).

reasonably low cost and its ability to function "for an indefinite period of time on one standard inexpensive battery" (Hochberg, 1979, p. 17).

The Desert Winds Project of the U.S. Geological Survey uses the available technology for a real-time data acquisition system that is specifically designed for the study and analysis of desert winds (McCauley et al., 1979). The objectives of the project include the establishment of satellite-relay meteorological stations (Data Collection Platform), capable of monitoring critical environmental factors that determine the capability of wind to erode and to transport various types of materials in deserts. Specifically, the aim is to quantify the relative efficiencies of wind versus running water as geologic agents at these stations. The planned studies are expected to provide a quantitative climatic data base. The long-term objective is to establish a better understanding of desert processes, including the possible effects of past and future climatic changes in desert (McCauley et al., 1979).

The remote meteorological station undergoing testing at the U.S. Geological Survey in Flagstaff (Figure 9) includes sensors for wind speed, wind direction, precipitation, air temperature, relative humidity, soil temperature, and barometric pressure. It is powered by solar panels. Data are stored in the collection platform within an enclosure, and transmitted to the GOES-1 satellite via an antenna. A U.S. Geological Survey poster (McCauley et al., 1979) describes these units as follows:

"- Wind Speed - The wind speed sensor is a three-cup anemometer. The metal cup assembly includes a permanent magnet that operates a sealed magnetic reed switch. As the cup assembly rotates, it closes the reed switch. The frequency of the switch closures is proportional to the wind speed. The anemometer is mounted on the crossarm at the top of the 20-foot mast.

"- Wind Direction - The wind direction sensor is a light-weight airfoil metal vane. The vane is coupled to a precision microtorque potentiometer for low-threshold operation. The wire-wound potentiometer produces an output that varies proportionally to the wind direction. The wind direction vane is mounted on the crossarm at the top of the 20-foot mast.

"- Precipitation Sensor - The precipitation sensor measures precipitation in the form of rain. It uses an 8-inch-diameter orifice to collect the water, which is directed through a funnel to a tipping bucket mechanism. Coupled to the tipping bucket mechanism is a mercury switch that momentarily closes each time the bucket tips.

Figure 9
The test model of the machine used by the Desert Winds Project
of the U.S. Geological Survey, Flagstaff, Arizona.

When 0.01 inch of precipitation has been collected, the bucket tips over, draining the collected water out the bottom of the gauge and placing the second bucket into position to collect more. The sensor is placed on the corner of the platform 6 feet above the ground.

"- Air Temperature - The air temperature sensor is a solid-state linear thermistor and precision resistor network, placed in a shockproof 3/8-inch-diameter stainless steel housing. The sensor is positioned in a vane-aspired radiation shield to reduce the effects of solar radiation upon the temperature data. The output is a resistance proportional to the ambient temperature. The sensor is positioned at 5.5 feet from the ground surface.

"- Relative Humidity - The relative humidity sensor is a thin-film polymer capacitor contained in a protective housing with a 216- m sintered bronze filter. Water vapor is absorbed into the polymer after the vapors pass through the upper metal electrode. The result is that the capacitance changes linearly as the moisture increases or decreases. The sensor is positioned on a crossarm approximately 4 feet above the ground.

"- Soil Temperature - The soil temperature sensor is a solid-state linear thermistor and precision resistor network placed in a 1 x 6-inch dowel. The output of the sensor is a resistance proportional to the ground temperature. Arrows indicate placement of sensor and cable route. The sensor is placed in the ground at a depth of 6 inches.

"- Barometric Pressure - The barometric pressure sensor is a sensitive aneroid barometer that provides a resistance output proportional to barometric pressure. It uses an evacuated bellows that is sensitive to changes in absolute pressure. It is housed inside a weatherproof enclosure to protect the sensor.

"- Solar Panels - The power source consists of solar panels that are used to recharge two gel-cell batteries inside the enclosure. The panels receive direct sunlight and provide a charge with the use of regulators. They are positioned on a horizontal arm of the structure 6 feet above the ground.

"- Enclosure - The enclosure contains the Convertible Data Collection Platform (CDCP), Signal Conditioning Unit (SCU), barometric pressure sensor, SCU/interface chassis, and batteries. The CDCP is a complete microprocessor-

controlled data acquisition system. It turns on the sensors at programmed times, accepts sensor data, and converts the voltage output of the analog sensors into digital equivalents. It then stores the data and transmits them at programmed times. The SCU is the housing for the sensor signal-conditioning modules and sensor cards that provide outputs scaled from 0 to 5 volts analog over the dynamic range of each sensor.

"- Antenna - The antenna is specifically designed for use as a transmitting antenna to the GOES satellite. It consists of two quadrature-phased cross-element yagi antennas mounted on a common boom. The antenna is mounted on the tower on the extended vertical section of the basic structure."

RECOMMENDATIONS

There is no doubt that it is necessary to interpret and analyze data on wind directions and velocity in order to understand and possibly predict the rates and directions of transport of particulate material in desert environments. However, it is clear from the above discussion that there are presently severe limitations to such interpretations. These situations may be remedied by the following:

1. Careful analysis of the wind data, keeping in mind such important factors as the control by topography of the wind direction.
2. Increase of the number of wind stations and placing some in the open desert to monitor the environments of both erosion and deposition by the wind.
3. Utilization of available technology in either the automated recording of the data or their transmittal in real time to receiving stations for analysis and synthesis as in the case of the Desert Winds Project.
4. Performing additional research on the type of parameters to be measured, particularly on the relationship of the height at which wind velocity measurements are made to the capacity of the wind to transport particulate material on the surface.

REFERENCES

- Bagnold, R.A. (1933) A Further Journey Through the Libyan Desert. *Geographical Journal* 82: 103-129.
- Bagnold, R.A. (1941) *The Physics of Blown Sand and Desert Dunes*. Methuen, London.

- El-Baz, F. and Wolfe, R.W. (1980) Wind Patterns in the Western Desert. In: Contributions to Planetary Geology: Desert Landforms of Egypt, in press.
- El-Baz, F., Breed, C.S., Grolier, M.J., and McCauley, J.F. (1979a) Eolian Features in the Western Desert of Egypt and Some Applications to Mars: Journal of Geophysical Research 84: 8205-8221.
- El-Baz, F., Slezak, M.H. and Maxwell, T.A. (1979b) Preliminary Analysis of Color Variations of Sand Deposits in the Western Desert of Egypt. In: Apollo-Soyuz Test Project Summary Science Report. NASA SP-412, v. II, p. 237-262.
- Fryberger, S.G. (1979) Dune Forms and Wind Regime. In: A Study of Global Sand Seas. U.S. Geol. Surv. Prof. Pap. 1052, p. 137-169.
- Hochberg, R.C. (1979) A Low Power Data Acquisition System. Masters Thesis, Kansas State University, Manhattan, Kansas.
- Manent, L.S. and El-Baz, F. (1980) Effects of Topography on Dune Orientation in the Farafra region, Western Desert of Egypt, and Implications to Mars. Reports of Planetary Geology Program, NASA Tech. Memo. 82385, p. 298-300.
- McCauley, J.F., Ward, A.W., Breed C.S., Grolier, M.J., Machinnon, D.J., Helm, P., Witbeck, N., and Nez, B. (1979) Desert Winds Project. U.S. Geol. Surv., Branch of Astrogeologic Studies, Flagstaff, Arizona, poster.
- McKee, E.D., Breed, C.S. and Fryberger, S.G. (1977) Desert Sand Seas. In: Skylab Explores the Earth. NASA SP-380, p. 5-47.
- Paulson, R.W. (1978) Use of Earth Satellite for Automation of Hydrologic Data Collection. In: Collection, Storage, Retrieval, and Publication of Water Resources Data. U.S. Geol. Surv. Circ. 756, p. 8-14.
- Wilson, I.G. (1971) Desert Sandflow Basins and a Model for the Development of Ergs: Geographical Journal 137: 180-199.
- Wolfe, R.W. and El-Baz, F. (1979) The Wind Regime of the Western Desert of Egypt. Reports of Planetary Geology Program 1978-1979. NASA Tech. Memo. 80339, p. 299-301.

AEOLIAN ACTIVITY AS A PLANETARY PROCESS

By

Ronald Greeley

ABSTRACT

Any planet or satellite having a dynamic atmosphere and a solid surface has the potential for aeolian processes. Survey of the solar system shows that wind plays an important, and in some regions, the key role in surface modifications. Most deserts and many coastal areas on Earth are subject to aeolian processes. Seasonal dust storms sweep the surface of Mars, where aeolian activity appears to dominate. Measurements of wind speeds on Venus, observations of its surface, and estimates of particle threshold wind velocities in the venusian environment suggest that aeolian processes operate there as well. Recent discoveries of the predominantly nitrogen atmosphere of Titan raise the possibility of wind activity on this, the largest satellite of Saturn and the only moon known to have an appreciable atmosphere. From the extremely dense, hot atmosphere of Venus to the low atmospheric density of Mars and the extremely cold environment of Titan, there is the opportunity to study a single geological process under a wide range of environments to derive fundamental knowledge of how aeolian processes operate.

INTRODUCTION

Many physical and chemical processes modify planetary surfaces. Aeolian is defined (Gary et al., 1972) as Pertaining to the wind; esp. said of rocks, soils, and deposits (such as loess, dune sand, and some volcanic tuffs) whose constituents

were transported (blown) and laid down by atmospheric currents, or of landforms produced or eroded by the wind, or of sedimentary structures (such as ripple marks) made by the wind, or of geologic processes (such as erosion and deposition) accomplished by the wind. Thus, any planet or satellite having a dynamic atmosphere and a solid surface is subject to aeolian processes. A survey of the solar system shows that Earth, Mars, Venus, and possibly Titan meet these criteria (Table 1). These planets afford the opportunity to study a basic geological process--aeolian activity--in a comparative sense (Murray et al., 1981) with each of the planets being a vast natural laboratory having a strikingly different environment. Because terrestrial processes and features have been studied for many years, Earth is the primary data base. However, because surface processes are much more complicated on Earth--primarily because of the presence of liquid water and vegetation--many aspects of aeolian processes that are difficult to assess on Earth are more easily studied on other planets. For example, cohesion in silt and clay particles has a large effect on the wind speed needed to set them into motion, but particle cohesive effects resulting from water are difficult to separate from those resulting from electrostatic charges on Earth; in the comparatively dry atmosphere of Mars, the water effects are negligible and the general problem is made simpler.

Wind blowing across a planetary surface has the potential for directly eroding material and redistributing it to other areas. However, most aeolian erosion occurs through abrasion caused by windborne particles of sand size or smaller. The most important factor in aeolian processes is the threshold windspeed, or the minimum speed to mobilize particles of different sizes, densities and shapes. The ability of wind to attain threshold is a function of atmospheric density, viscosity, composition, and temperature (Iversen et al., 1976a), as shown in Figure 1 for Earth, Mars, and Venus. Thus, the very low density Martian atmosphere (surface pressure is about 1/200 that of Earth) requires wind speeds that are about an order of magnitude stronger than on Earth. It can be thought of partly in terms of the number of gas molecules impinging on and passing over the particles to be moved; in a low-density atmosphere (fewer molecules), for the same amount of work to be done, the wind must be moving faster to achieve the same effective flux of molecules. Although this is an oversimplification, it demonstrates to a first-order the relationship between atmospheric density and wind velocity for threshold.

RELEVANCE OF AEOLIAN RESEARCH TO PLANETOLOGY

Aeolian processes are capable of redistributing enormous quantities of sediment over planetary surfaces, resulting in

Figure 1

Comparison of threshold friction speed versus particle diameter for Mars, Earth, and Venus and in liquid; ρ_p/ρ is ratio of particle density to fluid density, ν is kinematic viscosity, g is gravitational acceleration, U_F is terminal speed, u_{*t} is friction speed at threshold; arrow indicates approximate transition between grains in saltation (larger particles) and those in suspension (after Iversen et al., 1976a).

the formation of landforms large enough to be seen from orbit and deposition of windblown sediments that can be hundreds of meters thick. Any process capable of effecting these changes is relevant to understanding the present and past geological environment of the planets involved. Furthermore, because aeolian processes involve the interaction of the atmosphere and lithosphere, an understanding of aeolian activity sheds light on meteorological problems. Aeolian activity is discussed in this section in terms of large-scale modifications, small-scale modifications, and as observable active processes in the form of dust storms and changing surface patterns.

Large-scale modifications are defined as features that can be observed from ranges of orbiting spacecraft and are classified as either erosional or depositional. One of the most useful aeolian features for interpretation of aeolian processes are dunes, a depositional landform. A recent document edited by McKee (1979) describes the major sand seas on Earth and discusses various dune forms; because much of the book is based on spacecraft images, it is particularly relevant to planetology. Both the planimetric shape and cross-sectional profile of dunes can reflect the prevailing winds in a given area. Thus, if certain dune shapes and/or slopes can be determined from orbital data, local wind patterns can be inferred. Repetitive viewing of the same dunes as a function of season may reveal seasonal wind patterns. Although seldom do dunes change significantly in plan form with season, the slip face on the dune crest can be observed on images to show alternating wind directions.

Identification of dunes signals the presence of sand size (60 to 2,000 micrometers) particles because sands are the only materials known to accumulate in dune deposits on Earth (Bagnold, 1941), and there is nothing in the physics of windblown sediments to suggest that it would be different on Mars, Venus, or Titan.

On Earth great quantities of silt and clay size particles are transported in dust storms. This material is eventually deposited as vast sheets, or mantling units of loess. Thick loess deposits are found throughout much of the geological column. Even where relatively young and well-exposed on the surface, loess deposits are nearly impossible to identify by remote sensing methods. Yet, identification of such deposits could be very important in understanding planetary surfaces. For example, substantial areas of Mars are interpreted to be mantled with aeolian sediments. However, other processes could lead to similar appearing terrains and a definitive means for remotely identifying fine-grained aeolian sediments is greatly needed.

Large-scale aeolian erosional features include (1) pits and hollows (blowouts) that form by deflation, or removal of

loose particles, (2) wind-sculptured hills such as yardangs, (3) windgaps, or wind-eroded notches in ridges, (4) and general, unclassified wind-eroded landforms.

Impact crater frequency distributions are widely used in planetary geology to obtain relative dates for different surfaces. On planets having active aeolian processes, the erasure of craters by erosion or burial by aeolian sediments can drastically alter the crater record and invalidate crater-derived ages. Thus, knowledge of rates of aeolian erosion and deposition for a wide range of planetary environments is required in order to assess the possible effects on the impact crater record.

Small-scale aeolian features include ventifacts (wind-shaped rocks) and aeolian sedimentary structures--features that can be observed only directly from the surface or inferred from remote sensing data. Ventifacts can provide information about local wind directions and the length of time a surface has been exposed.

Crater streaks commonly occur as light and dark forms, although "mixed" forms are found in which both light and dark streaks occur in association with the same crater. The origin of crater streaks is a matter of debate, with several models having been proposed. Disagreement may stem from oversimplifications and from trying to apply a single origin to a class of features with members having more than one origin. Nearly all investigators agree that streaks represent a surface manifestation of windblown processes, such as relatively thin ($\sim$ cm) deposits of particles that shift in response to winds

Several different models (Greeley et al., 1978) of light and dark streak formation can be postulated as functions of wind characteristics (wind velocities, turbulence, etc.), particle sizes, and particle compositions. In addition, the origin of streak material (from a source within the crater or part of the general surface around the crater) may control the development of streaks. Nearly all models must take into account the flow patterns generated by winds blowing over and around craters. Wind tunnel simulations and limited field studies show that a horseshoe vortex (Figure 2) wraps around the crater rim and creates an erosive zone in the wake of the crater and a depositional zone in the immediate lee of the crater rim (Greeley et al., 1974). The size and shape of zones of erosion and deposition are functions of crater geometry, wind speeds, time, and other parameters (Iversen et al., 1976b). For example, high wind speeds appear to produce longer streaks, although experimental data are limited and quantitative expressions have not yet been derived to make extrapolations to Mars. Identification of ventifacts has relevance to other aspects of planetology. For example, rocks at the Martian Viking landing sites that show pitted surfaces have

Figure 2

(A to C) Sequential photographs (light from upper left, wind from left to right as indicated by the arrow) of a 17.8-cm crater modeled in loose sand and placed in the wind tunnel with a wind velocity of about 420 cm/sec. The crater became ovoid in outline, pointing upwind, and developed two erosional depressions corresponding to the trailing components of the horseshoe vortex shown in I. (D) Small Martian crater in Mare Tyrrhenum showing similar outline and dark zones off the leeward edge of the crater rim that are interpreted to be the result of erosion. (E to G) Sequential photographs of a 17.8-cm crater modeled in solid wood (nondeformable), partly buried by loose sand (E) and subjected to a wind of 850 cm/sec until relatively stable conditions ensued (G) in which the model surface was swept free of loose sand except in zones of relative deposition, shown by the white trilobate pattern and the white patch on the windward rim. (H) A 2-km Martian crater in the region northwest of Memnonia showing a similar trilobate pattern in the immediate lee of the crater and a white zone on the windward rim that are interpreted to be aeolian deposits, and a large dark zone in the crater wake area interpreted to have resulted from erosion (Mariner 9 shading-corrected image). (I) Flow field pattern for wind blowing across a raised-rim crater showing zones of preferential erosion and deposition (from Greeley et al., 1974).

been interpreted as vesicular igneous rocks and are part of the basis for identifying the surrounding plains as volcanic; alternatively, the pitted rocks could be solely the result of aeolian erosion and not igneous.

Aeolian processes can both mix and sort sediments. Deposits consisting of a wide range of particle sizes, such as river sediments or glacial deposits, when subjected to winds may leave the coarser particles behind which may lead to "desert pavement" surfaces. Conversely, windblown dust derived from a wide range of rocks may become compositionally "homogenized" in dust storms and settle on wide-spread surfaces. Remote sensing of either of these cases (wind-sorted, or wind-mixed deposits) could lead to erroneous conclusions about the surface composition of the areas observed. Knowledge of how aeolian processes operate under a wide range of planetary environments and recognition of identifying characteristics are critical to the problem.

Observations of some aeolian features provide information directly on the atmosphere. For example, crater streaks on Mars are albedo surface patterns that show surface wind direction; they occur in great numbers over much of Mars. Repetitive imaging of these and other variable feature patterns has shown that many of them disappear, reappear, or change their size, shape, or position with time. Mapping the orientations of variable features has been used to derive a near-surface atmospheric circulation model.

In this paper, aeolian activity is reviewed and discussed in the planetary context. The Earth is not treated separately, but is compared with Mars, Venus, and Titan. The reader is referred to other sections in this volume for details of terrestrial aeolian processes.

MARS

Windstorms were suspected to occur on Mars even before the successful space probes in the early 1970's returned conclusive evidence of aeolian activity. Earth-based observations made in the nineteenth century showed seasonal albedo patterns that were attributed to a variety of processes. Some of the earliest interpretations of these patterns as dust storms were those of Dean McLaughlin, as reviewed by Veverka and Sagan (1974).

As knowledge of the composition and density of the Martian atmosphere became better defined, predictions based on theory were made as to the wind velocities required to set particles in motion. Because of the low atmospheric density on Mars, the estimated minimum wind speeds are about an order of magnitude higher than on Earth (Sagan and Pollack, 1969). Wind tunnel tests conducted under low atmospheric pressure in a

Figure 3
Shaded relief map of Mars (to 65°N and 65°S) showing Solis Planum, Noachis, and Hellespontus, areas where dust storms are observed to originate.

Martian simulation substantiate these estimates (Greeley et al., 1976, 1977, 1980).

With the arrival of Mariner 9 and the Soviet spacecraft Mars 2 and 3 at Mars in 1971 during a major global dust storm, the speculations and predictions of martian aeolian processes were amply verified. After the dust cleared, the Mariner 9 cameras revealed abundant features attributed to aeolian activity, including dunes, yardangs (wind-sculpted hills), and various pits and grooves considered to be features of deflation (McCauley, 1973). The Viking mission (1976-1981) has added substantially to the catalog of Martian aeolian features and has provided high-resolution details not previously observed (Figures 3 to 6), and for the first time, showed images of the surface of Mars from two landing sites (Figure 7). The surface consists of rock fragments up to several meters in diameter and very fine-grained material described as aeolian drift deposits (Arvidson et al., 1978). Some of the rocks are pitted and sculpted, suggestive of ventifacts.

Dunes

One of the most impressive discoveries of the Viking mission is the vast erg of the circumpolar region of Mars (Cutts et al., 1976). Studied in detail by Tsoar et al. (1979), Breed et al. (1979), and others, the field covers more than $7 \times 10^5 \text{ km}^2$, larger than Rub Al Khali in Arabia, the largest active erg on Earth. All of the dunes are either transverse or barchan (Figure 8). Mapping the dune morphologies and other indicators of wind directions have enabled regional maps of the wind circulation pattern to be derived. Two major wind directions are suggested, off-pole winds that become easterly due to coriolis forces during summer, and on-pole winds that become westerly during winter. These wind patterns compare favorably with those based on models of the atmosphere. The low albedo (i.e., dark) appearance of the dunes suggests a composition other than quartz, an observation fitting with the apparent lack of silicic materials on Mars. Because basaltic lavas are very common over much of Mars, including the smooth plains south of the dune field, it is suggested that the north polar dunes are composed of windblown particles.

Yardangs

The primary wind erosional features observed on Mars are yardangs. Summarizing from Ward (1979), most yardangs on Mars occur in equatorial regions, notably in the Amazonis region, Aeolis region, Ares Valles, and Iapygia. Some of the largest features are interpreted to be early-stage yardangs; they are 50 km long, 1 km wide, and 200 m high, and appear to have

Figure 4

Dune field (40 km by 40 km) on the floor of an ancient impact crater in the region south of Sinus Sabaeus, centered at 46°S , 339°W . Craters and other topographic depressions are natural traps for windblown sediments and the crater shown here is typical of many that have been photographed from orbit. Crater is about 150 km across, (Viking Orbiter frame 94A42).

Figure 5

High-resolution image of cratered terrain south of Elysium Planitia at 14.4° and 190°W showing numerous transverse dunes on the plains; the position and orientation of the dunes appear to be controlled by the surrounding topography; area of photograph is about 12.5 by 12.5 km (Viking Orbiter frame 763A10).

Figure 6

a. A Martian sand dune field (upper right) some 50 km long on the floor of Gangis Chasma, one of the branches of Vallis Marineris. Walls of the canyon are fluted by wind erosion.

Figure 6

b. Enlargement of the dune field shown in (a); individual dunes are 500 m across. The wind appears to have been blowing from the left; some of the dunes appear to climb the canyon wall. In this region, the walls have been modified by landslides. Debris flows are numerous as are jumbled masses of debris below the cliffs. Wind may be an effective agent in removing such debris that has slumped into the canyon. The canyon thus enlarges itself by the combined processes of slumping and wind deflation.

Figure 7

Views of Martian surface from Viking. (a) Panoramic view of the surface of Mars from Viking Lander 1 in Chryse Planitia (see Fig. 2). This image was taken in early morning (about 7:30 local time), August 3, 1977, and covers the view from the northeast at the left, to the southeast at the right. The large boulder at the left is about 1 by 3 m and is about 8 m from the spacecraft. The light-colored materials are deposits of windblown particles. The shape of the deposits suggest erosion by winds blowing from the upper left to the lower right; (b) View of Martian surface from the Viking Lander 2 site in Utopia Planitia (see Fig. 2), showing the rock-littered terrain and accumulations of fine-grained, windblown material. The horizon, which is some 3 km distant, is actually level; the tilt in the picture reflects the 8° tilt of the spacecraft and the manner in which the panoramic pictures are obtained.

developed from the erosion of mesas (Figure 9). From studies of terrestrial yardangs and comparisons with Mars, Ward (1979) concludes that the Martian features are geologically young (on a Martian time-scale) and probably are composed of friable rocks such as ignimbrites (many of the yardang localities are near known volcanic craters), or indurated regolith (regolith in this sense being fragmental debris generated by impact cratering). On Earth yardangs develop by erosion of grains that are loosened by weathering processes involving liquid water; Ward suggests that on Mars (in the absence of liquid water) exfoliation, salt weathering, or freeze-thaw processes may operate, but that the net weathering rate would be slower than on Earth.

Variable Features

By far the most abundant aeolian features on Mars are variable features, so named from the Mariner 9 mission (Sagan et al., 1972) for albedo patterns that changed their size, shape, and position with time (Figure 10). Crater streaks are the most common of the variable features (Figure 11). They have been studied extensively by Arvidson (1974), Veverka et al. (1976, 1977, 1978), Greeley et al. (1974, 1978), Thomas and Veverka (1979), and others.

Some elongated bright streaks associated with craters in the south polar region appear to be accumulations of CO₂ frost. Based on their form and seasonal behavior, they have been interpreted as accumulations of wind-transported frost in the lee of craters, similar to the formation of bright crater streaks involving dust and sand particles. (Thomas et al., 1979). Because frost particles would have a fairly low density, threshold wind speeds would be lower than for sand or dust, and frost streaks therefore would be more active.

Crater streaks of all types have been used to map surface wind patterns and applied to atmospheric circulation models (Thomas and Veverka, 1979). Comparisons of crater streaks observed from orbit with wind measurements obtained at the Viking Lander sites (Greeley et al., 1978) shows that dark streak orientations coincide with maximum wind directions (Figure 12).

Dust Storms

More than one hundred years of observations of Mars have produced an impressive catalog documenting active dust storms (Briggs et al., 1979). Recent spacecraft observations of Mars have greatly enhanced our understanding of these storms (Figure 13). Mariner 9 arrived in 1971 at a time when the planet's surface was obscured by dust (Leovy et al., 1972), and observed the waning stages of a major global storm. The

Figure 8
Viking Orbiter views of the north polar region of Mars (a) showing the CO_2 ice cap; the dark region surrounding the cap consists of sand dunes (VO frame 814A02), (b) isolated dunes and dune complexes; image covers an area about 41 by 45 km (VO frame 544B07).

Figure 9
High-resolution image of yardangs in the Aeolis region; area shown is about 10 km across; sun is from the left of image. (Viking Orbiter frame 732A56).

Figure 10
Mariner 9 images of a 17 km crater in the Daedalia region near Solis Lacus. Images early in the orbiting mission, showing only a small dark zone and the same feature 38 days later showing the growth of the dark zone due to aeolian activity; these and similar surface patterns are termed variable features (from Sagan et al., 1972).

Figure 11

Examples of several types of dark and light streaks associated with craters on Mars. Most dark streaks are generally considered to result from erosion of particles from the surface; some light streaks appear to be deposits of windblown particles (from Greeley et al., 1978).

Viking spacecraft, however, provided not only orbital viewing of all stages of dust storms (Briggs et al., 1979), but also provided ground-based meteorological data (Ryan and Henry, 1979). These observations show that major storms typically begin during the southern hemisphere spring or early summer (close to perihelion). In a given year the major storms vary in intensity and number from none to two. The major dust storms have been observed to originate from areas in both the southern (Hellas, Tharsis, and Noachis, and Solis Planum) and northern hemispheres (Isidis Planitia).

The major storms appear to go through three phases (Gierasch, 1974). In phase I, numerous local dust storms occur in the southern hemisphere, associated with the sublimating south polar cap and strong surface thermal contrasts (Peterfreund and Kieffer, 1979). These local storms contribute dust to the global atmosphere increasing diurnal and thermal tides. This increases the likelihood of a global storm because of the energy absorbed by the atmosphere from the suspended dust (Gierasch and Goody, 1973). Phase II is the expansion of a local storm into a global event. This occurs rapidly, typically involving 3 to 7 days. Phase III marks the decay of the storm and lasts from 50 to 100 days. The first areas to clear are the poles and topographically high regions, such as the summits of the shield volcanoes.

Mariner 9 and Viking results show that average particle size in the atmosphere is less than 2 micrometers, or about the same as the particles carried over the Atlantic by major Saharan dust storms. The dust on Mars was found to be well-mixed in the atmosphere to heights of 30 to 40 km and had the effect of raising the atmospheric temperature by as much as 50 K. Calculation of dust deposition from the storms suggests significant mantling of the surface of Mars over geological time.

In summary, aeolian processes appear to be the dominant process currently active on the surface of Mars, and have also played an important role in the geological past.

VENUS

The atmosphere of Venus is composed primarily of CO₂ with minor amounts of hydrochloric, hydrofluoric, and sulfuric acids. With a surface pressure of more than 90 bar, it has the highest atmospheric density of all the terrestrial planets (Table 1). Venus is completely enveloped in a perpetual shroud of clouds that hide the surface from viewing. Repetitive pictures of the cloud tops obtained over a period of 8 days during the flyby of Mariner 10 in 1974 showed circulation patterns and allowed wind speeds to be determined for the upper atmosphere (Murray et al., 1974). Although speeds of about 100 m s⁻¹ were obtained for the upper clouds in the equatorial

Figure 12

Rose diagram showing light and dark streak orientations in Chryse Planitia (region of Viking Lander 1) and hodograph of winds measured via the lander; dark streaks are oriented with the wind directions (from Greeley et al., 1978).

Figure 13

High-altitude view of Solis Planum region during a relatively clear period (left) and during dust storm activity (right) in which a sharply defined dust cloud 600 km across is seen. Estimate based on shadow measurements place the eastern edge (right side) of the dust cloud as high as 12 km above the surface (from Briggs et al., 1979).

zone, when extrapolations were made to the surface, the winds were estimated to be very sluggish.

The Soviet Venus landers, Venera 9 and 10, measured wind speeds near the surface for two sites on Venus of 0.5 to 1 m s⁻¹ at the height of the wind sensors (1 to 2 m above surface). More recent measurements of wind speeds obtained by the Pioneer-Venus atmospheric probes have been extrapolated to the surface and yield values of 1 to 2 m s⁻¹ (Counselman et al., 1979). These values are well within the range predicted for particle threshold (Figure 1), based on a combination of theory (Hess, 1975; Sagan, 1975) and extrapolations of wind tunnel experiments (Iversen et al., 1976a). Venera 9 and 10 images (Figure 14) of the surface of Venus shows rock fragments several cm and larger set in a mass of fine (<1 cm) material interpreted to be sand size or smaller (Florensky et al., 1977). This bimodal size distribution is indicative of fluid transport and because liquid water cannot exist at the extremely high temperature on Venus, it is assumed that the fluid involved is the atmosphere, or wind. Thus, it is likely that aeolian processes are active at present on Venus, and probably have been active in the geological past.

Questions to be addressed through future work involve rates of sediment transport by the slow, dense atmosphere, aeolian erosion, and the nature of aeolian landforms.

TITAN

In the Fall of 1980, the Voyager 1 spacecraft flew past Saturn and made observations of some of the satellites. One of the most important observations was the radio occultation of the atmosphere of Titan, Saturn's largest moon. The atmosphere was found to be predominantly nitrogen with a density that suggests surface pressures of 2 bar. Estimates of surface temperatures suggest solid nitrogen on the surface. Thus, there is the potential for aeolian processes on this, the only satellite known to have a substantial atmosphere. Whether the atmosphere is dynamic, and whether granular particles exist on the surface are unanswered questions. Dunes composed of ice particles being blown about the dense, extremely cold nitrogen atmosphere borders on the realm of science fiction but nonetheless remains a possibility.

APPROACH FOR INVESTIGATING AEOLIAN PROCESSES

Aeolian processes incorporate elements of geology, meteorology, physics, and chemistry. A unified study, therefore, requires knowledge in all these areas, that is, a multidisciplinary approach. The approach commonly used is to isolate parts of the aeolian process for detailed study. The pioneering work of Bagnold (1941), who analyzed the physics of

Figure 14

Venera 9 and 10 images of the surface of Venus showing cobble-to-boulder size rocks set in fine-grained, presumably aeolian sediments,

Figure 15

High-resolution image of the Chryse Planitia region of Mars near Viking Lander 1 showing impact craters smaller than 80 m across; the preservation of these craters signals a very ancient surface (perhaps hundreds of millions of years old); area shown is about 6 km by 13 km.

windblown sand, is an example of this approach. Once the fundamental principles are understood, it is then possible to extrapolate to a wide range of conditions, i.e., other planetary environments. Before this can be done, however, it is necessary to fit the results from studies of isolated parts of the problem back into the whole system: for example, Bagnold's work on threshold wind speeds for particle movement were carried out principally in wind tunnel studies; before making generalizations, however, he field tested the results under natural conditions.

In the study of planetary aeolian processes, there is seldom the opportunity to field test the extraterrestrial models. Thus, we must rely on a somewhat different method, as follows:

1. Identification of the general problem and isolation of specific parameters for study, (e.g., wind threshold speeds for particles of different sizes on Mars).
2. Investigate the problem under laboratory conditions where various parameters can be controlled for the "Earth case", (e.g., wind tunnel test of particle threshold).
3. Field test the laboratory results under natural conditions to verify that the simulations were done correctly, (e.g., threshold tests in the field).
4. Correct, modify, and/or calibrate the laboratory simulations to take into account the field results.
5. Carry out laboratory experiments for the extraterrestrial case duplicating or simulating as nearly as possible the planetary environment involved (e.g., threshold tests under Martian conditions).
6. Extrapolate the results to the planetary case using the laboratory results and theory (for parameters that cannot be duplicated).
7. Field test the extrapolation via spacecraft observations and apply the results toward the solution of problems involving aeolian processes.

A benefit of this approach is not only to provide a logical means for understanding extraterrestrial problems, but to contribute toward solving aeolian problems on Earth as well. As an example of the application of this approach, let us consider the problem of wind abrasion and erosion on Mars.

Rates of Wind Abrasion on Mars

It was generally concluded after the Mariner 9 mission that the rate of aeolian erosion is very high on Mars (Sagan, 1973) based on three factors: (1) the high wind speeds required for particle motion on Mars, (2) the high frequency

of dust storms as viewed telescopically from Earth, and (3) the variety and large number of features attributed to wind erosion and deposition. It was reasoned that sand grains, once set into motion by the wind, would be accelerated to high speeds and would be very effective in "sandblasting" the surface.

The Viking mission has caused a reassessment of wind erosion rates on Mars. Although Viking has verified the widespread occurrence of aeolian features, several lines of evidence suggest that the rates of aeolian activity may not be as high as previously thought. For example, more than 4 years of monitoring wind speeds at the two Viking landing sites show that the near-surface winds seldom attain threshold speeds. More importantly, the Viking Orbiters reveal numerous surfaces that have small (~ 10 m), fresh-appearing impact craters (Figure 15); their presence signals surfaces at least hundreds of millions of years old that have been little modified by erosion of any type, including aeolian erosion. Thus, there appears to be a conflict between the predicted high rate of aeolian erosion on Mars and the constraints posed by the Viking results. Resolving this conflict is important for understanding Mars for several reasons, including the following questions:

1. Do the dunes and other aeolian features represent the current aeolian regime, or are they relict features formed at a time when the climate on Mars was more favorable for aeolian activity (e.g., higher atmospheric density)? This question has a direct bearing on understanding the climatic history of Mars.
2. What is the lifetime of topographic features on Mars, especially impact craters? Relative dating of surfaces on Mars is based primarily on number and morphology of superposed impact craters. Is aeolian erosion, or burial by aeolian deposits significant? If so, what sizes of craters are effected and in what manner?
3. What is the role of aeolian processes in surficial modification? Is aeolian erosion rapid enough to constantly expose "fresh" bedrock? Does windblown dust cause a desert varnish-like coating on Martian rocks as it does on Earth? These questions are intimately linked to knowledge of rates and styles of aeolian erosion, and have a direct bearing on attempts to characterize the composition of the Martian surface via remote sensing.

In order to determine rates of aeolian erosion on Mars, knowledge of three factors is required (Figure 16): (1) various particle parameters such as the velocities of par-

Figure 16

Diagram showing the three primary factors required to determine rates of aeolian abrasion: (1) wind parameters (strength, frequency, direction), (2) particle parameters (number, size, shape, density, velocity as function of wind speed and height above the surface), and (3) susceptibility to erosion of different materials.

Figure 17

Schematic diagram of the velocimeter used to measure particle speed; device consists of two light-sensitive phototransistors and a light source; as the particle interrupts the light beam, a voltage signal is generated that can be measured; from the time between the two signals and the geometry of the device the velocity can be determined directly (from Greeley et al., 1981).

ticles as functions of wind speeds and heights above the surface, (2) the susceptibility to erosion of various rocks and minerals, and (3) wind speeds and frequencies on Mars, which can be obtained from Viking meteorological measurements and Martian global circulation models. Ideally, the three factors needed to determine rates of aeolian erosion would be measured directly on Mars. Except for the meteorological data, this is not feasible. Even the meteorological data are limited to the two lander sites, and in order to derive wind velocity, wind frequency and surface pressure data required for the calculation, theoretical global circulation models must be employed.

Thus, without in situ measurements, the next best approach for the particle flux parameters and the erosion susceptibility factors is to simulate the Martian aeolian environment as closely as possible. The various simulations can be validated partly by comparing the experiments run under terrestrial conditions with field results, then adjusting the procedure or modifying the apparatus until there is a good correlation with natural conditions. The experiments can then be rerun under Martian or near-Martian conditions. Because not all Martian parameters (such as gravity) can be simulated, some of the final results used in the calculations must be adjusted using theory.

Determinations of various particle flux parameters can be accomplished by direct measurement in the wind tunnel under both Earth and Martian (as nearly as possible) conditions. Particle flux (number of grains passing through a given cross section per unit time) is obtained by collectors placed in the wind tunnel and is determined as a function of freestream wind speed and height above the surface. Particle velocity can be measured directly (Figure 17) as functions of freestream wind speed, height above the surface, and particle size; saltation trajectories for one atmosphere (Earth case) can be analyzed using high-speed motion pictures. Where possible, the experimental results are compared with published field measurements. Drawing on previous work and the experiments, a numerical model can be formulated that allows correction for gravity and temperature differences for Mars for the final calculation of martian rates of aeolian erosion.

Most wind tunnels cannot be used to determine the susceptibility to erosion of various rocks and materials because they can be run only for relatively short periods of time for Martian conditions, and the amount of erosion would be too small to be measured accurately. Several different designs of erosion machines have been used in which erosion of rocks is assessed as functions of impact particle velocity, angle, size, and composition, all for a range of atmospheric densities (Figures 18, 19).

Based on the approach outlined above, values for the three parameters required for determining rates of erosion have been

Figure 18
View of the Martian Surface Wind Tunnel at NASA-Ames Research Center, California. Used for conducting experiments dealing with the physics of aeolian materials under simulated Martian conditions (from Greeley et al., 1977).

Figure 19

Photograph of the apparatus used to determine susceptibilities to erosion of different materials. Test samples (1) are placed around the rim of the chamber; particles stored in a hopper flow into a rotating arm (2) and are slung outward to impact the samples (3). Chamber can be operated at atmospheric pressure ranging from 10^3 mb (Earth case) down to 6 mb (Mars case); other parameters that can be varied are particle size, velocity, angle of impact, and composition and sample configuration.

Figure 20

Diagram of the Venusian Wind Tunnel, a facility to be used to carry out research on aeolian processes under simulated Venusian conditions.

obtained and applied to Mars (Greeley et al., 1981). The result yields extremely high rates of abrasion. Even when using very conservative values for the various cases that can be considered, the surface of Mars ought to be reduced to a smooth plain. Several explanations can be considered:

1. Surface rocks are extremely resistant to abrasion. If this were the case, the rocks would have to be of some substance completely foreign to terrestrial petrology. Chemical analyses obtained on the surface of Mars by the Viking landers, and comparisons with various petrologic models suggests this is extremely unlikely.
2. Particles are in very short supply or are very ineffective as windblown abrasors. Theoretical studies and laboratory simulations of windblown particles show that sand grains on Mars once set into saltation have very short lifetimes (described as the kamikaze effect by Sagan et al., 1977; kamikaze in Japanese means "divine wind"), because they impact the ground and each other at velocities high enough to fragment into very fine debris. Nonetheless, the presence of dunes suggests that sand size particles do exist on Mars, so particle effectiveness cannot fully explain the very high rate of abrasion.
3. Ancient surfaces on Mars may have been protected from erosion. Photogeological studies suggest that vast parts of Mars have been mantled by sediments and exhumed. Such surfaces would have been protected from wind erosion (Sagan et al., 1977). Not all surfaces, however, show signs of mantling, yet are in areas where dust storms occur frequently; these surfaces display ancient impact craters and must be old.
4. Climatic effects. The present conditions may be more favorable for aeolian activity than in the past. Thus, rates of abrasion based on current wind speeds and frequencies would be anomalously high. It is well known that Mars experiences orbital cycles that would cause the surface temperature to vary sufficiently to release and store vast quantities of volatiles. Because the threshold wind speed is intimately linked to atmospheric pressure, variations should be expected in aeolian activity. However, most lines of evidence point toward higher atmospheric densities in the geological past.

Thus, although there is no unique answer to the general question of wind abrasion on Mars, the possible answers have provided insight into general related aspects of Martian science and in the course of the investigation parameters

regarding wind abrasion on Earth have become better understood.

SUMMARY

Aeolian processes play an important role in the modification of the surface of Earth and Mars. Indirect evidence suggests that Venus and perhaps Titan, one of the satellites of Saturn, also may experience aeolian activity. Study of aeolian activity in a planetary context thus affords the opportunity to examine a fundamental process under a wide range of environmental conditions--each planet can be viewed as a vast natural laboratory.

The key question for future work can be posed as follows:

1. How are dust storms initiated? This question applies to both Earth and Mars. Dust storms involve very fine grained materials (typically a few micrometers in size); an examination of the threshold curve (Figure 1) shows that extremely strong winds are required to move material this fine and it is unlikely that grains are placed into suspension by simply having a strong wind blow across the surface. The explanation frequently given is that sand size grains (relatively easily moved) placed into saltation impact dust particles, setting them into motion where they can easily be wafted aloft. This does not explain frequent dust storms on Earth that begin in areas lacking sand. Thus, there may be other factors involved, perhaps related to electrostatic effects, phase changes of absorbed volatiles, or other triggering mechanisms. This is a general problem that has received little attention, despite its importance on both Earth and Mars.
2. What are rates of aeolian erosion in various planetary environments? Despite much interest in aeolian erosion (both deflation and abrasion), relatively few studies have been undertaken even on Earth that provide quantitative data, much less data that are relevant to other planets. There are two parts to the problem, one part deals with erosion on a small scale (e.g., individual rocks), the other part concerns erosion of landforms (e.g., crater rims).
3. What is the evolution of aeolian landforms? Many studies have been carried out that define stages of development for various aeolian features on Earth. In some cases, there is disagreement over the results and the interpretations that are involved. Various models of evolution should be tested on Earth and studies done to determine how to apply them to extraterrestrial

environments.

4. What are the characteristics of aeolian sediments on planets? On Earth, most aeolian sediments are quartz sand and silts, various clay minerals, plus minor amounts of particles of other composition, (gypsum sands, calcite sands, etc.), and most of the knowledge of the physics of particle motion and aeolian features such as dunes deals with these materials. On Titan, if windblown particles exist, they may be composed of nitrogen ice or other frozen volatiles; composition of aeolian particles on Venus is completely open to speculation; on Mars, quartz is probably absent or is only a minor constituent because major source rocks for quartz, such as granite, are apparently absent. Basaltic sands have been proposed for Mars, given the widespread occurrence of basaltic lava flows, the low albedo of Martian dunes, and the prevalence of mafic to ultra-mafic compositions derived from spectral studies. How do windblown basaltic sands behave? What is their evolution? How efficient are they as agents of abrasion in the aeolian environment? Do they accumulate in deposits having morphologies different from quartz particles? These and other questions are intimately linked with understanding the general sedimentary cycle on Mars; similar sets of questions can be applied to Venus and possibly Titan as more information becomes available.
5. What are aeolian conditions on Venus? The physics of windflow and particle movement in the high temperature and high pressure and density atmosphere of Venus involves flow regimes not previously investigated. Even such fundamental questions as threshold wind speeds can only be crudely estimated. Other aeolian problems such as mass flux of windblown particles, rates of erosion, landforms, etc., require knowledge of threshold and additional study appropriate to the Venusian environment (Figure 20).

Because aeolian activity involves the interaction of the atmosphere with the surface, the analysis of various wind-related features can provide clues to the nature of the atmosphere. Understanding rates of aeolian erosion and analysis of ancient aeolian features have the potential for providing information on the climatic conditions for the past, and for helping to unravel the geological history of the solar system.

Acknowledgements

I thank M. Hassan for the opportunity to participate in the UNESCO workshop on the Physics of Desertification. A.

Peterfreund and J. Garvin read parts of the manuscript and made key suggestions for its improvement. This work was supported by the Planetary Geology Program, National Aeronautics and Space Administration, Washington, D.C.

REFERENCES

- Arvidson, R.E. (1974) Wind-Blown Streaks, Splotches, and Associated Craters on Mars: Statistical Analysis of Mariner 9 Photographs, *Icarus*, 21, 12-27.
- Arvidson, R.E., Binder, A.B. and Jones, K.L. (1978) The Surface of Mars. *Scientific American*, 238, n. 3, 76-89.
- Bagnold, R.A. (1941) *The Physics of Blown Sand and Desert Dunes*. Methuen and Co. London, 265 p.
- Breed, C.S., Grolier, M.J. and McCauley, J.F. (1979) Morphology and Distribution of Common "Sand" Dunes on Mars: Comparison With the Earth, *Jour. Geophys. Res.* 84, 8183-8204.
- Briggs, G.A., Baum, W.A. and Barnes, J. (1979) Viking Orbiter Imaging Observations of Dust in the Martian Atmosphere. *Jour. Geophys. Res.* 84, 2795-2820.
- Counselman, C.C., Gourevitch, S.A., King, R.W., Lorient, G.B., and Prinn, R.G. (1979) Venus Winds and Zonal and Retrograde Below the Clouds. *Science*, 205, 85-87.
- Cutts, J.A., Blasius, K.R., Briggs, G.A., Carr, M.H., Greeley, R., and Masursky, H. (1976) North Polar Region of Mars: Imaging Results From Viking 2. *Science*, 194, 1329-1337.
- Florensky, C.P., Ronca, L.B., Basilevsky, A.R., Burba, G.A., Nikolaeva, O.V., Pronin, A.A., Trakhtman, A.M., Volkov, V.P., and Zazetsky (1977) The Surface of Venus as Revealed by Soviet Venera 9 and 10. *Geol. Soc. of Amer. Bull.*, 88, 1537-1545.
- Gary, M., McAfee, R. Jr., and Wolf, C.L., eds. (1972) *Glossary of Geology*. Amer. Geol. Inst., Washington, 805 p.
- Gierasch, P.J. and Goody, R.M. (1973) A Model of a Martian Great Dust Storm. *J. Atmos. Sci.*, 30, 169-179.
- Gierasch, P.J. (1974) Martian Dust Storms. *Rev. Geophys. Space Phys.* 12, 730-734.
- Greeley, R., Iversen, J.D., Pollack, J.B., Udovich, N., and White, B. (1974) Wind Tunnel Simulations of Light and Dark Streaks on Mars. *Science*, 183, 847-849.
- Greeley, R., White, B., Leach, R., Iversen, J., and Pollack, J. (1976) Mars: Wind Friction Speeds for Particle Movement. *Geophys. Res. Lett.*, 3, 417-420.
- Greeley, R., White, B.R., Pollack, J.B., Iversen, J.D., and Leach, R.N. (1977) Dust Storms on Mars: Considerations and Simulations. NASA Tech. Memo, TM78423, 29 p.

- Greeley, R., Papson, R. and Veverka, J. (1978) Crater Streaks in the Chryse Planitia Region of Mars: Early Viking Results. *Icarus*, 34, 556-567.
- Greeley, R., Leach, R., White, B., Iversen, J., and Pollack, J. (1980) Threshold Windspeeds for Sand on Mars: Wind Tunnel Simulations. *Geophys. Res. Lett.*, 7, 121-124.
- Greeley, R., Leach, R.N., Williams, S.H., White, B.R., Pollack, J.C., Krinsley, D.H., and Marshall, J.C. (1981) Rate of Wind Abrasion on Mars. NASA CR (in press).
- Hess, S.L. (1975) Dust on Venus. *J. Atmos. Sci.* 32, 1076-1078.
- Iversen, J.D., Greeley, R., and Pollack, J.B. (1976a) Windblown Dust on Earth, Mars, and Venus. *J. Atmos. Sci.*, 33, 2425-2429.
- Iversen, J.D., Greeley, R., White, B.R., and Pollack, J.B. (1976b) The Effect of Vertical Distortion in the Modeling of Sedimentation Phenomena: Martian Crater Wake Streaks. *Jour. Geophys. Res.*, 81, 4846-4856.
- Leovy, C.G., Briggs, G.A., Young, A.T., Smith, B.A., Pollack, J.B., Shipley, E.N., and Wildey, R.L. (1972) The Martian Atmosphere: Mariner 9 Television Experiment Progress Report. *Icarus*, 17, 373-393.
- McCauley, J.F. (1973) Mariner 9 Evidence for Wind Erosion in the Equatorial and Mid-Latitude Regions of Mars. *J. Geophys. Res.*, 78, 4123-4137.
- McKee, E.D., ed. (1979) A Study of Global Sand Seas. U.S. Geol. Surv. Prof. Pap. 1052, 429 p.
- Murray, B.C., Belton, M.J.S., Danielson, G.E., Davies, M.E., Gault, D., Hapke, B., O'Leary, B., Strom, R.G., Soumi, V., and Trask, N. (1974) Venus: Atmospheric Motion and Structure From Mariner 10 Pictures. *Science*, 183, 21-29.
- Murray, B.C., Malin, M.C. and Greeley, R. (1981) Earthlike Planets. W.H. Freeman and Co., San Francisco, 387 p.
- Peterfreund, A.R. and Kiefer, H.H. (1979) Thermal Infrared Properties of the Martian Atmosphere 3, Local Dust Clouds. *J. Geophys. Res.*, 84, 2853-2864.
- Ryan, J.A. and Henry, R.M. (1979) Mars Atmospheric Phenomena During Major Dust Storms, as Measured at Surface. *J. Geophys. Res.*, 84, 2821-2829.
- Sagan, C. and Pollack, J.B. (1969) Windblown Dust on Mars. *Nature*, 223, 791-794.
- Sagan, C., Veverka, J., Fox, P., Dubisch, R., Lederberg, J., Levinthal, E., Quam, L., Tucker, R., Pollack, J.B., and Smith, B.A. (1972) Variable Features on Mars: Preliminary Mariner 9 Television Results. *Icarus*, 17, 346-372.
- Sagan, C. (1973) Sandstorms and Aeolian Erosion on Mars. *Jour. Geophys. Res.*, 78, 4155-4161.
- Sagan, C. (1975) Windblown Dust on Venus. *J. Atmos. Sci.*, 32, 1079-1083.

- Sagan, C., Pieri, D., Fox, P., Arvidson, R.E., and Guinness, E.A. (1977) Particle Motion on Mars Inferred From the Viking Lander cameras. *Jour. Geophys. Res.*, 82, 4430-4438.
- Thomas, P. and Veverka, J. (1979) Seasonal and Secular Variation of Wind Streaks on Mars: An Analysis of Mariner 9 and Viking Data. *Jour. Geophys. Res.*, 84, 8131-8146.
- Thomas, P., Veverka, J. and Campos-Marquetti, R. (1979) Frost Streaks in the South Polar Cap of Mars. *Jour. Geophys. Res.*, 84, 4621-4633.
- Tsoar, H., Greeley, R. and Peterfreund, A.R. (1979) Mars: The North Polar Sand Sea and Related Wind Patterns. *Jour. Geophys. Res.*, 84, 8167-8180.
- Veverka, J. and Sagan, C. and Greeley, R. (1976) Variable Features on Mars. VI. An Unusual Crater Streak in Mesogaea, *Icarus*, 27, 241-253.
- Veverka, J., Thomas, P., and Greeley, R. (1977) A Study of Variable Features on Mars During the Viking Primary Mission. *Jour. Geophys. Res.*, 82, 4167-4187.
- Veverka, J., Cook, and Gougen, J. (1978) A Statistical Study of Crater-Associated Wind Streaks in the North Equatorial Zone of Mars. *Icarus*, 33, 466-482.
- Ward, A.W. (1979) Yardangs on Mars: Evidence of Recent Wind Erosion. *Jour. Geophys. Res.*, 84, 8147-8166.

ON THE REDDENING OF QUARTZ GRAINS IN DUNE SAND

by

Farouk El-Baz

SUMMARY

The color of desert surfaces is indicative of the composition of the exposed rock, rock rubble, soil, and sand. Deposits of particulate material in the desert are often light-toned with predominantly yellow and yellowish-red colors. The reddish component in desert sands has been attributed to the presence of hematite on individual sand grains. This hematite coating was previously thought to have originated in lateritic soils of humid climates. Today, however, many investigators agree that the red coating is acquired in the aeolian environment itself.

Observations from space have confirmed that the red color of dune sand increases with age and as the distance from the sand source increases. Visual observations by astronauts and color photographs taken from orbit show transitions from less red to redder sands in the same field in numerous deserts.

In general the color of dune sand depends on the composition of the source rock. For example, white dunes in Bermuda are made of calcite, resulting from the erosion of limestone and sea shells, and dunes of the White Sands in New Mexico are made up of nearly pure gypsum grains. Some dunes near the Red Sea coast are made up of pink coral grains, and others are covered with black magnetite particles. Furthermore, dune sand that is composed mostly of quartz may display color variations depending on the amount of lighter or darker components. However, in most cases the color of quartz sand depends largely on the red coating on individual grains.

Study of the amount and nature of reddish coatings on sand grains can be performed by simple microscopic examination. The stain on the quartz grains appears as a frost of yellowish-red color. Darker and thicker coatings occur within grooves or indentations on grains. The mineralogy, texture, and chemical composition of both types of coatings can be determined by a combination of X-ray diffraction techniques and scanning electron microscopy.

Samples so studied from the Western Desert of Egypt show that the coating is basically made of platelets of kaolinite, which are covered by submicroscopic particles of hematite. The importance of the clay component in the coatings suggests a method of formation similar to that responsible for desert varnish on rocks in aeolian environments. Furthermore, the coatings on quartz grains in samples from three locations in the Great Sand Sea increase in thickness from about 0.5 micrometers in the north to about 1.5 micrometers in the middle, to between 2 and 5 micrometers in the south.

These results confirm the observation of reddening of the sands as the transport distance increases. They also suggest that the quartz grains acquire the coatings during aeolian transport. This property can be used to determine the relative ages of color zones in the same field, and thus can be indicative of the transport direction of the sand.

INTRODUCTION

Desert color is an indication of the composition of the outer surfaces of exposed rock and particulate material. In addition to bright-yellow colors, the most widespread color in desert environments is red. Field investigations have shown that the red hue in desert sands increases with the distance of transport and with increasing age (Norris, 1969).

The red color in desert sands has been attributed to the presence of hematite (iron oxide) coatings on individual grains. Reddened sands have been observed in deserts throughout the world, but their mode of formation is controversial (Glennie, 1970; Folk, 1976). There are two major hypotheses concerning the origin of the red color caused by hematite in desert sands. According to the first, the hematite is detrital, having been formed in lateritic soils of hot, humid climates and later transported to desert basins (Van Houton, 1973). Advocates of the second hypothesis contend that the hematite coating is postdepositional and results from the weathering of iron-bearing minerals (Walker, 1967 and 1979; Norris, 1969; Glennie, 1970).

Although the origin of the red color is controversial, the fact that this color increases with time has been established in many localities. This paper will discuss occurrences of reddened aeolian sands with emphasis on recent studies of

samples from the Western Desert of Egypt. The significance of this is that the degree of reddening may be used to determine the relative age of the sands within the same dune field.

Because of its importance in desert sands, observations of the red color were conducted on manned space missions. A special effort was made on the Apollo-Soyuz mission of 1975 to study color variations in several sand seas as part of the Earth Observations and Photography Experiment (El-Baz and Mitchell, 1976).

Orbiting astronauts were able to detect color variations that were not easy to detect on film. This is in agreement with the fact that under laboratory conditions, the human eye is estimated to be able to distinguish 7.5 million color differences. This precision is two or three times better than that of most photoelectric spectrophotometers (Committee on Colorimetry, 1963, p. 136).

To make observations of color variations of desert surfaces from orbit, a color wheel was prepared for the Apollo-Soyuz astronauts (El-Baz, 1977, p. 49). The color wheel was fabricated in a "doughnut" shape with a 20.3-cm (8-inch) diameter and a 12.7-cm (5-inch) central hole (Figure 1). It was constructed of a 3-mm (1/8-inch)-thick aluminum ring with double rows of Munsell Standard Color chips. Each color chip was identified by "A" or "B" and a number that indicated its position in the row. A total of 54 colors were fastened on one side of the wheel (Figure 1).

The doughnut shape allowed the astronauts to conveniently rotate the wheel until the proper color in either row A or B matched the scene of the Earth's surface (El-Baz, 1977, p. 54). This shape was also found to be practical in field work. The wheel has been used in field comparisons and documentation of the color of desert surfaces. Figure 1 is intended to allow any investigator to make an identical device for making such standard color measurements.

REDDENED AEOLIAN SANDS

North American Deserts

Walker (1967) studied two stratigraphic sequences containing red beds, including Holocene deposits in the Sonora Desert of Mexico and late Paleozoic deposits in Colorado. He provided some of the more convincing evidence for the post-depositional formation of these red sands. Glennie (1970, p. 190) concurs in part with Walker (1967) but also believed that "much of the oxidation of iron in solution takes place at the interface between the ground-water saturated sediment below the water table and the air-filled pore space above."

Although some workers have questioned whether the modern desert environment can produce dune reddening, "an extensive

Figure 1
 Left, Munsell color designations of the Apollo-Soyuz wheel of 27 desert colors. Each designation indicates hue, value, and chroma; Right, black-and-white photograph of color wheel (8 inches in diameter) on top of sand showing the arrangement of the Munsell standard color chips in two rows marked A and B, which have the same hue and value but vary in chroma.

array of investigators offer evidence that red soils and sands do in fact form today in desert environments" and that "reddening of desert and subtropical dune sands clearly progresses with the flow of time" (Norris, 1969, p. 8 and 9).

In a study of the Algodones dunes of southern California, Norris and Norris (1961) presented the important factors in the development and persistence of hematite coatings on individual grains. In these dunes, Norris and Norris (1961, p. 611) found that the intensity of sand color increases from north to south; 25 to 60 percent of the grains are hematite-coated, with maximum values to the south. Since the more southerly dunes are older, the degree of reddening may be used to determine the relative age of the sands.

Southwest Africa Desert

Color photographs taken from space provide an additional tool for the study of desert color by: (1) documenting regional variations in the color of desert sands; (2) aiding in the location of areas for ground investigations; (3) providing evidence of transport directions and potential source areas; and (4) aiding in the determination of the relationship between dune morphology and desert color.

One example of the utility of orbital photographs in confirming and extending field observations is that of the Namib Desert of Southwest Africa. Logan (1960, p. 136) studied the linear dunes there, which have migrated from sources along stream courses. He found that the sands farthest inland are much redder in color and of greater age. Skylab 4 photographs of the same region portray color zones in the dune sand. In these photographs younger sands near the coast appear brighter than the redder zones farther inland (McKee and Breed, 1974).

It is important to note that although Landsat data are well-suited to classify major features of dune areas (McKee and Breed, 1976), they show few density and brightness variations in sand seas. True-color photographs are more suitable for the study of desert color because: (1) the photo-interpreter does not have to mentally translate one color into another, as necessary when working with Landsat false-color composites; (2) natural-color photographs can be used as such, unlike Landsat data that require enhancement by computers; and (3) the cost of digitally enhancing Landsat data is prohibitive when compared to the cost of a photographic print from a color negative (El-Baz, 1978).

Australian Deserts

Using a bracket-mounted camera, the Apollo-Soyuz astronauts obtained a stereo strip of photographs of the Sturt Desert in southern Australia. In addition, they took pho-

tographs with a handheld camera of the Simpson Desert in central Australia (El-Baz, 1978; Breed and Breed, 1979).

Photographs of the Lake Blanche area show the lake basin as the probable source of closely spaced linear dunes. This dry lake is 45 km long and covers an area of about 840 km². Sand from the playa deposit extends northward into the western edge of the Sturt Desert (Figure 2). The color photographs indicate three distinct zones, each about 40 km wide: a light-colored zone close to the sand source; a dark red zone farthest from the lake; and an intermediate zone in between. This sequence confirms dune sand reddening with time, or at least as the distance from the source increases.

Darkening of dune sand with time in this region is also known from field investigations. Norris (1969, p. 9) states that: "H. Wopfner (1965, written communication), in connection with his studies of the Lake Eyre Basin, Australia, also has observed a consistent darkening of sand grains with increasing distance from sources and with increased time of exposure." Folk (1969) studied dunes in the Simpson Desert and considered them to have attained the redness from dust that initially was red and was derived from red lateritic soils developed under humid climates of the geological past.

The Apollo-Soyuz astronauts photographed the area of the Simpson Desert just north of Lake Eyre (Breed and Breed, 1979). The photographs show dune reddening with the increase of distance from the sand source. In this case, the source of the sand appears to be the dry Lake Eyre and numerous small lakes north of it. The dunes are closely spaced linear ridges; the sand is brighter near the lakes and darker (redder) to the north. This direction is also that of the prevailing wind. According to Madigan (1936, p. 209), the mean wind direction in this part of Australia is due north; it changes slightly to north-northwest in January. Thus, the Apollo-Soyuz photographs of the Simpson Desert constitute another example of dune reddening with time and along the prevailing wind direction.

It must be stated however, that detailed field studies may reveal additional features not observed in the space photographs. For example, Breed and Breed (1979) found that dunes deep within the main body of the Simpson Desert "have the reddest hue and most intense chroma of any of the dunes observed in Australia." These are redder than the dunes at the edge of the desert. Thus, they attribute such color variations to "differences in the sources of sand and/or different stages of dune development at these sites" (Breed and Breed, 1979, p. 336).

Arabian Desert

Nearly one-half of the interior of the Arabian Peninsula is blanketed by aeolian sands (El-Baz, 1979). Ar-Rub al-Khali

Figure 2

Linear dunes emanating from the dry Lake Blanche (A; 45 km in length) near the western edge of the Sturt Desert in southeastern Australia. The fine dunes north and northeast of the lake are brighter near the sand source and darker in color farther out. The topography near the left edge is part of the North Flinders Ranges. Arrow points towards the direction of sand transport (Apollo-Soyuz photograph 1133).

Figure 3

Desert terrain just west of the Nile Delta showing three color zones: (1) Fertile soil in a triangular wedge near the Mediterranean Sea; (2) A bright zone of active sand; and (3) A rocky plain covered by sand dunes, sand sheets, and desert-varnished pebbles. Arrow points towards the direction of sand transport (Apollo-Soyuz photograph 1256).

is the largest sand sea and occupies approximately 600,000 km². Apollo-Soyuz photographs of this desert show color variations in Ar-Rub al-Khali sand sea, particularly in its northern extension known as Uruq ar-Rumaylah (El-Baz, 1979, p. 295).

Similarly, red color variations were noted in the Ad-Dahna Desert, which connects An-Nafud to the north and Ar-Rub al-Khali to the south. The southerly extensions of the dunes in this corridor are redder in color than their northern counterparts. Because the prevailing wind direction is toward the south in this part of the Arabian Peninsula, the red color here also increases with the transport distance.

Eastern Sahara Deserts

The eastern part of the North African Sahara includes numerous sand deposits, dune fields, and sand seas that display reddened sands as exemplified by sand deposits in the western parts of Libya and Egypt.

Walker (1979) describes outstanding examples of modern dune sands that have reddened with age and with distance of transport in the Awbari and Marzuq deserts of western Libya. He provides convincing evidence of the successive buildup of the red pigment as derived internally from the dunes. He concludes that: "Each step in the sequence of events leading to the accumulation of the stain and its gradual reddening can be documented in deposits that are forming in the desert today. This important fact indicates that the reddening is a product of the modern desert environment, not the product of an alleged earlier moist climate" (Walker, 1979, p. 63).

Similarly, sand deposits in the Western Desert of Egypt show an increase in reddening as the distance of aeolian transport increases (El-Baz, 1978). In this desert two components affect the sand color, the composition of sand-sized grains and the coatings on individual quartz grains. For example, in the part of the desert just west of the Nile Delta it was found that three distinct color zones, as depicted in one Apollo-Soyuz photograph (Figure 3), represents the following: (1) dark, arable soil with a mottled appearance that is composed of quartz sand, clay, and calcium carbonate particles; (2) yellow, relatively active sand with or without sparse vegetation, and with increased reddening toward the north; and (3) desert pavement with relatively inactive sand that is mixed with dark, desert-varnished gravel (El-Baz and El-Etr, 1979, p. 203).

The overall pattern of sand transport in the Western Desert of Egypt was established by Gifford et al. (1979). Compositional variations that affected the overall color were also studied (El-Baz et al., 1979). It was found that the northern region of the Western Desert has the highest percent-

tage of calcareous grains originating from local limestone outcrops. Samples from the central part of the Western Desert contain numerous iron-rich grains originating from the iron deposits and are locally enriched in shale fragments. Sands of the Great Sand Sea (Figure 4) are relatively homogeneous, quartz-rich deposits that vary little in their percentage of minor components (El-Baz et al., 1979).

Because the sand samples of the Great Sand Sea were uniformly composed of quartz grains, their color variations depended largely on coatings on the grains. These coatings were studied in detail in the laboratory as discussed below.

COATING ON GREAT SAND SEA QUARTZ

Sample Locations

Quartz grains from three samples from the Great Sand Sea were examined. The samples were taken from areas separated by nearly 400 km. The first sample was taken from the dune mass at the northern part of the Great Sand Sea just south of the Siwa depression. The second sample was obtained from the linear dunes southwest of the Farafra depression. The third sample was obtained from the southern extensions of the Great Sand Sea in southwestern Egypt (Figure 4).

Laboratory Techniques

Quartz sand grains were initially examined with a binocular microscope to study the main textures and overgrowths. Single, small, heavily coated grains were examined by X-ray powder diffractometry to study the mineralogy of the coatings. Because of the thin nature of the coating, only quartz appeared in the diffraction pattern. Heavily coated sand grains and those with coatings concentrated in pits or cracks were chosen for X-ray powder diffraction using the Gandolfi camera. The coating was gently scraped from many individual grains in order to collect sufficient material for analysis. The Gandolfi camera geometry allows precession of the sample in the X-ray beam so that crystals in all orientations can be brought into the diffraction condition, thereby enhancing diffraction lines produced by a small number of crystals. Fe K α X-radiation was used to minimize fluorescence from iron in the sample.

Additional grains were fractured under liquid Freon to expose the interior of the grains and a cross section of the coating. The grains were examined by scanning electron microscopy (SEM) at magnifications from 100X to 200,000X. Energy dispersive X-ray analysis (EDXA) was used to qualitatively determine the chemistry of the coating and attached crystals (Prestel and El-Baz, 1979; Prestel et al., 1980).

Figure 4
Sketch map of Egypt showing the oases depressions and the dune belts in the Western Desert.

Microscopic Characteristics

Microscopic examination of the grains indicated that the reddish-brown color of the sand grains may be attributed to: (1) an extremely thin coating or stain of reddish-brown material entirely covering the grains except on freshly broken surfaces, (2) small clumps of reddish-brown granules in pits on grain surfaces, (3) a reddish-brown material concentrated within cracks in the grains, and (4) detrital dark brown grains occurring within the sand grains, which were probably present in pore spaces when the parent sandstone was lithified.

Grains having either a relatively thick coating or a high concentration of material in pits or cracks were selected for X-ray diffraction analysis. The material concentrated in cracks was removed for detailed examination. The SEM study of the morphology of the coating in depressions compared to that on exposed grain surfaces (and associated EDXA chemical analyses) supported the assumption that the material concentrated in the cracks was of the same composition as that on the grain surfaces.

Mineralogical Composition

Analysis of the X-ray diffraction patterns (Prestel and El-Baz, 1979) indicates that the coating on the quartz grains is composed of kaolinite and hematite. Diffraction lines from kaolinite suggest a well-crystallized material. In contrast, hematite displays diffuse lines, which suggest poor crystallinity and very fine grain size. A quantitative estimate of the amount of each mineral present cannot be made from these data. However, the greater number of diffraction lines corresponding to kaolinite lines suggests that kaolinite is either more abundant or better crystallized.

Each of the quartz grains examined had an irregular, pitted surface with a ubiquitous coating of fine-grained material (Figure 5) and a well-defined contact between the coating and its underlying quartz surface. The coating conforms closely to grain morphology.

The coating consists, at higher magnifications, of a variety of submicrometer-size particles and granules. Many of the particles are obviously crystals and frequently are hexagonal platelets (Figure 5). These platelets occur as randomly oriented single grains or as books typical of kaolinite (Keller, 1978). Individual platelets range in size from a few hundred angstroms to about 2 micrometers. Individual platelets have a powdery surface, with small platelets occurring on the surfaces of larger ones. This may indicate that books of platelets are formed as the result of in situ growth of small platelets on larger, preexisting ones (Prestel and El-Baz, 1979).

Figure 5

Scanning electron microscope photographs of sand grains from the Western Desert of Egypt: (a) sand sample from longitudinal dune at 30X; (b) view of rounded quartz grain with a pit, marked "Y" in Figure (a); (c) oblique view of coating on quartz grain at 1,500X showing concoidal fracture; and (d) hexagonal platelets and books of platelets that form the coating on the quartz grain, 20,000X (a and b photographed by Marie Slezak; c and d photographed by Donna Prestel).

The coating appears to develop gradually during aeolian transport, and it shows little evidence of successive layers of deposition over most of the surface area of the grain. The lack of successive layering most likely precludes episodic growth periods with intervening erosion and is consistent with gradual, continuous growth. No systematic changes in morphology (crystal size or orientation), mineralogy, or chemistry were observed in the coatings. However, in a few pitted regions, some preferred orientation of platelets in response to grain morphology was seen.

The hexagonal platelet nature of the coating agrees well with the presence of kaolinite and hematite, since both minerals are known to have a hexagonal, platy habit. Chemical analyses by EDXA indicate that the coating is much richer in Al than Fe (Figure 6). This suggests that the coating is composed predominantly of a clay material with very finely disseminated hematite. The two minerals are probably growing either simultaneously or successively (El-Baz and Prestel, 1980).

Coating Thickness

The thickness of the coatings on quartz grains was determined at from one to twelve locations on each measured grain. This was done by fracturing a number of representative grains from each sample. The grains were then ultrasonically cleaned and examined with the SEM. Stereo photographs were used to position the sample with the fracture plane normal to the viewing direction.

These thickness measurements revealed that the northernmost Siwa samples contain the thinnest coating with a thickness of 0.8 micrometer. The samples from the middle part of the Great Sand Sea west of Farafra have a thickness of 1.5 micrometers. The thickest coatings were measured on quartz grains from the southern terminae of the Great Sand Sea at the Gilf Kebir plateau, where the thickness of coatings varied between 2 and 5 micrometers. These measurements add proof to the increase of the thickness of the red coating on quartz grains with the transport distance.

COATING MECHANISMS

It is clear that the coatings on quartz grains from the Great Sand Sea are made of kaolinite crystals that are covered by submicroscopic hematite powder. The association of clays and iron oxide makes these coatings similar to desert varnish as described by Potter and Rossman (1977 and 1979).

The association of the red pigment with clay has also been described by Folk (1969) in the Australian dunes and by Walker (1979) in the Libyan dunes. Walker (1979, p. 72) cites evidence indicating that the clay coatings consist of airborne

Figure 6
X-ray spectrum of overall coating on quartz sand grains are from the conductive coating on the sample. Note the sharp Al peak and the more diffuse Fe peak (analysis performed by Donna Prestel).

dust emplaced on sand grains by mechanical infiltration of particles into the dune rather than inherited from matrix clays from the source sandstones, as follows:

- "1. Clay coatings are not present on sand grains of the beach from which coastal dune sands are derived, but they are common on grains in the dunes located a short distance inland from the beach. Thus clay coatings must be acquired as the sand migrates inland.
- "2. Clay coatings on sand grains of dunes throughout the region characteristically are composed of aggregates of tiny platelets, which display clastic textures that are similar to those produced experimentally by infiltrating clay-bearing water into sand (Crone, 1974; 1975). These textures, in turn, are similar to those produced naturally by clay which has been infiltrated into sand by influent seepage of surface water (Walker and Crone, 1974; Crone, 1975; Walker, 1976). Such similarities indicate that the clay is composed of mechanically infiltrated particles, and influent rainwater is the most logical agent to accomplish the infiltration.
- "3. The mineralogy of the clay in the coatings is similar to that expected of the airborne dust, suggesting that dust is the probable source of the clay."

The evidence for the origin of the clay as airborne dust is quite convincing. Most of the aeolian transport of quartz sands is accompanied by the transport of very fine dust in the desert. The mechanical buildup of the coating and the importance of water in the process are debatable, particularly in the case of the extremely dry area of the Great Sand Sea in the Western Desert of Egypt.

According to the Aridity Index Map of Henning and Flohn (1977), the Western Desert of Egypt is part of the largest expanse of hyperarid terrain on Earth. Only the very much smaller Atacama and Peruvian Deserts of South America have a comparable aridity index. The Budyko ratio that is the basis for this map has been called the "radiational index of dryness." It expresses the relationship between net radiation and precipitation at the surface. Both quantities are taken in energy units and applied to long-term yearly means values. This ratio thus expresses the number of times the annual net radiation could evaporate the precipitation at a given place. The Budyko ratio is 200 for southwestern Egypt, whereas it is only 7 for the most arid parts of the southwestern United States and 50 or less for most of the Sahara in the western parts of North Africa.

Because of the unusual dryness of the Western Desert of Egypt, it is necessary to think of another mechanism for the accumulation of the kaolinite crystals and hematite powder on quartz grains. I would like to propose the generation of electrostatic charges by the collision of bouncing grains during their aeolian transport.

The author has seen sparks generated on steel wires during sandstorms of the dry spring season in the Western Desert of Egypt. If the bouncing quartz grains generate enough electric charges to attract fine kaolinite dust, then this would also allow the growth of the coatings during aeolian transport.

To check this hypothesis, W. D. Keller (1980; pers. comm.) suggested to: (1) dry some clean quartz in a simulated hot desert atmosphere; (2) roll the quartz grains against each other, and to simulate drift pour them out in a thin stream, and check for any produced negative charges; and (3) gently blow light dust of desert-dry, fine kaolin towards the stream of poured sand and determine if it is attracted and/or held by the clean quartz.

The same procedure can be undertaken with hematite. Furthermore, one could check if the kaolin that may be attracted by an electric charge from the bouncing quartz grains remains as-is or recrystallizes. If it recrystallizes, then a small amount of water, such as from dew, may be necessary to the process of coating formation. It is hoped that such experiments will be conducted in the future.

The effects of these coatings on the reflectance properties of sand accumulations should also be studied. The chemistry and thickness of the coatings affect the reflectance of the desert surfaces as measured by remote sensing, e.g., by the Landsat multispectral sensors. Attempts should be made to relate a measure of reflectance in the Landsat data to the nature of coatings on sand grains.

In addition, there is a possible correlation between the reddish-yellow sands of the extremely dry Western Desert and the yellowish-brown grains on the surface of Mars. Results of the Viking mission to Mars indicate that the color of the surface is consistent with an abundance of Fe^{+3} -rich weathering products, notably nontronite (Huck and others, 1977, p. 4401). Thus, the study of the process of kaolinite/hematite coating on quartz grains in the Western Desert of Egypt may help us understand the mechanism of reddening of the Martian regolith.

CONCLUSIONS

1. Desert sands are reddened in modern aeolian environments, by the formation of a red coating on individual quartz grains. The coating thickness increases with time and with transport distance.

2. Color photographs taken from space depict such reddening within the same sand deposit or dune field. Color zoning in these photographs can be used as a measure of relative age of the zones; the redder the zone the older the sand or the farther away from its source.
3. The coating on quartz grains in the Great Sand Sea is made of five kaolinite platelets that are covered by submicroscopic powder of hematite. Both may have originated as fine dust in the aeolian regime.
4. The presence of both clay and iron oxide in the coating on the quartz grains makes it similar, at least chemically, to desert varnish on desert rocks.
5. The mechanisms of deposition of the kaolinite and hematite on the quartz grains are not yet fully understood, particularly whether or not water is a significant factor.
6. The reddened sands in the extremely dry Western Desert of Egypt may shed some light on the reddened soils of Mars. Correlations should be made of both environments to better understand the processes of reddening of desert soils.

REFERENCES

- Breed, C.S. and Breed, W.J. (1979) Dunes and Other Windforms of Central Australia (and Comparisons With Linear Dunes on the Moenkopi Plateau, Arizona). In: Apollo-Soyuz Test Project Summary Science Report, V. II. NASA SP-412, p. 319-358.
- Committee on Colortimetry (1963) The Science of Color. The Optical Society of America, Washington, D.C.
- Crone, A.J. (1974) Experimental Studies of Mechanically Infiltrated Clay Matrix in Sand. Geol. Soc. Amer. Abs. with Programs, v. 6, p. 701.
- Crone, A.J. (1975) Laboratory and Field Studies of Mechanically Infiltrated Matrix Clay in Arid Fluvial Sediments. Colorado University Ph.D. Thesis, Boulder, Colorado.
- El-Baz, F. (1977) Astronaut Observations From the Apollo-Soyuz mission. Smithsonian Institution Press, Washington, D.C.
- El-Baz, F. (1978) The Meaning of Desert Color in Earth Orbital Photographs. Photogramm. Eng. Rem. Sens., v. 44, p. 69-75.
- El-Baz, F. (1979) Color of Desert Surfaces in the Arabian Peninsula. In: Apollo-Soyuz Test Project Summary Science Report, v. II. NASA SP-412, p. 285-299.

- El-Baz, F. and El-Etr, H. (1979) Color Zoning in the Western Desert of Egypt. In: Apollo-Soyuz Test Project Summary Science Report, v. II. NASA SP-412, p. 203-217.
- El-Baz, F. and Mitchell, D.A. (1976) Earth Observations and Photography, Experiment MA-136. Sec. 10 of Apollo-Soyuz Test Project Summary Science Report. NASA TMX-58173.
- El-Baz, F. and Prestel, D.J. (1980) Desert Varnish on Sand Grains From the Western Desert of Egypt: Importance of the Clay Component and Implications to Mars. Lunar and Planetary Science XI, Lunar and Planetary Institute, Houston, Texas, p. 254-256.
- El-Baz, F., Slezak, M.H. and Maxwell, T.A. (1979) Preliminary Analysis of Color Variations of Sand Deposits in the Western Desert of Egypt. In: Apollo-Soyuz Test Project Summary Science Report, v. II. NASA SP-412, p. 237-262.
- Folk, R.L. (1969) Grain Shape and Diagenesis in the Simpson Desert, Northern Territory, Australia. Geol. Soc. Amer. Abs. with Programs, pt. 7, p. 68-69.
- Folk, R.L. (1976) Reddening of Desert Sands: Simpson Desert, N.T. Australia. J. Sed. Petrol., v. 46, p. 604-615.
- Gifford, A.W., Warner, D.M. and El-Baz, F. (1979) Orbital Observations of Sand Distribution in the Western Desert of Egypt. In: Apollo-Soyuz Test Project Summary Science Report, v. II. NASA SP-412, p. 219-236.
- Glennie, K.W. (1970) Desert Sedimentary Environments. Elsevier, New York.
- Henning, D. and Flohn, H. (1977) Climate Aridity Index Map. UN Conference on Desertification, Nairobi.
- Huck, F.O. et al. (1977) Spectrophotometric and Color Estimates of the Viking Lander Sites. Jour. Geophys. Res., v. 82, p. 4401-4411.
- Keller, W.D. (1978) Classification of Kaolians Exemplified by Their Textures in Scan Electra Micrographs. Clays and Clay Minerals, v. 26, p. 1-2.
- Logan, R.F. (1960) The Central Namib Desert. National Academy of Sciences, National Research Council Publication 785.
- Madigan, C.T. (1936) The Australian Sand Ridge Deserts. Geogr. Rev., v. 26, p. 205-227.
- McKee, E.D. and Breed, C.S. (1974) Preliminary Report on Dunes. Sec. 9 in Skylab 4 Visual Observations Project Report. NASA TM S-58142.
- McKee, E.D. and Breed, C.S. (1976) Sand Seas of the World. In: ERTS-1, A Window on Our Planet. U.S. Geol. Surv. Prof. Pap. 929, p. 81-88.
- Norris, R.M. (1969) Dune Reddening and Time. J. Sed. Petrol., v. 39, p. 7-11.
- Norris, R.M. and Norris, K.S. (1961) Algodones Dunes of Southeastern California. Geol. Soc. Amer. Bull., v. 72, p. 605-620.

- Potter, R.M. and Rossman, J.R. (1977) Desert Varnish: The Importance of Clay Minerals. *Science*, v. 196, p. 1446-1448.
- Potter, R.M. and Rossman, J.R. (1979) The Manganese and Iron Oxide Mineralogy of Desert Varnish. *Chemical Geology*, v. 25, p. 79-94.
- Prestel, D., and El-Baz, F. (1979) Microscopic Characteristics of Quartz Sand From the Arid Environment of the Gilf Kebir, Southwest Egypt. *Reports of Planetary Geology Program, 1978-1979, NASA TM-80339*, p. 293-295.
- Prestel, D., Wainwright, J.E. and El-Baz, F. (1980) Studies of Coatings on Sand Grains From the Gilf Kebir, Southwest Egypt. *Reports of Planetary Geology Program, 1979-1980, NASA TM-81776*, p. 238-240.
- Van Houten, F.B. (1973) Origin of Red Beds: A Review, 1961-1972. *Ann. Rev. Earth and Plan. Sci.*, v. 1, p. 39-61.
- Walker, T.R. (1967) Formation of Red Beds in Modern and Ancient Deserts. *Geol. Soc. Amer. Bull.*, v. 78, p. 353-368.
- Walker, T.R. (1976) Diagenetic Origin of Continental Red Beds. In: *The Continental Permian in West, Central and South Europe: NASA Advanced Study Institute, Mainz, Germany*, p. 240-282.
- Walker, T.R. (1979) Red Color in Dune Sand. In: *Study of Global Sand Seas, U.S. Geol. Surv. Prof. Pap. 1052*, p. 61-81.
- Walker, T.R. and Crone, A.J. (1974) Mechanically Infiltrated Clay Matrix in Desert Alluvium. *Geol. Soc. Amer. Abs. with Programs*, v. 6, p. 998.

THE WIND AND DESERTIFICATION PROCESSES IN THE SAHARO-SAHELIAN AND SAHELIAN REGIONS

By

Monique Mainguet

INTRODUCTION

We consider that aeolian action is an agent of desertification when it is responsible for a massive invasion of sand, and conversely, when it reactivates the surface of a sand sheet in an ancient dune system. In the south of the Sahara, the southern limit of the aeolian actions extend more and more southward gradually as one travels from the Atlantic coast towards the central and eastern part of the continent (Figures 1 and 2).

In Mali, the aeolian actions extend more southward than in Senegal. At the level of the Bandiagara plateau they reach the latitude of 12°N. The effects on the ground of the sand-carrying winds, as they can be seen on the Landsat satellite images from 1973 to 1976, extend to 11°30'N in both Nigeria and Upper Volta. The effects of the winds can be seen as strips, which are the result of either of deflation, or of a severe drying out of the vegetation. These are more pronounced in 1973 than in 1976. On the flat rocky plateaus (Ader Doutchi, south of Adrar of Ifoghas), the aeolian dynamic manifests itself through giant barchans larger than 1 kilometer wide. The steppic vegetation of these dunes and the layer of soil seem to indicate an ancient origin. However, their axis of symmetry is completely consistent with the overall trajectory of the dominant winds of today. After the climatic crisis, their vegetation cover has been partially degraded; they have become reservoirs of sand mobilized by the actual winds.

Figure 1
Current and paleoclimatic limits of ergs.

Figure 2
Dune types and average yearly precipitations

Active dunes: semi-fixed dunes
sparsely covered steppe,
with Aristida pungens and
Cornulaca monacantha

Fixed dunes

Bushy steppe: Cenchrus biflorus, Eragrostis tremula,
Acacia sp. Balanites aegyptiaca,
Guiera senegalensis.

In Chad, the effects of the wind extend beyond the southern border of this country; their southern effects reach the Central African Republic, at 11°N, and between 20 and 21°E. Here, the aeolian actions are combined with hydric action, which begins in Chad, at 16°N, because at these latitudes the Ennedi and the Ouaddai behave as water towers. The zone of the dominant hydric actions, combined with the degradation, begins south of an oblique line found at 12°30'N in the west and at 14°N in the east.

The aim of this work is not to study aeolian dynamics but rather to point out the role of the wind in desertification processes and, in particular, to determine if the wind is the only desertifying agent or whether it is merely the most apparent agent, after or together with other agents (natural and human), locally returning the appearance of the Sahara to the Sahel as it was before the sandy sheets re-covered it--locally accumulating up to 60 m thickness, as in the Maradi region of Niger. In the course of this paper we will see in the first part all the characteristic traits of the wind as well as those that participate in desertification, in the second part we will study desertification through four specific examples: (1) the Cap Vert peninsula, (2) the Upper Voltaic Sahel, (3) the Saharo-Sahelian transition zone in Mauritania, and finally (4) the Saharo-Sahelian and the Sahelian areas of Niger.

WIND CHARACTERISTICS AT THE SAHARO-SAHELIAN AND SAHELIAN MARGINS OF THE SAHARA

The desertification of the Sahelian regions, as observed for the last decade, is probably due to a combination of natural causes, the persistence of a serious rain deficiency, and to human activities. The regime of Sahelian winds is well known, even though the anemometric data are often mediocre and not always reliable. The efficient winds are rare, but yet they erode, move, transport, accumulate, and model dunes; and they form and carve the rocky outcrops. The wind, as a desertification factor, does not manifest itself significantly in an environment unless it has been modified beforehand. Contrary to what is commonly written, it is not the Sahara that progresses but it is the Sahel that locally takes on certain Saharan aspects.

Delineation of the Area Under Study

There are many and varying definitions of the Sahel, depending on the criteria used. Etymologically "Sahil" means shore. This broad meaning is good insofar as it defines the Sahel as the annex of the Sahara. From the climatological point of view, the Sahel receives between 50 and 500 mm of

precipitation. The phytogeographical criterion is not very precise; the Sahel comprises, between the northern limits of the area, Cenchrus biflorus and Butyrospermum parkii, limits that correspond roughly to the lines of the isohyets at 150 to 750 mm, respectively. Bioeconomical criteria only allow the consideration of exploitable parts of the Sahel, namely the regions in which the agropastoral activities are possible, thanks to sufficient pluviometry that seems to be sufficient but aleatory. The exploitable Sahel would be limited in the north by the 250-mm isohyet; we deal therefore with an area of movable limits, the extension of which is strictly connected to the year-to-year rainfall variations. Paleogeographic criteria leads us to conclude that the southern limit of Sahel corresponds closely to the limit of the ergs, which are presently fixed, ever since their establishment at the time of the arid pleistocene period, and thus remain to the extreme limit of the widest part of the paleo-Sahara, except in Nigeria, where an ancient dune system is now covered by a heavily forested savanna. These different criteria give us a better understanding of the diversity of the Sahelian environment.

The Air Masses

The northern part of Africa, and the Saharo-Sahelian zone in particular, are under the influence of two main air masses: the tropical boreal air (TBA) and the tropical austral air (ATA), the contact of which causes the materialization of the Intertropical Convergence Front (ITC). Above these two air masses is the equatorial altitude air (EA); in the higher layers of the troposphere the ITC splits in two (see the meridional cut of the West African troposphere, Figure 3).

The meridional displacement of these air masses, just like the centers of action that they activate, is linked in a more or less strict connection with the planetary balance associated with the apparent motion of the sun. The movement of these air masses from the south to the north and vice versa causes the shifting of the ground traces of the ITC; the ITC oscillates around its southern extreme position from December to February (2° - 3° N), and around its northern extreme position from July to September (22° - 24° N). It therefore sweeps all of continental West Africa. The migration towards the north lasts for six months, and the one towards the south takes place in four.

The ground trace of the FIT appears as a large convex undulation, with the axis situated on the meridian from 0° to 8° W, the eastern and western sides of which bend towards the meridian from 10 - 12° W and 10 - 14° E. This particular aspect can be linked to the dynamics of boreal flexures and to the more or less pronounced lowering of the continental depressions.

Wind and Air Masses

The winds of the tropical boreal air mass (TBA) are the winds of anticyclonic origin, impulsed by the permanent anticyclone of the Azores, perhaps associated with the thermal and seasonal Maghreb anticyclone; the trade winds that flow toward the south up to the ITC, following variable trajectories. The oceanic trade winds, the continentalized trade winds, and the continental trade winds can be distinguished. The oceanic trade winds of the sector WNW to N affect all of the Atlantic coast, from southern Morocco to the northwest of Guinea. The continentalized trade winds of the sector NW to NNW—which originally are oceanic trade winds—are obvious until several hundreds of kilometers into the interior of the western and southwestern Sahara.

The continental trade winds of the N to NE sector in the Saharan part of their trajectory are at the end of their course from the NE to the ESE sector. The winds of the E sector are called harmattan; they are dry and hot.

The winds of the tropical austral air mass:

These comprise the austral trade winds and the monsoon, issued from the Saint Helen anticyclone, which differentiate under the influence of dynamic and thermal factors. The West African monsoon is, by definition, meteorological; the trade winds flux is deflected, after passing the equator, by the transequatorial declivity between the high austral oceanic pressures and the low boreal continental pressures. South of the ground trace of the ITC, the monsoon is a variable wind from the sector S to W.

The nonanticyclonic winds: the winds associated with the "grain" lines

The lines of grain appear as "fronts" and move from east to west at an average speed of 60 to 80 km/hr, both well above the continent and the Atlantic Ocean. As they develop in the zone C1 (see the meridional cut of the West-African troposphere, Figure 3), they consist of storms joined together, or of "grains" that spread out N-S over more than 1,000 km and the width of which (E-W) can reach from 50 to 150 km. The core of the perturbation consists of cumulonimbus clouds, of large vertical development (>12,000 m). Every grain line (25 of them per year on the average) is preceded and accompanied by violent winds, blowing blasts, from E to the S sector. The sand winds, which are trade winds or winds associated with the grain lines and the vortices, will be analyzed later.

This list of characteristic fluxes shows that there is no aerological Sahel, since the winds that manifest themselves

Figure 3
Schematic cut of the troposphere along the meridial origin
(July) derived from ASECA documents.

there have their origin either in the Maghrebo-Saharan region, above the Atlantic at subtropical latitudes, or above the southeast Atlantic Ocean. The Sahel is a zone in which the winds manifest themselves more or less permanently and more or less efficiently, according to the seasons and the years, as a function of particular meteo-dynamical conditions.

It is also interesting to note that the trade winds blew continuously for about 10 months of the year in the region situated around 14° - 16° N during the recent drought (1968-70), but during the 1950's they prevailed for only about 6 months, accompanied by excessive rainfall.

The Anemometric Measurements

The first wind studies were undertaken by the "Service meteorologique de l'Afrique Occidentale Française" (Meteorological Survey of French West Africa) followed in 1960 by the "Agence pour la Sécurité de la Navigation Aérienne" (Agency for the Security of Aerial Navigation), ASECNA. ASECNA is in Dakar, equipped with a center for computer analysis. The anemometric measurements are performed for the most part in the stations of the West African network using either recording anemometers or chronototalizers, which can measure the distance covered per unit of time.

Moreover, in every station equipped with these recording instruments, periodic observations of speed and direction are made. We specify here that the instantaneous speed is in fact, in the meteorological data, the average of the wind speeds during the 10 minutes preceding the observation time. According to the station, the measurements are performed every hour or every 6 hours.

Our knowledge of the wind is far from being perfect; only scanning and analysis may add the necessary data to the lacking information. On the other hand, the measuring instruments are placed between 6 and 15 meters above the ground; it is therefore difficult to measure the wind speed and efficiency at ground level. The location of some stations has been very badly chosen in certain cases--at the foot of a relief, in a valley, on a promontory--although it is well known that the topography of a site can decrease or increase wind speed or modify its trajectory. The main stations: Dakar, Niamey, and Bamako also perform soundings, which enable us to establish complete tropospheric cuts.

Wind speed

As a general rule the trade winds are winds of low speed and even irregular, as it can be seen from the high percentage of calms. One should however distinguish the oceanic trade winds (the fast and regular ones that are of interest to the

Atlantic seaboard) from the continental trade winds, that are weaker and very irregular (see Tables 1, 2, 3, and 4). This contrast has both geographical and meteorological causes: the spatial distribution of the centers of action, the thermodynamic characteristics of the lower layers of the troposphere, and the nature of the surfaces swept by the wind.

Regarding the continental trade winds, their speed decreases regularly from the Central Sahara to the Sahel, where it is seldom higher than 2 to 3 m/s (Figure 4).

There is therefore a gradient in the speed decrease, correlated with a gradient in the decrease of the efficiency which, as we shall discuss in more detail later, is the determining factor of the desertifying and morphological effects of wind. The monsoon winds contrary to what is generally assumed, are regular, and their speed is far from negligible. The winds associated with the grain lines are of short duration (15 to 30 minutes), but violent and turbulent.

Efficiency of the Wind

The efficiency of the wind is a function of its speed, but also, and above all, of the nature of the medium on which it exerts its action. One should therefore distinguish the theoretical efficiency, which is only a function of the speed, from the real efficiency. According to Bagnold (1965), a wind is morphologically efficient when its speed is higher or equal to 5 m/s; such a wind can displace sand grains of an average diameter of 500 micrometers by rolling them, uplift those of a diameter of 200 micrometers (saltation) and transport by air (suspension) all those whose diameter are smaller than 200 micrometers.

The main factor that limits the dynamic efficiency of the wind is vegetation; the real efficiency is therefore larger in the Sahara than in the Sahelian regions, where vegetation is more abundant. In the Sahel, although the flora composition of the vegetation is more homogeneous, the landscapes and the vegetal groupings, in agreement with the variety of the biotopes, vary considerably. The sandy plains and the pleistocene dunes, covered by bushy steppes or trees, are opposed to the rust-covered pediments, and to the slopes and the open areas of scattered vegetal covering. The natural environment consists of a juxtaposition between the areas of deflation, of transit, or erosion, and of protected areas. The grass cover, where it exists, protects the soil and even favors the accumulation of particles by trapping them.

The Sahelian environment is in a situation of precarious equilibrium, and this equilibrium, during the last ten years, has been seriously affected by an important lack of rain and by human activities. The vegetation has been seriously degraded; if the drought has caused the death of large numbers of

Table 1.
DAKAR STATION--WIND DIRECTIONS FOR 1950-1960
(Frequency Expressed in %)

Directions	N	NNE	NE	ENE	E	ESE	SE	SSE	S	SSW	SW	WSW	W	WNW	NW	NNW
Dakar	26.59	15.98	4.82	1.28	1.63	0.86	1.06	1.01	1.77	2.11	2.45	2.97	4.34	5.39	8.39	19.28

Table 2.
DAKAR STATION--SPEED OF WINDS IN THE DRY SEASON, 1966-1972
(Frequency of Winds for Which the Speed Equals 0-5 m/sec, etc.,
Deficient or Excessive in Relationship to Average 1900-1972.)

Speed Year	0-5 m/sec	6-10	11-15	>15	In mm	Deficient or Excessive
1966-67	5.63	52.30	38.65	3.38	595.0	+8
1968-69	6.48	53.17	37.64	2.62	259.4	-53
1969-70	6.93	48.34	40.66	3.79	751.3	+37
1970-71	4.92	37.27	50.52	7.21	177.0	-68
1971-72	1.51	17.93	60.33	19.74	366.7	-33

Table 3.
CONTINENTAL SAHEL--DIRECTION OF WINDS FOR 1953-1960
(Frequency Expressed in %)

Directions	N	NNE	NE	ENE	E	ESE	SE	SSE	S	SSW	SW	WSW	W	WNW	NW	NNW	Calm
Bilma	7.3	11.80	14.36	8.98	11.40	10.22	7.39	4.43	4.63	3.42	2.07	1.64	2.13	2.38	4.34	3.36	42.2
N'Guigmi	7.55	1.18	5.68	4.68	16.89	5.52	12.24	4.90	13.74	1.28	9.11	1.49	10.68	0.87	3.31	0.81	70.36
Tillabéri	11.93	0.34	13.61	1.20	18.56	0.76	9.51	1.08	18.51	0.95	6.67	0.51	8.81	0.57	6.29	0.61	17.22
Tahoua	4.21	1.48	16.74	4.91	16.26	2.92	7.10	1.77	11.46	4.37	10.82	2.64	8.19	1.85	4.45	0.75	7.71
Dori	11.03	1.91	7.19	6.11	17.86	3.59	1.67	7.07	16.54	4.67	1.91	3.71	9.23	3.11	0.8	3.47	29.3

Table 4.
SPEED OF WINDS IN THE DRY SEASON FOR 1953-1960
(Frequency of the Winds of a Speed Equal to 0-1.9 m/sec, etc., Expressed in %)

Speed in m/sec	0-1.9	2-4.9	5-6.9	7-14.9	15-20.9	>21
Bilma	37.05	43.27	10.87	8.35	0.52	0
N'Guigmi	42.81	49.55	6.17	1.43	0.02	0
Tillabéri	16.50	61.92	17.36	4.15	0.05	0
Tahoua	8.61	62.92	20.15	8.30	0	0

Figure 4
Average annual isotachs of the wind.

woody plants (trees and bushes), the crops, the grazing, the trampling of cattle, the pruning and cutting of trees has locally destroyed the natural protection of the soil in a proportion varying from 20 to 70%. It follows that nowadays, as the rainfall is still showing a deficit, that the wind is the decisive factor of desertification, as will be shown later in this article.

Analysis of the Anemometric Data From Some Sahelian Stations

Dakar Station--Atlantic Sahel. The oceanic trade winds are typical of this station. Coming from the NW to NNE sector, they represent 70% of the total of the measured winds, and blow for the entire dry season (from October to May). The monsoon winds (S to W) are clearly of minor importance, being only 13% of its total (Table 1).

From the systematic study of the winds in the dry season between 1968 and 1972, performed by M. F. Courel (1974), it follows that:

- 90% of the measured winds are morphologically effective; (Table 2)
- The winds coming from the N to NE sector are of higher speed than those coming from the sector NW to NNW
- The (theoretical) efficiency of the winds is maximal during February, March and April and between 12 and 3 p.m.

It appears moreover that the average speed of the trade winds is higher during the period of drought than during periods of rain. This could be a consequence of the persistence of the Azores Anticyclone at an average latitude that is abnormally southern; this is just a hypothesis; which to be confirmed, would require a comparative analysis of the major centers of action that rule the dynamics of the air masses and the characteristics of the air fluxes above West Africa.

Stations at Bilma, Tahoua, N'Guigmi, Tillaberi (Niger) Dori (Upper Volta), Tombouctou (Mali), Continental Sahel. The wind roses (Figures 5a and b) are very distinctly different from those of Dakar: the continental trade winds from the sector NNW to NE, the harmattan from the sector ENE or ESE and the monsoon from the sector S to W are all well-characterized. For example, at Tillaberi the respective frequencies of these winds are 26, 30 and 35%. The monthly directions vary from year to year, but the monthly wind roses conserve the same fundamental characteristics (see Table 3 and Figures 4 and 5).

Speed and efficiency

In the continental Sahel the wind velocity is lower and it is quite rare that the efficient winds are more than 25% of

Figure 5a
Average wind roses at Bilma from 1966-1971 and 1975.

Figure 5b
Average wind roses at N'guigmi from 1966-1975.

the total of the measured winds (Table 4). Analysis of the data of the Tombouctou and Dori stations (Years 1972-73) gives the following details (Tables 5 to 7).

At Tombouctou the efficient winds (19% of the measured winds) are more frequent just before and during the rainy season (21.50%) than during the dry season (17.4%). They are from the sector S-SW (20%) during the first period and from the sector NNW to ENE (80%) during the second.

At Dori efficient winds are rare; 5% of the measured winds. They prevail during the rainy season, and come from the sector ESE to SSE. During the dry season the effective winds are from the sector N to NNE.

Comparison of these two stations, one of which (Tombouctou) is at the border of the Sahara, the other (Dori) is in the center of the Sahel, is a good illustration of what has been stated previously, namely that the speed and the theoretical efficiency decrease regularly from north to south (Figure 4).

Winds and Currents of Air That Are Always Efficient: Sand Winds and Vortices

Sand winds. Sand winds manifest themselves mainly in the Saharan regions where the yearly rains are less than 100 mm, but their secondary effects spread into the Sahel. Among these the most important is the phenomenon of dry mist. Fine particles (aerosols), which may be clay, microcrystals of evaporite, and quartz particles, load the low and medium layers of the troposphere up to 3,000-4,000 m altitude, considerably reducing insulation and visibility.

The aerosols are dragged over very large distances by the high-altitude winds from the E sector; one part of them is deposited on the continent, and the rest in the Atlantic Ocean, from the African coast to the Gulf of Mexico. The sand winds (or sand-driving) are also found in the Sahel, and one has noticed their outbreak during the recent drought, (see Table VIII; Mainguet and Chemin, 1979). Although at latitudes higher than 18°N their origin is tied to the strengthening of the continental trade winds during January, February, and March, at the lower latitudes sand winds are in fact the winds associated with the grain lines from just before the rainy season (April, May, and June).

The winds are less frequent, of a shorter duration, but as efficient as those of the Sahara. They can in fact uplift the dust to more than 5,000 m and also can uplift particles of a diameter from 80 to 100 micrometers to a few tens of meters. The speed of these winds can reach and even exceed 22 m/sec. The same grain line can sweep some hundreds of square kilometers from Chad to Senegal. The winds associated with it in

Table 5.
STATIONS OF TOMBOUCTOU (MALI) AND DORI (UPPER VOLTA)
MONTHLY DISTRIBUTION OF EFFICIENT WINDS (Expressed in %)

Station	Month	J	F	M	A	M	J	J	O	S	O	N	D
Tom- Bouctou	1972	0.27	0.83	2.67	5.9	10.24	15.97	16.10	15.42	8.2	9.23	6.64	8.49
	1973	8.50	8.24	10.62	10.15	8.83	13.45	17.15	7.25	7.71	2.11	4.15	1.78
Dori	1972	8.12	8.12	25.62	7.5	4.30	7.5	11.25	2.5	5.0	5.0	6.25	8.75
	1973	5.00	4.56	8.12	7.61	11.16	27.41	14.21	4.56	3.55	1.01	12.69	0

Table 6.
STATION OF DORI--ANEMOMETRICAL DATA 1972-1973
(Frequency Expressed in %)

Directions of Winds in Tens of Degrees	2	4	6	8	10	12	14	16	18	20	22	24	26	28	30	32	34	36
Annual Winds	7.74	10.19	7.16	3.43	0	4.16	8.15	11.06	7.45	4.47	6.7	4.45	1.96	0	2.37	7.31	5.42	7.91
Efficient Winds (>5 m/sec) Nov 1972-Sept 1973 inclusive	9.70	16.50	8.25	1.45	0	12.13	11.16	12.13	4.42	2.91	0.9	2.91	0.9	0	1.94	4.36	0.9	11.16
Efficient Winds (Dry season: Nov-May inclusive)	17.47	25.24	7.76	1.94	0	4.85	4.85	5.82	0.9	0.9	0.9	1.94	0.9	0	1.94	6.79	1.94	15.53
Efficient Winds (Rainy season: June-Oct inclusive)	1.94	7.76	8.73	0.9	0	19.41	17.47	18.44	3.88	4.85	0.9	3.88	0.9	0	1.94	1.94	0	6.79

Table 7.
STATION OF TOMBOUCTOU--ANEMOMETRICAL DATA 1972-1973
(Frequency Expressed in %)

Directions of Winds in Tens of Degrees	2	4	6	8	10	12	14	16	18	20	22	24	26	28	30	32	34	36
Annual Winds	15.14	13.46	14.81	9.34	9.93	2.70	1.05	0.96	1.09	0.8	1.19	1.15	0.39	0.4	1.21	1.97	10.5	14.48
Efficient Winds (>5 m/sec) Nov 1972-Sept 1973 inclusive	9.4	8.8	9.17	7.21	5.5	1.89	2.62	2.99	7.58	3.79	13.88	3.91	3.60	2.87	1.83	3.05	4.77	6.97
Efficient Winds (Dry season: Nov-May inclusive)	16.19	14.51	16.31	10.46	8.21	2.69	0.7	0.7	1.46	1.12	1.68	1.34	0.89	0.2	1.34	3.82	7.87	10.23
Efficient Winds (Rainy season: June-Oct inclusive)	1.47	2.01	0.6	3.35	2.27	0.9	4.82	5.63	14.87	6.97	28.41	6.97	6.05	6.03	2.41	2.14	1.07	3.08

this way can transport considerable quantities of aerosols and sand grains as a whole over shorter distances.

Vortices. Although vortices are not winds in the meteorological sense, we believe they are as frequent in the Sahara as in the Sahel. They load the lower layers of the troposphere with aerosols and move the sand locally. As they are of thermal origin, they can only be formed on bare surfaces; their movement can be one of three types: circular, ascending, or linear. The original ascendance is due to a hyperadiabatic gradient of the air above an overheated substratum of a limited surface. (Note: The caloric accumulation power of the soil is a function of its nature and its albedo; therefore over an iron crust the air temperature is higher than over a clay sand soil, or over a steppe-covered soil.)

The rotational and ascending speed of the air is high, as it allows sand and dust to uplift and to transport. The ascendance can be between a few meters and several tens of meters. The vortices have not been carefully and extensively studied yet. The vortices that cover several hundreds of meters have a duration that seldom exceeds 5 minutes. They take place during the hottest hours of the day, between 12 and 3 p.m.

Wind Trajectories at Ground Level

Ground Indicators of the Aeolian Transport of Sand. For a study of winds as desertification agents, it is interesting to complement the data on the direction, obtained from the anemometric stations, with observations at ground level, essentially in order to estimate more correctly the role of topography in the modification of the fluxes. The effects of topography can be detected in satellite images, on aerial photographs, and on the ground, in part from indicators, which do in fact trace out the lines of circulation of the sand and enable us to follow them. On the satellite images, the transport axes appear as bundles of clear lines, because they correspond to trails of sand, or because they result in a clear corrosion effect on rocks, covered by a dark desert coating.

In the Sahara the sandstones are tinted by desert varnish or coats of patina. The sand winds wear down the patinas and the clear sandstone beneath appears according to preferential strips, clearly visible in the satellite images in the Egyptian desert and north of Borkou (Northern Chad). These strips, parallel to the dominant wind, enable the wind trajectories to materialize and are one of the best criteria in the diagnosis of aeolian trajectories (Mainguet and Canon, 1976). This aeolian erosion, called corrasion, having a gouge effect, results not only in isolated furrows but also in deep periodical systems of crests and corridors; the most beautiful of

which are visible today; the Kaluts in Iran (Dresch, 1968) and the crests-corridor systems east and southwest of Tibesti (Mainguet, 1972b). The shelter effect exerted by certain isolated reliefs--an inselberg dominating vast pediments--can also be a good indicator of the aeolian direction (Mainguet, 1976). In fact, leeward of certain residual reliefs, such as those one can find north of Hoggar or downwind of the residual plateaus of Koutous and Tanout in the Sahel or Niger, satellite images show very dark triangular forms that correspond to areas of rocky outcrops, covered by a dark brown coating. The triangles lean on the leeward slope of the hill, with the bisecting line of the triangle indicating the direction of the average or the dominant wind.

These spearhead-shaped forms are therefore also a criterion for diagnostics of the aeolian currents. They constitute an especially interesting tool, since the bare rock, with its ferruginous coat that is not very reflective, contrasts with the very reflective clear color of the sand in the neighboring areas. Leeward (namely SW) of the small Koutous massif ($14^{\circ}30'N-10^{\circ}E$), which constitutes an impassable obstacle to the trajectories of sand, the longest shadow effect of aeolian action reaches 230 km, with a width of 25 km. These areas are repellent for the farmers, since the rocky substratum outcrops are barely covered by a fine veneer of sand.

From the images of NOAA 3, 4, 5, Meteosat, and Landsat, it has become possible to determine that the wind, across the Sahara and up to the southern border of the Sahel, consists of a flux, drawing continuous trajectories, along which the local transport of sandy material is discontinuous (Mainguet, 1972a; Mainguet and Canon, 1976). These large trajectories are represented in Figure 6. Areas sometimes dominated by deposition and sometimes by transport relieve one another along these trajectories, as dictated by topographical features (essentially the great mountain groups Eglab, Hoggar, Air, Tibesti).

Large Sand Transport Trajectory, Its Meeting With Tibesti Mountain and Its Evolution in the Sahel. In order to show the effect of a mountain on aeolian transport, we have chosen to present one of these trajectories, precisely the one that splits into two branches when approaching Tibesti (Mainguet and Callot, 1978).

One of the branches, after having turned around Tibesti from the east, sweeps to the north of Chad, and ends up in the area of deposition located to the northeast of Lake Chad--the Manga erg. The other turns around the mountain from the west, with a path that begins almost N-S, but afterwards turns towards the SW. It comes into the Fachi-Bilma erg leeward of the mountain, crosses the erg and ends at the area of the Haoussa erg, an erg of Niger, which is fixed by vegetation. This air current, which is trans-Saharan, therefore does not

Figure 6
Trajectories of sand transport in the Sahara (remote sensing of images from NOAA-3):

1. Atlantic sand current
2. Western sand current
3. Cellular circulation in the central Sahara
4. Eastern sand current
5. Obstacles upwind zones without sandy deposits
6. Ergs or surfaces of sandy borders: (a) Admer; (b) Chech; (c) Fachi-Bilma; (d) Iguidi; (e) Mourzouk; (f) Western; (g) Eastern; (h) Rebiana; and (i) Makteir.

stop at the southern border of the Sahara, i.e., at the 150-mm isohyete, but continues across the Sahel until the 625 mm isohyete, showing clearly the existing interdependence, persistent although diminished, between the arid and Sahelian areas, in the same way as during the time of the great aridity when the Sahara and the Sahel were forming only one paleo-Sahara. This occurred precisely in the strip of land between these two isohyetes, where one finds the sectors of reactivation of the sandy mantle.

The NE-SW circulation of winds at ground level follows the direction of the trade wind until the beginning of the Sahel. Inside the Sahel, a very distinct E-W direction appears; we have been able to study this phenomenon in the Niger Sahel, between the Air and the Niger River.

The Wind Action on the Ground Between the Air and the Niger River. The marks on the ground resulting from the sand-transporting winds, and reflected by the axes of the barchans and by the striations of corrasion and deflation strips, have a general NE-SW orientation: the direction of the trade wind, as it has been described as leeward of Tibesti and of the Koutous massif (14°30'N - 10°E). In the real core of the Sahel, new directions appear in the west of the Ader Doutchi massif in Niger (5 to 6°E). The massif is locally carved into small yardangs, with an E-W orientation similar to those of Koutous, providing another indication of this direction: manifested in spearhead shapes, ranging from 60 to 100 km in length being created leeward of Ader Doutchi and flowing towards the dallol Maouri (4°10'E).

The sandy aeolian edifices covered with vegetation are oriented according to four directions, identified through photointerpretation, as it is difficult to determine the exact orientation at the ground level of a several kilometers long dune system.

Beginning around 14°15'N and 5°E, at the southwest foot of Ader Doutchi the direction ESE-WNW is shown by the alignment of sandy hills, and then further towards the west through the crest line of sand ridges, that characterize the Nigerian west. The aforementioned edifices are asymmetrical, strongly inclined towards NNE, with a gentle slope towards SSW. They begin on the western bank of the great dallols, dry emissaries of the valleys going down the Air and the Adrar of Ifoghas, and crossing the Continental Terminal plateau towards the west where they release barchanic edifices, oriented along an E-W axis. These sand ridges, as the hilly edifices with a NE-SW linear deposition are oblique with respect to the direction of the wind.

None of these edifices are completely cultivated with millet; a severe reactivation can be seen in the aerial photographs taken in 1975. The eastern part of the sandy ridges, situated on the western bank of the dallols and the Niger

river, is active; the sand is mobile on the surface and the vegetation is reduced to a few bushes. The sand ridges decrease in number towards the west, both between the Bosso dallo and the Niger River or in the region from the Niger River to Upper Volta. The surface of these ridges is characterized by the outcrop of the B horizon, producing a "moire" effect in certain areas on the aerial photographs.

In this region of west Nigeria, where sand is scarce, there is a sector of transverse edifices, oriented N-S, which is localized upwind of Ader Doutchi. These edifices in the Tahoua-Abalak region are strongly reactivated by wind erosion. Some tiny white lines, or some clear gray-to-white areas are visible on these structures through analysis of the aerial photographs. On the ground, the clear areas are sectors of reactivation of the wind affecting either the top of the edifices themselves or the interfluvial areas between the ravines that are sometimes fixed, sometimes functional, and that end at the foot of the slopes in fans of sand, that are able to be set into motion by the wind.

There are numerous barchan-like edifices between the Ader Doutchi and the Niger River. They always have the same orientation; their convexity turned towards the east and their E-W axis indicate that they were deposited by E-W winds. At the western foot of the Ader Doutchi, they are grouped into a "duck flight" shape, the sharp-pointed eastern end of which is directed towards the east, which is also situated about an E-W axis. The E-W direction is marked out by these duck flight shapes up to the Niger River, and by isolated structures on the tabular plateau of the Continental Terminal.

The barchan-like structures can most easily be identified through photointerpretation. A difference in the condition of the surface appeared between 1955 and 1975. In 1979, some of these edifices were cultivated with millet, but they were covered again by a thin layer of mobile sand that submitted to winnowing action, creating a concentration of coarse sand particles and small nebkas formed leeward of every millet stalk. In 1955 the fields were extended leeward by scratches: fine white striations were caused by the progression of the active sand upon the fixed dunes. In 1975, in the aerial photographs, the areas of cultivation regressed and the edifices were scratched with N-S stripes with alternate white and dark gray belts etched in a pattern similar to "akles." Such is the case of the deposits on the western foot of Ader Doutchi.

Other barchanic edifices are left as a very poor fallow field. The entire horizon of mobile sand that could be used in millet cultivation has been exported; the red, lightly aggregated, B horizon consists of outcrops covered with a "glaze." It is sculpted into small aerodynamic forms of the yardang type. This is the case for most of the sandy edifices on the Continental Terminal plateau at the NE of Tillaberi (Goldgom

hill, 14°26'N-1°21'E). Because of their size these dunes can hardly be seen on the satellite images and therefore cannot be analyzed.

This example of the northwest of Niger shows that the actual mechanisms of desertification due to the wind consist of a remobilization of sand under the effect of specific manifestations of the arid zone; it is not the desert that breaks into Sahel, but rather mechanisms of the arid zone extending and rehandling the sandy material of the ancient dunes.

DESERTIFICATION--SOME CASE STUDIES

The Cap Vert Peninsula

The Cap Vert Peninsula, located between 14°30 and 15°N and 17° and 17°31'W, has a surface 600 kilometers², two-thirds of which is covered with dunes, some of which are fixed. Courel (1974), based upon morphological, sedimentological, pedological and stratigraphic criteria, has distinguished between the following five dune systems:

1. Camberene I-middle Pleistocene
2. Ogolian--28,000-10,000 BP
3. Camberene II
4. Yellow dunes <°3,000 BP
5. White dunes sub-present and present

Only coastal white dunes are active; they extend along the coast between Dakar and Saint-Louis and their width ranges from 0.5 to 1.8 kilometers. The natural vegetation is reduced to just a few discontinuous colonies of Cyperus maritimus, Schizachyrium pulchellum, Alternanthera maritima and Ipomea pescaprae. This grouping consists of parabolic edifices with a NNE-SSW orientation and barchanic edifices, the crest lines of which trend ESE-WNW.

Present Morphological Evolution

The coastal dunes.

The morphogenesis and the dynamics of these dunes are independent of climatic data and of aridity. At every latitude one finds coastal dunes; their model, which varies from one place to another, is obviously dependant only on the wind regime. Everywhere man tries to limit their extension by planting them with suitable herbaceous or woody species of plants. Since the Second World War, the General Government of French West Africa has started the creation of perimeters of filaos (Casuarina equisetifolia) in order to limit the extension of the active dunes. Afforestation has been continued since 1960 by the Water and Forest Authority of Senegal. These belts are rectangular in shape, but are separated by

considerable distances, stretched parallel to the coast for about 20 km between Youi and Kayar, covering a total area of about 500 hectares. This plantation had the aim of stabilizing the sand, but it does not seem to have reached its goal; every year the extension and the progression of the dunes increases.

The different origins of this phenomenon add to each other:

- The planted belts do not form a continuous windbreak and the intervals between two parallel belts are too wide as compared with the height of the trees
- The dunes are continuously fed from the beach, which is a permanent source of sand. This is formally proven by the ration of the granulometric indices of the sand of the beach and of the coastal dunes
- The trade winds, efficient in the coastal sector, were very dynamic during the years 1968-78
- The sand is fine; on the average, 210 micrometers, as shown by the categories of sand, which result from the analysis of 25 samples.

It follows that the sand is particularly mobile. According to M. Sall, the quantity of sand moved by the trade winds over a width of 1 meter should have been, in March 1970, on the order of from 1.3 to 2.9 tons.

The dunes progress; their progression varies between 2 and 7.5 m per year (Courel, 1974). They gain gradually over the interdune depressions, the naay, whose organic soils support gardening intended for market sales, or over lacustrine depressions (lakes Yout, Ourouaye, M'Beubeuss, Malika, and Karar). In only the past few years has one been able to observe the coastal sands covering the fixed dunes.

In the case of Cap Vert there is, therefore, a desertification caused by the shifting of active dunes.

Fixed Dunes (Camberene I and II Systems, Yellow Dunes and Ogolian)

Until about 1960 the dunes in the interior of the peninsula were effectively protected by herbaceous, shrubby and/or arboreous vegetation, that covered the peninsula. However, it has to be emphasized that there was an excess of rainfall in the 1950's.

From 1960 to 1978 we have seen the progressive degradation (locally irreversible), of this natural protection, especially between Dakar and Rufisque.

This evolution is due to the extension of the Dakar urban agglomeration and to the human settlements, to the trampling by herds walking towards the refrigerated slaughterhouses of Dakar or towards the Bambilor ranch (north of Dakar), and to the 1968-78 drought, which has caused the lowering of the

underground water levels in the interdunal depressions. The action of man however has been the main deteriorating cause.

The creation of large residential centers on both sides of the Dakar-Yoff highway, between Camberene and Pikine and between Hann Bay and the Dakar-Thies railroad, the opening of several sandpits has been required, as well as the leveling of large dunes and the partial filling of several depressions. The sand, uncovered in this way and mobilized by the trade winds, is shaped into transverse dunes; the deflation is such that the streets of the recent districts of Pikine and Camberene have to be regularly cleaned of the shifting sand.

This desertification, which affects several thousands of hectares, does not stop its growth, and no practical counter-measure has been taken to date. It seems that the creation of a wind-breaking system, including *Euphorbia balsamifera* and *Anacardium senegalensis* would suffice to stop the desertification.

The Voltaic Sahel

The causes of desertification are more abundant in northern Upper Volta, where the rainfall decreases from 600 to 350 mm between 13°30' and 15°N, than in the Gourma plain in Mali, where permanent water holes are quite rare. This is a result of several morphological, climatological, and human factors.

The Sahel in Upper Volta appears as a wide plain, sprinkled with modest reliefs, hills and schistous secondary mountain chains, granitic or gabbroitic inselbergs, at an altitude from 50 to 200 meters, in relation to the geological units of the precambrian shelf. The most remarkable morphological element is the dune relief, consisting of large sand ridges, oriented from ESE to WNW, several hundreds of kilometers long, from 1 to 4 kilometers wide, from 20 to 30 meters high, and spaced from 10 to 50 kilometers from each other. These pleistocene dunes, in the moment of their formation, have dammed most of the affluents and subaffluents of the middle-east Niger River that created several perennial or semipermanent pools in the north of Upper Volta.

The interdune plains, where a thin layer of soil partly covers the substratum or the gravelly iron crust, are covered by a sparse vegetation, which strongly contrasts the bushy and arboreous steppe covering the dunes. The discontinuity of the vegetation cover is used by the morphodynamic processes--water and wind. During the rainy season the interdune spaces are the location of dense rilles and gullies, and during the dry season, they are the location of deflation. They are privileged areas for the transit of the sand and the formation of vortices.

In contrast, the dunes, when their steppic cover remains non-degraded, manifest an astonishing resistance against these agents. More than the drought of the last decade, it has been man who, directly or indirectly, has destroyed the equilibrium of the dune biotope, especially close to the watering holes.

The main causes of the degradation of the steppe are:

1. Crops, and their very slow rotation; the clearing of land; the cutting away of the branches of the ligneous trees, and fires (very rare)

2. Grazing and (even more), trampling. When the steppe offers excellent grazing conditions, it is only little exploited; its mechanical degradation is essentially due to the twice-daily passing of very large herds, which go watering at the pools

3. The rain deficit, even when prolonged, has only a little effect on the natural regeneration of the gramineae. It reduces the grain production, but this decrease is compensated for by the germination of ancient grains (phenomenon of "dormancy"). Drought however, causes the mortality of the ligneous vegetation in the plain or on the slopes of the reliefs.

As a consequence, in the Upper-Voltaic Sahel the desertification manifests itself at first close to the villages situated on the dunes and at the outskirts of the pools, before extending to the whole cultivated area and the transit area of the herds. As soon as the gramineous carpet is destroyed, the dune soils deflate. The wind effects are favored by the granulometrical characters of the dune sands. They are in fact fine sands (on the average: 140 to 160 micrometers), well classified, and rather poor in clay. The particles which have a diameter of between 50 and 250 micrometers represent already 70 to 80% of the total. It follows that the winds of a speed from 3 to 4 m/sec are enough to move these particles; the very rare, efficient winds give gregarious active forms concentrated in akle. Observations at ground level show that the active dunes result from the remodelling of the sand inside the limits of the space previously bared.

Two cases of desertification have been studied in Upper Volta; they are the Menegou-Bidi sector, north of the semipermanent pool of Tassamakot, from 14°22'N and 0°18'W (Courel and Chamard, in press), and the Oursi sector northeast of the permanent pool with the same name, at 14°40'N and 0°27'W. The degradation of plant cover, already perceptible in 1955, has expanded during the recent drought. Thus, at Menegou-Bidi, the surfaces of degraded steppe and active sand have passed (between 1955 and 1974) from 181 to 390 hectares and from 56 to 150 hectares, respectively. At Oursi, where the average yearly rainfall is 50 mm less than at Menegou, the total surface of the dune active dune field has increased tenfold between 1955 and 1975; increasing from 22 to 250 hectares.

For the entire Voltaic Sahel, about 3,000 hectares of arable land have been lost for the agro-pastoral economy, namely the equivalent of 800 tons of millet or 4,500 tons of straw. Desertification makes it necessary to provide a series of arrangements, which moreover are not so onerous; complete protection of active dune areas, creation of corridors for the transit of herds, systematic planting of hedges of *Euphorbia* (*Euphorbia balsamifera*) in the whole cultivated area.

The Saharo-Sahelian Transition Zone in Mauritania.

This zone, the boundaries, of which are difficult to define, is interesting to the extent that it allows observation of the interdependent relationships that exist between the active ergs and the semifixed ergs, and to appreciate the morphological and desertifying role of the wind.

The akles. The akles that surround the large ergs: Majabat-al Kubra or Makteir, are interdependent upon the aforementioned qualities from the sedimentological point of view and their genesis appears to be linked to the obvious increase of the aridity in the last two millennia.

Although the body of an akle is hardly mobile, it feeds and emits barchans, the movement of which is a function of the wind speed and the nature of the circulation area can reach several tens of meters per year. The barchans, which move collectively in families (or "rebt"), can produce seifs through coalescence. In this case one can say that the desert advances.

The seifs can also begin on an ancient dune system, from a generally fine sand, which is created by the winnowing of an heterometric sandy stock. The orientation of these recent dunes, which testify to an increase in aridity, is different in general from that of the sand ridge on which they were established. In the Makteir erg their orientation is NNE-SSW, while the large ridges are generally NE-SW. This difference can only be explained by a modification of the overall wind regime, but not necessarily of the directions.

The Makteir erg in its central part (20°30'N) appears as a rhythmic wavy formation of NE-SW dunes lacking vegetation. Very frequently efficient winds move and winnow the sand at the surface. One can see that the particles have an average size of 630 micrometers at the top of the undulation, and 800 to 1,000 micrometers inside the large waves that cover the axis of the interdunal spaces where the canalized wind has a larger speed. At the depth of 1 meter the average size is lower: 250 to 400 micrometers. The gradual increase of the size of quartz sand at the surface gives a firm stability to the dune and to the whole erg with respect to the very frequently efficient winds, whose speed is generally between 5

and 11 m/sec. The mobilization and the winnowing of sand are only efficient beyond the last speed.

In the southern part of the erg, where the rainfall averages from 50 to 70 mm/year, the dunes are covered with a very sparse vegetation, of perennial tufts of *Aristida pungens*. Their thick bases trap sand, which becomes the support of the nebkas and can reach from 3 to 5 meters. The material of the nebka is a fine to a medium sand (on the average, 200 to 315 micrometers), which represents the lower fraction of the winnowed sand upwind of this part of the erg. This proves that a vegetation cover, even when discontinuous, can slow down the progression of sand.

Similar observations have been made in the ergs Hammami, Akchar, and Azefal at the western border of Ouarane, close to Chinguetti. Near this oasis, the progression of active sands is especially remarkable: between 1968 and 1972 the deeply embanked bed of the oued Chinguetti was occupied by a rebt evolving into an akle and the rocky surface of the plateau, west of the oued, covered in the same time by a sandy mantle, from 0.10 to 0.40 m thick and more than 1 kilometer wide. In the Mauritanian Sahel, properly so-called, the reactivation of the dunes proceeds by the same agents as those already described at Senegal or Upper Volta. In the fixed erg of Trarza, especially near Boutilimit and Mederdra, the pressure of cattle has been such that it originated massive active dunes.

In summary, the Sahara advances progressively on the border of the active ergs: in the Sahel the degradation of the vegetal cover and the wind suffice to create Saharan landscapes again, even where the yearly rainfall reaches 600 mm (Chamard, 1970; Chamard and Courel, in press).

The Saharo-Sahelian and Sahelian Zones of Niger

We have mentioned the Saharo-Sahelian relations in the past, and we have seen that the overflow of clayey-sandy material can still be relevant in the form of the dust-blowing wind and as barchanic arrival, as is the case of Morocco, where dunes, pushed by the winds called "Saheli" and "Chergui" progress toward the palm tree zones. The same localized phenomena occurs in Mauritania and Senegal, but nothing similar is observed in Upper Volta nor in the Nigerian Sahel, where the geomorphological mechanisms of desertification in which the wind takes part, are deflation and corrosion, sand sprinkling and an avalanche effect, construction of edifices of the nebka type, or a seif or akle surface. The invasion of sandy edifices caused by an allocthonous deposit of sand such as barchans, only rarely occurs. Conversely, the appearance of autocthonous active sandy edifices, the genesis of which corresponds to the incursion of typical arid zone mechanisms

like nebkas, seifs, and akle surfaces, are all quite frequent.

Aeolian Winnowing and Seifs At the North of the Saharo-Sahelian Zone.

In this steppe zone, where barrenness increases towards 16°N, desertification manifests itself by the effects of winnowing and the appearance of active seifs; in other words, this is an extension of the Saharan external mechanisms, into the Sahelian area.

The winnowing of the fine particles, having as a primary consequence the coarsening of the surface texture of the sandy edifice, constitutes the most elementary aeolian mechanism. We have observed this action at the northern border of the Saharo-Sahelian steppe, in the center of Niger, but also more southward, in the Sahelian region itself, namely in the region of Maradi; the millet granary of the Niger, where the soil around the villages lays above sandy quaternary hillock-shaped sheets that is subjected nowadays to a severe impoverishment due to the exporation of particles smaller than 360 micrometers.

The seifs are elongated sandy edifices that are rectilinear as a whole, more or less sinuous in detail, and possessing a triangular cross section, on which the sand behaves as if along a transport rail. At Niger, passing from the arid to the Sahelian zone, immediately to the east of Termit (16°N), the appearance of the seifs on the sand ridges, fixed by a pavement of coarse particles and covered by a Saharo-Sahelian steppe, is the most alarming manifestation of the progression of the desert-like landscapes. These seifs are formed at the expense of the sandy substratum, and are fed by the finest sand, sifted by the wind. A comparative granulometric analysis of both the sand of the seifs, and the sand ridges around Termit in Niger at the northern border of the Saharo-Sahelian zone, shows:

1. Frequency and cumulative frequency curves with similar profiles;
2. Fundamental differences concerning the size of the material: the median size in the seifs is 160 micrometers, and on the surface of the sand ridges it is 630 and 800 micrometers.

The sand ridge at a depth of 10 cm below the surface, consists of sand 46% of which is >630 micrometers and 6% of which is of an even finer material, <250 micrometers. There is therefore the presence of a predominant stock of coarse sand mixed with a finer stock taken off the surface, comprising material 50% of which is >800 micrometers, with no particles <400 micrometers. It is a pavement perfectly sifted by the wind. Only the coarse material remains in place, while the

finer material is exported by the aeolian currents and fed to the downstream seifs of the functioning aeolian system.

As far as the seifs are concerned, more than 60% of the material measures <160 micrometers and no grains are >250 micrometers. This is therefore a material easily moved by the wind.

If the seifs on the northern fringe of the Saharo-Sahelian zone are the local mark of desertification, their fine material, which is especially movable by the wind, represents a danger for the leeward regions. This is the case in the heart of Niger for the real Sahelian areas cultivated with millet.

The Desertification Indicators in the Real Sahelian Area of Niger.

No type of dunal edifice, whatever its shape, can be sheltered against wind erosion, nevertheless the largest fragility appears in the transverse chains. For example, in the sector around the Koutous massif of Niger. We notice also that, as soon as sand is present and the rainfall is higher than 200 mm, there is human activity. All the sandy Sahel is therefore occupied by man. It is this that causes the difficulty in isolating the degradation attributable only to the wind. At Niger one can see that the domain of the cereal crops extends in the zone of the Saharo-Sahelian steppe as soon as there is sand.

It is only at the extreme northern border of the Sahel that one can observe an action uniquely due to the wind. The other areas, however, as soon as they are sandy, are so densely populated that it is impossible to disentangle the actions that are solely aeolian. The degradation is more pronounced in the cultivated areas than in those that are only used for grazing. In general, overgrazing produces an effect that is reversible whenever a type of defense is installed. Nevertheless, in 1979, in the southern part of the Ekrafane ranch in Niger (30 km north of Filingue), some barchan-like edifices were observed, progressing and constituting a case of irreversible degradation due to overgrazing. On the land cultivated with cereals, even after abandoning the cultivation of cereals, the effects of the degradation can be seen for quite a long period before returning to the original state.

On the Landsat photographs taken at 1:1,000,000 scale of Niger one can see:

- A areolar deflation: areas of reactivation on the hillock formations and the transverse edifices. A structure with white-centered areas, with a diameter ranging from 2 to 4 km, sometimes coalescing appears particularly well on channel 7. The central black point corresponds to a village, and the white halo corresponds to the agriculturally cultivated

soil around it. The reactivation linked to the villages and the agricultural activity is so intensive that it can be even detected on satellite images at 1:1,000,000.

- A longitudinal deflation: the reactivation occurs along lines, but in a discontinuous way, and this makes it difficult to detect on aerial photographs at a scale of 1:50,000, but more easily on the satellite images, especially on channel 7 of Landsat.

A special photographic mission at 1:10,000 carried out in December 1976 enabled the recognition of more subtle reactivation mechanisms attributed to sand overloading in the Maradi region. Several bushes are surrounded by a halo of trapped active sand, which is a proof of a transport--probably subautochthonous--and a deposition of sand in the heart of the Sahel. The hedges that surround the plots of millet are covered by sand in a comb-effect, often elevating the height of the hedge more than 40 cm. We observed the effect of so-called avalanches i.e., the spreading of the sand over the hedge, perhaps at the expense of the sand trapped by the hedge. Other phenomena of sand removal exist, producing deflation craters around large trees, with an asymmetrical aeolian overhollowing, which finally denudes the tree roots.

The mechanisms of erosion and degradation of the landscapes by reactivation of the sandy sheets of the Nigerian Sahel were clearly visible on the aerial photographs of 1975 and on the satellite images after the 1968-1973 climatic crisis. However, (especially between 6-8°N), they are imperceptible or hardly existing on the aerial photographs taken in 1957.

Rather than a progression of a total front from the desert border, on the aerial photos expressions of desertification are created by the human parameters; desertification appears in two different patterns both in the form of isolated patches and strips: the patches result from a trampling around precise points and the strips appear along traffic lines. These locations are the initial grain-points for desertification.

As far as the desertification due to physical aeolian parameters, water parameters, and their combination is concerned, its localization does not follow the parallels. In fact, the degradation due to water precedes the aeolian desertification at the north, and this last one may reach the 625 mm isohyete.

CONCLUSIONS

The wind is certainly one of the factors responsible for the recent desertification in the Saharo-Sahelian and Sahelian regions, but the consequences of the aeolian actions are different in these two zones:

In the arid transition Saharo-Sahelian region, the wind, frequently efficient, expresses itself more easily as long as

the vegetation is very sparse or absent. It uplifts, sifts, moves or accumulates the sands and dusts: it models mobile forms--individual or in association: barchans, seifs, akle, found either at the border or even in the middle of the large, fixed Pleistocene sands. In this way a real progression of the Sahara and, a progressive but unequal appearance of mobile sand in the landscapes of the northern Sahel can be seen.

In the heart of the Sahel, the speed and the average efficiency of the winds, trade winds, or monsoons are weak. It is probably no longer the wind that starts desertification, but sedentary or nomadic man. The wind actions are the mechanisms of desertification. The plant cover, subjected to overexploitation and the effects of the recent drought, is locally very degraded if not destroyed. The wind, in contrast, after the destruction of equilibrium of the dunal biotope mobilizes the sand and creates dunes and fields, which each year become more vast. The most decisive agent of desertification and sterilization of the dune soils can therefore be found in the central Sahel.

It does not seem that the desertification of the useful Sahel must be irreversible. Man cannot modify the climate and in the Sahel, as everywhere, he cannot hope to return to a normal pluviometry. Perhaps most of all man can, in waiting, not degrade his surroundings anymore, indeed he might even attempt to restore it.

REFERENCES

- Chamard, P. (1970) Le Bassin-Versant de la Sebkha de Chemchane (Adrar de Mauritanie). Etude Geomorphologique. These Doct., Universite de Dakar, Senegal.
- Chamard, P. and Courel, M. F. (in press) Contribution a l'Etude du Sahel Voltaique. Causes et Consequences de la Degradation du Couvert Vegetal des Dunes. Secteur de Menegou-Bidi (Department du Sahel - Sous Prefecture de l'Oudalan). Mem., Universite de Reims.
- Courel, M. F. (1974) Etude Geomorphologique des Systemes Dunaires du Cap Vert (Senegal). Mem., Maitrise Univesite Paris VII.
- Dresch, J. (1968) Reconnaissance dans le Lut (Iran). B. Assn. Geogr. franc. 362-3, p. 143-153.
- Mainquet, M. (1972a) Etude d'un erg (Fachi-Bilma, Sahara, central). Son alimentation sableuse et son insertion dans le paysage d'apres les photographies prises par satellites. C. R. Acad. Sc., Paris, Ser. D., 274, p. 1633-1636.
- _____, (1972b) Le Modele des Gres. Problemes Generaux. Etudes de Photo-Interpretation. These de Doctorat d'Etat, Inst. Geogr. Nation.

- _____, (1976) Etude geodynamique de l'effect d'obstacle topographique dans le transport eolien. Resultats obtenus a l'aide des images aeriennes et satellites au Sahara. Soc. fr. de Photogram., B, 61, p. 31-38.
- Mainguet, M. and Canon, L. (1976) Vent et paleovents du Sahara. Tentative d'approche paleoclimatique. R. Geogr. physique Geol. dynam., 18, p. 241-250.
- Mainguet, M. and Callot, Y. (1978) L'Erg de Fachi-Bilma (Tchad-Niger). Contribution a la connaissance de la dynamique des ergs et des dunes des zones arides chaudes. Mem et Doc., 18, CNRS, Paris.
- Mainguet, M. and Chemin, M.C. (1979) Precipitations et Vents dans le Sahel de la Republique du Niger. Etude des variations dues a une crise climatique (1968-1973). Rapport No. 4. DGRST et GERDAT, Paris.

THE ADVANCE MECHANISM OF LONGITUDINAL DUNES

By

Haim Tsoar

ABSTRACT

Results of field work studying the mechanism of movement and advance for longitudinal dunes shows that wind that encounters longitudinal dunes obliquely is deflected on the lee flank and blows along that flank parallel to the crest line. The magnitude of the deflected wind depends on the angle of incidence between the wind and the crest line. When the angle is acute ($<40^\circ$), sand is transported along the lee flank and when the angle is less acute ($>50^\circ$) the magnitude of the deflected wind abates, and deposition takes place on the lee side. The sinuous outline of the dune crest line accounts for the alternation in the angle of incidence that causes erosion and deposition along the lee flank. The wind regime in the study area is bidirectional with the two dominant winds blowing from both sides obliquely to the dune, thus maintaining its symmetry and sharp profile.

INTRODUCTION

Longitudinal sand dunes, which are the most widespread type of dunes in deserts, are characterized by a symmetrical cross profile, a sharp-edged crest and continuous sinuate elongation (Figure 1). They comprise 72% of the dunes in the Sahara desert and are also dominant in Australia and Arabia (Jordan, 1964; Mabbutt, 1968). In contrast to transverse dunes, whose mechanism of movement and advance are essen-

Figure 1
Oblique aerial view of longitudinal dunes, Sinai Desert.

Figure 2
Circular diagram of the effective annual wind direction (percentage). Dashed line indicates the dune axis.

tially well-known, (Bagnold, 1954), the longitudinal dunes still pose many unanswered questions (Lancaster, 1980).

Several theories have been put forward to explain some of their typical characteristics—elongation for many tens of kilometers, equal distances between the dunes and sinuosity of their crest-lines. There are two main theories on the genesis of the longitudinal dunes. In one, the wind drift theory, their elongation is explained as a consequence of excavation of a previously deposited sand mass, leaving residual wind drift dunes (King, 1960; Folk, 1971). The opposite view is that the longitudinal dunes (like the transverse ones) are the result of the mode of sand deposition (Wopfner and Twidale, 1967), which raises the question: How are the elongated strips (of sand) deposited? Two theories attempt to explain this: according to one, their form is due to bidirectional winds (McKee and Tibbitts, 1964), and in the other, due to unidirectional wind and lateral vortices (helical flow) (Hanna, 1969).

Despite the various theories, the processes that create the dunes and the dynamics of the dunes are still uncertain. All the above theories are speculative and some have no basis in reality.

RESEARCH METHOD AND APPROACH

In the present research, it was not intended to prove or disprove the theoretical premises proposed by other investigators, but rather to examine the actual processes and phenomena that occur on the dune itself.

The research was concentrated on a field study of the mechanisms of movement and advance of the dunes and on the reciprocal relationship between their morphology and dynamics.

Research methods included field measurements on a typical longitudinal dune in the Sinai Desert: a macro- and micro-wind direction and velocity measurements, grain size analysis, tracing the movement of sand grains by marking them with a fluorescent dye, analysis of the internal structure, measurements of the rate of advance and of profile changes during several seasons.

RESULTS

Results of the macro-wind measurement show that the longitudinal dune is under a bidirectional wind regime (Figure 2). Results of the micro-wind measurements show that when wind encounters a dune body perpendicular to the crest line, the streamlines are separated from the surface over the brink and are reattached to it a certain distance downward (Figure 3). This physical phenomenon is known in "Fluid Mechanics" as

Figure 3
Separation of flow above a profile of a longitudinal dune
(with wind profiles).

Figure 4
Plan of the changes in the horizontal wind component that is
perpendicular to the crest line. White arrows indicate wind
horizontal components and black, the actual wind direction.
(1) - The situation before the separation. (2) - The situation
in the reverse flow region. (3) - The situation on the line
of attachment. (4) and (5) - The situation outside the separa-
tion area.

"separation of flow." In the "space" that is formed between the point of separation and the point of attachment, we can usually find a light reverse flow that is the consequence of the vortex that is formed there. At the point of attachment, the velocity of the wind is zero, because from that point the streamlines are split into two contrary directions: streamlines that are moving to the crest and streamlines that continue in the general wind direction. The drastic fall of velocity on the lee side of the dune leads to deposition of sand there, and to a creation of a slip face, which is characteristic of transverse and barchan dunes. On the longitudinal dunes, we find the same process of separation, but in this case the wind is blowing obliquely to the dune, which makes the process of separation more complicated.

In this case, the separation will be performed, but the wind will have two components: one perpendicular to the crest line, and the other parallel to it (Figure 4). The separation is formed by the component that is perpendicular to the crest line. Therefore, the distance between the point of separation and the point of attachment is controlled by the magnitude of the wind and the angle it formed with the crest line. At the line of attachment, the component of the wind that is perpendicular to the crest has a zero velocity. Thus, in this case we can neglect the component of the wind that is perpendicular to the crest line, so that only the component that is parallel to it remains (Figure 4).

From this, it follows that when the wind encounters a longitudinal dune obliquely, the wind direction on the lee side is deflected at the point or the line of attachment, and blows along that line parallel to the crest line.

The trigonometric relation between the magnitude of the wind component parallel to the crest line at the attachment line and that of the incidence angle of the wind can be found according to the equation:

$$C_p(z) = V(z) \cos \alpha \quad (1)$$

where:

$C_p(z)$ = the magnitude of the wind component parallel to the crest line at the line of attachment at altitude z .

$V(z)$ = the magnitude of the general wind at the crest line at altitude z .

α = the incidence angle of the wind.

z = the altitude where the wind is measured above the level of the dune (found at the bottom section of the atmospheric boundary layer).

Therefore, it is clear that when the wind blows parallel to the crest line ($\alpha = 0^\circ$) $C_p = V(z)$, and when the wind is perpendicular, ($\alpha = 90^\circ$) $C_p = 0$.

Because of this separation phenomena, the air layer that was on the dune surface rises to a greater altitude and cuts off contact with the dune surface (Fig. 3). In this way, the dune surface resistance to the air movement is eliminated on one hand, and on the other, the separated flow gains momentum from faster layers of air above. Therefore, Eq. 1 should be expressed as follows:

$$C_p(z) = V(z+\Delta z)\cos\alpha \quad (2)$$

where:

Δz = the addition of altitude to the wind flow as a result of the separation.

When the wind returns to the ground level (altitude z) on the line of attachment, it has a velocity typical of a higher altitude ($z + z$), which originated from the additional momentum gained when it was above the separation vortex.

We know that:

$$V(z + z) V(z)$$

Therefore, in cases where the angle is small and the value of $\cos\alpha$ approaches 1, it is possible that:

$$C_p(z) V_z$$

Experimentally it was found that:

$$C_p(z) - V_z \text{ 2 meters/second.}$$

This value was obtained in cases where 40° and in wind measurement at an altitude of $z = 16$ cm, and the magnitudes of $V(z)$ were between 4 to 6 meters/second.

There is a relationship between z and the angle of incidence. The less acute the angle, the larger is the wind component perpendicular to the crest (which determines the separation vortex) and thus, the larger is z . When the angle approaches zero, the larger is the wind component parallel to the crest line, but the separation vortex is very small and z is small, and the above phenomenon is weak. Accordingly and as a result of the wind measurements, we can assume that the angle equals $30^\circ \pm 10^\circ$ is the optimal angle of incidence that causes the maximum increase in magnitude of the deflected wind on the lee flank.

The general wind regime in the area of the longitudinal dune on which the measurements were carried out is bidirectional--oblique to the dune axis from both sides (Figure 2). The wind that comes from the northern sector of the dune has a high degree of constancy of wind direction and velocity. This type of wind regime exists in the summer (sea breeze) and it is at an angle of 50° with the dune axis. The inflection of the dune lessens this angle to 20° to 30° --the optimal angle of incidence that brings about maximum erosion and movement in the lee flank.

THE MORPHODYNAMIC MODEL OF A LONGITUDINAL DUNE

We can not sum up the morphodynamic model of the longitudinal dune and realize that because of the separation process, we get, with wind that meets the crest line at an acute angle, high velocities of the deflected wind on the lee flank that blows parallel to the crest line. In these cases, the sand eroded from the windward flank is not deposited on the lee flank, but continues to move along the dune with the deflected wind. Because of the increase in the deflected wind magnitude, it also erodes sand from the lee flank and carries it along its path. This process exists as long as the wind magnitude at the lee flank continues to increase. Deposition begins only when the wind magnitude drops. This occurs when the dune meanders, and as a consequence the angle of incidence becomes less acute (Eq. 2).

The model of mechanism of movement of a longitudinal dune is the following: The wind is encountering the dune at an acute angle (usually 40 to 30°). Erosion takes place on both flanks (except for a minor depositional area on the lee flank near the crest line). When the dune is meandering and the crest line changes position with the wind direction, the velocity of the deflected wind on the lee flank decreases and deposition takes place (Figure 5). Later the crest line returns to its former position, and the angle once more becomes acute and erosion takes place again at the lee flank of the dune, and so on.

We can apply the morphodynamic model that has been developed to explain some of the relationships existing between transverse and longitudinal dunes.

The longitudinal dunes are not the creation of the initial response of the sand to the wind. The first forms are barchans and transverse dunes--initial aeolian bedforms from which the longitudinal dunes develop.

Longitudinal dunes are formed and developed only in cases where a bulk of sand piles up with a sharp profile--which is the one capable of elongation in response to oblique winds, through the processes described above. The initial contact between the wind and the sand forms rounded sand piles that

Figure 5
Plan of the longitudinal dune showing the depositional areas in two seasons. All other areas are subjected to erosion.

Figure 6
Longitudinal dunes initiated from transverse dunes and elongate obliquely to the transverse dunes in the direction of the resultant wind. Sinai Desert.

turn into barchans or transverse dunes in deserts (Bagnold, 1954). A longitudinal dune will be formed only at a later stage, after these aeolian forms were developed.

In cases where the longitudinal dunes develop from transverse dunes or barchans, their development is dependent on the crescent-like structure of the barchan and the transverse dune whose horns have a sharp profile, able to elongate under dynamic processes into longitudinal dunes (Figure 6).

In this case we have to aeolian bedforms in one system, each of which has a different movement mechanism. The volume of the transverse dune (or barchan) advances by means of erosion on the windward flanks and its deposition, through slides and avalanching, on the slip face. Therefore, the movement rate of the transverse dune is found in a direct relation to the wind and in an inverse relation to its height (Bagnold, 1954). The longitudinal dune does not advance, but elongates, its elongation dependant only on the wind velocity and the angle that the wind makes with the crest line. From the dynamic point of view, the longitudinal dune has an advantage over the transverse in that any wind direction from a sector of 180° , whose center is at the beginning of the dune, brings about its elongation. On the contrary, in transverse dunes, only winds whose direction coincides with the direction of the dune movement can bring about an optimal advance rate. Cross winds cause most of the morphological changes and considerable abatement in the advance rate (Tsoar, 1974). Longitudinal dunes "utilize" wind from all directions and elongate in any wind direction at all. This is the reason for the fact that longitudinal dunes are the most widespread type in the world's deserts (Jordan, 1964).

The result of the above process is that the longitudinal dune, after its formation (from a transverse dune or barchan) elongates faster than the advance rate of the transverse dune connected to it (Figure 6). Therefore, if a barchan field or transverse dunes begin to develop longitudinal dunes, it is because of the faster relative elongation rate of the longitudinal dunes, their dominance in the sand dune fields will be only a question of time.

REFERENCES

- Bagnold, R.A. (1954) *The Physics of Blown Sand and Desert Dunes*, 2nd edition. Chapman and Hall, London, 265 p.
- Folk, R.L. (1971) Longitudinal Dunes of the Northwestern Edge of the Simpson Desert, Northern Territory, Australia, 1. Geomorphology and Grain Size Relationships. *Sedimentology*, 16, 5-24.

- Hanna, S.R. (1969) The Formation of Longitudinal Sand Dunes by Large Helical Eddies in the Atmosphere. *Jour. Appl. Meteorology*, 8, 874-883.
- Jordan, W.M. (1964) Prevalence of Sand Dune Types in the Sahara Desert. *Geol. Soc. Am. Spec. Pap.*, 82, 104-105.
- King, D. (1960) The Sand Ridge Deserts of South Australia and Related Aeolian Landforms of the Quaternary Arid Cycles. *Trans. Roy. Soc. S. Australia*, 83, 99-108.
- Lancaster, N. (1989) Dune Forms and Processes in the Namib Sand Sea. *Namib Bulletin*, Supplement to the Transvaal Museum Bulletin, 3, 3-5.
- Mabbutt, J.A. (1968) Aeolian Landforms in Central Australia. *Australian Geog. Studies*, 6, 139-150.
- McKee, E.D. and Tibbitts, G.C. (1964) Primary Structures of a Seif Dune and Associated Deposits in Libya. *Jour. Sed. Petrol.*, 34, 5-17.
- Tsoar, H. (1974) Desert Dunes Morphology and Dynamics, El-Arish (Northern Sinai). *Zeit. Geom. Suppl.*, 20, 41-61.
- Wopfner, H. and Twidale, C.R. (1967) Geomorphological History of the Lake Eyre Basin, in J.N. Jennings and J.A. Mabbutt (eds.): *Landform Studies From Australia and New Guinea*. ANU Press, Canberra, 118-143.

PRODUCTION OF DUST

By

Dale A. Gillette

ABSTRACT

The driving force for dust production is the motion of sand-sized particles. A model for sand and dust production developed from scaling considerations is compared to measurements of dust production in the field. Production estimates from large-scale dust storms are found to be consistent with measurements in individual eroding areas. The size distribution of the produced dust is roughly bimodally lognormal.

DEFINITIONS OF SOIL PARTICLE MOTIONS

A summary of soil particle motions is given in Figure 1. This figure shows the creeping motion of a coarse particle close to threshold velocity. In Figure 1(b) the particle is lifted into the air by turbulent fluctuations and in Figure 1(c) the particle collides with the surface and bounces. In Figure 1(d) a collision is followed by creeping and another lift. Figure 1(e) shows the breaking off of smaller particles that were encrusted on the colliding particle's surface. In Figure 1(f) a coarse particle collision is followed by splashing of fine particles into the air. Taken together, Figures 1(e) and 1(f) illustrate sandblasting. In Figure 1(g) many of the above movements are combined and the fine particles emitted are carried high into the air. Figure 1(h) shows an example of a nonerodible element.

Figure 1.
Soil particle motions during wind erosion (after Gillette, 1980).

- a. Creeping motion of a particle moving with a wind speed slightly greater than threshold.
- b. A coarse particle lifted into the air by turbulent air fluctuations.
- c. An airborne particle collides with the surface and bounces (saltation motion).
- d. An airborne particle collides with the surface, rolls, and is then relifted by air fluctuations.
- e. A particle collision followed by breaking off of smaller particles that were encrusted on the colliding particle's surface.
- f. A particle collision followed by "splashing" of the soil.
- g. A combination of c, d, e, and f.
- h. A nonerodible element giving protection to the soil downwind.

MICROSCOPIC EVIDENCE FOR SANDBLASTING

Krinsley and Doornkamp (1973) have shown pitting of quartz grains on microscopic examination of quartz grains that have been in saltation (Figure 2). This actual breakage of crystalline quartz shows the likelihood of breakage of soil aggregates when put into saltation motion by the wind.

Another example of the effect of saltation on the breakage of clay aggregates is shown in Figure 3. The figure shows size distributions and scanning electron micrographs of loose clay pellets that were collected on the surface of a dry lake in western Texas (Randall clay soil). Figure 3(a) shows soil condition shortly after the soil dried from a thorough soaking and the shrinking clay (mixed layer montmorillonite) broke into rather well-defined smooth pellets with flat surfaces and straight edges. Figure 3(b) presents the soil condition on the same soil after a long dry winter with frequent strong winds. Aggregate disintegration changed the soil into a finer, loose mixture more subject to wind erosion. The particles larger than 200 micrometers (0.2) collected after the windy dry period show a much more rounded and irregular surface than the particles of the same size collected before the windy dry period, which exhibit rather straight edges and smooth surfaces. The windy drought effect has thus broken material from the large particles, to create a reserve of smaller (~ 0.1 mm) particles that break into even finer material in their movement by the wind. It is also possible that a supply of fine material ($d < 0.02$ mm) is created by erosion, and then ejected into the air upon saltation of the soil by a "splash" mechanism.

DIMENSIONAL ARGUMENTS LEADING TO A SIMPLE MODEL OF DUST PRODUCTION

Since wind plays a double role in both generating and transporting aerosol particles, it might be helpful to begin with a dimensional analysis.

In the turbulent field in which erosion is taking place, a more significant dimension than the roughness height of the particle, is the length of particle travel, ℓ . Unfortunately, this length is not normally measured; we can, however, measure flux of momentum, τ , and friction velocity, u_* .

Thus the dimension quantities are:

$$\text{velocity: } u_* = K[\bar{u}(z_2) - \bar{u}(z_1)] / n \frac{z_2}{z_1}$$

Figure 2.

Top: Surface smoothness and rounding. Sabha Sand Sea, South-Central Libya. Modern dune sand. Note surface smoothness and rounding on these large sand grains. The irregular shapes were probably inherited from the source rock.

Bottom: Irregular depressions on rounded grain. Sabha Sand Sea, South-Central Libya. Modern dune sand. Very well rounded large aeolian grain with several irregular depressions in the middle; these are probably due to mechanical chipping during particularly powerful sand storms. [After Krinsley and Doornkamp (1973)]

Figure 3.

Top: Scanning electron micrographs of particles and size distributions of loose particles on the surface of a Randall clay soil before a period of windy drought. Bottom: Scanning electron micrographs of particles and size distributions of loose particles on the surface of a Randall clay soil after a period of windy drought. (After Gillette (1978))

Figure 4.

Textures of sampled soils; total soil movement vs. wind friction velocity; ratio of vertical flux of particles smaller than 0.02 mm to total soil movement per unit area per time versus wind friction velocity (after Gillette, 1978).

where K is von Karman's constant, z_1 and z_2 are two heights in a neutrally stratified atmosphere that are well above any roughness elements in the field, and $u(z)$ is a mean wind speed at height z .

$$\text{time: } \frac{u_*}{g}$$

where g is the acceleration of gravity.

$$\text{length: } \frac{u_*^2}{g}$$

mass density: ρ , the mass density of air.

Now by definition, the momentum flux is ρu_*^2 .

By dimensional analysis, the horizontal flux of sand, q , having the dimensions of mass/length X time may be written

$$q = \frac{\rho u_*^3}{g}$$

An expression for the flux of kinetic energy onto the surface from hits by sand grain is

$$\text{Flux}_A = \rho u_*^3$$

Thus, if the vertical flux of fine particles is proportional to the flux of kinetic energy delivered to the soil surface, then the ratio of the vertical flux of fine material to the horizontal flux of sand should remain constant with wind speed. Physically, such a situation would reflect a soil surface uniformly susceptible to destruction by sandblasting. An example of such a surface would be a loose soil well mixed with fine and coarse particles. If the ratio of the vertical flux of fine material to the horizontal flux of sand increases with wind speed, the soil is probably being pulverized or disaggregated by energetic sandblasting, thereby producing a higher proportion of fine to coarse materials. If, on the other hand, the ratio of the vertical flux of fine material to the horizontal sand flux decreases with wind speed (i.e., the proportion of fine material to coarse material in the eroding soil decreases), it may mean that fine particles are being rapidly depleted at high wind speeds. An example of such a soil is one in which an unerodible layer is being reached but sandblasting continues from an upwind source of coarse material.

RELATION OF FINE DUST PRODUCTION TO SAND FLUX

Sand-size particles are important in wind input of fugitive dust since their availability as loose individual particles largely determines the threshold velocity of erosion and since they carry most of the mass of moving soil near the ground, acting as sandblasters of the surface. However, only a few centimeters above the surface, the concentration of sand decreases (Gillette and Goodwin, 1974) and the sand is not transported very far compared to dust smaller than 20 micrometers (Gillette, 1978).

The production of dust smaller than 20 micrometers and the total soil movement of a number of soil textures under natural conditions are shown in Figure 4. In the figure, F_a' is the vertical flux of particles smaller than 20 micrometers, q is the horizontal flux of all particles moving with the wind up to a height of 76 cm and F_a'/q is the ratio of vertical flux of fine particles to measured total movement of soil. (A method of calculating total soil movement for individual soils is given in Gillette, 1978.) The soil textures are shown in the figure. Total particle flux for all soils asymptotically approaches a law proposed by Bagnold (1941) (i.e., sand flux varies as the friction velocity cubed, u_*^3). The fine ($d < 20$ micrometers) particle flux is more variable. In the plot of F_a'/q it is shown that for sand loam soils, in an increasing part of the total soil movement is the fine particle movement as the wind speed increases. In general, soils having finer textures produced more fine dust per unit soil movement, except those fine-textured soils whose mineral components formed small aggregates that were hard enough to resist impact breakage at mean wind speeds up to 25 m s^{-1} . Fine particles are generated by the breakage of soil aggregates or by the splashing of loose fine material by saltating sand grains (Gillette and Walker, 1977).

Test soils 1, 2, 4, and 5 (Figure 4) showed no increase with wind speed in the ratio of vertical flux of fine particles to the horizontal flux of sand. Such a result was expected (see scaling considerations) for a loose soil of well-mixed fine and coarse particles that would not change proportions with increased sandblasting energy. Indeed, all of these soils were classified as having fine sand textures. For soils 3, 6, and 7 (loamy sand or sandy loam textures) the ratios of vertical flux of fine material to horizontal flux of sand increased with wind speed. This would be expected for aggregated soil that is easily broken and that changes its proportion of fine material to coarser aggregates as sandblasting energy is increased. The clay soil (No. 9) probably had aggregates of sufficient hardness that increasing wind speed did not disaggregate the soil enough to show an

DUST GENERATION

increase of fine to coarse particles with increasing wind speed.

The effect of increased dust particle flux with increasing fetch, up to a certain distance, was shown by Gillette (1978), as it had been shown earlier for sand flux by Chepil (1957). Briefly, the reason for such an effect is the efficiency of momentum transfer to the soil surface by saltating sand grains which, in turn, propels particles into the air. This fine and coarse particle emitter supplements direct aerodynamic pickup of soil grains. Since this "avalanche effect" must be related to the saltation length, it is a nonlinear function of wind speed (see Section on scaling considerations).

SIZE DISTRIBUTIONS OF SOIL-DERIVED DUST

The size distribution of soil-derived aerosol may be roughly characterized as a bimodal lognormal distribution, when particle volume is plotted versus particle size. The larger particles, having a mode around 50 micrometers, are sand grains that are moving close to the surface and whose sedimentation velocities are on the order of the vertical velocity fluctuations of the air. The finer particles, having a mode between 1 and 10 micrometers are particles produced by the sandblasting of sand grains on the soil surface (Patterson and Gillette, 1977).

LARGE-SCALE DUST INPUT

Total mass of dust produced in two dust storms in 1972 in the southwestern United States was computed from aircraft-obtained dust concentrations and meteorological analysis of dusty air (Gillette et al., 1978). Total mass produced by the two storms was estimated to be 0.8×10^{12} g, which is only 0.16% of the yearly production of dust estimated by Goldberg from consideration of pelagic sediment depositions (Gillette, 1980). The average vertical flux, computed from total mass, storm duration, and area of eroding land was consistent with measured values from other dust storms. Thus, if dust storm frequency were around 10 per year, the southwestern United States would be responsible for only about 4 million tons per year, or less than 1% of Goldberg's total estimated yearly production of 500 million tons.

REFERENCES

- Bagnold, R. A. (1941) The Physics of Blown Sand and Desert Dunes. Methuen and Co., London, 265 pp.

- Chepil, W. S. (1957) Sedimentary Characteristics of Dust Storms 3. Composition of Suspended Dust. *Amer. J. Sci.* 255, 206-213.
- Gillette, D. A. (1978) Fine Particulate Emissions Due to Wind Erosion. *Trans. ASAE* 20, 890-897.
- Gillette, D. A. (1978) A Wind Tunnel Simulation of the Erosion of Soil: Effect of Soil Texture, Sandblasting, Wind Speed, and Soil Consolidation on Dust Production. *Atmospheric Environment* 12, 1735.
- Gillette, D. (1980) Major Contribution of Natural Primary Continental Aerosols: Source Mechanisms. Reprinted from *Annals of the New York Academy of Sciences*, 338, 348-358.
- Gillette, D. and Goodwin, P.A. (1974) Microscale Transport of Sand-Sized Aggregates Eroded by Wind. *J. Geophys. Res.* 79, 4080-4084.
- Gillette, D. A. and Walker, T. R. (1977) Characteristics of Airborne Particles Produced by Wind Erosion of Sandy Soil, High Plains of West Texas. *Soil Sci.* 123, 97-110.
- Gillette, D. A., Clayton, R. N., Mayeda, T. K., Jackson, M. L., and Sridhar, K. (1978) Tropospheric Aerosols from Some Major Dust Storms of the Southwestern United States. *J. Appl. Met.* 17, 832-845.
- Krinsley, D. and Doornkamp, J. (1973) *Atlas of Quartz Sand Surfaces*. Cambridge, at the University Press, 92 p.
- Patterson, E. M. and Gillette, D. A. (1977) Commonalities in Measured Size Distributions for Aerosols Having a Soil-Derived Component. *J. Geophys. Res.* 82, 2074-2082.

SOIL EROSION BY WIND: AN OVERVIEW

By

E. L. Skidmore

ABSTRACT

Wind erosion is a serious problem in many parts of the world. It physically removes the most fertile portion of the soil from the field, pollutes the air, fills road ditches, reduces seedling survival and growth, and lowers the marketability of many vegetable crops. Wind erosion also creates new desert landforms and landscapes.

Wind erosion is generally worse in arid and semiarid climates. It can be a problem wherever soil, vegetative, and climatic conditions are conducive.

Soil particles erode when shear stress exerted by the wind against the ground surface exceeds the forces tending to hold the particles in place. The eroding particles may be transported in saltation, creep, or suspension.

Principles for controlling wind erosion include: stabilizing with various materials; producing a rough, cloudy surface; reducing field width with strips and barriers; and establishing and maintaining sufficient vegetative cover.

PROBLEM

Lands undergoing desertification become vulnerable to wind erosion (Secretariat, 1977, p. 14). In pastoral rangelands, composition of pastures subject to excessive grazing in dry periods deteriorates, the proportion of edible perennial plants decreases, and the proportion of annuals increases. The thinning and death of vegetation in dry seasons or droughts

increases the extent of bare ground. The surface soil conditions deteriorate. In rain-fed farming, removal of the original vegetation and fallow expose the soil to accelerated wind and water erosion.

Wind erosion is worse in arid and semiarid areas where the conditions conducive to wind erosion frequently occur. Those conditions are: (1) loose, dry, finely divided soil; (2) smooth soil surface devoid of vegetative cover; (3) large fields; and (4) strong winds (FAO, 1960).

The arid lands are extensive, comprising about one-third of the world's population (Dregne, 1976; Gore, 1979). The general areas most susceptible to wind erosion on agricultural land are much of North Africa and the Near East, parts of southern and eastern Asia, Siberian Plain, Australia and southern South America, and the semiarid and arid portions of North America (FAO, 1960).

Extensive soil erosion in the Great Plains, USA, during the last half of the 19th century and the 1920's in the prairie region of western Canada gave warning of impending disaster, and during the 1930's a prolonged dry spell culminated in dust storms and soil destruction of disastrous proportions of the prairie regions of both western Canada and the Great Plains of the United States (Anderson, 1975; Svobida, 1940; Malin, 1946a, b, c; Johnson, 1947).

Wind erosion physically removes the most fertile portion of the soil from the field and therefore lowers productivity of the land (Daniel and Langham, 1936; Lyles, 1975).

Some soil from damaged lands enters suspension and becomes part of the atmospheric dustload. Hagen and Woodruff (1973) estimated that eroding lands of the Great Plains contributed 244 and 77 million tons of dust per year to the atmosphere in the 1950's and 1960's, respectively. Jaenicke (1979) estimated the source strength of mineral dust from the Sahara at 260 million tons per year. Dust obscures visibility and pollutes the air, causes automobile accidents, fouls machinery, and irritates the housewife.

Blowing soil fills road ditches, reduces seedling survival and growth, lowers the marketability of vegetable crops like asparagus, green beans, and lettuce, and increases the susceptibility and transmission of some diseases (Hayes, 1965, 1966; Claflin, Stuteville, and Armbrust, 1973).

MECHANICS

Surface Wind. Movement of soil particles is caused by wind forces exerted against the surface of the ground. The average forward velocity of the wind near the ground increases exponentially with height above the ground surface. The change in velocity with height is known as the velocity gradient. It is

this gradient that determines the shear stress or drag force exerted on the ground surface.

The velocity gradient or the shape of the adiabatic windspeed profile is given by

$$\frac{\partial u}{\partial z} = \frac{u_*}{kz} \quad (1)$$

where u is mean windspeed at height z above the mean ground surface, k is the von Karman constant (0.4); and u_* is friction velocity further defined as $(\tau/\rho)^{1/2}$ where τ is surface shear (force per unit area) and ρ is fluid density. The surface shear then is

$$\tau = \rho u_*^2. \quad (2)$$

The surface shear associated with the decrease in wind velocity as the surface is approached is a vertical transfer of horizontal momentum. Momentum (the product of mass and velocity) decreases as the surface is approached. The eddy diffusion equation for steady state one-dimensional transport is

$$\tau = \rho K_m \partial u / \partial z \quad (3)$$

where K_m is momentum transfer coefficient. The integrated form of equation 1 over a rough surface becomes

$$u = \frac{u_*}{k} \ln \frac{z - z_d}{z_0}. \quad (4)$$

The parameter z_d , the effective displacement height, is the distance from the ground surface to the plane at which the momentum exchange coefficient extrapolates to zero. Roughness parameter, z_0 , is the distance from the displaced reference plane to the surface at which the wind profile extrapolates to zero.

Equilibrium Forces. In addition to surface shear, another force tending to dislodge a soil grain is a negative pressure on the top as compared to the bottom of the grain. This Bernoulli effect causes lift on the grain (Chepil, 1959).

Chepil (1959) analyzed the nature of forces of drag, lift, and gravity on soil grains at the threshold of their movement by wind. He found that equilibrium between these forces and the soil grains was influenced by the diameter, shape, and density of the grains, the angle of repose of the grains with respect to the mean drag level of the fluid, the closeness of packing of top grains, and the impulses of fluid turbulence associated with drage and lift. The relationship was

$$\bar{\tau}_c = 0.66 \, g d \rho \tan \phi n / (1 + 0.85 \tan \phi) T \quad (5)$$

where $\bar{\tau}_c$ is the mean critical drag per unit horizontal area, g is acceleration of gravity, d is diameter of spherical grain,

ρ is difference in density of grain and fluid, ϕ is angle of repose of the grain with respect to the mean drag level of the fluid, n is ratio of mean drag and lift per unit area to mean drag and lift per unit area on the top grain moved by the fluid, T is the ratio of maximum to mean drag and lift on the soil grain. Chepil (1959) experimentally determined the following values for the constants of equation 5: $T = 2.5$, $\tan \phi = 0.45$, and $n = 0.21$.

When the mean critical drag on a particle is exceeded, the particle dislodges and is transported by the wind. This occurs for loose grains with 0.25 mm diameter when the friction velocity u_* is 20 to 44 cm/sec (Lyles and Krauss, 1971; Chepil, 1959; Zingg, 1953; Bagnold, 1943), which corresponds to surface drag of 0.48 to 1.94 dynes/cm². The windspeed at initial particle movement is from 4.0 to 5.8 m/sec at 30 cm (Chepil, 1945b, c; Malina, 1941).

Initial Particle Motion. The windspeed at which sand movement starts, due to the direct pressure of the fluid was called "fluid threshold" by Bagnold (1943). Bagnold described the initial motion as "surface grains, previously at rest, began to be rolled along the surface by the direct pressure of the wind... A foot or so downwind of the point at which the rolling began, the grains could be seen to have gathered sufficient speed to start bouncing off the ground." Others (Bisal and Nielsen, 1962; Lyles and Krauss, 1971) observed that as the fluid threshold was approached, some particles began to vibrate, or rock back and forth. Erosive particles vibrated with increasing intensity as windspeed increased and then left the surface instantaneously as if ejected. Evidence supported the hypothesis that the particle-vibration frequency is related to the frequency band containing the maximum energy of the turbulent motion.

Saltation. The bouncing or ejection of the eroding particle off the surface bed into the airstream and subsequent forward movement is referred to as saltation. Fifty to 75 percent of the movement of soil particles takes place through saltation (Chepil, 1945a). In saltation the particles rise almost vertically, rotating from 20 to 1,000 revolutions per second, travel 10 to 15 times their height of rise, and return to the surface with an angle of descent of about 6 to 12 degrees from the horizontal (Chepil and Woodruff, 1963). On striking the surface they either rebound and continue their movement in saltation or impart most of their energy by striking other grains, causing these to rise upward or roll

along the surface. Most of the saltating particles range in size from 0.1 and 0.5 mm in diameter.

Creep. The rolling or sliding of larger particles with energy derived from saltating particles is called creep. Individual grains are knocked onward by the blow they receive from behind. Bagnold (1943) observed that at low windspeeds the grains move in jerks a few millimeters at a time, but as windspeed increases, the distance moved lengthens and more grains are set in continuous motion until in high winds the whole surface appears to be creeping forward.

Suspension. Particles smaller than about 0.1 mm may enter suspension and be carried to great heights by the eddies of the erosive winds. Movement of these fine particles is usually initiated by the impact of particles in saltation. The greatest amount of soil is moved by saltation and surface creep but that moved by suspension is the most spectacular and easily recognized from a distance.

Sorting. An eroding wind has been said to act on the soil like a fanning mill on grain, removing the finer and more porous particles and leaving the coarser and denser behind (Chepil, 1957a; Moss, 1935; Daniel, 1936). The coarser eroded material usually ends up in a soil drift whereas the finer enters suspension and is transported often times great distances before deposition. Chepil (1957a) observed that the most distinct feature in the sorting process was that the particles of peak diameter tend to remain in the wind-eroded fields, and particles smaller than this diameter tend to be carried far through the atmosphere in suspension.

Peak diameter of drifted material derived from fields composed of sand and loamy sand was about 0.4 mm, and that of drifted material from the finer-textured soils was about 0.6 mm. The drifted materials derived from fields of sand and loamy sand were composed principally of discrete, nonporous grains having an average bulk density of 2.37, whereas the materials drifted from the finer-textured soils were predominantly aggregates exhibiting a distinct degree of porosity and having an average bulk density of 1.70 (Chepil, 1957a).

By applying equation 5 for peak diameters and average bulk densities, I found that critical mean drag is about the same for both conditions; 1.7 and 1.8 dynes/cm² for the single grain and porous grains, respectively.

Very little sorting occurs on fine-textured soil derived from loess. Moss (1935) found that clay soils and the corresponding drifted materials were practically identical in composition. In some cases, wind erosion virtually removes the surface soil (Zingg, 1954; Chepil, 1957a, b). This nonselective removal by wind is associated primarily with loess which was already sorted and deposited from the atmosphere during past geologic eras.

CONTROL PRINCIPLES

Principles for controlling wind erosion include: stabilizing with various materials; producing a rough, cloddy surface; reducing effective field width with barriers; and establishing and maintaining sufficient vegetative cover.

Stabilizers. Various soil stabilizers have been evaluated to find suitable materials and methods to control wind erosion (Armbrust and Dickerson, 1971; Armbrust and Lyles, 1975; Chepil, 1955; Chepil and Woodruff, 1963; Chepil et al., 1963; Lyles et al., 1969; Lyles et al., 1974). Several tested products successfully controlled wind erosion for short periods of time but were often expensive as compared with equally effective wheat straw anchored with a rolling disk packer (Chepil et al., 1963). The following are criteria for surface soil stabilizers; (1) 100 percent of the soil surface must be covered, (2) the stabilizer must not adversely affect plant growth or emergence, (3) erosion must be prevented initially and reduced for at least 2 months, (4) the stabilizer should apply easily and without special equipment, and (5) cost must be low enough for profitable use (Armbrust and Lyles, 1975). Armbrust and Lyles (1975) found five polymers and one resin-in-water emulsion that met all those requirements. However, they added that before soil stabilizers can be used on agricultural lands, methods must be developed to apply large volumes rapidly. Also, reliable preemergent weed control chemicals to use on coarse-textured soils must be developed, as well as films resistant to raindrop impact, yet still allow water and plant penetration without adversely affecting the environment.

Rough, Cloddy Surface. Chepil and Milne (1941a), investigating the influence of drifting dune materials and cultivated soils found that the initial intensity of drifting was always much less over a ridged surface. Ridging cultivated soils reduced the severity of drifting, but ridging highly erosive dune materials was less effective because the ridges disappeared rapidly. The rate of flow varied inversely with surface roughness.

Armbrust et al. (1964) studied the effects of ridge roughness equivalent on total quantity of eroded material from three simulated, cultivated soils exposed to different friction velocities. From their data, a curve can be constructed showing the relationship between quantity of eroded material and ridge roughness equivalent. Presumably, this was the origin of the chart (Woodruff and Siddoway, Figure 4, 1965) showing a soil ridge roughness factor as a function of soil ridge roughness so that a ridge roughness equivalent of 6 cm reduces wind erosion 50 percent. As roughness increases to about 11 cm, the soil ridge roughness factor remains about constant, then with additional roughness, the effectiveness of ridges gradually decreases.

When ridges are mostly gone, vegetative cover is depleted, and the threat of wind erosion continues, a rough, cloddy surface resistant to the force of wind can be created on many cohesive soils with appropriate "emergency tillage." Listers, cultivators, one-ways with two or three disks removed at intervals, and pitting machines can be used to bring compact clods to the surface. Emergency tillage is most effective when done at right angles to the prevailing wind direction. Since the clods eventually disintegrate (sometimes rapidly), emergency tillage offers, at best, only temporary wind erosion control (Woodruff et al., 1957).

Residue. Living vegetation or residue from harvested crops protects the soil against wind erosion. Standing crop residues provide nonerodible elements that absorb much of the shear stress in the boundary layer. When vegetation and crop residues are sufficiently high and dense to prevent intervening soil surface drag from exceeding threshold drag, soil will not erode. Rows perpendicular to wind direction control wind erosion more effectively than rows parallel to wind direction (Englehorn et al., 1952; Skidmore et al., 1966). Flattened stubble, though not as effective as standing, also protects the soil from wind erosion (Chepil et al., 1955).

Studies (Chepil, 1944; Chepil et al., 1955; Siddoway et al., 1965) to quantify specific properties of vegetative covers influencing wind erosion led to the relationship presented by Woodruff and Siddoway (1965) showing the influence of an equivalent vegetative cover of small grain and sorghum stubble for various orientations (flat, standing, height).

Efforts have continued to evaluate the protective role of additional crops (Craig and Turelle, 1964; Lyles and Allison, 1981), range grasses (Lyles and Allison, 1980), feedlot manure (Woodruff et al., 1974), and the protective requirements of equivalent residue needed to control wind erosion (Skidmore and Siddoway, 1978; Skidmore et al., 1979).

Barrier. Reducing the field width or the distance that wind travels in crossing the field reduces wind erosion. Chepil and Milne (1941b) reported that the rate of soil movement began with zero on the windward side of fields or field strips and increased with distance downwind. Later Chepil (1946) found that the cumulative rate of soil movement with distance away from the windward edge of eroding fields was the main cause of increasing abrasion and gradual decrease in surface roughness along the direction of the wind. He called the increase in rate of flow with distance downwind "avalanching."

"Rate of soil flow increased with distance downwind across an eroding field until, if the field was large enough, it reached a maximum that a wind of a given velocity can carry. Beyond that point the rate of flow remained essentially constant"--(Chepil, 1957c).

Use of wind barriers is an effective method of reducing field width. Barriers have long been recognized as valuable for controlling wind erosion (Bates, 1911). Hagen (1976) and Skidmore and Hagen (1977) have presented a model which, when used with local wind data, showed wind-barrier effectiveness in reducing wind erosion forces: Barriers reduce wind forces more than they do windspeed; a properly oriented barriers, when winds predominate from a single direction, will decrease wind erosion forces by more than 50 percent from the barrier leeward to 20 times its height; the decrease is greater for shorter distances from the barrier.

Different combinations of trees, shrubs, tall growing crops, and grasses can reduce wind erosion. Besides the more conventional tree windbreak (Ferber, 1969; Read, 1964; Woodruff et al., 1976), many other barrier systems are used to control wind erosion including annual crops like small grains, corn, sorghum, sudangrass, sunflowers (Carreker, 1966; Fryrear, 1963, 1969; Hagen et al., 1972; Hoag and Geiszler, 1971), tall wheatgrass (Aase et al., 1976; Black and Siddoway, 1971), sugarcane, and rye strips on sands in Florida (Griffin, SCS Agronomist, personal communication, 1975).

However, most barrier systems for controlling wind erosion occupy space that could otherwise be used to produce crops. Perennial barriers grow slowly and are often established with difficulty (Dickerson et al., 1976; Woodruff et al., 1976). These barriers also compete with the crop for water and plant nutrients. Thus, the net effect for many tree barrier systems is that production may not be benefited from their use (Frank et al., 1977; McMartin et al., 1974; Skidmore et al., 1975; Skidmore et al., 1974; Staple and Lehan, 1955). Perhaps, the tree-barrier systems could be designed so that they become a useful crop, furnishing nuts, fruit and wood.

Strip cropping. The practice of farming land in narrow strips on which the crop alternates with fallow is an effective aid in controlling wind erosion. Strips are most effective when they are at right angles to the prevailing wind erosion direction but also provide some protection from winds that are not perpendicular to the strip. Strip cropping reduces wind erosion damage in the following ways: it reduces the distance the wind travels across exposed soil; localizes drifting that starts at a focal point; and reduces wind velocity across the strip when adjacent fields are covered with tall stubble or crops.

Although each method to control wind erosion has merit and application, when feasible, establishing and maintaining vegetative cover remains the best defense against wind erosion. However, this becomes a difficult challenge as pressure increases to use the crop residues for livestock feed and fuel for cooking.

REFERENCES

- Aase, J. K., Siddoway, F. H. and Black, A. L. (1976) Perennial Grass Barriers for Wind Erosion Control, Snow Management, and Crop Production. In Shelterbelts on the Great Plains, Proceedings of the Symposium. Great Plains Agr. Pub. No. 78, p. 69-78.
- Anderson, C. H. (1975) A History of Soil Erosion by Wind in the Palliser Triangle of Western Canada. Historical Series No. 8, 25 p., Research Branch, Canada Dept. Agr.
- Armburst, D. V., Chepil, W. S. and Siddoway, F. H. (1964) Effects of Ridges on Erosion of Soil by Wind. Soil Sci. Soc. Amer. Proc. 28, p. 557-560.
- Armburst, D. V. and Dickerson, J. D. (1971) Temporary Wind Erosion Control: Cost and Effectiveness of 34 Commercial Materials. J. Soil and Water Conserv., v. 26, p. 154-157.
- Armburst, D. V. and Lyles, L. (1975) Soil Stabilizers to Control Wind Erosion. Soil Conditioners, v. 7, p. 77-82.
- Bagnold, R. A. (1943) The Physics of Blown Sand and Desert Dunes. William Morrow & Co., New York, New York, 265 p.
- Bates, C. G. (1911) Windbreaks: Their Influence and Value. USDA Forest Service Bul., v. 86, 100 p.
- Bisal, F. and Nielsen, K. F. (1962) Movement of Soil Particles in Saltation. Canadian J. Soil Sci., v. 42, p. 81-86.
- Black, A. L. and Siddoway, F. H. (1971) Tall Wheatgrass Barriers for Soil Erosion Control and Water Conservation. J. Soil and Water Conserv., v. 26, p. 107-110.
- Carreker, J. R. (1966) Wind Erosion in the Southeast. J. Soil and Water Conserv., v. 21, p. 86-88.
- Chepil, W. S. (1944) Utilization of Crop Residues for Wind Erosion Control. Sci. Agr., v. 24, p. 307-319.
- Chepil, W. S. (1945a) Dynamics of Wind Erosion: I. Nature of Movement of Soil by Wind. Soil Sci., v. 60, p. 305-320.
- Chepil, W. S. (1945b) Dynamics of Wind Erosion: II. Initiation of Soil Movement. Soil Sci., v. 60, p. 397-411.
- Chepil, W. S. (1945c) Dynamics of Wind Erosion: III. The Transport Capacity of the Wind. Soil Sci., v. 60, p. 475-480.
- Chepil, W. S. (1946) Dynamics of Wind Erosion: V. Cumulative Intensity of Soil Drifting Across Eroding Fields. Soil Sci., v. 61, p. 257-263.
- Chepil, W. S. (1955) Effects of Asphalt on Some Phases of Soil Structure and Erodibility by Wind. Soil Sci. Soc. Amer. Proc., v. 19, p. 125-128.
- Chepil, W. S. (1957a) Sedimentary Characteristics of Dust Storms: I. Sorting of Wind-Eroded Soil Material. Am. J. Sci., v. 255, p. 12-22.
- Chepil, W. S. (1957b) Sedimentary Characteristics of Dust Storms: III. Composition of Suspended Dust. Am. J. Sci., v. 255, p. 206-213.

- Chepil, W. S. (1957c) Width of Field Strip to Control Wind Erosion. Kansas Agr. Expt. Sta. Tech. Bul. 92.
- Chepil, W. S. (1959) Equilibrium of Soil Grains at the Threshold of Movement by Wind. Soil Sci. Soc. Am. Proc., v. 23, p. 422-428.
- Chepil, W. S. and Milne, R. A. (1941a) Wind Erosion of Soil in Relation to Roughness of Surface. Soil Sci., v. 52, p. 417-433.
- Chepil, W. S. and Milne, R. A. (1941b) Wind Erosion of Soils in Relation to Side and Nature of the Exposed Area. Sci. Agr., v. 21, p. 479-487.
- Chepil, W. S. and Woodruff, N. P. (1963) The Physics of Wind Erosion and Its Control. Advances in Agron. v. 15, p. 211-302.
- Chepil, W. S., Woodruff, N. P., Siddoway, F. H., Fryrear, D. W. and Armbrust, D. V. (1963) Vegetative and Nonvegetative Materials to Control Wind and Water Erosion. Soil Sci. Soc. Am. Proc., v. 27, p. 86-89.
- Chepil, W. S., Woodruff, N. P. and Zingg, A. W. (1955) Field Study of Wind Erosion in Western Texas. USDA, SCS-TP-125, 60 p.
- Claflin, L. E., Stuteville, D. L. and Armbrust, D. V. (1973) Windblown Soil in the Epidemiology of Bacterial Leaf Spot of Alfalfa and Common Blight of Beans. Phytopathology, v. 63, p. 1417-1419.
- Craig, D. G. and Turelle, J. W. (1964) Guide for Wind Erosion Control on Cropland in the Great Plains States. USDA, SCS, Washington, D. C., 104 p.
- Daniel, H. A. (1936) Physical Changes in Soils of the Southern High Plains Due to Cropping and Wind Erosion and the Relation Between the (Sand and Silt)/Clay Ratios in These Soils. J. Am. Soc. Agron., v. 28, p. 570-580.
- Daniel, H. A. and Langham, W. H. (1936) The Effect of Wind Erosion and Cultivation on the Total Nitrogen and Organic Matter Content of Soils in the Southern High Plains. J. Am. Soc. Agron., v. 28, p. 587-596.
- Dickerson, J. D., Woodruff, N. P. and Banbury, E. E. (1976) Techniques for Improving Survival and Growth of Trees in Semiarid Areas. J. Soil and Water Conserv., v. 31, p. 63-66.
- Dregne, H. E. (1976) Soils of the Arid Regions. Elsevier Scientific Pub. Co., 237 p.
- Englehorn, C. L., Zingg, A. W. and Woodruff, N. P. (1952) The Effects of Plant Residue Cover and Clod Structure on Soil Losses by Wind. Soil Sci. Soc. Am. Proc., v. 16, p. 29-33.
- Ferber, A. E. (1969) Windbreaks for Conservation. U. S. Dept. of Agr. SCS, Agr. Inf. Bul. 339.
- Food and Agricultural Organization of the United Nations (FAO) (1960). Soil Erosion by Wind and Measures for Its Control on Agricultural Lands. FAO Agr. Development Paper No. 71.

- Frank, A. B., Harris, D. G. and Willis, W. O. (1977) Growth and Yields of Spring Wheat as Influenced by Shelter and Soil Water. *Agron. J.*, v. 69, p. 903-906.
- Fryrear, S. W. (1963) Annual Crops as Wind Barriers. *Trans. ASAE*, v. 6. p. 340-342, 352.
- Fryrear, D. W. (1969) Reducing Wind Erosion in the Southern Great Plains. *Texas A&M Univ.*, MP-929, Sept.
- Gore, R. (1979) The Desert: an Age-Old Challenge Grows. *National Geographic*, v. 156, p. 586-639.
- Hagen, L. J. (1976) Windbreak Design for Optimum Wind Erosion Control. In *Shelterbelts on the Great Plains - Proceedings of the Symposium*. Great Plains Agr. Pub. No. 78, p. 31-36.
- Hagen, L. J. and Woodruff, N. P. (1973) Air Pollution From Duststorms in the Great Plains. *Atmos. Environ.*, v. 7, p. 323-332.
- Hagen, L. J., Skidmore, E. L. and Dickerson, J. D. (1972) Designing Narrow Strip Barrier Systems to Control Wind
- Hayes, W. A. (1965) Wind Erosion Equation Useful in Designing Northeastern Crop Protection. *J. Soil and Water Conserv.*, v. 20, p. 153-155.
- Hayes, W. A. (1966) Guide for Wind Erosion Control in the Northeastern States. *USDA, SCS*.
- Hoag, B. K. and Geiszler, G. N. (1971) Sunflower Rows to Protect Fallow From Wind Erosion. *North Dakota Farm Research*, v. 28, p. 7-12.
- Jaenicke, R. (1979) Monitoring and Critical Review of the Estimated Source Strength of Mineral Dust From the Sahara. In *Christer Morales (ed.) Saharan Dust. Mobilization, Transport, Deposition*. Scope 14, p. 233-242, John Wiley and Sons.
- Johnson, V. (1947) *Heaven's Tableland - the Dust Bowl Story*. New York, Farrar-Straus, p. 155-157.
- Lyles, L. (1975) Possible Effects of Wind Erosion on Soil Productivity. *J. Soil and Water Conserv.*, v. 30, p. 279-283.
- Lyles, L. and Allison, B. E. (1980) Range Grasses and Their Small Grain Equivalents for Wind Erosion Control. *J. Range Mgmt.*, v. 33, p. 143-146.
- Lyles, L. and Allison, B. E. (1981) Equivalent Wind-Erosion Protection From Selected Crop Residues. *Trans. ASAE (in press)*.
- Lyles, L., Armbrust, D. V., Dickerson, J. D., and Woodruff, N. P. (1969) Spray-On Adhesives for Temporary Wind Erosion Control. *J. Soil and Water Conserv.*, v. 24, p. 190-193.
- Lyles, L. and Krauss, R. K. (1971) Threshold Velocities and Initial Particle Motion as Influenced by Air Turbulence. *Trans. ASAE*, v. 14, p. 563-566.
- Lyles, L., Schrandt, R. L. and Schmeidler, N. F. (1974) How Aerodynamic Roughness Elements Control Sand Movement. *Trans. ASAE* 17 (1), p. 134-139.

- Malin, J. C. (1946a) Dust Storms - Part One, 1850-1860. The Kansas Historical Quarterly, v. 14, p. 129-144.
- Malin, J. C. (1946b) Dust Storms - Part Two, 1861-1880. The Kansas Historical Quarterly, v. 14, p. 265-296.
- Malin, J. C. (1946c) Dust Storms - Part Three, 1881-1900. The Kansas Historical Quarterly, v. 14, p. 391-413.
- Malina, F. J. (1941) Recent Developments in the Dynamics of Wind Erosion. Am. Geophys. Union Trans., p. 262-284.
- Malin, F. J. (1941) Recent Developments in the Dynamics of Wind Erosion. Am. Geophys. Union Trans., p. 262-284.
- McMartin, W., Frank A. B. and Heintz, R. H. (1974) Economics of Shelterbelt Influence on Wheat Yields in North Dakota. J. Soil and Water Conserv., v. 29, p. 87-91.
- Moss, H. C. (1935) Some Field and Laboratory Studies of Soil Drifting in Saskatchewan. Sci. Agr., v. 15, p. 665-675.
- Read, R. A. (1964) Tree Windbreaks for the Central Great Plains. USDA, Agr. Handbook No. 250.
- Secretariat (1977) Desertification: Its Causes and Consequences. Compiled and Edited by the Secretariat of the United Nations Conference on Desertification, Nairobi. Pergamon Press.
- Siddoway, F. H., Chepil, W. S. and Armbrust, D. V. (1965) Effect of Kind, Amount, and Placement of Residue on Wind Erosion Control. Trans. ASAE, v. 8, p. 327-331.
- Skidmore, E. L. and Hagen, L. J. (1977) Reducing Wind Erosion With Barriers. Trans. ASAE, v. 20, p. 911-915.
- Skidmore, E. L., Hagen, L. J., Naylor, D. G., and Teare, I. E. (1974) Winter Wheat Response to Barrier-Induced Microclimate. Agron. J., v. 66, p. 501-505.
- Skidmore, E. L., Hagen, L. J. and Teare, I. D. (1975) Wind Barriers Most Beneficial at Intermediate Stress. Crop Sci., v. 15, p. 443-445.
- Skidmore, E. L., Kumar, M. and Larson, W. E. (1979) Crop Residue Management for Wind Erosion Control in the Great Plains. J. Soil and Water Conserv., v. 34, p. 90-96.
- Skidmore, E. L., Nossaman, N. L. and Woodruff, N. P. (1966) Wind Erosion as Influenced by Row Spacing, Row Direction, And Grain Sorghum Population. Soil Sci. Soc. Am. Proc., v. 30, p. 505-509.
- Skidmore, E. L. and Siddoway, F. H. (1979) Crop Residue Requirements to Control Wind Erosion. In W. R. Oschwald (ed.) Crop Residue Management Systems. ASA Special Pub. No. 31, p. 17-33.
- Staple, W. J. and Lehane, J. H. (1955) The Influence of Field Shelter-Belts on Wind Velocity, Evaporation, Soil Moisture, and Crop Yields. Canadian J. Agr. Sci., v. 35, p. 440-453.
- Svobida, L. (1940) An Empire of Dust. Caxton Printers, Ltd., Caldwell, Idaho, 203 p.

- Woodruff, N. P., Chepil, W. S. and Lynch, R. D. (1957) Emergency Chiseling to Control Wind Erosion. Kans. Agr. Expt. Sta. Tech. Bul. 90.
- Woodruff, N. P., Dickerson, J. D., Banbury, E. E., Erhart, A. B., and Lundquist, M. C. (1976) Selected Trees and Shrubs Evaluated for Single-Row Windbreaks in the Central Great Plains. USDA, ARS, NC-37.
- Woodruff, N. P., Lyles, L., Dickerson, J. D., and Armbrust, D. V. (1974) Using Cattle Feedlot Manure to Control Wind Erosion. J. Soil and Water Conserv., v. 29, p. 127-129.
- Woodruff, N. P. and Siddoway, F. H. (1965) A Wind Erosion Equation. Soil Sci. Soc. Amer. Proc., v. 29, p. 602-608.
- Zingg, A. W. (1953) Wind-Tunnel Studies of the Movement of Sedimentary Materials. In Proc. Fifth Hydraulic Conference, Iowa Institute of Hydraulics, Res. Bul. 34, p. 111-135, John Wiley and Sons, New York, New York.
- Zingg, A. W. (1954) The Wind Erosion Problem in the Great Plains. Trans. Am. Geophys. Union, v. 35, p. 252-258.

WIND EROSION IN THE INTERIOR DELTA OF THE NIGER IN MALI

By

H. K. Barth

ABSTRACT

Because of its azonal hydrological run-off regime, the inland delta of the Niger (Figure 1) occupies a special place in the semiarid region on the southern edge of the Sahara. The ecosystems there are marked by a periodic rhythm of flooding followed by recession of the floodwater, accompanied by partial drying out of the delta area (Figure 3). This hydrologic cycle and the relatively favorable soils of high clay content with a humic topsoil make the large-scale cultivation of rice feasible. In fact, at present, a number of projects are underway, planning 32,000 hectares of rice fields in the eastern part of the delta alone. Altogether the present area under rice cultivation in the delta is about 100,000 hectares and it is intended to double by 1985.

It is evident that such extensive exploitation has far-reaching effects on the natural balance of the ecosystem. One of the consequences is the particularly serious threat of the erosion of soil through the action of wind. The annual wind distribution pattern for Mopti (Figure 4) is derived from the seasonal shifts in the ITC under the varying influence of the air flows from the trade winds and the monsoon. Northerly to easterly winds prevail from December to April and southerly to westerly winds dominate from June to September. According to Tables 1 and 2 there are peak values in wind speed of up to 54 m/sec. Assuming that wind-induced erosion occurs at wind velocities greater than 6 m/sec^{-1} , then there have been 27 (1971) days and 30 days (1972) with these conditions.

Figure 1
Survey of interior delta of the Niger

In the areas of rice cultivation, large fields are ploughed when the once-flooded basins have drained and dried out. When this happens, even normal harmattan winds can give rise to deflation, and the wind erosion attains massive proportions when the ITC shifts northwards in June and maximum wind speeds are reached. The load picked up by the wind causes turbidity in the atmosphere, and visibility can drop to less than 5 m. In one dust storm, measurable erosion occurs in the alluvial sandy-silty clay soils. The consequences of this deflation can be detected in profiles (Figures 2 and 3) although large-scale rice cultivation did not start here until 1974.

The degraded profile does not contain any humic topsoil with its dark grey to greyish-blue color; instead, the surface is formed by a coarse-sand bleached horizon. In fact, the humus content is less than 0.5%, which is much lower than that of the nondegraded topsoil which has a humus content of 4%. There is a clear increase in the coarse fraction in the grain size distribution, and a pronounced drop in the silt-clay fraction.

The degradation of the hydromorphic soils of the delta as a result of wind erosion, i.e., the decrease in the humus and nutrient content, the changes in soil structure and soil texture as well as the associated changes in the water economy all have grave consequences in the productivity of rice. In fact, there has been a reduction in yield from 4.6 t/ha in 1974 to 2.8 t/ha in 1978.

If greater change to the natural equilibrium of the delta regions is to be prevented, measures for the restraint of wind erosion are unavoidable and must be carried out immediately.

1. AREA STUDIED AND STATEMENT OF THE PROBLEM

Because of its azonal hydrological runoff regime, the inland delta of the Niger River in Mali occupies a special place in the semiarid region on the southern edge of the Sahara, and the geoecological structure of the area is marked by corresponding special features. Instead of being "Sahelian", the main characteristics of the region are determined by the periodic rhythm of flooding followed by recession of the floodwater, and then low water accompanied by partial drying out of the soil.

At first, the Niger exhibits a tropical semihumid runoff regime, since its headstreams are in the marginal highland region of Fouta Djallon, a zone of high precipitation. As the terrain rapidly levels out downstream from Bamako, the Niger flows as an exotic stream in a northeasterly direction into the broad alluvial plains of the inland delta where vast areas are flooded. At high water the flooded regions form an amphibious river landscape consisting of countless river branches,

Figure 2
Soil profiles in different areas of Niger Delta

Figure 3
Soil profiles before and after dust storm

lakes, and swampy regions that represent Mali's "lifeline" in the dry regions of the Sahel. In the midst of the dry regions, which are the scene of extensive nomadism and sparse agricultural activity during the wet season, this particular zone opens up many economic opportunities of which, in addition to inland fishing, the growing of rice is particularly important. In fact, at present a number of projects are underway that plan for the cultivation of approximately 32,000 hectares of rice fields in the eastern part of the delta alone. Altogether, the present area growing rice in the delta is estimated to be about 100,000 hectares, and this figure is intended to double by 1985.

It is already evident that such extensive exploitation of the natural potential of the region will have far-reaching effects on the natural ecosystems. The erosion of the soil through the action of the wind constitutes a particularly serious threat to the existing agricultural potential. This paper reports on the observations and studies that have been carried out since 1973 on the extent and effects of aeolian soil erosion.

2. COMPONENTS OF AEOLIAN MORPHODYNAMICS

Based on the cycle of erosion conceived by Chepil (1961) (see diagram in Figure 4), there are a number of factors responsible for the process of deflation in the inland delta of the Niger. Various erosion-producing conditions result in destabilization of the soil and thus give rise to aeolian erosion.

3. RUN-OFF REGIME IN THE DELTA

The semihumid, tropical precipitation cycle in the catchment area and the geomorphological conditions on the northern slope of the Guinean ridge or in the delta itself determine the water economy and run off in the delta region. The increased quantities of precipitation falling in the summer months in the region of the Guinean ridge release a flood wave that reaches Bamako in September. As the floodwater enters the delta, the particularly high retentive capacity and the slight longitudinal gradient of the sedimentary filling of the basin zone cause a considerable phase shift in the further progress of the floodwater. After countless existing and abandoned river branches as well as numerous mares ("Mare,"--French word meaning "puddle," "stagnant pool"--is a shallow, periodically or perennially water-bearing terminal basin) and lakes fill up, the water spreads over the flood plains. In this way the highest water levels are recorded in Koulikoro in mid-September, in Mopti in mid-November and in Dire only in December (Figure 5).

Figure 4
Chepil's (1961) cycle of erosion

Figure 5
Agricultural activity as the cause of wind erosion of the soil

4. WIND CONDITIONS

The annual wind distribution pattern for Mopti as depicted in Figure 6, is derived from the seasonal shifts in the ITC, under the varying influence of the air flows of the trade winds and the monsoon. According to this diagram, northerly to easterly winds prevail in the months from December to April, and southerly to westerly winds dominate from June to September--the wind directions vary in the other months. Consequently, during the low-water period in the delta (between February and June), which is the period when wind erosion occurs, the harmattan blows from February to April, and the southwesterly monsoon winds prevail in June.

However, the decisive factor from the point of view of the aeolian dynamics is the wind speeds that occur during this period. In judging the wind erosion, it must be assumed according to Woodruff et al. (1972), that wind-induced erosion occurs at wind speeds greater than 5.7 m/sec^{-1} at a height of 0.3 m above the ground. The values listed in Tables 1 and 2 show that from February to June, on 27 days (1971) and on 30 days (1972) the wind speed was $\geq 6 \text{ m/sec}^{-1}$ and on 10 days (1971) and 25 days (1972) the wind speed was $\geq 16 \text{ m/sec}^{-1}$. The absolute peak values during this season were between 12 and 54 m/sec^{-1} . These data are derived from ASECNA Mali in Bamako. (Tables 1, 2, 3). They were recorded at Mopti airport at a height of 8 m above the ground, and they indicate that there is a considerable risk of wind erosion in the delta region near Mopti, even if we convert the wind speeds to those prevailing at 0.3 m above the ground (see Roth and Tuchenhausen, 1975) using other surface roughness parameters than those actually existing at the measurement site. Taking the roughness parameter at the measurement site to be $1 \times 10^{-4} \text{ m}$ (open airfield; see Eagleston, 1970), and that in the flood regions of the delta to be $1 \times 10^{-5} \text{ m}$ (open cultivated areas on the alluvial soils of the Niger), the values shown in Table 3 are obtained for the effective wind speeds. A comparison between the figures in Tables 1 and 2 indicates that the wind speeds calculated for the cultivated surfaces differ only slightly from the measured values.

5. SOILS AND VEGETATION

Although at first sight, the landscape of the interior delta of the Niger appears very monotonous, it actually contains a surprising wealth of different soil and vegetation types that accompany the kaleidoscopic variations in the meso-macrorelief of the flood zones. Differences in height amounting merely to a few centimeters cause considerable differences in the hydromorphically structured environment. An example of this is provided by the profile of the flood zone

Figure 6
The annual wind distribution for Mopti

Table 1
Wind Conditions in 1971 at Mopti*

	u (Max) m/sec ⁻¹	Numbers of Days When u $\geq$ 6 m/sec ⁻¹	Numbers of Days When u $\geq$ 16 m/sec ⁻¹
January	11	1	-
February	12	3	-
March	15	3	-
April	21	5	1
May	54	7	4
June	25	9	5
July	25	17	4
August	16	12	2
September	19	13	3
October	25	3	1
November	14	4	-
December	13	2	-
Total		79	0

*Data obtained from ASECNA, Bamako

Table 2
Wind Conditions in 1972 at Mopti*

	u (Max) m/sec ⁻¹	Numbers of Days When u $\geq$ 6 m/sec ⁻¹	Numbers of Days When u $\geq$ 16 m/sec ⁻¹
January	15	4	-
February	14	2	-
March	19	5	2
April	25	6	2
May	25	8	5
June	29	9	6
July	36	11	7
August	26	15	9
September	21	8	4
October	14	7	-
November	11	2	-
December	10	3	-
Total		80	35

*Data obtained from ASECNA, Bamako

Table 3

Effective Wind Speeds at a Height of 0.3 M Above the Ground in the Area Studied (Calculated According to the Method of Roth and Tuchtenhagen, 1975)

U in m/sec^{-1} (measured value)	u (eff.) in m^{-1} (calculated value)
7.5	7.1
10	9.2
12.5	12.3
15	14.2
17.5	16.4
20	18.5
22.5	20.6
25	22.7
27.5	24.9
30	27.1

Figure 7

Dam with a floodgate in the flooded plains of the eastern interior delta of the Niger near Gania during the high water period. The natural hydromorphic grasses (Vetiveria, Echinochloa, Eragrostis associations) are replaced by many thousands of hectares of rice fields

near Gania, south of Mopti, which is shown in Figure 2. The area studied represents a characteristic section of the amphibious river landscape of the Niger. Flat basin zones, which are divided up into many sub-units by barely perceptible differences in height and which are intersected by old branches of the river and by mares, extend to both sides of the river between modern and fossil levees. The annual flooding that occurs here happens either through an intact system of abandoned channels that are still connected with the river or--and this is most frequently the case--through decantation over the system of levees that accompanies the river.

The areas that are all completely flooded during the high water period gradually emerge in stages as the water recedes. This means that some parts of the land remain under water longer than others, and as a result a clear topomorphic stockwork structure forms in individual areas. The different degrees of waterlogging expressed above are all part of the formation of the soils and the vegetation.

Without being able to go into the different profiles of the gley-like alluvial soils in any more detail at this point (see Barth, 1977), it is necessary to draw attention to the grain size distribution, which is very important as far as the problem of wind erosion is concerned. As the grain size diagrams show (Figure 2), the soils in question are sandy-silty clay soils with a relatively homogeneous vertical composition. Their clay content increases proportionately with increasing distance from the river towards the interior of the flood basin. This fact is explained by the process of sedimentation that occurs with every incidence of flooding; Gallais (1967) has calculated a mean sedimentation value of 0.07 to 0.2 mm for the entire area of the "delta vif" (in the central and lower regions of the delta, the rate of deposition of fine sediments and suspended matter amounts to several millimeters per annum (Gallais, 1967)). Depending on the vegetation and the flow conditions this process of sedimentation follows very different patterns: In the areas where there is dense growth of Oryza and Echinochloa, high sedimentation values are recorded. In the associated basin zones, only very fine material, which has diffused through the filter of Echinochloa settles out. On the other hand, stands of Andropogon and Vetiveria are less effective in encouraging sedimentation to occur, and so only the larger grain sizes settle out.

6. AGRICULTURAL DEVELOPMENT

The inland delta of the Niger occupies a special position in the Sahel region of Mali, because the individual geocomponents permit various possible forms of exploitation. For example, the favorable conditions encountered in this region with its azonal hydrological runoff regime have led to a con-

centration of population, and to the blossoming of economic activities in a way that is unparalleled anywhere else in the Sahel zone (Figure 7). In recent decades, large areas have been developed for the cultivation of rice, thus adding to the areas taken under cultivation by the Office du Niger during the period of French colonial occupation. In the vicinity of Mopti alone, the cultivated areas were increased to more than 20,000 hectares in 1975 as part of the Operation Riz project, and a further 12,000 hectares for rice-growing were added by 1980, according to information supplied by the Ministère de la Production (Rapport de la factibilité de l'Operation Riz, Bamako 1971).

7. EXTENT AND EFFECTS OF WIND EROSION

Rice-growing involves the ploughing of enormous fields, once the flooded areas have drained and dried out. When this happens, even normal harmattan winds can give rise to deflation, and the erosion attains massive proportions when the ITC shifts northwards in June, and the peak wind speeds are reached. An additional factor is the frequent practice of burning the rice straw following the harvest. Apart from the effects of the fire itself, the burning process also reduces the amount of humus in the soil. This is then accompanied by many undesirable consequences on the soil structure, soil fauna, mineral content, and soil water content. The removal of the ash residues by the wind causes a particularly severe loss to the soils. In the days before cultivation commenced, the flood zones were well protected from wind erosion, even when certain areas of the basin had dried out. This protection was provided by the relatively dense associations of Vetiveria, Echinochloa, and Oryza. However, under the conditions that now prevail, the wind can act on the soil completely unimpeded, particularly since very few windbreaks have been constructed. The peak wind speeds measured in storms are in excess of 40 m/sec^{-1} . The load picked up by the wind causes turbidity in the atmosphere, and visibility can drop to less than 5 m. In one such storm, measurable soil erosion occurred. In the fields investigated at Gania-Kouakourou, where 10 measuring rods were distributed over a 100 m^2 area, the deflation amounted in certain areas from 1.8 to 2.1 cm of the cultivated topsoil.

But even normal winds, much less violent than a singular event such as a storm, can have an erosive effect. On days when wind speeds do not exceed $5 \text{ to } 10 \text{ m/sec}^{-1}$, the wind load can cause turbidity with visibility reduced to less than 1,000 m. Normal winds prevail for an average of 30 days in the five months from February to June alone.

The extent of this deflation can be detected in the soil profiles from the cultivated area under investigation near

Gania, although large-scale rice production did not start until 1974 in this region. In the basin zone of Gania-Kouakourou, which is adjacent to the levees the upper sections of the profiles are clearly different from those of the soils supporting associations of Eragrostis, Echinochloa, and Vetiveria (Figure 3). For example, the clearly degraded profile does not contain any humic topsoil with its characteristic dark grey to greyish-blue color, (10 YR-6-3; color coded according to the scale of the Munsell Soil Color Chart, 1954 edition) and its friable structure. Instead, the surface is formed by a partly polyhedrally structured or coarse-sand bleached horizon (2, 5 Y-7-2), with only a few roots passing through it. In fact the humus content is less than 1%, which is much lower than that of the nondegraded topsoil, which has a humus content of >4%. The deflation which is a consequence of the rice cultivation has resulted in a considerable reduction in the height of the profiles of between 22 and 32 cm. Taking into account the annual sedimentation rate, it can thus be assumed that approximately 10 cm/annum of topsoil in the rice-growing areas is eroded by wind. Furthermore, there is a clear increase in the coast fraction in the grain size distribution, and a pronounced drop in the silty-clayey fraction. In places it is even possible to observe surface enrichment of pisolitic FeO concretions, which are loosely scattered over the surface of the fields. This change in the grain size composition is without a doubt a result of the selective action of the aeolian dynamics. However, an additional factor is certainly the fact that under the new conditions created by the rice-growing activity, the sedimentation that occurs during the annually recurring flooding also undergoes a change. While the dense stands of the grasses Eragrostis, Echinochloa, and Vetiveria offered optimum sedimentation conditions, especially for the very fine grain fractions, the rice plants are less retentive and the water flow values are higher. Thus, either no sedimentation occurs, or else only coarse grain sizes settle out. This in turn has consequences for fertility of the soils. When these basin zones drain and dry out it is possible to detach from the surface of the soil 5 to 10-cm-thick mats of flattened grass embedded in the clay and silt. The natural regeneration of the flooded soils that would ordinarily be brought about by the sedimentation is lacking, and increased amounts of fertilizer must be applied to compensate for this deficiency.

All deflation goes together with the correlative phenomenon of accumulation. As soon as the air flow diverges or slows down, some of the wind load is deposited. While very fine mineral particles are transported over very great distances, the sands are generally deposited just a short distance away. Areas covered by sand in this manner, as well as dune formations, are regularly encountered in the inland delta of the

Niger. The sand accumulates, for example on fossil dune systems in the delta. These old systems, although not very elevated, are the only rough surface features in the region. Thus, they generate an increase in friction and turbulence in the uniform airstream, causing the wind load to fall as sediment. The longitudinal shapes that show up in satellite images and give the appearance of recent mobile dunes are in fact these old dune formations that have become covered with sand in recent years.

In addition to the degradation of the hydromorphic soils of the delta that result from wind erosion, i.e., the drop in the humus and mineral content, there are many other serious changes in the soil structure and texture, and also associated changes in the water economy. There are a number of other consequences that are not mentioned in this paper but are important.

Apart from the cultivation of rice, fishing is another extremely important factor in Mali's economy. When enormous areas of land were put under rice, large sections of the "Bourgou" (Echinochloa stagnina) stands, which had been serving as spawning and feeding grounds for the fish, were eliminated. In fact, fish catches have been dwindling steadily ever since the extreme overfishing of the waters of the Niger delta that began with the dry period between 1967 and 1973.

ACKNOWLEDGEMENTS.

I am grateful to the Director of the Service Meteorologique, Monsieur Gologou, for permitting me to inspect the meteorological statistics, and for his friendly support.

REFERENCES

- Bagnold, R.A. (1954) The Physics of Blown Sand and Desert Dunes. London.
- Barth, H.K. (1977) Der Geokomplex Sahel. Untersuchungen zur Landschaftsökologie im Sahel Malis als Grundlage agrar- und weidewirtschaftlicher Entwicklungsplanung (The Geokomplex of the Sahel. Studies of the Geoecology of the Sahel of Mali as the Basis for Agricultural and Pasture Resources Planning). Tubinger Georg. Stud., 71, Sonderband 12, Tübingen.
- Chepil, W.S. (1946) Dynamics of Wind Erosion: VI. Sorting of Soil Material by the Wind. Soil Science, 61; pp. 331-340. Washington, D.C.
- Chepil, W.S. (1961) The Cycle of Wind Erosion. Trans. of 7th Intern. Congr. of Soil Science, Madison, Wisc., U.S.A., 1960, 1; pp. 225-231 Amsterdam.
- Eagleson, P.S. (1970) Dynamic Hydrology. New York.

- Gallais, J. (1967) Le delta interieur de Niger et ses bordures. Etude geomorphologique. Mem. et Doc. du Centre Nat. des Rech. Scient. Paris.
- Glander, W. (1956): Bodenerosion und ihre Bekämpfung (Soil Erosion and its Prevention). Berlin.
- Richter, G. (1965) Bodenerosion. Schaden und gefährdete Gebiete in der Bundesrepublik Deutschland (Soil Erosion. Damage and Endangered Areas in the Federal Republic of Germany). Forsch. z. dt. Landeskunde 152,. Bad Godesberg.
- Roschke, G. (1974) Land und Bodenabtrag durch boigen Wind (Erosion of Land and Soil by Gusty Winds). Zeitschr. f. Kulturtechnik u. Flurbereinigung, 15; pp. 6-20 Berlin/Hamburg.
- Roth, R. and M. Tuchtenhagen (1975) Die Abschätzung von bodennahen Windgeschwindigkeiten zue Beurteilung von Winderosionsgefahr (The Estimating of Wind Speeds Close to the Ground as a Means of Judging the Risk of Wind Erosion). Meteorol. Rundsch., 28; pp. 121-125. Berlin/Stuttgart.
- Woodruff, N.P., L. Lyles, F.H. Siddoway, and D.W. Fryrear (1972) How to Control Wind Erosion, Agric. Inf. Bull. 354; U.S. Dept. Agric., Washington, D.C.

AEOLIAN PROCESSES AND FEATURES
AT AMBOY LAVA FIELD, CALIFORNIA

By

Ronald Greeley

and

James D. Iversen

ABSTRACT

The Amboy lava field in the Mojave Desert affords the opportunity for the study of aeolian processes in a volcanic terrain. Atmospheric experiments show that the surface roughness of the pahoehoe lavas has a pronounced effect on the atmospheric boundary layer and increases the threshold velocity. Experiments also show that the presence of the Amboy cinder cone generates a zone of atmospheric turbulence that keeps the surface swept free of particles in the wake of the cone. Wind tunnel simulations compare well with field measurements and enable data to be obtained for other parts of the lava flow not studied in the field.

INTRODUCTION

The Amboy lava field in the Mojave Desert of southern California (Figures 1, 2) affords the opportunity to assess several aspects of aeolian processes. The Amboy lava field covers about 70 km² and consists primarily of vesicular pahoehoe lava. The field is situated in an alluvial-full valley between the Bullion Mountains to the southwest and the

Figure 1
General locations of Amboy test site in southern California.

Figure 2
Amboy lava field and other features in the immediate area.

Bristol Mountains to the northeast; within the valley, it lies between Bagdad Dry Lake to the west and Bristol Dry Lake to the east--both are playa lakes typical of the Mojave Desert. The prevailing strong winds are from the northwest across the smooth valley; sands and other particles are transported across the lava field resulting in a wide variety of aeolian features.

The objectives of this study were to determine: (1) the origin of the dark, apparently wind-related streak associated with the Amboy cinder cone (Figure 3); (2) the effect of the rough lava on the near-surface wind velocities and the transport of aeolian particles; (3) possible changes (size, shape, etc.) in windblown particles across the lava field, and (4) the origins and characteristics of various aeolian features such as ventifacts.

Cone Streak

When viewed from the air, one of the most striking aspects of the Amboy lava field is the dark streak that extends 4 km downward from the Amboy cinder cone. Hypotheses for the origin of the streak (Greeley and Iversen, 1978) include: (1) the cone-and-crater forms an obstruction ('wind shadow') that blocks particles from reaching the leeward area, leaving the basalt uncovered by particles; (2) the dark zone is composed of basaltic particles eroded from the cone and deposited downwind; or (3) the cone-and-crater causes a turbulent wind flow in the wake zone, creating an area of erosion that keeps the lavas swept free of particles. These hypotheses were tested using a combination of field studies, wind velocity measurements, and wind tunnel simulations.

Surface Roughness

It is well-known that surface roughness (the topographic relief of a surface and the spacing of the roughness elements) affects the ability of a given wind to move particles because of the influence of roughness on the structure of the lower part of the atmospheric boundary layer. Surface roughness has been studied in boundary layer wind tunnels and in field experiments for agricultural applications, but full-scale quantitative tests in geological environments are mostly lacking. The evaluation of the roughness parameter is of first-order importance in determining aeolian erosion rates, zones of preferential erosion versus deposition, and in the general expressions for threshold speeds for particle movement by the wind. Knowledge of these factors is critical for understanding the geology of all surfaces subject to aeolian processes.

Figure 3a
 High-altitude aerial photograph showing Amboy lava field (Amboy cone is the dark spot), community of Amboy, and the alluvial-fill valley in which the field is located. Prevailing winds sweep out of the northwest (left side of photograph). Area shown is about 25 km wide; north is toward the upper left.

Figure 3b
 Vertical aerial photograph showing part of lava field; meteorological towers used in the field experiments, sample locations (numbers), and desert pavement test sites (a, b) are identified.

In order to assess the affect of surface roughness under field conditions, two 15.24-m towers, each with an array of five cup anemometers and a wind vane were aligned with the dominant wind direction. Tower 1 was established upwind of the lava field (Figure 3) to measure the wind velocity profile for winds blowing across the smooth plain. Tower 2 was placed about 4 km downwind on the rough lava field. Concurrent measurements at the two towers were made of wind velocity and direction to determine the effect of the lava surface on the wind velocity profile.

Aeolian Particles

Samples of windblown material were collected in a line parallel to the prominent wind direction across the field (Figure 3) and analyzed in the laboratory to determine possible changes in various parameters such as grain size and percentage of basalt particles with distance of transport.

Miscellaneous Aeolian Features

Pits, grooves, and other ventifact features were studied on the basalt surfaces to assess the dominant long-term wind directions and to study the general formation of ventifacts in vesicular rocks. Other aeolian features and the development of desert pavement were also studied.

The study was initiated because the area appears to be analogous to regions on Mars that are wind-swept basaltic plains. As such, the Amboy lava field has been the site of NASA-sponsored field trips (Greeley and Iversen, 1978) and as an engineering test site (Blanchard et al., 1977; Greeley and Bunch, 1976) for planetary applications.

Previous geological studies include those of Bassett and Kupfer (1964) who provide a geological reconnaissance of the southeastern Mojave Desert and the playas, Parker (1959, 1963) who studied volcanism at Amboy Crater, and Hatheway (1971) who included the Amboy lava field in part of an extensive study of basalt 'collapse' depressions. A series of open file reports (Watkins, 1965, 1966a, 1966b) of the U.S. Geological Survey include data on Amboy Crater. Studies of aeolian processes in the general area are summarized by Smith (1967).

GENERAL GEOLOGY

Figure 4 is a geological map after Hatheway (1971) showing the main flow units of the Amboy lava field. Most of the field is composed of undifferentiated flow units of relatively dense, 'degassed' pahoehoe (Swanson, 1973) lavas that form a hummocky terrain. The surface relief on this unit ranges from 2 to 5 m. The flow is characterized by abundant tumuli (small

Figure 4
Geological sketch map (after Hatheway, 1971) of Amboy lava field; sites 1-4 are locations where NASA spacecraft experiments were conducted.

Figure 5
Vesicular basalt showing pits and grooves resulting from aeolian erosion. (Photograph 434, R. Greeley, 1976).

mounds) and pressure ridges and, as typical for this type of flow, a fractured surface. Lava tubes are not present in any of the flows, nor are blisters or shelly-type pahoehoe; only a few lava channels are present. Low-lying areas on the flow are filled with aeolian sediments that range from a few centimeters to more than 1 m thick. The thickness of the aeolian deposits increases toward the southwest part of the field. Ventifacts (Figure 5 and 6) in the form of pits and grooves on the exposed basalts are prevalent over the entire flow and wind-faceted pebbles of basalt are also common.

The oldest flows occur in the eastern and southeastern parts of the field. They are characterized by numerous collapse depressions (circular pits) up to about 10 m diameter and several meters deep (Figure 7). Although the name would imply that 'collapse' depressions formed by the collapse of a crust over fluid lava, observations by Holcomb (personal communication) of active lava flows in Hawaii suggest that at least some depressions may form as a result of inflation of an emplaced (but still plastic) crust by molten lava around a general void in the flow. This mechanism would be more in concert with the general degassed nature of the hummocky pahoehoe.

Two other types of flows occur in the field: platform units and vent lavas (Figure 8). Platform units are isolated zones of uniform basalt that have relatively flat surfaces. These areas appear to represent stagnant parts of basalt flows that solidified in place with relatively little lateral movement after emplacement. The surfaces of the platform units consist of a layer of fist-sized cobbles and smaller fragments of basalt weathered in situ, and wind-blown sediments. Vent lavas are so named because they appear to be situated over some of the vents for the Amboy lava field. Vent lavas are topographically raised areas about 10 m high that consist of relatively dense pahoehoe. The flows appear to have cooled in relatively stagnant ponds that either drained back down the vent, causing slight crustal collapse on the surface, or broke out through subsequent flows around the edges of the pond. The surfaces are relatively flat and unfractured except around the edges of the pond. Individual collapse craters that could represent drain-back in the vents occur in the middle of the ponds. Like the platform areas, the surfaces of the vent lavas have weathered basalt fragments and aeolian particles to a depth of several centimeters.

Amboy Crater (Parker, 1963) is a prominent cinder cone (Figure 3) in the northeastern quadrant of the lava field. The cone rises 75 m above the surrounding lava flows and is about 460 m in basal diameter. It is composed of a loose accumulation of volcanic ejecta with secondary amounts of agglutinated ejecta and flows. The ejecta include angular scoriaceous tephra plus ropy, ribbon- and almond-shaped bombs. Some

Figure 6

View of desert pavement surface between basalt outcrops on the lava field, showing footprints resulting from dark basalt particles (pavement surface) being pushed into the underlying sand and silt. (Photograph 450, R. Greeley, 1976).

Figure 7

Low-altitude oblique aerial photograph showing 'collapse' depressions on the surface of the basalt flows; most depressions have been filled partly with aeolian sediments; jeep track indicates scale. (Photograph 448, R. Greeley, 1976).

Figure 8
Low-altitude oblique aerial photograph of one of the vent areas showing small crater (~70 m across) and dark wind-generated streaks associated with topography. (Photograph 430, R. Greeley, 1976).

Figure 9
View of two desert pavement test plots located in the streak zone (see Figure 3); each plot is 3 m square, (a) plot was cleared of layer of basalt particles into the underlying sand and silt (Nov., 1979); (c) view of both plots taken two months later, showing that the 'pressed' surface was reestablished very rapidly by winds removing the fine-grained material, reexposing the basaltic particles.

Figure 8
Low-altitude oblique aerial photograph of one of the vent areas showing small crater (~70 m across) and dark wind-generated streaks associated with topography. (Photograph 430, R. Greeley, 1976).

Figure 9
View of two desert pavement test plots located in the streak zone (see Figure 3); each plot is 3 m square, (a) plot was cleared of layer of basalt particles into the underlying sand and silt (Nov., 1979); (c) view of both plots taken two months later, showing that the 'pressed' surface was reestablished very rapidly by winds removing the fine-grained material, reexposing the basaltic particles.

lithic nonvesicular accessory basaltic ejecta are present, but included xenolithic fragments are absent.

The Amboy Crater is not a single cone, but is composed of at least four nearly coaxial nested cones (Parker, 1963). Although the outer slopes of the main cone are slightly gullied by erosion, particularly on the east side, they average about 30° , typical for fresh cinder cones. Within the main outer cone, there is a remnant of a second cone on the west side; both cones are breached on the west side. In addition to the two main cones, there are two relatively undisturbed cone walls within the main crater. These innermost conelets are composed almost entirely of angular scoriaceous cinders.

Cone Streak

The dark streak extending toward the southeast from Amboy cinder cone is about 4 km long and 0.5 km wide. The streak zone was examined in the field and found to consist primarily of exposed basalts and patches of desert pavement, and virtually no ash deposits or similar material derived from the cone. A study by Zimbelman and Kieffer (1980) shows that within the streak zone, rock concentrations reach a maximum of 90 percent per unit surface area, whereas adjacent surfaces outside the zone are typically 20 to 30 percent exposed rocks (i.e., 80 to 70 percent of the surface is mantled with aeolian sediments).

It was observed in the field that the zone of desert pavement (both inside and outside the streak area) consisted of gravel-sized basalt fragments underlain by very fine silt, leaving bright footprints. However, the original surface was reestablished after only one or two strong winds. To assess the development of pavement surfaces in relation to the cone streak, experimental plots 3 by 3 m were established both inside the streak zone and outside, as shown in Figures 3 and 9.

ATMOSPHERIC EXPERIMENTS

Data were taken at three meteorological towers from January 3 to February 28, 1976. Five cup anemometers were placed on Towers 1 and 2, and two anemometers were placed on Tower 3 at the beginning of testing. On Towers 1 and 2, the anemometers were placed at a distance of 0.73, 1.52, 3.14, 6.63 and 14.17 m and on Tower 3 at 0.94 and 13.64 m above the surface. On February 22, 1976, the five Tower 1 anemometers were placed on Tower 3 at heights above the surface of 0.76, 1.58, 3.17, 6.68, and 14.14 m. The twelve anemometers recorded accumulated travel of the wind, and therefore, only the average wind speed during the time from one instrument reading

lithic nonvesicular accessory basaltic ejecta are present, but included xenolithic fragments are absent.

The Amboy Crater is not a single cone, but is composed of at least four nearly coaxial nested cones (Parker, 1963). Although the outer slopes of the main cone are slightly gullied by erosion, particularly on the east side, they average about 30°, typical for fresh cinder cones. Within the main outer cone, there is a remnant of a second cone on the west side; both cones are breached on the west side. In addition to the two main cones, there are two relatively undisturbed cone walls within the main crater. These innermost conelets are composed almost entirely of angular scoriaceous cinders.

Cone Streak

The dark streak extending toward the southeast from Amboy cinder cone is about 4 km long and 0.5 km wide. The streak zone was examined in the field and found to consist primarily of exposed basalts and patches of desert pavement, and virtually no ash deposits or similar material derived from the cone. A study by Zimbelman and Kieffer (1980) shows that within the streak zone, rock concentrations reach a maximum of 90 percent per unit surface area, whereas adjacent surfaces outside the zone are typically 20 to 30 percent exposed rocks (i.e., 80 to 70 percent of the surface is mantled with aeolian sediments).

It was observed in the field that the zone of desert pavement (both inside and outside the streak area) consisted of gravel-sized basalt fragments underlain by very fine silt, leaving bright footprints. However, the original surface was reestablished after only one or two strong winds. To assess the development of pavement surfaces in relation to the cone streak, experimental plots 3 by 3 m were established both inside the streak zone and outside, as shown in Figures 3 and 9.

ATMOSPHERIC EXPERIMENTS

Data were taken at three meteorological towers from January 3 to February 28, 1976. Five cup anemometers were placed on Towers 1 and 2, and two anemometers were placed on Tower 3 at the beginning of testing. On Towers 1 and 2, the anemometers were placed at a distance of 0.73, 1.52, 3.14, 6.63 and 14.17 m and on Tower 3 at 0.94 and 13.64 m above the surface. On February 22, 1976, the five Tower 1 anemometers were placed on Tower 3 at heights above the surface of 0.76, 1.58, 3.17, 6.68, and 14.14 m. The twelve anemometers recorded accumulated travel of the wind, and therefore, only the average wind speed during the time from one instrument reading

to the next could be determined. Wind direction information was also recorded during the experiments with one direction indicator on each of Towers 1 and 2. Wind roses for Towers 1 and 2 are illustrated in Figure 10.

In total, about 120 wind periods were measured on Tower 1, 130 on Tower 2, 60 on Tower 3 with two anemometers, and 16 on Tower 3 with five anemometers. Of these, the wind was blowing fairly strongly (4.5 m/s or greater at the top anemometer) and was also nearly unidirectional from the northwest (parallel to the streak direction) 29 times for Tower 1, 24 times for Tower 2, 12 times for Tower 3, and 5 times for Tower 3. Some of these time periods were overnight, and all but three of these were discarded in calculating the composite wind speed profiles discussed below. Finally, 15 sets of data were chosen for consistency and validity of results for Tower 1, 16 sets for Tower 2, and four sets for Tower 3 (Table 1). The average measurement periods for these three sets were 5.49, 6.48, and 6.19 hrs., respectively. Because the relative heights of the anemometers above the mean terrain surface surrounding each tower was not unknown, the wind velocity logarithmic profile is written including a zero-plane displacement height, D , as follows:

$$u = \frac{u_*}{0.4} \ln \frac{z + D}{Z_0} \quad (1)$$

where u_* = friction speed.

It is assumed in writing Eq. 1 that the atmosphere is neutral or very near neutral, an assumption borne out by the closeness of fit to the data shown below. In calculating the values of D and equivalent roughness height, Z_0 , first several values of D were assumed and the profile data for each run were then used to find the values of u_* and Z_0 , which resulted in a least squares best fit (linear regression). The value of zero-plane displacement height, D , was then interpolated to find the fit to the data that gave the smallest sum of residuals to the least squares fit (Table 1).

The average wind speeds corresponding to the sums of the 15, 16, and 4 data sets for Towers 1, 2, and 3, respectively (Table 2), were also fit to Eq. 1 by the same technique. The data show that the only direction from which the wind blows fairly strongly other than the northwest is from the opposite direction, the southeast. The only time a southeast wind blew above 7 m/s at the top anemometer (threshold), however, was during a rainstorm, so there was little or no movement of sand with the wind blowing from the southeast during the period of the experiment. In contrast to the amount of wind from the northwest (29 runs for Tower 1 unidirectional and greater than 4.5 m/s), strong winds from the southeast were recorded only 9

Figure 10
Windrose diagrams for Tower 1 (a) and Tower 2 (b), note that most of the winds stronger than 4 m/s are in the streak direction.

Figure 10
Windrose diagrams for Tower 1 (a) and Tower 2 (b), note that most of the winds stronger than 4 m/s are in the streak direction.

Table 1(a)
Northwest Wind, Tower 1 data

Run No.	Period, hr	D(m) ^a	Z ₀ (m) ^b	Average Speed, Top Anemometer, m/c
10	8.48	-0.143	0.0014	6.69
53	10.18	0.109	0.0042	4.84
69	5.80	0.076	0.0068	5.62
91	5.73	-0.168	0.0009	8.76
272A	0.25	-0.223	0.0010	5.72
273A	0.75	-0.119	0.0047	8.28
274A	0.25	-0.512	0.0000	5.90
275A	0.75	-0.387	0.0002	4.59
285*	17.00	0.110	0.0005	7.07
286	5.00	-0.402	0.0001	4.55
295	3.50	-0.396	0.0002	6.99
300	3.67	-0.290	0.0002	4.84
314	10.42	-0.235	0.0005	6.57
320	6.00	-0.512	0.0000	8.78
322	4.50	0.341	0.0151	5.43

^aZero-plane displacement height

^bRoughness height

Table 1(b)
Northwest Wind, Tower 2 data

Run No.	Period, hr	D(m)	Z ₀ (m)	Average Speed, Top Anemometer, m/s
11	9.53	0.143	0.0035	6.17
30	6.17	0.494	0.0108	8.06
33	5.92	0.384	0.0156	5.17
265	3.58	0.366	0.0051	7.17
268	1.42	-0.021	0.0008	9.03
279	1.00	0.290	0.0034	7.51
281	1.00	0.139	0.0016	6.12
283*	15.67	0.756	0.0277	7.37
288	7.42	0.006	0.0011	4.36
297	3.42	0.061	0.0007	7.55
316	7.50	0.287	0.0044	7.80
318*	15.67	0.777	0.0245	9.11
321	5.83	0.320	0.0048	9.70
343	6.08	0.049	0.0007	6.04
344	5.58	0.872	0.0245	4.58
366	7.83	0.329	0.0041	5.71

*Overnight runs.

Table 1(c)

Northwest Wind, Tower 3 data

Run No.	Period, hr	D(m)*	Z ₀ (m)*	Average Speed, Top
				Anemometer, m/s
342	5.50	3.182	0.886	7.07
345	6.17	5.182	2.151	4.79
356	5.25	3.447	1.103	3.84
367	7.83	3.591	1.181	6.24

*These numbers are actually not very meaningful because the Tower 3 wind profile is not logarithmic.

Table 2

Wind speed profiles

Tower No.	Z ₀ (m)	D(m)	u*/U @ 15.24 m	S/u*	No. of Runs
Summary--Northwest Wind					
1	0.00173	-0.113	0.0441	0.275	15
2	0.00847	0.436	0.0534	0.174	16
3	0.1007	0	0.0797	0.331	4
Summary--Southeast Wind					
1	0.00375	-0.0091	0.0481	0.159	8
2	0.00686	-0.0732	0.0519	0.138	12
3	0.03664	0.0	0.0663	0.300	1

Table 3

Wind and friction speed ratios

	NW Wind	SE Wind
Summary--Amboy Tower Data		
U ₂ */U ₁ * =	1.200	1.019
U ₃ */U ₁ * =	2.133	1.403
U ₃ */U ₂ * =		
@ Z + D = 15.24 m	1.733	1.347
U ₂ /U ₁ =	0.987	0.950
U ₃ /U ₁ =	1.177	1.018
U ₃ /U ₂ =	1.160	1.054

Table 4

Wind tunnel--field comparison

		Tower 2
Roughness-Z ₀ (m)	Wind Tunnel	0.016
	Field	0.009
u*/u ₁₅	Wind Tunnel	0.058
	Field	0.053

times on Tower 1, 11 times on Tower 2, and 7 times and 1 time for Tower 3. Of these, sums were made for 8 runs (Tower 1), 12 runs (Tower 2), and 1 run (Tower 3), and the zero-plane displacement (D) and equivalent roughness height (Z_0) were calculated for the northwest wind (Table 2).

The values of Z_0 and friction speed ratio, u^*/U (at 15.24 m), for Tower 3 are not very accurate because this tower was located adjacent to the cinder cone and thus was not located in an area of planar homogeneity or with a long uniform upwind fetch as were the other two towers. The values of Z_0 and u^* for Tower 3 were calculated assuming a zero-plane displacement value of $D = 0$ and are included only for comparison and cannot be regarded as reliable. The values of S/u^* are proportional to the square root of the sum of the residuals to the least-squares fit and thus represent relative goodness of fit.

The Z_0 values of Table 2 show that for the northwest wind, Tower 2 was located in a plain (lava flow) with an equivalent roughness height (Z_0) 4.9 times as great as Tower 1 (alluvial plain). The relative surface friction speed is 21 percent greater at Tower 2 than at Tower 1 (compared by equating wind speeds at $Z + D = 15.24$ m). The roughness height and friction speed calculated for Tower 3 (not expected to be quantitatively accurate) give still higher values. If the southeast wind is compared with the northeast wind, a smoother terrain is indicated for Tower 3 with the southeast wind (the cone is downwind for the southeast wind), about the same roughness for Tower 2 (still in the midst of the lava flow), and greater roughness for Tower 1 with the southeast wind (the upwind fetch for the southeast wind now consists of lava flow until close to the tower).

Figure 11 shows 9, 12, and 14 individual northwest wind profiles for Towers 1, 2, and 3, respectively. The lines drawn in these figures are curve fits of Eq. 1 using the values of Z_0 and D from Table 2. For Towers 1 and 2 in Figure 11, the logarithmic curves fit the experimental data quite well. For Tower 3 only the top three data points fit the equation closely. The curvature of the profiles in Figure 11c may be due to the complex wind regime in which the tower is located, i.e., in the cinder cone wake. The composite profiles used to calculate the values of Table 2 are shown for Towers 1, 2, and 3 in Figure 12. Again, for Towers 1 and 2 the data are fit very well by Eq. 1, whereas the data for Tower 3 exhibit the curvature.

Using the values of Z_0 and D from Table 2, values of the friction speed (u^*) were calculated for 32 runs, and then u^* values from Tower 2 were plotted against u^* values from Tower 1 taken during nearly the same time period (Figure 13). Although there is some scatter (partly caused by not quite coincidental time periods; it took some time for the data gatherer to travel from tower to tower to record the data), a

Figure 11
Wind speed profiles for various wind speeds (U); solid lines are least-squares fit to composites of several runs; symbols are individual run data points. Tower 1 (a) on alluvial plain, 3,290 m upwind of Amboy cone; Tower 2 (b), 400 m upwind of cone; Tower 3 (c) in the cone streak zone, 400 m downwind.

Figure 12

Wind speed profiles for composite runs; solid line, least-squares fit, symbols are composite data points; Tower 1 (a) for 15 run-composite; Tower 2 (b) for 16 run-composite; Tower 3 (c) for 4 run-composite.

Figure 13

Friction speed ratio Tower 2 to Tower 1; each data point represents an approximately concurrent time period. Because Tower 2 is in rougher terrain than Tower 1, the average ratio is greater.

Figure 14

Friction speed ratio Tower 3 to Tower 1; each data point represents an approximately concurrent time period. Tower 3 values of friction speed are more questionable than Tower 1 or 2 because of its nonlogarithmic profile.

Figure 15

Friction speed ratio Tower 3 to Tower 2.

least-squares fit results in an average ratio u_{*2}/u_{*1} of 1.20. The ratio u_{*3}/u_{*1} (Figure 14, 16 runs) is 2.133 and the ratio u_{*3}/u_{*2} (Figure 15, 18 runs) is 1.733. Most of the data points obtained in Figure 15 were for Tower 3 when only two anemometers were in place.

Final summaries of the data illustrated in Figures 13 to 15 are listed in Table 3 as well as data for the southeast wind, which are not illustrated. In summary, the most reliable conclusion regarding friction speed is that obtained from the comparison of Tower 1 and Tower 2 data. For the northwest wind, the lava surface is obviously rougher, giving a 20 percent increase in the surface friction speed. For the southeast wind, there is little difference because Tower 1 is downwind of the rough lava flow. Tower 3 values of surface friction are greater for both wind directions, but the southeast wind gives smaller values of surface friction than the northwest wind. Of interest are extrapolated values of wind speed at the top of each 15.24-m tower, which should be reliable for all three towers. These data were obtained for the same runs as illustrated in Figures 14 and 15. Of special interest is the fact that the ratios U_3/U_1 and U_3/U_2 for the northwest wind are well over 1.0, illustrating the little-known fact that wind speeds at certain places within an obstacle wake in the boundary layer can be higher than unobstructed upstream values. The enhancement of wind speed is apparently attributable to the existence of the horizontal trailing vortex system rather than the acceleration of air over the top of the crater, because the top of Tower 3 was about 60 m below the highest point on the crater rim.

WIND TUNNEL EXPERIMENTS

A scale model of the Amboy crater area (Greeley and Iversen, 1978; Iversen and Greeley, 1978) was constructed to a geometric scale (vertically and horizontally) of 1/1,000, based on U.S. Geological Survey 7 1/2' Amboy quadrangle (20-foot contour interval). The area built to scale represents a portion of the lava flow 1,219 m by 3,658 m (4,000 ft by 12,000 ft) with the long side in the direction of the streak, as shown in Figure 16 (Iversen and Greeley, 1978). The windward rim on the scale model is 2.33 m (about 7.8 rim diameters) downstream of the windward edge of the modelled region. The total fetch modelled in the wind tunnel upwind of the windward rim was 4.52 m, with the first 2.19 m of the tunnel floor upwind of the modelled area and covered with 2 cm rocks placed randomly on the floor about 4 cm apart to provide a thick, fully turbulent boundary layer over the modelled region (Cermak, 1975). Also shown in Figures 3 and 16 are the locations of the three meteorological towers erected in the field.

Figure 16
Topographic map of the wind tunnel model showing the area modeled (1,219 m by 3,658 m) to a geometric scale of 1:1,000. Location of Towers 1, 2, and 3 indicated by circled numbers.

Figure 17
Wind speed profile for Tower 2 comparing full-scale (field) and wind tunnel (model) data. Vertical distance is measured with respect to the local surface.

Velocity profiles were measured in the wind tunnel at 52 different locations. The comparisons between wind tunnel and atmospheric velocity profiles for Tower 2 and Tower 3 locations are shown in Figures 17 and 18. The wind tunnel was operated at free-stream speed of about 28 m/s, resulting in a value of fetchlength Reynolds number of about 9 million. The value of equivalent roughness height obtained from the curve in Figure 17 is 0.0016 cm, representing a full-scale roughness height of 0.016 m, which is about 1.8 times the value found in the atmosphere. Thus, the wind tunnel model was aerodynamically somewhat rougher than the atmosphere, which is not an entirely unexpected result for a geometric length scale ratio as large as 1,000. This accounts for the slight discrepancy in profile in Figure 17. The comparison for roughness and friction speed for Tower 2 are shown in Table 4.

Values of surface friction speed could be determined from most of the wind tunnel velocity profiles (Figure 19). Of particular interest are the high values of surface friction speed (0.051 average ratio) in the crater wake locations 7A, 8A, 25A, 26A, 33A, 45A, Tower 3; the extremely high values (average 0.074) on the lateral flanks of the cinder cone (locations 11A, 12A), the low values (0.034, 0.033) just upwind (location 18A) and downwind (location 44A) of the cinder cone, and the relatively uniform values (average 0.040) everywhere else. Thus, the average friction speed in the wake is 27 percent higher than the upstream value whereas the peak values on the cone flanks are 85 percent higher.

In summary, the field and wind tunnel experiments show that the maximum surface shear stress over the field occurs on the surface of the cinder cone and in its wake. This area of greater surface shear stress corresponds to the dark streak and thus illustrates that the dark streak is primarily caused by erosion of the light-colored surficial sand from the region of the streak and is not a so-called shadow zone where wind speeds are low. However, there is a small 'shadow' zone, a small triangular region starting at the crater base and extending downwind about a crater diameter or so.

Additional saltation experiments in the wind tunnel served to show that irregularities in the planform shape of the streak are caused by minor variations in topography. Regions of topographic prominencies are generally dark because of locally higher surface shear stress so that a greater percentage of basalt is exposed and desert pavement areas are kept swept free of sediments, and topographic depressions are lighter in color because sand is trapped in these regions.

SEDIMENT ANALYSIS

Samples of windblown material were collected along a traverse approximately parallel to the cone-streak across the

Figure 18
Wind speed profile for Tower 3 comparing field and wind tunnel data.

Figure 19

Locations of the wind tunnel profiles; numbers in rectangles are the ratios of surface friction speed to wind tunnel free stream speed, u_*/u_∞ .

lava field, as indicated in Figure 3b. Samples were acquired from drifts of material that occur primarily in low-lying 'pockets' among the basaltic hummocks; lack of vegetation on the drifts was taken as evidence that the grains are currently active. Samples of about 300 grams were taken with a hollow tube pushed into the drift in order to obtain a relatively homogeneous sample representative of the average active sediments.

In the laboratory, the samples were sieved and analyzed to determine grain size distribution, sorting, and compositions as functions of position on the lava field. Particle size distribution in phi units are shown in Figure 20, along with coefficients of sorting indicated by standard deviations. In general, all samples consist of fine-to medium-size sand and are moderately well-sorted, typical of aeolian sands. Maximum sorting occurs about 2/3 the distance downwind across the field. However, the differences are so slight that they may not be significant.

The samples were examined microscopically and classified in four groups: (1) light-colored grains (mostly quartz, plus feldspar), (2) basalt grains, (3) agglutinated grains (sand-size clumps consisting of fine particles cemented together), and (4) other grains. Of primary interest is the potential increase in the fraction of basalt grains as a function of distance from the windward edge of the field. As shown in Figure 20 there is a trend toward increased basalt grains with distance in transport across the field. Evidently as the mass of windblown material moves across the field, basaltic grains are added to the mass. Sample 11 (Figure 3) was acquired on the alluvial plain immediately off the lava field; the basaltic grains contained in the sample (Figure 16) probably were derived from weathering of the flow and accumulation along the flow front.

In general, the windblown sediments appear to change very little in the short (~8 km) transport across the field. The quartz and other grains derived from the alluvium adjacent to the field are picked up by the wind and moved onto the lava flow. Thus, sorting occurs in the initial stage of passing from the mixed fluvial/aeolian regime onto the lava field where aeolian processes dominate. The primary change across the field is the addition of basalt grains.

AEOLIAN FEATURES

Because desert pavement (Figure 6) appears to play an important role in the prominent appearance of the cone-streak, the development of pavement surfaces was monitored in the field. Experimental plots were established in the cone-streak zone and on the lava flow outside the streak zone (Figure 3b). In both localities, two plots were established in 1979; in

Figure 20

Histograms in particle sizes for samples collected across the lava field (numbers at the top of each data set are keyed to sample locations shown in Figure 3b); scale on left is percentage for each size range. For each sample, data for five sizes of particles are given, from left to right, bar 1 is <1.0 and ≥0.5 mm, bar 2 is <0.5 and ≥0.25 mm, bar 3 is <0.25 and ≥0.125 mm, bar 4 is <0.125 and ≥0.062 mm, and bar 5 is <0.062. Black bars indicate percentage of basalt particles of all sizes, keyed to the scale on the right. Standard deviations given at the bottom of each sample set indicates sorting.

Type A plots, the pavement surface of basalt particles was raked clean, leaving a bare sandy clay surface; in Type B plots, the basalt particles were pressed into the sand clay (Figure 9). The plots were then monitored monthly for more than a year. After only one month, the Type B plots had nearly reestablished the pavement surface through the winnowing of the fine grains to reexpose the basalt particles. Type A plots showed relatively little change, even after 1 year. There was no pronounced difference between the streak locality and the locality outside the streak zone.

In addition to the desert pavement surfaces, the vesicular basaltic rocks on the lava field were examined to assess their modifications by the wind. Most rocks facing the prevailing winds have been sculpted into ventifact surfaces. Although a rigorous study was not carried out, there does not seem to be any relation between the vesicles and pits and grooves abraded by windblown particles.

SUMMARY AND CONCLUSIONS

Combined field studies and wind tunnel experiments for the Amboy lava field lead to several results:

1. The prominent streak associated with Amboy cinder cone appears to be the result of turbulent winds generated by the topographic obstruction of the cone; the turbulence continues downwind from the cone and keeps the zone free of aeolian particles, leaving the basaltic bedrock exposed and creating desert pavement surfaces. Other streaks on the lava field appear to be generated in a similar fashion.
2. The greater surface roughness created by the lava flow in comparison to the alluvial valley is manifested by a change in the atmospheric boundary layer near the surface such that energy is removed from the wind by increased friction. The net result is a greater deposition of windblown particles on the lava flow. Presumably, the lava field would continue to act as a trap until it reached a surface roughness equivalent to the surrounding plains, at which 'equilibrium' time, the field would cause no effective change in the aeolian regime.
3. Analyses of sediments collected on the lava field and from the surrounding plains show an increase in sorting of the sands from the fluvial/aeolian environment as they pass onto the predominantly aeolian environment on the lava field. As the particles traverse the field, however, there is little change in their character other than for the mass as a whole to increase slightly in the percentage of basalt grains present.

ACKNOWLEDGMENTS

Since our studies began in 1972, at Amboy, many students have been involved in various aspects of the investigation. We particularly wish to thank Curt Wilbur and Bob Miller, who manned the meteorological towers during the collection of atmospheric data. Sedimentary analyses was carried out by Nancy Jackson, Bob Miller, and Maureen Smeltzer. Steve Williams assisted in establishing the desert pavement sites and monitored them during the experiment. Michael McClure constructed the wind tunnel model and assisted in the wind tunnel testing.

This study was supported by the National Aeronautics and Space Administration through the Planetary Geology Program Office.

REFERENCES

- Bassett, A.M. and Kupfer, D.H. (1964) A Geologic Reconnaissance in the Southeastern Mojave Desert. Special Report 83, California Division Mines and Geology.
- Blanchard, M., Bunch, T., Davis, A., Shade, H., Erlichman, J., and Polkowski, G. (1977) Results of Analyses Performed on Basalt Adjacent to Penetrators Emplaced Into Volcanic Rock at Amboy, California, April 1976. NASA Tech. Paper 1026, 15 p.
- Cermak, J.E. (1975) Applications of Fluid Mechanics to Wind Engineering--A Freeman Scholar Lecture. J. Fluids Eng. 97, 9-38.
- Greeley, R. and Bunch, T.E. (1976) Basalt Models for the Mars Penetrator Mission: Geology of the Amboy Lava Field, California. NASA, TM X-73, 125, 53p.
- Greeley, R. and Iversen, J.D. (1978) Field Guide to Amboy Lava Flows, San Bernardino County, California. In: Aeolian Features of Southern California: A Comparative Planetary Geology Guidebook, NASA, 24-52.
- Hatheway, A.W. (1971) Lava Tubes and Collapse Depressions. Univ. Arizona, Ph.D. Thesis.
- Iversen, J.D. and Greeley, R. (1978) Atmospheric and Wind Tunnel Experiments of the Amboy Crater Sand-Covered Lava Flow. Engineering Research Institute, Iowa State University, Report 78235, 79 p.
- Parker, R.B. (1959) Magmatic Differentiation at Amboy Crater, California. Am. Mineralogist, 44, 656-658.
- Parker, R.B. (1963) Recent Volcanism at Amboy Crater, San Bernardino County, California. Special Report 76, California Div. Mines and Geology.

- Smith, H.T.U. (1967) Past Versus Present Wind Action in the Mojave Desert Region, California. Report 67-0683, Air Force Cambridge Research Lab.
- Swanson, D.A. (1973) Pahoe-hoe Flows From the 1969-1971 Mauna Ulu Eruption, Kilauea Volcano, Hawaii. Geol. Soc. Amer. Bull., 84, 615-626.
- Watkins, J.S., ed. (1965) Investigation of In Situ Physical Properties of Surface and Subsurface Site Materials by Engineering Geophysical Techniques. U.S. Geol. Surv. Proj. Rept. to NASA for Fiscal Year 1965.
- Watkins, J.S., ed. (1966a) Investigation of In Situ Physical Properties of Surface and Subsurfaces Site Materials by Engineering Geophysical Techniques. U.S. Geol. Surv. Quarterly Rept., 1 Oct. 1965 to 31 Dec. 1965.
- Watkins, J.S., ed. (1966b) Investigation of In Situ Physical Properties of Surface and Subsurface Site Materials by Engineering Geophysical Techniques. U.S. Geol. Proj. Rept. to NASA for Fiscal Year 1966.
- Zimbelman, J. and Kieffer, H. (1980) Field Investigations of Dark Streaks. Bull. Amer. Astron. Soc. 12, 723.

AEOLIAN PROCESSES IN THE ENVIRONMENTAL WIND TUNNEL
AND IN THE ATMOSPHERE

By

J. D. Iversen

ABSTRACT

The environmental wind tunnel is a useful tool for simulating atmospheric winds near the Earth's surface and their effects. It has been used to determine surface pressure on, and flow patterns around buildings and other boundary layer obstructions. Several wind tunnels have been constructed for modeling saltation phenomena, i.e., blowing sand and snow at small scale.

DESCRIPTION

Probably the most important criterion in the successful small-scale physical modeling of events caused by winds within the atmospheric boundary layer is to correctly reproduce the upwind aerodynamic roughness. This criterion, according to Jensen (1958), is important in establishing the appropriate boundary layer velocity profile and turbulent spectral characteristics (Cermak, 1971, 1975).

The environmental wind tunnel is generally characterized by a long test section designed to develop a turbulent boundary layer whose velocity and turbulence characteristics match those of the atmospheric boundary layer that is modeled. Even with long test sections, however, it is sometimes necessary to

augment turbulence intensity and/or boundary layer depth with upstream turbulence generators.

Several wind tunnels have been constructed for the purpose of studying the basic phenomena of saltating sand or snow and/or for modeling saltation phenomena at small scale. The primary facility for studying the transport of sand by wind was that of Bagnold (1941). Other wind tunnels for the study of soil and sand transport by wind have been built by Malina (1941), Ford (1957), Chepil and Woodruff (1963), Iversen et al. (1973), Greeley et al. (1974a, 1974b), and Nickling and Ecclestone (1980). Those constructed for simulating snowdrifting include those of Becker (1944), Finney (1939), Gerdel and Strom (1961), Brier (1972), Kind (1976), Møller (1980), and Anno (1980). Tabler (1979) has also conducted small-scale experiments in the atmosphere by using real snow as the modeling material.

The possible applications of the aeolian boundary layer wind tunnel are many and varied. The basic phenomena that can be studied include: (1) determination of saltation threshold (discussed later); (2) particle trajectory (White and Schulz, 1977); and (3) mass transport (White, 1979). Modeling of surface topographic features is useful in the determination of local wind speeds (see Figures 1, 2, and 3) and in the determination of qualitative and quantitative aspects of erosional and depositional patterns due to saltating material.

Flow visualization can be a useful tool in the investigation of flow field characteristics. Different methods include the use of steam, smoke, tufts, and surface films (Merzkirch, 1974; Brandt and Iversen, 1977). There are also many aeolian effects in nature that can lead to increased understanding of local flow fields (Iversen, 1980).

REFERENCES

- Anno, Y., Hokkaido, Japan (1980) Private Communication.
Bagnold, R.A. (1941) *The Physics of Blown Sand and Desert Dunes*. Methuen, London.
Becker, A. (1944) Natural Snow Fences Along Roads. *Bautechnik* 22: 37-42.
Brandt, S.A. and Iversen, J.D. (1977) Merging of Aircraft Trailing Vortices. *Journal of Aircraft* 14: 1212-1220.
Brier, F.W. (1972) Snowdrift Control Techniques and Procedures for Polar Facilities, Technical Report R767, Naval Civil Engineering Laboratory, Port Heuneme, California.
Cermak, J.E. (1971) Laboratory Simulation of the Atmospheric Boundary Layer. *AIAA Journal* 9: 1746.

Figure 1

Surface pressure distribution in a model crater in an environmental wind tunnel. Deposition of erodible material will generally occur where the pressure coefficient is positive or in separated flow regions where it is slightly negative. Erosion will occur generally in nonseparated flow regions where the pressure coefficient has a large negative number.

Figure 2

Isotachs (constant wind speed) for flow in the center plane over a model crater. Note the high speed over the windward rim and the low speeds forward of the windward rim.

Figure 3

Wind speed profiles over a model crater.

- Cermak, J.E. (1975) Applications of Fluid Mechanics to Wind Engineering--A Freeman Scholar Lecture. *Journal of Fluid Engineering*, 97:9-38.
- Chepil, W.S. and Woodruff, N.P. (1963) The Physics of Wind Erosion and Its Control. *Advances in Agronomy* 15: 211-302.
- Finney, E.A. (1939) Snowdrift Control by Highway Design. Bulletin No. 86, Michigan Engineering Experiment Station, East Lansing, Michigan.
- Ford, E.F. (1957) The Transport of Sand by Wind. *Trans. American Geophys. Union* 38: 171-174.
- Gerdel, R.W. and Strom, G.H. (1961) Scale Simulation of a Blowing Snow Environment. *Proc. Inst. Environ. Sci.* 53: 53-63.
- Greeley, R., Iversen, J.D., Pollack, J.B., Udovich, N., and White, B.R. (1974a) Wind Tunnel Simulations of Light and Dark Streaks on Mars. *Science* 183: 847-849.
- Greeley, R., Iversen, J.D., Pollack, J.B., Udovich, N. and White, B.R. (1974b) Wind Tunnel Studies of Martian Aeolian Processes. *Proc. Royal Soc. London, Series A* 341: 331-360; also NASA TMX-62297.
- Iversen, J.D., Greeley, R., Pollack, J.B. and White, B.R. (1973) Simulation of Martian Aeolian Phenomena in the Atmospheric Wind Tunnel, AIAA/NASA/AST/IES 7th Space Simulation Conference, NASA SP-336: 191-213.
- Iversen, J.D. (1980) Photographic Determination of Transport Rate Effects in Blowing Sand and Snow. *Proc. Internat. Symposium on Flow Visualization*, Bochum, West Germany.
- Jensen, M. (1958) The Model Law for Phenomena in Natural Wind. *Ingenioren* 2: 121.
- Kind, R.J. (1976) A Critical Examination of the Requirements for Model Simulation of Wind-Induced Erosion/Deposition Phenomena Such as Snowdrifting. *Atmos. Environ.* 10: 219-227.
- Malina, F.J. (1941) Recent Developments in Dynamics of Wind Erosion. *Trans. American Geophys. Union*, Pt. II: 262.
- Merzkirch, W. (1974) *Flow Visualization*. Academic Press, New York.
- Møller, J. (1980) Aarhus, Denmark. Private Communication.
- Nickling, W.G. and Ecclestone, M. (1980) A Technique for Detecting Grain Motion in Wind Tunnels and Flumes. *J. Sedimentary Petrology* 50(2).
- Tabler, R. (1979) Natural Geometric Scaling of Snowdrifts Suggest Possibility for Outdoor Modeling. Preprint, Scientific Symposium on Snow in Motion, Ft. Collins, Colorado.
- White, B.R. and Schultz, J.C. (1977) Magnus Effect in Saltation, *J. Fluid Mech.* 81 (3): 497-512.
- White, B.R. (1979) Soil Transport by Winds on Mars. *J. Geophys. Res.* 84: 4643-4651.

THRESHOLD VELOCITIES FOR DUST PRODUCTION

By

Dale A. Gillette

ABSTRACT

There exists a minimum wind stress that corresponds to a "threshold velocity" needed to initiate wind erosion on desert surfaces. For both undisturbed and disturbed soils, threshold velocity is strongly related to the mode of the aggregate soil size distribution. For undisturbed soils, clay percentage is also related to threshold velocity due to its effectiveness in forming protective crusts. Other influences such as presence of nonerodible elements and special destabilizing mechanisms are discussed.

PHYSICAL PROPERTIES OF THE SOIL

The wind speed at which the generation of dust begins is that for which aerodynamic forces are sufficient to overcome those factors holding individual particles in the soil. A number of theoretical and experimental studies have been made for idealized particle systems (e.g., Punjrath and Heldman, 1971; Iversen et al., 1976; Bagnold, 1941; Chepil, 1951; and Greeley et al., 1973). These studies have not considered natural soil surfaces, however, nor man's impact upon soil surfaces and aggregate storage. Gillette et al. (in preparation) measured threshold velocities and other soil parameters on dry soils in both undisturbed and disturbed conditions. The disturbance was a pulverization of the surface by the wheels of a 3/4-ton pickup truck. A priority of threshold velocities versus soil types is given in Figure 1.

Figure 1
Wind speed for initiation of dust production versus
undisturbed or disturbed soil type.

Approximate range of

u_* cm/s ⁻¹	
20-40	1. Disturbed soils having less than 50% clay and less than 20% pebble (<1 cm diameter) cover.
25-40	2. Tilled bare sandy soils.
40-65	3. Disturbed pebbly soils.
45-70	4. Bare clay soils that have been disaggregated by natural forces.
40-150	5. Disturbed soils having a high salt content or more than 50% clay.
140-200	6. Undisturbed sandy soils having a crust and soils covered with fine gravel.
>150	7. Undisturbed soils having more than 50% clay and surface crusts and salt-crusts soils.
>180	8. Soils covered by coarse (>5 cm) pebbles.

Summary of Regression Analyses

The variable having the highest correlation with the threshold velocities that were measured is the mode of the dry aggregate size distribution. For all tests in which a threshold velocity was reached (i.e., all tests except undisturbed soils of clay content >20%)

$$u_{*t} = 64 + 0.0055 \text{ MODE} \quad (r = 0.64, \text{ d.f.} = 63)$$

where MODE is the mode of the soils dry aggregate size distribution in micrometers, r is the correlation coefficient, and d.f. are the number of degrees of freedom.

A summary formula for undisturbed soils is

$$u_{*t} > 200 \text{ cm/sec}$$

for a clay content greater than 20% and

$$u_{*t} = 390 - 3.3 (\text{sand } \%) \quad (r = 0.72, \text{ d.f.} = 25)$$

for a clay content less than 20%.

A summary formula for disturbed soils is

$$20 < u_{*t} < 60 \text{ cm/sec}$$

for a sand content greater than 90% and

$$u_{*t} = 14.5 + 0.0071 \text{ MODE} \quad (r = 0.76, \text{ d.f.} = 12)$$

where % colloidal clay is the percentage of soil mass in particles smaller than 0.5 micrometer.

The relationships for undisturbed soils imply that the threshold velocities of undisturbed soils depend on soil texture, fine soils having high threshold velocities, and sandy soils having lower threshold velocities. The relationships for disturbed soils state that the threshold velocities for finer pulverized soils is lower, while coarser pulverized soils having higher fine clay content have larger threshold velocities.

THE EFFECT OF NONERODIBLE ROUGHNESS ELEMENTS

The ordering of threshold velocities suggests that the threshold velocity is determined by the availability of loose, erodible, sand-sized particles at the surface. It also suggests that the threshold velocity is increased (erosion decreased) by the effect of aerodynamic partitioning of wind stress by nonerodible elements such as pebbles and larger

objects, and by the cementation of the soil by clay and salts.

Marshall (1971) studied the effect of pebble-like nonerrodible elements in absorption of wind stress. He found that for a configuration in which the silhouette area of the nonerrodible roughness element per floor area per element (L_c) was greater than 0.1, most of the wind momentum was absorbed by the elements and little stress was available for movement of fine particles lying between nonerrodible elements. Lyles and Allison (1976), studying the threshold velocity of sand between nonerrodible dowel rods, found a similar effect for $L_c > 0.1$. Thus, it would seem that even a relatively sparse pebble, brush, or boulder cover would give a large amount of protection from fugitive dust production by the wind.

A wind tunnel experiment by Gillette (1978) investigated the effects of wind and sandblasting on consolidated and unconsolidated samples of six different dry soil textures. Surface crusts had the effect of raising threshold velocities to values beyond the wind tunnel range. Sandblasting with spherical glass beads was sufficient to break the surface crust of a sample that had surface textures of fine sand and loamy fine sand; erosion was initiated within a few minutes of exposure. The surface crust samples having finer surface textures were not disintegrated by sandblasting in the wind tunnel exposures (10 minutes at a friction velocity of about 100 cm s^{-1}). Threshold velocities of the loose soil samples varied, with sandy and powdery soils eroding at lower velocities than soils having some remaining aggregation.

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

$$\ln V_T = 6.0438 + 0.2332 C$$

where V_T is the threshold velocity in cm s^{-1} at 30.5 cm, and C is the percentage of soil mass in a sample of aggregates larger than 1 mm. On the basis of analysis, I believe that the effect observed by Bisal and Ferguson (1970) is a result, not of the partitioning of momentum by nonerrodible elements, but of increased soil cohesion, which requires larger wind forces for the onset of wind erosion.

Since finer textured soils in general tend to have more stable large soil aggregates, we would expect finer soil textures to correlate with higher threshold velocities. Mineralogical effects such as the cracking and breaking of montmorillonitic clays would tend to reduce aggregate size and threshold velocity, however. The presence of soluble salts increases threshold velocity as does the presence of decaying vegetation that feeds algae and fungus, which create a network of thread-like cells that tend to stabilize a soil.

SPECIAL MECHANISMS LEADING TO DUST PRODUCTION

There exist special mechanisms by which dust may be produced other than by disturbance of the soil and breakage of its crust:

1. Sand may be carried by the wind to initiate erosion in other wise nonerodible soils.
2. Certain salty soils form crusty domes that break open and expose loose, fluffy, and erodible soil below.
3. Water may drain into a dry lake basin and suspend the sediment by wind-wave action. The thin coating of sediment upon drying will often peel and blow in light wind.

REFERENCES

- Bagnold, R.A. (1941) The Physics of Blown Sand and Desert Dunes. Methuen, London, 265 p.
- Bisal, F. and Ferguson, W. (1970) Effect of Nonerodible Aggregates and Wheat Stubble on Initiation of Soil Drifting. Can. J. Soil Sci. 50, 31-34.
- Chepil, W.S. (1951) Properties of Soil Which Influence Wind Erosion, 4, State of Dry Aggregate Structure. Soil Sci. 72, 387-401.
- Gillette, D.A. (1978) A Wind Tunnel Simulation of the Erosion of the Soil: Effect of Soil Texture, Sandblasting, Wind Speed, and Soil Consolidation on Dust Production. Atmospheric Environment 12, 1735.
- Gillette, D.A., Adams, J., Endo, A., and Smith, D. (in preparation) Threshold Wind Velocities for Natural Desert Soil Surfaces Before and After Disturbances by Off-Road Vehicles.
- Greeley, R., Iversen, J.D., Pollack, J.B., Udovich, N., and White, B. (1973) Wind Tunnel Studies of Martian Aerolian Processes. NASA Tech. Memorandum 62:297.
- Iversen, J.D., Pollack, J.B., Greeley, R., and White, B.R. (1976) Saltation Threshold on Mars; the Effect of Interparticle Force, Surface Roughness and Low Atmospheric Density. Icarus 29, 281-393.
- Lyles, L. and Allison, B. (1976) Wind Erosion. The Protective Role of Simulated Standing Stubble. Trans. of ASAE 19, 61-64.
- Marshall, J. (1971) Drag Measurements in Roughness Arrays of Varying Density and Distribution. Agr. Meteor. 8, 269-292.
- Punjath, J.S. and Heldman, D.R. (1971) Mechanisms of Small Particle Reentrainment From Flat Surfaces. J. Aerosol Sci. 3, 429-440.

FORMATION OF RIPPLES, DUNES, AND ANTIDUNES IN RIVER BEDS

By

Jorgen Fredsøe

ABSTRACT

The present paper describes briefly the different bed configurations that develop in a river bed. Furthermore, the formation of different bed forms are explained by use of stability analysis. The importance of different physical mechanisms is reviewed.

1. BED CONFIGURATIONS

A plane stream bed will usually be unstable. It tends to break up and form one of the bed configurations illustrated in Figure 1, which is a picture based on the first extensive and systematic investigations made in a sufficiently wide flume (Fort Collins, USA).

The most important bed forms are ripples, dunes, plane beds, and antidunes; these will be further mentioned in what follows.

RIPPLES

When the critical tractive force is exceeded so that sediment transport starts, the bed will be unstable. In case of fine sediment, ripples are formed, while coarse sediments usually will form dunes.

Ripples are the notation of small triangular-shaped sand waves, usually shorter than about 0.6 meters and not higher than about 60 mm.

When the flow velocity is small a viscous sublayer is

Figure 1.
Typical bed forms in order of increased stream power.

formed, which gradually disappears as velocity increases. It is observed that ripples are formed, if a viscous layer is present, when the critical tractive force is just surpassed, while dunes are formed if the bed is hydraulic rough.

The ripple length depends on the sediment size (as well as other parameters) but is essentially independent of the water depth.

DUNES

Dunes are the large, more or less irregular sand waves usually formed in natural streams. It is by far the most important bed form in practical river engineering.

The longitudinal profile of a dune is roughly triangular with a mild and slightly curved upstream surface and a downstream slope approximately equal to the angle of repose (Figure 2). Flow separation occurs at the crest, reattachment at the trough, so bottom rollers are formed at the lee side of each dune.

On the stream side of the dune, the shear stress moves sediment particles uphill until they pass the crest and eventually become buried in the bed for a while. As sediment is moved from the stream side and deposited at the lee side of the dune, the result is a slow continuous downstream migration of dune pattern.

TRANSITION AND PLANE BED

For increased stream power, the dunes may tend to wash out, i.e., become much longer and flatter and finally disappear. This happens for Froude numbers between, say 0.6 - 1.0, depending on the grain size, hence at subcritical flow. This change from dunes to flat bed means a rather drastic reduction of the hydraulic resistance and water depth and is a great practical problem for river navigation.

I shall later revert to the problem of finding a physical explanation for the transition.

ANTIDUNES

Further increase in stream power leads from transition and plane bed to the so-called upper flow regime, the formation of antidunes and related configurations. In this case, the longitudinal bed profile is nearly sinusoidal, and so is the water surface but usually with a much larger amplitude [Figure 1(f) and (g)].

At higher Froude numbers the amplitude of the surface profiles often tend to grow until breaking occurs. After breaking, the amplitude may be small for a while and then the process of increase and breaking is repeated.

Figure 2.
Definition sketch of migrating dunes.

Figure 3.
Definition sketch, perturbation of sand bed.

The name "antidune" indicates the fact that the bed and surface profiles are moving upstream, particularly immediately before breaking.

Figure 1(h) illustrates an extreme form of antidunes occurring at high Froude numbers.

2. THE FORMATION OF BED FORMS IN ALLUVIAL CHANNELS

It has been a continuous challenge to scientists as to why the bed forms grow and change from one form to another, yet remain stable for a given set of flow and sediment conditions.

Most scientists today agree that the problem of sand wave formation is a problem of instability. Suppose that an originally plane bed is slightly perturbed so that the flow and sediment transport is disturbed. Then there will be the following two main possibilities:

1. The changes in flow pattern and sediment transport will tend to attenuate the amplitude of the perturbation, so that the bed goes back to original plane state. This means that the plane bed is stable.

2. The second possibility is that the flow causes the perturbation of the bed to increase in time, which corresponds to the unstable situation, ultimately leading to the formation of ripples, dunes and antidunes.

In the following, this will be discussed by putting this line of thinking in a simple mathematical form.

Figure 3 shows a longitudinal section through the channel. The sand bed has been given a small sinusoidal perturbation with the amplitude h_0 , so that the surface of the sand bed is described by

$$h = h_0 \exp[ik(x - ct)], \quad (1)$$

where k is the wave number ($=\pi/L$, where L is the wave length), x is the coordinate in the main flow direction, t is time and c is the migration velocity of the sand wave, while i is the imaginary unit $\sqrt{-1}$.

Equation (1) may alternatively be written

$$h = h_0[\cos k(x - ct) + i \sin k(x - ct)]$$

and only the real part of this expression is supposed to have physical significance.

As in most theories of hydrodynamic instability the idea is to assume the migration velocity c to be complex, i.e.

$$c = c_r + i c_i \quad (2)$$

so that the exponent in equation (1) becomes

$$\begin{aligned} ik(x - ct) &= ik[x - (c_r + i c_i)t] \\ &= ik(x - c_r t) + kc_i t. \end{aligned}$$

Then equation (1) becomes

$$h = h_0 e^{kc_i t} \exp[ik(x - c_r t)]. \quad (3)$$

From this it is realized that if c_i can be calculated, it may be used to determine whether the flow situation is stable or not. If c_i is positive, the factor

$$e^{kc_i t}$$

indicates that the bed waves grow exponentially in time. On the other hand, if c_i is negative the perturbation will attenuate and the bed becomes plane again.

We can proceed a little further by considering the well-known continuity equation for the sediment

$$\frac{\partial q_s}{\partial x} = - (1 - n) \frac{\partial h}{\partial t} \quad (4)$$

Here q_s is the rate of sediment transport and n the porosity of the sand bed. It is easily seen, that this equation can be satisfied with equation (1) and the following expression for the transport rate:

$$q_s = q_{s0} + \Delta q_s \exp[ik(x - ct - \delta)], \quad (5)$$

where δ is a phase lag between the local rate of sediment transport and the bed form. q_{s0} is the average (undisturbed) transport rate. With equations (4) and (5) the continuity equation is satisfied, provided

$$\Delta q_s e^{-ik\delta} = (1 - n)h_0 c$$

or

$$c = \frac{\Delta q_s}{(1 - n)h_0} (\cos k\delta - i \sin k\delta). \quad (6)$$

The imaginary part of c_i is found to be

$$c_i = \frac{\Delta q_s}{(1 - n)h_0} (-\sin k\delta). \quad (7)$$

3. THE FORMATION OF DUNES

Let us consider the subcritical flow situation ($\Delta q_s > 0$). This is defined by the criterion for the Froude number F :

$$F^2 < \frac{\tanh(kD)}{kD} \quad (8)$$

(For long waves, $kD \ll 1$, this becomes equal to $F < 1$). In this case the depth and the bed form are 180° out of phase (Figure 4). Consequently, the sediment transport rate is maximal near the crest and minimal at the trough, apart from the lag between the local rate of sediment and the bed form. Hence Δq_s is positive, so because of equation (7) the stability condition can be expressed in terms of the phase shift δ . If δ is negative the flow is unstable, but if $\delta > 0$, stability can be assumed.

The difficulties appear when δ has to be calculated from the laws of motion of (turbulent) fluid and sediment particles. The main conclusion that can be drawn from the available research is that δ depends on the following items, quoted in the order of decreasing importance:

- A) Fluid friction
- B) Rate of suspended sediment (versus bed load)
- C) Gravity forces on moving bed load
- D) Inertia of sediment particles
- E) Percolation in the river bed

The importance of these five items will be discussed in the following.

A) Fluid friction

The contribution from the fluid friction to δ is the most important at low shear stresses, where the sediment transport mainly occurs as bed load. Disregarding item (D) above, the bed load will respond immediately on changes in bed shear stress. In the subcritical area the flow upstream is converging, while the flow from crest to trough is diverging.

Now flow in converging channels gives larger fluid friction on the wall than uniform flows. Diverging flow, on the other hand, gives smaller fluid friction than uniform flow, in fact if the divergence angle is about 5° friction drops to zero and separation occurs. As a consequence of this, bed friction is not just in phase with velocity (and bed form), but there is a negative phase shift, or a negative contribution to δ .

CASE I
SUBCRITICAL FLOW

$$Fr^2 < \frac{\tanh(kD)}{kD}$$

CASE II
SUPERCritical FLOW

$$\frac{\tanh(kD)}{kD} < Fr^2 < \frac{\coth(kD)}{kD}$$

CASE III
SUPERCritical FLOW

$$Fr^2 > \frac{\coth(kD)}{kD}$$

Figure 4.
Variation in water surface over a wavy bed with Froude number.

The perturbed flow can be calculated from the vorticity transport equation

$$\frac{d\omega}{dt} = \epsilon \nabla^2 \omega$$

where ω is the vorticity, t the time and ϵ the eddy viscosity.

Taking the eddy viscosity as a constant it is possible to find that maximum bed shear occurs about 10^0 before maximum bed level.

Just by including this contribution to δ , maximum growth rate for the wavelength was found to occur at very small wave lengths. (We will return to this question in item C.)

B) Suspended sediment

At higher bed shear stresses the rate of suspended sediment increases, which contributes to δ with the opposite sign of that originating from the fluid friction. This lag can be found by calculating the non-uniform sediment distribution by a continuity equation for the sediment, expressing equilibrium between convection, settling and diffusion:

$$\frac{dc}{dt} = w \frac{\partial c}{\partial y} + \epsilon \nabla^2 c \quad (9)$$

where w = fall velocity, c = concentration of sediment, ϵ = eddy viscosity, t = time and y = vertical coordinate.

Because it takes some time for the suspended sediment to settle after being picked up from the bed, the result of the analysis yields that maximum transport of suspended sediment occurs downstream of the maximum bed shear stress, so in case of a subcritical flow, the suspended sediment gives a positive contribution to δ .

This mechanism can be used to explain a transition to plane bed in a subcritical flow: for increasing shear stresses, where the ratio between suspended load and bed load increases, the positive contribution to δ from the suspended sediment ultimately surpasses the negative contribution from friction and causes stability of the bed.

By this theory it is also possible to explain that a decrease in temperature may cause transition from dunes to plane bed: in case of a temperature drop, the ratio between suspended load increases due to the decrease in fall velocity, while the amount of bed load remains essentially unchanged. At sufficiently high bed shear stresses, this may cause the dunes to disappear as illustrated in Figure 5, which shows measurements of the Missouri River bed during the autumn of 1966. Here a temperature drop from about 25°C to about 5°C led to a stretching of the dunes until they finally disappeared. As

Figure 5.

The development of the bed in the Missouri River near Omaha during the autumn 1966.

Figure 6.

The variation with temperature in the upper limit to existence of dunes, the data being based on measurements taken in Missouri River near Omaha: $D = 3.2$ m and $d_m = 0.21$ mm.

shown in Figure 6, this is consistent with the prediction from the stability theory (Engelund and Fredsøe, 1974).

C) Gravity forces

The effect of a local bed slope emerging from the perturbation of the bed is quite significant for the result of the stability analysis of low bed shear stress. A phase between local bed shear stress and local bed load transport is introduced because it is more difficult to move the bed load uphill than downhill.

By incorporating the effect of gravity in a stability analysis, it was possible to obtain a maximum growth rate for wave lengths, which was much longer and more consistent with those observed in nature than those predicted earlier as shown in Figure 7, as derived by Fredsøe (1974).

D) Inertia of sediment particles

The contribution to δ from sediment inertia in a subcritical flow is positive, due to maximum bed load transport which occurs downstream of the maximum bed shear stress. The effect of inertia can easily be incorporated in a stability analysis, expressing that some of the bed shear stress is used to accelerate the transport of bed load

$$\rho_s \frac{dq_b}{dt} = \tau_{eff} - \tau_{tr}, \quad (10)$$

in which τ_{eff} is the effective bed shear stress, q_b the instantaneous bed load transport and τ_{tr} that bed shear stress, which in an equilibrium situation will be able to transport the amount q_b .

The effect is shown in Figure 8: because of the inertia, the most unstable perturbation tends to have a longer wave length. However, the contribution from inertia of a bed load is only of minor importance in most rivers, as the contribution from suspension dominates the contribution from inertia.

E) Percolation in the river bed

Because of the variation in pressure along a wavy bed, groundwater flow is introduced in the bed as sketched in Figure 9. Hereby, the mobility of the bed load varies slightly along a wavy bed and a contribution to δ arises. No stability analysis has been carried out to the author's knowledge, including this mechanism, but experiments indicate that this mechanism usually is unimportant concerning bed instability as this effect arises when very large flow velocities are combined with very coarse material.

Figure 7.
Effect of gravity on stability limits for dune development in open channel flow: in the upper figure, the effect of gravity is not included, as it is in the lower figure.

Figure 8.
Effect of inertia on stability limits for dune development in closed channel flow: on the left side inertia is not included, on the right side inertia is included.

4. THE FORMATION OF ANTIDUNES

If we return to Figure 4, it is seen that when the Froude number IF lies in the interval

$$\frac{\tanh(kD)}{kD} < F^2 < \frac{\coth(kD)}{kD}, \quad (11)$$

it is known from traditional potential theory that the water surface amplitude is larger than the bed wave amplitude, case II, Figure 4. Hence, sediment transport is minimal at the crest, which means that Δq_s is negative.

As the contribution to δ at these high flow velocities is dominated by the contribution that originates from the suspended load, δ here is normally positive. Now we obtain from equation (7), that c_1 is positive indicating instability.

Case III, Figure 4, comprises the largest values of F

$$F^2 > \frac{\coth(kD)}{kD}$$

In this case the flow is still supercritical, but the surface amplitude is smaller than the bed wave amplitude, so that again Δq_s is positive and the flow's stable.

From equation (8) it is seen that the migration velocity c_r of the bed waves is negative in case II, corresponding to what we call antidunes.

In Figure 10 is shown the areas corresponding to the three cases. The points correspond to all the antidune observations from Fort Collins (circles) and some data from J. F. Kennedy. The data confirm the instability of the area corresponding to case II and the stability of case III.

5. THE FORMATION OF RIPPLES

Until recently, stability analysis was connected to the formation of dunes, where flow depth is an important parameter, rather than to the formation of ripples. Kelvin Richards (1980), extended previous stability analysis by making a more advanced description of the flow over a wavy bed. Introducing the transport equation for turbulent kinetic energy, Kelvin Richards improved the flow description, compared to those made earlier, by including the memory effects in the turbulence. Now, a stability analysis yields that the bed at low bed shear stress is still unstable, and there is not only one maximum growth rate, but rather two local maxima, as seen from Figure 11.

Only one of these maxima depends strongly on the flow depth, while the other is essentially unaffected by changes in depth. The latter, Richards related to the instability of

Figure 9.
Phase shift introduced between bed level and flow velocity because of percolation in bed. Ho and Gelhar (1973).

Figure 10.
Stability limits for formation of antidunes.

Figure 11.
Variation in amplification factor A with wave number. Z_0 = roughness of bed. —: $D/Z_0 = 10^5$. ----: $D/Z_0 = 10^4$ after Richards (1980).

Figure 12.
The development of bed form with increasing flow velocity in closed channel flow.

ripples, while the former yields the instability of dunes. The instability limits for dunes are coherent with what has been found earlier in the works of Engelund and Fredsøe (1974), and also that of Fredsøe (1974).

Kelvin Richards found that the preferred initial length of ripples was proportional with the roughness of the bed.

The model mentioned above is carried out with flow over a hydraulic rough bed. No similar analysis has been carried out until now over a hydraulic smooth bed. Normally the existence of ripples are observed to disappear when the flow becomes hydraulic rough.

6. THE FORMATION OF BED WAVES IN CLOSED CONDUITS

In the case of a closed conduit, the flow will always converge before the top of the bed perturbation and diverge downstream of the top, so Δq_s is always positive. This involves that dunes occur as in open channel flow at low flow velocities but these dunes will disappear at higher bed shear stress where the contribution to δ from suspended sediment increases. At sufficiently high flow velocities, the bed will be flat as the contribution to δ from suspended sediment always acts to stabilize.

The result of a stability analysis is shown in Figure 8 (on the right) and the observed behavior is depicted in Figure 12, both figures are taken from Fredsøe and Engelund (1975).

REFERENCES

- Engelund, F. and Fredsøe, J. (1974) Transition From Dunes to Plane Bed in Alluvial Channels. Series Paper, No. 4, ISVA, Tech. Univ. Denmark.
- Fredsøe, J. (1974) On the Development of Dunes in Erodible Channels. J. Fluid Mech. 64.
- Fredsøe, J. and Engelund, F. (1975) Bed Configurations in Open and Closed Alluvial Channels. Series Paper, No. 8, ISVA, Tech. Univ. Denmark.
- Richards, K. J. (1980) The Formation of Ripples and Dunes on an Erodible Bed. J. Fluid Mech. 99.

SALTATION THRESHOLD MECHANICS

By

James D. Iversen

ABSTRACT

Observations of dust storms on Mars via telescope and spacecraft have led to new investigations and understanding of particle threshold phenomena. Wind tunnel testing with particles of different density and with atmospheric pressure down to the level of the Martian surface (500 Pa), has led to improved formulations for threshold prediction, including better understanding of the effects of particle cohesion. The effect of nonerodible roughness on threshold is discussed as well as the physics of threshold of particle motion due to an atmospheric vortex (dust devil).

INTRODUCTION

Observations of dust storms and their effects on the surface of Mars, particularly from NASA's Mariner 9 and Viking spacecraft, have led to speculation about the magnitude of winds necessary to initiate movement of surface particles in the low-density atmosphere of Mars. Subsequent experimental determination of particle threshold speeds for a variety of particle and atmospheric densities has led to a new understanding of threshold phenomena.

The well-known logarithmic velocity profile, valid within the atmospheric boundary layer close to the surface of flat homogeneous terrain under neutrally stratified conditions, is

$$\frac{u}{u_*} = \frac{1}{K} \ln \frac{z}{z_0}$$

where u is the wind speed magnitude, k is von Karman's constant (~ 0.4), z the distance above the surface, z_0 the equivalent roughness height, and u_* is the so-called friction speed, equal to the square root of surface shear stress divided by air density. Bagnold (1941) derived an empirical curve for the dimensionless threshold friction speed $A = u_{*t} (\rho/\rho_p g D_p)^{1/2}$ as a function of friction Reynolds number $B = u_{*t} D_p/\nu$ by determining the friction speed at threshold for sand particle groups of various average diameters.

Later investigators, in estimating threshold speeds on Mars, assumed that Bagnold's A-B curve was universal and extrapolated the results to Martian conditions. Among these predictions were those by Ryan (1964), Sagan and Pollack (1969), Arvidson (1972), Iversen et al. (1973), Hess (1973, 1975), Wood, Weaver and Henry (1974), Greeley et al. (1974), Maegley and Diederich (1975) and Maegley (1976). These predictions all resulted in rather high estimates of minimum thresholds for Mars (and therefore high surface wind speeds) with large values of optimum diameter.

EXPERIMENTS WITH PARTICLES OF DIFFERING DENSITY

In order to obtain the effect of the particle to fluid density ratio, a series of experiments was undertaken to determine threshold friction speed, using particles of differing density, from 210 kg/m^3 to $11,350 \text{ kg/m}^3$. The results from the Iowa State University wind tunnel are shown in Figures 1 and 2 (from Iversen et al., 1976). The effect of particle diameter and density is well illustrated in Figure 1. The optimum particle diameter (for minimum threshold speed) ranges from about 100 micrometers for a specific gravity of about 1, to approximately 50 micrometers for the heaviest particles. In Figure 2, the data are collapsed into approximately a single curve, by replotting in terms of threshold parameter A as a function of friction Reynolds number B . The data scatter for a smaller B is partly due to differences in particle density, since as shown later, for small particles, A is a function of diameter (interparticle forces of cohesion) as well as B (viscous aerodynamic forces). Also plotted in Figure 2 are the experimental curves of Bagnold (1941), Chepil (1945, 1959), and Zingg (1953). The two curves from Chepil's data are different because of differences in particle size distribution.

Figure 3 depicts an idealized erodible particle at rest on top of particles. The forces on the particle include drag and lift forces D and L , aerodynamic moment M , weight W , and interparticle force I_p . At threshold, the particle forces are assumed to be in equilibrium about the point P . The moment arms a , b and c are shown in the figure. Thus, the equation of

u_{*t} VS $\sqrt{D_p}$

Figure 1.

Threshold friction speed, u_{*t} versus D_p .

THRESHOLD FRICTION SPEED A VS B

Figure 2.
Threshold friction speed parameter, A versus B.

Figure 3.
Particle force geometry.

equilibrium, obtained by summation of moments about the point of contact P, is

$$Da + Lb + M = Wb + I_p c$$

By writing appropriate expressions for D, L, M, W, and I_p , it is possible to solve for a general expression for threshold friction speed u_{*t} . It is assumed that

$$D = K_D \rho u_{*t}^2 D_p^2$$

$$M = K_M \rho u_{*t}^2 D_p^3$$

$$L = K_L \rho u_{*t}^2 D_p^2$$

$$I_p = 6 K_p c D_p^n / b$$

The expression for threshold parameter A becomes, if K_D , K_M , K_L are all functions of B,

$$A = f(B) [1 + K_p / (\rho_p g D_p^{3-n})]^{1/2}.$$

Since the coefficients K_D , K_M , K_L , K_p , n and moment arm geometry are unknown, the function $f(B)$, coefficient K_p and exponent n must be determined empirically. Although the particle to fluid density ratio range in Figures 1 and 2 was not sufficient to establish accurate values of K_p or n, there is a noticeable change in A with change in diameter D_p with B held constant as shown in Figure 4.

It was discovered that a few data points existed for threshold friction speeds at very low atmospheric pressure (Weinberger and Adlon, 1971). These data are plotted in Figure 5 (Iversen et al., 1976b) along with one atmosphere data and empirical equations that take into account the effects of cohesion. The effect of ignoring cohesion effects in predicting the threshold on Mars is shown in Figure 6. The upper curve, labeled (1), is the prediction made assuming that Bagnold's A-B curve is universal. The lower curve (2) takes into account the interparticle force term of the preceding equation. A large discrepancy exists for particles less than about 500 micrometers in size.

Sagan (1975) and Sagan and Bagnold (1975) also realized that the existence of an optimum diameter (minimum threshold friction speed) might be due to the variation with diameter of particle cohesion forces instead of or in addition to viscous forces (Reynolds number), and made threshold predictions for

Figure 4.

Threshold friction speed parameter, A versus particle density, extrapolated to $B = 0.6$.

Figure 5.

Dimensionless threshold friction speed versus dimensionless particle diameter in air. Equation 1 is for particles with cohesion, 2 for cohesionless particles (Iversen et al., 1976b).

Figure 6.

Prediction of threshold friction speed on Mars: (1) assuming the A-B curve is universal, and (2) accounting for the effects of cohesion.

Figure 7.

Dimensionless threshold friction speed versus dimensionless particle diameter in liquid. Comparison of data with cohesionless equations for air (Equation (2) Iversen et al. (1976b) and Sagan and Bagnold (1975)).

Mars and Venus. These predictions by Sagan and Bagnold were made first by extrapolating a cohesionless threshold estimate for air on Earth from White's (1970) water data. Then they assumed that $A = A(B, \rho_p/\rho)$ and made their predictions for Mars and Venus by extrapolating to the Martian and Venusian values of density ratio ρ_p/ρ , Figure 7. Unfortunately, since the forces of cohesion are not linearly proportional to fluid density (being primarily a function of particle diameter), the threshold coefficient A cannot be written as an explicit function of the density ratio ρ_p/ρ . Miller and Komar (1977) also made threshold speed estimates based on a cohesionless particle with no minimum diameter. Low air density data (shown below) clearly show the existence of the optimum diameter however, and it is thus expected that cohesion forces and the resulting minimum threshold also exist on Mars.

THE LOW-AIR-DENSITY WIND TUNNEL

Besides having only three data points of somewhat questionable accuracy for threshold friction speed for large values of density ratio ρ_p/ρ , there was also a large gap in the data available ($10^4 \lesssim \rho_p/\rho \lesssim 10^6$). In order to close that gap, a boundary layer wind tunnel was built and placed in the altitude chamber at NASA Ames Research Center, Moffett Field, California. Threshold friction speeds for values of ρ_p/ρ from 900 to 630,000 were obtained using materials with densities of 1,100 and 2,650 kg/m³ and particle diameters of $37(10)^{-6}$ to $673(10)^{-6}$ m. Both air and carbon dioxide were used as gaseous fluid media.

The data were taken by placing a sample of material on the wind tunnel floor, lowering the chamber pressure to Martian values (300 to 500 Pa), determining the threshold speed optically, and then increasing the chamber pressure one step at a time and repeating the process. A description of the wind tunnel facilities, experimental procedures, and the results of the experiments are reported in Greeley et al. (1976, 1980). The results of a typical series of experiments are shown in Figure 8. This curve illustrates the isolated effect of the Reynolds number, as the data are for a constant particle diameter. The effect of particle diameter is shown in Figure 9. These data were obtained by extrapolating threshold data at constant values of Reynolds number B, and so each curve in Figure 9 illustrates the isolated effect of particle diameter. The A versus B curve for cohesionless particles is illustrated in Figure 10. Here the threshold parameter A^2 is divided by the interparticle force term to eliminate that effect. The upward swing for small values of B in the curves of Figure 2

Figure 8.

Threshold parameter A as a function of particle Reynolds number B. These curves are for constant values of particle diameter and density, but with differing values of fluid density. Data for 154-micrometer sand particles with air as gas.

Figure 9.

Threshold parameter A^2 as a function of particle diameter for constant values of Reynolds number B.

Figure 10.

Threshold parameter A^2 normalized by particle cohesion function versus friction Reynolds number B .

Figure 11.

Threshold friction speed comparisons for Earth, Mars, and Venus. Predictions made using equations in the text.

is eliminated. The equations resulting from empirical curve fits of the data in Figures 8, 9, and 10 are

$$A = 0.129 [(1 + 0.006/\rho_p g D_p^{2.5})^{0.5} / (1.928 B^{0.092} - 1)^{0.5}]$$

for $0.03 \leq B \leq 10$

and

$$A = 0.120 \left\{ [(1 + 0.006/\rho_p g D_p^{2.5})^{0.5} / [1 - 0.0858 e^{-0.0617(B-10)}]] \right\}$$

for $B \geq 10$.

Predictions of the threshold for Earth, Mars, and Venus resulting from these equations are shown in Figure 11.

EFFECT OF NONERODIBLE ROUGHNESS

It has frequently been observed (e.g., Chepil and Woodruff, 1963) that the presence of large nonerodible elements distributed through a field of loose soil or sand has the effect of increasing threshold friction speed. A comparison of rough surface data for glass spheres with data from Figure 2 is shown in Figure 12. The threshold friction speed with nonerodible roughness is considerably higher than for the erodible surface only. Curves plotted in terms of free-stream wind tunnel speed or in terms of geostrophic wind in the atmosphere would be somewhat closer together. Predictions of geostrophic wind speeds on Mars based on threshold equations of Iversen et al. (1976) were made by Pollack et al. (1976).

DUST DEVIL THRESHOLD

In order to assess the relative effectiveness of swirling motion winds as a means of achieving particle threshold, a series of exploratory experiments was conducted at one atmosphere pressure to determine the strength and size of vortices needed to lift erodible particles from the surface (Greeley et al., 1977). Research involving a vortex generator (Hsu and Fattahi, 1975) has been conducted at the Iowa State University Aerospace Engineering Department for the past several years. The vortex, formed with its axis perpendicular to the ground or surface plane, is produced by passing air through a rotating honeycomb and stationary nozzle located some distance above the surface. A vortex can be formed by the passage of air either upwards or downwards through the honeycomb, or with no air passage at all. The vortex formed is an

THRESHOLD FRICTION SPEED~ROUGH AND SMOOTH SURFACES

Figure 12.

Threshold friction speed parameter for glass spheres, rough and smooth surfaces.

Figure 13.

Characteristic radius of the simulated dust devil vortex as a function of rotor speed.

approximate small-scale model of atmospheric vortices such as the tornado or dust devil.

The swirling motion due to the atmospheric vortex (dust devil) causes a radial pressure gradient at the surface

$$\frac{dp}{dr} = \frac{\rho u_{\theta}^2}{r} \sim \rho r \omega_o^2$$

or

$$\rho \frac{u_{\theta}^2}{r} \Delta r \approx \Delta p = \rho r_o^2 \omega_o^2$$

or

$$r_o = \left(\frac{\Delta p}{\rho \omega_o^2} \right)^{1/2}$$

where p is pressure, r is radial distance from the vortex center, ρ is air density, u_{θ} is tangential (swirl) speed, ω_o is the angular speed of the generator honeycomb, and the characteristic radius r_o is as defined in equations (4), where Δp is the maximum surface pressure difference from the center of the vortex to that at a large radius.

The maximum pressure difference, Δp , was measured on the surface plane of the vortex generator with a differential pressure transducer for the range of angular speeds ω_o of the generator. The values of characteristic radius r_o calculated from the measured values of Δp are shown in Figure 13. A linear fit of the data results in an equation for r_o

$$r_o = 0.0092 \omega_o + 6.87$$

where r_o is in cm and ω_o in rad/sec.

The mechanism for entrainment of particles from the surface by a dust devil is quite different than for a uniform wind. A primary cause for liftoff is the pressure difference, created by the swirling flow, between the minimum pressure point within the vortex and that existing at the surface before the dust devil came along. That pressure difference does not exist for the uniform flow. Assume that the top layer

of particles of thickness equal to particle diameter D_p is lifted by the vortex at threshold,

Δp represents the instantaneous pressure difference between the lower and upper surfaces of the particle layer, caused by the vortex moving over the part of the layer where particles are observed to lift off the surface. The factor α is included to account for pressure relief that might occur prior to particle lift-off; σ_I is the stress due to interparticle force, τ is an upward normal stress due to the effects of viscosity, $\rho_p g D_p$ is the weight of the particle layer per unit area, and α is an unknown constant of proportionality. If the upper layer were completely sealed α would be 1; if there were leakage α would be less than 1. If the surface stress τ is defined as $\tau = C \rho r_o^2 \omega_o^2$, then the value $r_o \omega_o$ at threshold where forces are in equilibrium can be written as

$$r_o \omega_o = \frac{1}{(\alpha + C\tau)^{1/2}} \left(\frac{\rho_p g D_p}{\rho} \right)^{1/2} \left(1 + \frac{\sigma_I}{\rho_p g D_p} \right)^{1/2}$$

The characteristic threshold speed $r_o \omega_o$ is not a true speed but is proportional to the square root of the pressure difference Δp (i.e., $r_o \omega_o = (\Delta p / \rho)^{1/2}$).

Unfortunately neither the function $\sigma_I(D_p)$ nor the coefficient $C\tau$ nor the ratio α are known, and sufficient data to determine these functions empirically are not yet available. The values of $r_o \omega_o$ at threshold are shown as a function of particle diameter in Figure 14 and as a function of $(\rho_p g D_p / \rho)^{1/2}$ in Figure 15. The particles tested are the same as those in Figures 1 and 2. Figure 15 does show a trend for increasing angular threshold speed for larger values of this function. However, since the effects of Reynolds number $r_o^2 \omega_o / \nu$ and interparticle force are not accounted for, there is a large amount of data scatter for most of the range of the data.

The entrainment mechanism for raising surface particles with a vertical vortex (dust devil) is obviously quite different from the uniform wind boundary-layer case. The variation with particle diameter is much less pronounced for vortex threshold than it is for boundary-layer threshold. The

Figure 14.

Characteristic vortex threshold speed $r_0\omega_0$ as a function of particle diameter.

Figure 15.

Characteristic vortex threshold speed $r_0\omega_0 = (\Delta p/\rho)^{1/2}$ as a function of particle parameter. Note that for large particles the pressure difference Δp is nearly equal to the weight per unit area of a single layer of particles. For small particles the pressure difference Δp ranges up to nearly ten times the weight per unit area of a single layer. One explanation for the difference is small particle cohesion (interparticle stress σ_I).

important points to note from these exploratory experiments are that size and density seem to have little importance in the swirling motion motion threshold, except for the two extreme cases of large lead and bronze particles. Thus, cyclonic motion winds would appear to be very effective in raising erodible particles of all types and sizes. Once lifted above the surface, much lower winds than 'normal' threshold could continue to carry particles along.

REFERENCES

- Arvidson, R. E. (1972) Aeolian Processes on Mars: Erosive Velocities, Settling Velocities, and Yellow Clouds, *Geol. Soc. Am. Bull.* 83: p. 1503-1558.
- Bagnold, R. A. (1941) *The Physics of Blown Sand and Desert Dunes*. Methuen, London.
- Chepil, W. S. (1945) Dynamics of Wind Erosion, II. Initiation of Soil Movement. *Soil Sci.* 60, p. 397-411.
- Chepil, W. S. (1959) Equilibrium of Soil Grains at the Threshold of Movement by Wind, *Soil Science Society Proceedings* 23: p. 422-428.
- Chepil, W. S. and Woodruff, N. P. (1963) The Physics of Wind Erosion and Its Control. *Advances in Agronomy* 15: 211-302.
- Greeley, R., Iversen, J. D., Pollack, J. B., Udovich, N., and White, B. R. (1974) Wind Tunnel Studies of Martian Aeolian Processes. *Proc. R. Soc. London A341*: p. 331-336.
- Greeley, R., White, B. R., Leach, R., Iversen, J. D., Pollack, J. B. (1976) Mars: Wind Friction Speeds for Particle Movement. *Geophys. Res. Lett.* 3: p. 417-420.
- Greeley, R., Iversen, J. D., White, B. R., Pollack, J. B., and Leach, R. N. (1977) Dust Storms on Mars: Considerations and Simulations. *AAS Desert Dust Symposium*, American Association for the Advancement of Science, annual meeting, Denver, Colorado. NASA TM 78423.
- Greeley, R., Leach, R., White, B. R., Iversen, J. D., and Pollack, J. B. (1980) Threshold Windspeeds for Sand on Mars: Wind Tunnel Simulations. *Geophys. Res. Lett.*, v. 7, p. 121-124.
- Hess, S. L. (1973) Martian Wind and Dust Clouds. *Planet. Space Sci.*, v. 21, p. 1549-1557.
- Hess, S. L. (1975) Dust on Venus. *J. Atmos. Sci.*, v. 32, p. 1076-1087.
- Hsu, C. T. and Fattahi, B. (1975) Tornado Tunnel Formation From a Tornado Cyclone. *Proc. 9th Conf. on Severe Local Storms*, American Meteorological Society, Norman, Oklahoma, October.
- Iversen, J. D., Greeley, R., Pollack, J. B., and White, B. R. (1973) Simulation of Martian Aeolian Phenomena in the

- Atmospheric Wind Tunnel. Proc. 7th Conference Space Simulation. NASA SP-336, p. 191-213 (NTIS Accession No. N74-10232).
- Iversen, J. D., Pollack, J. B., Greeley, R., and White, B. R. (1976a) Saltation Threshold on Mars: The Effect of Interparticle Force Surface Roughness, and Low Atmospheric Density. *Icarus*, v. 29, p. 318-393.
- Iversen, J. D., Greeley, R., and Pollack, J. B. (1976b) Windblown Dust on Earth, Mars and Venus. *J. Atmos. Sci.*, v. 33, p. 2425-2429.
- Maegley, W. J. (1976) Saltation and Martian Sandstorms. *Rev. Geophys. Space Phys.*, v. 14, p. 135-142.
- Maegley, W. J. and Diederich, D. P. (1975) Martian Sandstorms and Effects on the 1975 Viking Lander System. *J. Test. Eval.*, v. 3, p. 380-388.
- Miller, M. D. and Komar, P. D. (1977) The Development of Sediment Threshold Curves for Unusual Environments (Mars) and for Inadequately Studied Materials (Foram Sands). *Sedimentology*, v. 24, p. 709-721.
- Pollack, J. B., Haberle, R., Iversen, J. D., and Greeley, R. (1976) Estimates of the Minimum Wind Needed to Raise Dust on Mars. *Icarus*, v. 29, p. 395-417.
- Ryan, J. A. (1964) Notes on the Martian Yellow Clouds. *J. Geophys. Res.*, v. 69, p. 3759-3770.
- Sagan, C. (1975) Windblown Dust on Venus. *J. Atmos. Sci.*, v. 32, p. 1079-1083.
- Sagan, C. and Bagnold, R. A. (1975) Fluid Transport on Earth and Aeolian Transport on Mars. *Icarus*, v. 26, p. 209-218.
- Sagan, C. and Pollack, J. B. (1969) Windblown Dust on Mars. *Nature*, v. 223, p. 791-794.
- Weinberger, R. K. and Adlon, G. L. (1971) Particle Development and Entrainment by a Low Density Airstream Flowing Over a Surface. NASA CR 111924 (NTIS Accession No. N71-32551).
- White, S. J. (1970) Plane Bed Thresholds of Fine Grained Solids. *Nature*, v. 22, p. 152-153.
- Wood, C. P., Weaver, W. R., and Henry, R. M. (1974) The Minimum Free-Stream Wind for Initiating Motion of Surface Material on Mars. NASA TMX-71959, Langley Research Center (NTIS Accession No. N74-23184).
- Zingg, A. W. (1953) Wind Tunnel Studies of the Movement of Sedimentary Material. Proceedings of the Fifth Hydraulic Conference, Bulletin 34, University of Iowa Studies in Engineering.

DUST PRODUCTION BY WIND EROSION:
NECESSARY CONDITIONS AND ESTIMATES OF VERTICAL
FLUXES OF DUST AND VISIBILITY REDUCTION BY DUST

By

Dale A. Gillette

ABSTRACT

Differing meteorological conditions that supply sufficient momentum to the surface to sustain wind erosion are organized with respect to the ratio of boundary layer height to Monin-Obukof length, L . Conditions for large-scale dust storms include strong surface heating and a strong upper level momentum source.

Visibility is related to mass concentration by an equation $MV = C$, where visibility, C , is a constant and M is the mass concentration of dust in the air. Different studies of visibility versus dust concentration are related to distance from the source region.

A. Meteorological Conditions for Dust Production

For the production of dust we must have:

1. The appropriate meteorological conditions to supply momentum to the surface.
2. The appropriate soil conditions to allow dust production.

For the transport of momentum to the surface, we must have turbulent mixing. In almost every case, this mixing is con-

Figure 1
Height versus stability diagram of the boundary layer, indicating various observable phenomena. The height scale is logarithmic to enable the dimensions of the roughness elements near the surface to be shown. However, detail in the inversion layer is lost. (After Businger, 1975).

Figure 2
Scheme of the typical 500 mb situation favorable to straight transport of dust. (After Prodi and Fea, 1978).

nected with a heat flux so that strong heating of the ground is required. Figure 1 shows a height versus stability diagram from Businger (1975), which shows the state of the atmosphere described by a small number of parameters. In the case of a heat flux from the ground to the atmosphere in which we are interested, the height of the portion of the atmosphere into which dust may mix (boundary layer height) is determined by the strength of heat flux, the history of the heat flux, and the thermal structure of the atmosphere above the boundary layer. This height is denoted Z_1 in Figure 1. The parameter L is defined by Lumley and Panofsky (1964) as

$$L = \tau \frac{U_*^3 C_p \rho T}{KgH}$$

Where H is heat flux (ergs/cm²/sec)

U_* is 'friction velocity'

$$= \sqrt{\tau/\rho}$$

Where τ is surface stress (dynes/cm²)

ρ is density of the air (grams/cm³)

g is gravitational acceleration

C_p is specific heat capacity of air at constant pressure

T is temperature

K is Von Karman's constant (above 0.4).

For the sake of illustrating meteorological conditions of dust production, we may specify a constant temperature and then write

$$\frac{Z_1}{L} = - \frac{Z_1}{3} \frac{H}{U_*^3} \times \text{constant.}$$

Large values of the ratio occur for very large heat transfers from the ground, large boundary layer heights, and for small values of friction velocity. From Figure 1 we notice that these are accompanying conditions for 'dust devils' in which wind speed at the ground is locally great, due to large upward velocities due to heating of air in an otherwise fairly low wind speed environment. This production of dust is rather localized, and since there is little wind, dust is not carried far from its point of origin although it may be mixed into a deep layer of atmosphere.

A larger scale event is a 'haboob' or dust containing downwash from a convective storm. This event is larger in scale compared to dust devils, but is still a rather limited

event insofar as long-range transport of dust is concerned. Dust devils and haboobs are quite important disturbances of soils; however, and do locally produce great quantities of dust.

When strong heating of the ground and near-surface air is added to a source of upper-level strong winds, a situation is developed for large-scale dust production and large-scale transport. In this case, Z_1/L begins with large values due to heating, but when the thermal structure of the atmosphere becomes well mixed due to this heating, high momentum is brought to the surface and the value of Z_1/L approaches 1 due to the increased wind speed brought to the surface. This high wind condition is applied to a very large area, which creates intense large-scale wind erosion, mixing of dust to large heights of the atmosphere, and large-scale transport potential.

Other conditions leading to large-scale dust storms besides strong surface heating and an upper level momentum source are an increasing horizontal pressure gradient and large-scale convergence that produces ascending trajectories. These conditions are most probable in the spring when solar heating produces hot surface temperatures in arid regions and when, upstream from these regions, vigorous cyclonic storm growth generates strong upper-level jet streams. Major dust storms are produced when the upper-level jet and the strong surface heating interact (Jackson et al., 1973).

Examples of such interaction are given by Prodi and Fea (1978) for transport of Saharan dust to Europe and by Jackson et al. (1973) for transport of Southwestern U.S. dust to the Atlantic Ocean. Figure 2 shows a scheme of the typical 500 mb situation favorable to straight transport of dust after Prodi and Fea (1978) and Figure 3 shows three-dimensional trajectories which supply momentum for large-scale dust storms after Jackson et al. (1973). Figure 4 shows a cross-sectional analysis of wind speed and potential temperature (temperature of the air when at 1,000 mb pressure). The thermal mixing causing high wind speed to be brought to the surface is clearly shown in this figure (after Jackson et al., 1973).

B. The Relation of Visibility to Dust Concentration (Condensed from Patterson and Gillette, 1977).

Visibility is defined as the greatest distance in a given direction at which a prominent dark object can be seen against the horizon with the unaided eye; the prevailing visibility is defined as the greatest horizontal visibility that is equalled or surpassed through half (not necessarily a continuous half) of the horizon circle. The concept of visibility is made more quantitative by defining a visual range V as the distance from an observer at which the contrast ratio between a black object

Figure 3
Three-dimensional trajectories that supply momentum for large-scale dust storms. (After Jackson et al., 1973).

Figure 4
Vertical cross section for midnight (Greenwich time) April 22, 1963. Solid lines are $\theta^{\circ}\text{K}$ at 4° intervals; dashed lines are isotachs (knots); J = jets; crosshatching = clouds; radioactivity, dpm/SCF = disintegration per minute per standard cubic foot of air. (After Jackson, et al., 1973).

and a bright background is equal to 0.02. This contrast ratio of 0.02 is taken as the minimum contrast ratio that can be distinguished by the unaided human eye, and this defined visual range will be taken as the measure of the visibility.

The visual range is related to the atmospheric extinction coefficient by

$$V = \left| \frac{\ln 0.02}{\sigma} \right| \quad (1a)$$

which becomes

$$V = \frac{3.912}{\sigma} \quad (1b)$$

where σ is composed of two parts -- done due to extinction by the air molecules, σ_A and the other due to extinction by atmospheric particles, σ_p ; but for visibilities of less than a few kilometers, $\sigma_p \gg \sigma_A$ and the extinction is completely determined by σ_p so $\sigma = \sigma_p$. In general, it is assumed that the distribution of atmospheric particles is uniform, leading to uniform σ_p . Under this assumption, measurement or calculation of σ_p at a point should lead to a calculated point visibility that is equal to the observed visual range.

The extinction coefficient may be calculated from Mie theory, which describes scattering by spherical particles. Although our particles are not perfect spheres, electron microscope pictures indicate that they are compact, amorphous solids and the usual assumption is that Mie theory approximately describes the scattering. Assuming that the aerosol particles are uniform homogeneous spheres, we can use the formalism of Mie scattering. The particle extinction cross section is commonly written in terms of the cross-sectional area πr^2 of the particle times an efficiency factor for extinction, Q_{ext} .

The total extinction due to a polydisperse distribution of particles is given by the integral over the size distribution

$$\sigma_p = \int_{r_1}^{r_2} \pi r^2 Q_{extn}(r) dr, \quad (2)$$

where $n(r)$ is a continuous function defined within the radius interval from r_1 to r_2 representing the concentration of particles per unit volume and per unit increment of dr .

The extinction efficiency Q_{ext} is shown in Figure 5.

Figure 5

Extinction efficiency Q_{ext} vs Mie size parameter $2r/\lambda$ where r is the radius of the particle and λ is the wavelength of the light.

C. Visibility and Mass Concentration Relations:

The ratio of the mass concentration, M , to the extinction due to particles, σ_p , may be expressed by means of equation (3)

$$\frac{M}{\sigma_p} = \frac{d_p \int_{r_1}^{r_2} \frac{4}{3} \pi r^3 n(r) dr}{\int_{r_1}^{r_2} \pi r^2 Q_{\text{extn}}(r) dr} \quad (3)$$

$$= \frac{4 d_p \int_{r_2}^{r_1} r^3 n(r) dr}{3 \int_{r_1}^{r_2} r^2 Q_{\text{extn}}(r) dr}$$

with d_p the particle density and Q_{ext} the Mie extinction efficiency.

The ratio M/σ_p is independent of the total number of particles and depends only on the number-fraction frequency distribution, (n) ; the ratio is designated by K :

$$\frac{M}{\sigma_p} = K \quad (4)$$

By equating σ_p to the total extinction σ (σ includes gaseous as well as particulate optical effects), which is valid for conditions of low visibility, and substituting this relation into equation (2) we obtain

$$V = \frac{3.912K}{M} = \frac{C}{M} \quad (5)$$

where V = visibility and $C = 3.912 K$, which leads to the relation

$$MV = C \quad (6)$$

Different relative particle size distributions will result in different values for K and consequently for C ; if the particle size distribution changes in a consistent way with

increasing or decreasing visibility, an empirical fit to the data relating mass concentration and visibility may result in an equation of the form used by Bertrand et al. (1974) and by Chepil and Woodruff (1957) to describe their measurements:

$$MV^{\gamma} = C. \quad (7)$$

A summary of values from several visibility studies fitted to equation 7 is given in Table 1.

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

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

REFERENCES

- Bertrand, J., Baudet, J. and Drochon, A. (1974) Importance des aerosols naturels en Afrique de L'ouest. *J. Recherches Atmos.*, 8, p. 845-860.
- Businger, J. (1975) The Need for Boundary-Layer Measurements and Their Significance for Solving Atmospheric Problems.
- Charlson, R.J. (1969) Atmospheric Visibility Related to Aerosol Mass Concentration. *Environ. Sci. Technol.*, 3, p. 913-918.
- Chepil, W.S. and Woodruff, N.P. (1957) Sedimentary Characteristics of Dust Storms -- II. Visibility and Dust Concentration. *Am. J. Sci.* 255, p.104-114.
- Jackson, M.L., Gillette, E.F., Danielson, E.F., Blifford, I.H., Bryson, R.A., and Syers, J.F. (1973) Global Dustfall During the Quaternary as Related to Environments. *Soil Science*, 116, p. 135-145.

- Lumley, J. and Panofsky, H. (1964) The Structure of Atmospheric Turbulence. Wiley Interscience, New York, p. 102.
- Pilat, M.J. and Ensor, D.S. (1970) Plume Opacity and Particulate Mass Concentration. Atmospheric Environment, 4, p. 163-173.
- Patterson and Gillette (1977) Measurements of Visibility VS Mass-Concentration for Airborne Soil Particles. Atmos. Env., 11, p. 193-196.
- Prodi, F. and Fea, G. (1978) Transport and Deposition of Saharan Dust Over Alps, 15. Internationale Tagung fuer Alpine Meteorologie, Grinde l wald 19-23, Sept. 1978.

Table 1

Study	C_{mean} (g m ⁻³ km)	C_{fitted} (g m ⁻³ km)	γ
General Urban (Charlson, 1969)	---	1.8×10^{-3}	1
Bertrand et al. (1974)	---	1.4×10^{-3}	1.05
NCAR (Total mass)	2.3×10^{-2}	2.0×10^{-2}	1.07
(r < 20 μm)	1.7×10^{-2}	1.3×10^{-2}	1.08
(r < 10 μm)	1.3×10^{-2}	1.1×10^{-2}	0.95
Chepil and Woodruff			
All data	---	5.6×10^{-2}	1.25
No local erosion	2.0×10^{-2}	---	---

THE AEOLIAN WIND TUNNEL - SALTATION SIMILITUDE

By

J. D. Iversen

ABSTRACT

The similitude requirements for modeling saltation phenomena at small scale are exceedingly complex. This paper reviews the basic dimensionless parameters and how theoretical means, namely the effects of aerodynamic roughness in saltation, mass transport rate, and the particle equations of motion are used to group the basic parameters and thus somewhat simplify the similitude problem. Two examples illustrate the use of mass transport rate and roughness parameters to correlate model data.

INTRODUCTION

The deposition and erosion of fine particles blowing over a complex terrain presents a very intricate phenomenon to attempt to physically simulate on a smaller geometric scale. The erodibility of soil or sand is a complicated function of mean wind speed, frequency and intensity of gusts, and particle size distribution, density and shape, and topographic surface roughness. An optimum size occurs, for soil, sand, or snow on earth, for which the threshold speed is a minimum. On Mars, the optimum particle size and threshold speeds are predicted to be larger, mostly because of lower atmospheric density.

The list of important physical parameters in the aeolian process includes:

D_p	Particle diameter
L	Major terrain feature length (such as crater diameter)
ℓ	All other terrain feature lengths
z_0	Roughness length
z_0'	Roughness length in saltation
h	specified altitude for reference speed
L^*	Monin-Obhukov length (atmospheric stability parameter)
λ	Ripple wavelength
$U(h)$	reference wind velocity
U_F	particle terminal speed
u_*	friction speed
u_{*t}	Threshold friction speed
ρ	Atmospheric density
ρ_p	Particle density
ν	Kinematic viscosity
t	Time (duration of phenomenon)
e	Coefficient of restitution

These parameters can be arranged in the form of dimensionless similitude parameters.

1. $\frac{D_p}{L}$ Length ratio
2. $U(h)/U_F$ Reference to particle terminal speed ratio
3. $\left\{ U(h) \right\}^2 / gL$ Froude number

4. e	Coefficient of restitution
5. l/L	Topographic geometric similarity
6. z_0/L	Roughness similitude
7. z_0/L	Roughness similitude in saltation
8. h/L	Reference height similitude
9. z_0/L^*	Stability parameter
10. λ/L	Ripple length similitude
11. U_F/U_{*t}	Particle property similitude
12. $U_{*t} D_p/\nu$	Particle friction Reynolds number
13. u_*/u_{*t}	Friction speed ratio
14. ρ/ρ_p	Density ratio
15. $U(h)t/L$	Time scale

It is impossible to satisfy all of these parameters in the aeolian wind tunnel, in particular, numbers 1, 3, and perhaps 7 and 10 are extremely difficult unless the length scale ratio is close to unity. The relative importance of these parameters is discussed by Gerdel and Strom (1961), Odar (1962), Isyumov (1971), Iversen et al. (1973), Greeley et al. (1974), and Iversen (1980a).

A series of experiments with geometrically similar model craters (10, 20, and 30 cm diameter) was undertaken (Iversen et al., 1975) to attempt to establish similitude prediction factors, necessary when a small-scale model is distorted (i.e., the preceding dimensionless parameters are not the same for the model as for the full scale). The results for a variety of particle densities and diameters and wind speeds are shown in Figure 1. This figure displays the amount of area in the crater wake from which material has been eroded as a function of time. Obviously, a prediction of full-scale phenomena cannot be made from this figure alone.

Particle Equations of Motion

In order to obtain prediction factors, it is necessary to use additional means other than simple dimensional analysis so that the preceding dimensionless terms can be grouped. Particle equations of motion and the theoretical expression

for mass transport rate are used to do this. The particle equations of motion, in dimensionless form can be written

$$\frac{d^2X}{dT^2} = \frac{3}{4} C_D \frac{\rho L}{\rho_p D_p} \left\{ \frac{1}{0.4} \ln \frac{hZ}{z'_0} + \frac{u'}{u_*} - \frac{dX}{dT} \right\}$$

$$U - \frac{dX}{dT}$$

$$\frac{d^2Z}{dT^2} = \frac{3}{4} C_D \frac{\rho L}{\rho_p D_p} \left\{ \frac{1}{0.4} \ln \frac{hZ}{z'_0} + \frac{u'}{u_*} - \frac{dX}{dT} \right\}$$

$$\left\{ W - \frac{dZ}{dT} \right\} - \frac{gL^2}{u_*^2}$$

where

$$U - \frac{dX}{dT} = \frac{1}{0.4} \ln \frac{hZ}{z'_0} + \frac{u'}{u_*} - \frac{dX}{dT}$$

$$W - \frac{dZ}{dT} = W_0 - 4W_0 X^2 - \frac{dZ}{dT}$$

$$X = x/L, Z = z/h, T = U_* t/L$$

Thus, to ensure dynamic similarity, the parameters

$$\frac{z'_0}{h}, C_D \frac{\rho L}{\rho_p D_p}, \text{ and } \frac{gL^2}{u_*^2 h}$$

should be the same in model and full scale.

The equivalent roughness height in saltation, z'_0 , has not generally been determined. However, Owen (1964) shows that on earth it is at least approximately proportional to u_*^2/g , i.e.,

$$z'_0 \sim \frac{u_*^2}{2g}$$

Recent calculations of particle trajectories by White et al. (1976) for various particle densities and for both terrestrial and Martian atmospheres show that the maximum height in saltation, to which z_0 is probably proportional, involves the ratio of atmosphere to particle density. Thus a more likely valid expression is

$$z'_0 \sim \frac{\rho u_*^2}{\rho_p g}$$

Figure 1.
Model crater wake area eroded as a function of time. $2.42 \leq v_f/u_{*t} \leq 14.2$, $950 \leq \rho_p/\rho \leq 2,270$, $5.2 \leq v^2/gD_c \leq 56$.

Figure 2.
Erosion rate similitude based on mass transport rate.

The roughness parameter thus becomes

$$\frac{z_o'}{h} \sim \frac{\rho u_*^2}{\rho_p g h} = \frac{\rho}{\rho_p g D_p} \left(\frac{u_*}{u_{*t}} \right)^2 u_{*t}^2 \frac{D_p}{h}$$

But $u_{*t}^2 = A_1^2 \rho_p g D_p / \rho$, so

$$\frac{z_o'}{h} \sim \frac{\rho u_*^2}{\rho_p g h} = A_1^2 \frac{D_p}{h} \left(\frac{u_*}{u_{*t}} \right)^2$$

Since the preceding combinations of parameters have been obtained from particle trajectory equations, their satisfaction in modeling is probably necessary for small-scale aeolian bed form features, such as ripples, to be properly modeled. Experimental correlation of gross erosional and depositional features has been obtained however (Iversen et al., 1975), by basing a similitude on mass transport rate rather than on particle trajectory parameters.

Derivation of Erosion Rate Parameter

Bagnold (1941) derived the expression for momentum loss of the air due to sand in saltation as

$$\frac{q_s(u_2 - u_1)}{l} \sim \frac{q_s u_2}{l} = \tau = \rho u_*^2.$$

This quantity represents the momentum loss per unit time per unit length of travel per unit lateral dimension, i.e., momentum loss per unit area per unit time, which is equal to force per unit area or surface stress τ . Mass of sand per unit lateral dimension per unit time is q_s , u_2 is final horizontal velocity on impact, u_1 initial horizontal velocity, assumed small, l is distance traveled per grain, τ is surface stress, ρ the air density, and u_* the surface friction speed in saltation. The value of u_* some distance upstream of the crater may be more or less than the threshold value, since local friction speed is higher in some areas in the vicinity of the crater. The mass transport q_s here refers to the average mass flow rate in the eroded portion of the crater wake.

From ten particle trajectory calculations for a wide variety of conditions, Bagnold's assumption that

$$u_2/l \sim g/w_1$$

was found to hold very well, where g is gravitational acceleration and w_1 is initial vertical velocity. Thus

$$q_s \sim \rho u_*^2 w_1 / g.$$

Now, the mass flow q_s has the dimensions of mass per unit lateral dimension per unit time. Thus, the mass flow involved in the formation of a crater wake is proportional to $q_s D_c$, where D_c is the crater rim diameter, the lateral length characteristic of wake width. In the wind tunnel modeling experiments, the cleared wake area A was recorded as a function of time,

$$\frac{dA}{dt} = \frac{\text{area}}{\text{time}} = \frac{\text{area mass}}{\text{mass time}}.$$

The mass per unit area of material cleared is given by the product of particle density ρ_p times the initial material depth h . Thus

$$dA/dt = q_s D_c / \rho_p h \sim \rho u_*^2 w_1 D_c / \rho_p g h$$

or

$$d(A/A_c)/dt \sim \rho u_*^2 w_1 / \rho_p g h D_c$$

or

$$A/A_c \sim \rho u_*^2 w_1 t / \rho_p g h D_c,$$

if the derivative is assumed to be constant. In each wind tunnel test, the ratio h/D_c was set at a constant 1/30. Thus

$$A/A_c \sim \rho u_*^2 w_1 t / \rho_p g D_c^2.$$

Bagnold assumed that the initial vertical velocity w_1 is proportional to friction speed u_* . However, for the values of u_* in the neighborhood of threshold friction speed u_{*t} , it should also be a function of the difference $u_* - u_{*t}$. In the wind tunnel, with no model present, the friction speed difference $u_* - u_{*t}$ is proportional to $U_\infty - U_0$, and U_* is approximately proportional to U_∞ , where U_∞ is the free-stream and U_0 is the highest value of U_∞ for no particle motion with the crater in place. With similitude based only on erosion rate, the final form of the dimensionless time is as shown in Figure 2. Although there is significant scatter, the effect of using the mass transport rate similitude is to provide a prediction factor, based upon the degree of distortion, which enables one to estimate full-scale erosional or depositional effects. It should be emphasized that the greater the length scale ratio, the greater the degree of distortion, and thus the more difficult it is to make accurate predictions from model data.

It was later found that, for the model crater tests, there were subtle effects of Reynolds number and the roughness ratio

z_0'/h (Iversen et al., 1976). The improvement in correlation of the data by accounting for these effects is shown in Figure 3.

A second series of experiments was performed for the Iowa Department of Transportation in order to determine the best form of snow-drift control for overpass bridges of superhighways (Ring et al., 1979). Initial data for uncontrolled models, this time for material deposited on the roadway as a function of time (rather than material eroded as in the crater wake), is displayed in Figure 4 (Iversen, 1980a, 1980b). Three different particles (of different density and diameter), two models (one vertically distorted geometrically), and a variety of wind speeds are represented by the data. The selection of modeling material is discussed by Iversen (1979). The result of application of the mass transport rate and roughness parameters is shown in Figure 5. From this correlation it is possible to estimate the amount of roadway covered by snow for a given full-scale snowstorm, duration and wind speed. Quantitative comparison of the uncontrolled roadway and different methods of simulated hedgerow control can be made using the correlation parametric time as shown in Figure 6. In this case hedgerow plan L-20 is an obviously quite superior control to plan T-20 or to the uncontrolled 'bare' model.

A scale model of a Wyoming Highway Department snowfence (Tabler, 1979) was constructed and tested in the wind tunnel in order to see if model drift geometry conforms to full-scale measurements (length scale of 120 to one). Partial results are shown in Figure 7. This figure seems to indicate that, after 16 hours of wind tunnel time, the model drift volume was approaching the full scale equilibrium volume asymptotically. Other aspects of drift geometry were also in good agreement with full scale (Iversen, 1980c, 1980d).

In addition to fence and hedgerow types of controls it is also possible to trap blowing particulate material by means of distributed roughness (Chepil and Woodruff, 1963; Tabler and Jairell, 1980). It should be as feasible to compare methods of controlling blowing sand or soil by modeling in the aeolian wind tunnel as it is to compare snowdrift control methods.

REFERENCES

- Bagnold, R. A. (1941) *The Physics of Blown Sand and Desert Dunes*. Methuen, London, 265 p.
- Chepil, W. S. and Woodruff, N. P. (1963) *The Physics of Wind Erosion and Its Control*. *Advances in Agronomy* 15, 211-302.
- Gerdell, R. W. and Strom, G. H. (1961) *Scale Simulation of a Blowing Snow Environment*. *Proc. Inst. Environ. Sci.* 53, 53-63.

$$\frac{U_F}{U_*} \left(\frac{U_*}{U_c} \right)^2 \left(\frac{U_*}{U_c} \right)^2 \left(\frac{U_*}{U_c} - 1 \right) \frac{U_* \Delta t}{U_c} = 25 + 6.42(10)^5 A_1^2 \frac{U_*}{U_c} \left(\frac{U_*}{U_c} \right)^{4/3} \left(\frac{U_*}{U_c} - 1 \right)^{2/3}$$

Figure 3.
Erosion rate similitude based on mass transport rate and aerodynamic roughness in saltation.

Figure 4.
Snowdrift deposition area as a function of time on a section of superhighway under an overpass bridge.

Figure 5. Snowdrift deposition area as a function of parametric time defined on the basis of mass transport rate and aerodynamic roughness.

Figure 6.

Comparison of deposition area for the uncontrolled roadway and for two hedgerow control configurations.

Figure 7.

Drift volume collected downwind of a model (1/120 scale) Wyoming Highway Department snow fence.

- Greeley, R., Iversen, J. D., Pollack, J. B., Udovich, N., and White, B. R. (1974) Wind Tunnel Studies of Martian Aeolian Processes. Proc., Roy. Soc. London, Series A, 341, 331-360.
- Isyumov, N. (1971) An Approach to the Prediction of Snow Loads. Ph.D. Dissertation, University of Western Ontario, London, Ontario, Canada.
- Iversen, J. D., Greeley, R., Pollack, J. B. and White, B. R. (1973) Simulation of Martian Aeolian Phenomena in the Atmospheric Wind Tunnel. Space Simulation, NASA Spec. Publ. 336, 191-213.
- Iversen, J. D., Greeley, R., White, B. R., and Pollack, J. B. (1975) Aeolian Erosion on the Martian Surface, Part 1, Erosion Rate Similitude. Icarus 26, 321-331.
- Iversen, J. D., Greeley, R., White, R. B., and Pollack, J. B. (1976) The Effect of Vertical Distortion in the Modeling of Sedimentation Phenomena: Martian Crater Wake Streaks. J. Geophys. Res. 81, 4846-4856.
- Iversen, J. D. (1979) Drifting Snow Similitude. Journal of the Hydraulics Division, Proc. of Amer. Soc. Civil Eng. 105 (HY6), 737-753.
- Iversen, J. D. (1980a) Drifting Snow Similitude - Drift Deposit Rate Correlation. Wind Engineering, J. Cermak, ed., Pergamon Press, Oxford, England, 1035-1047.
- Iversen, J. D. (1980b) Drifting Snow Similitude - Transport Rate and Roughness Modeling, in press. Journal of Glaciology.
- Iversen, J. D. (1980c) Wind-Tunnel Modeling of Snow Fences and Natural Snow-Drift Controls. 37th Eastern Snow Conference, Peterborough, Ontario, Canada, ERI 80225.
- Iversen, J. D. (1980d) Comparison of Wind Tunnel Model and Full-Scale Drifts. Submitted to Journal of Industrial Aerodynamics.
- Odar, F. (1962) Scale Factors for Simulation of Drifting Snow. J. Engineering Mechanics Division, ASCE 88 (No. EM2), 1-16.
- Owen, P. R. (1964) Saltation of Uniform Grains in Air. J. Fluid Mechanics 20 (Pt. 2), 225-242.
- Ring, S. L., Iversen, J. D., Sinatra, J. B., and Benson, J. D. (1979) Wind Tunnel Analysis of the Effects of Planting at Highway Grade Separation Structures. ISU-ERI-Ames-79221, Final Report and Supplement, June 1979, Iowa State University.
- Tabler, R. D. (1979) Geometry and Density of Drifts Formed by Snow Fences. (Abstract) Snow in Motion Symposium, Ft. Collins, Colorado.
- Tabler, R. D. and Jairell, R. L. (1980) Studying Snowdrifting Problems with Small-Scale Model Outdoors. Western Snow conference, April 15-17, Laramie, Wyoming.

White, B. R., Iversen, J. D., Greeley, R., and Pollack, J. B.
(1976) Estimated Grain Saltation in a Martian Atmosphere.
Jour. Geophys. Res. 81, 5643-5650.

SHAPE AND DIMENSIONS OF DUNES IN OPEN CHANNEL FLOW

By

Jørgen Fredsøe

ABSTRACT

The shape of a dune formed in a river bed by flowing water is calculated from the equation of continuity for sediment combined with the measured distribution of bed shear stress, downstream a rearward-facing step. From this calculation, the ratio between dune-length and height is obtained. The absolute value of the dune-height is found from a perturbation analysis, from which it turns out that only one single value of the dune-height is stable.

Finally, the flow-resistance for a dune-covered bed is found from the dune-dimensions, by calculating the expansion-loss behind the individual dunes as a Carnot loss.

INTRODUCTION

Dimensions of dunes (length L and height L) are of essential interest for sediment transport in rivers, because the dunes contribute significantly to the total flow resistance due to the expansion-loss behind the dune crest.

In this lecture, some ideas are outlined which explain several of the important features of dunes; like why dunes grow in height with bed shear stress at small bed shear stress, while the dune height decreases with increasing bed shear stress at larger bed shear stress. Further on in this paper, the stretching of dunes at a large bed shear stress is explained.

The main idea is that the equation of continuity for sediment yields a relationship between local sediment transport and local bed level. By use of a sediment transport formula, we find out how the bed level varies with the variation in bed shear stress. This bed shear stress is determined from measurements behind a rear-facing step.

The absolute value of the dune height is found by a perturbation analysis.

1. EQUATION OF CONTINUITY FOR SEDIMENT

Let us make the simplifying assumption that the dunes are moving at a constant velocity a , and without any change in shape. This is of course not true in detail, but it represents all the essential features of the problem with sufficient accuracy. Under these conditions the shape of the bed is described by an expression of the form

$$h = Q(x - at) \quad (1)$$

in which t is the time, and h is the local height of the dune above an x -axis placed through the troughs (Figure 1). Next, we consider the sediment discharge q_T at two consecutive sections, spaced a unit length in the x -direction.

The net outflow of sediment is

$$\frac{\partial q_T}{\partial x},$$

which must equal the change in bed elevation when correction for the porosity, n , of the bed material is considered. Hence, we get the equation of continuity:

$$\frac{\partial q_T}{\partial x} = - (1 - n) \frac{\partial h}{\partial t} \quad (2)$$

If equation (1) is substituted into equation (2) it becomes obvious that the equations are satisfied if we put

$$q_T = q_T(x) = q_0 + a(1 - n) h. \quad (3)$$

The quantity q_0 is a constant, immediately interpreted as the value of q_T for $h = 0$, i.e. at the troughs where the bed load vanishes. q_0 is the part of the total sediment discharge that does not take part in the bed process, and it is consequently identified with the suspended load.

Figure 1
Definition sketch of migrating dunes.

Figure 2
Variation in bed shear stress downstream of a rearward-facing step (Engleund and Fredsøe, 1974).

For the bed load we then obtain the interesting relationship

$$q_B = a(1 - n)h \quad (4)$$

from which we see that the local intensity of bed load transport is proportional to the local height of the bed above the plane through the troughs.

The migration velocity a is thus determined by

$$a = \frac{q_{top}}{(1-n)H} \quad (5)$$

Equation (5) states that the sediment is deposited on the front just downstream of the crest of the dunes, which is the case as long as the sediment is transported mainly as bed load.

From equation (4), the local bed level can be found as a function of the local bed shear stress ϕ^* , because the non-dimensional bed load ϕ is a function of the non-dimensional bed shear stress (θ^*) only

$$\phi = \phi(\theta^*) \quad \phi = q/\sqrt{(s-1)gd^3} \quad (6)$$

$$\theta = \frac{\tau}{\rho(s-1)gd}$$

where g is the acceleration of gravity, s the relative density, ρ the water density and τ the local bed shear stress.

As a transport formula, the Meyer-Peter formula

$$\phi = 8(\theta^* - 0.047)^{3/2} \quad (6a)$$

can be used.

Combining equations (4) and (5), we obtain

$$\frac{h}{H} = \frac{\phi(\theta^*)}{\phi(\theta_{top})} \quad (7)$$

Hence, if the variation in θ^* is known, the local variation in the bed level can also be found from equations (6) and (7).

2. BED SHEAR DISTRIBUTION PAST A DUNE

The non-uniform flow past a dune changes from flow in a very strong adverse pressure gradient just behind the dune top, to a slightly favorable pressure gradient along the mild positive slope of the dune. Disregarding this favorable gradient, the variation in bed shear stress along the dune is very similar to the bed shear distribution behind a rear-facing step. As long as the dune height is not very small when

compared to the boundary layer thickness, it will be characterized by a local maximum in bed shear stress, about 15-18 times the step height downstream (Figure 2). This maximum can be explained as follows: just after the point of reattachment, the boundary shear stress increases due to flow acceleration. Further downstream, the decrease in bed shear stress caused by the increase in the boundary layer thickness is larger than the increase due to flow acceleration, so the bed shear stress decreases slightly.

As mentioned above, flow over dunes is slightly different from flow past a step due to the favorable pressure gradient created by the presence of the dune itself. The bed shear stress distribution on a triangular element downstream of a step has been measured as shown in Figure 3. In this case, local maximum in bed shear stress is also observed as being about 15 times the step height downstream, but further downstream, there is yet another increase in bed shear stress. This is due to the fact that in a converging flow, the boundary layer thickness is not increasing in the downstream direction. A similar increase in flow over real dunes will not take place, as the dune shape is not triangular, but rather flat at the dune top. Hence, the experiment indicates that even in the case of a slightly favorable pressure gradient, a local maximum in bed shear stress is observed.

The measurements described in Figures 2 and 3 yield that the bed shear stress past a dune can be written as

$$\frac{\theta^*}{\theta^*_{\text{top}}} = \left(\frac{1 - H/D}{1 - h/D} \right)^2 \frac{f(x)}{H} \quad (8)$$

The first factor on the right-hand side of equation (8) takes into account the changes in cross-section, while the second is the non-dimensional function given in Figure 2.

3. CALCULATION OF SHAPE AT LOW SEDIMENT TRANSPORT RATE

For a given dune height, H , water depth D and maximum bed shear stress θ^*_{top} , the local bed height h , and hence the form of the dunes can be calculated from equations (7) and (8) as illustrated in Figure 4, where such calculations are shown for two different values of the relative step height H/D .

These calculations of the dune shape can be carried out for an arbitrary value of the step height, so nothing can be said about the value of H . However, in the following, it will be demonstrated that only one single value of H/D will be stable while at all other values of H/D , the dune form, calculated as demonstrated in Figure 4, will either increase or decrease with time if the dune is superposed by a small disturbance, h (Figure 5). The calculations follow the idea

Figure 3
Variation in bed drag on a triangular element downstream of a rearward-facing step.

Figure 4
Calculated dune form for different step height.
a: $H/D = 0.1$ b: $H/D = 0.3$ $\theta_{\text{top}} = 0.3$

Figure 5
Definition sketch of the perturbation of the dune form, presented in Figure 4.

outlined by Fredsøe (1980) concerning the calculation of dune behavior in an unsteady flow.

When we perturb the original bed form h by a perturbation $\tilde{h}$ the change in the total bed height at the dune top can be calculated by

$$\frac{d(h + \tilde{h})}{dt} = \frac{\partial}{\partial t} (h + \tilde{h}) + a \frac{\partial(h + \tilde{h})}{\partial x} \quad (9)$$

where a is the migration velocity of the perturbed dune, given by

$$a = \frac{1}{1 - n} \frac{q + \tilde{q}}{h + \tilde{h}} \quad (10)$$

cf. equation (5) q is the local change in sediment transport, due to the perturbation h of the dune. In equations (9) and (10), all the values shall be taken from the top of the dune.

Equation (10) can be linearized to be

$$a = \frac{1}{1 - n} q_{top} + \frac{q}{q_{top}} - \frac{\tilde{h}}{H} \quad (11)$$

assuming the perturbation to be small. The first term on the right-hand side of equation (9) can be found from the equation of continuity for sediment

$$\frac{\partial(h + \tilde{h})}{\partial t} = - \frac{1}{1 - n} \frac{\partial}{\partial x} (q + \tilde{q}). \quad (12)$$

The longitudinal variation in sediment transport is related to the longitudinal variation in bed shear stress by equation (8), assuming that the friction factor remains constant in the perturbed flow. Further, the longitudinal variation in bed shear stress is related to the local bed height by equation (6), in which h now must be substituted by $h + \tilde{h}$. H remains unchanged in equation (6), as the step downstream of the dune is fixed. Now, the last term in equation (12) can be calculated as

$$\frac{\partial}{\partial x} (\tilde{q}) = \frac{\partial}{\partial x} \left(\frac{\partial q}{\partial \theta} \frac{\partial \theta}{\partial h} \tilde{h} \right) = \frac{\partial q}{\partial \theta} \frac{\partial \theta}{\partial h} \frac{\partial \tilde{h}}{\partial x} \quad (13)$$

$$\frac{\partial \theta}{\partial h} = 2\theta \cdot f \cdot \frac{x}{H} \cdot \frac{H/D}{1 - H/D} \quad (14)$$

The function f is defined in equation (8).

Inserting equations (11), (12), (13), and (14) into equation (9), we obtain, after some simple rearrangement

$$(1 - n) \frac{d\tilde{h}}{dt} = \gamma \frac{\partial}{\partial x} \left(\frac{\tilde{h}}{h} \right);$$

where

$$\gamma = q - 2 \frac{dq}{d\theta} \frac{H/D}{1 - H/D} \quad (15)$$

From equation (15) it is seen, that if γ is positive, the perturbation of the dune will increase if the longitudinal gradient of the perturbation is positive. If γ is negative, the dune form is still unstable, namely to perturbations with negative longitudinal gradients in the ratio between the perturbation and the original bed level. Hence, the only stable dune form occurs for $\gamma = 0$, which yields the following relationship from dune height to water depth

$$\frac{H}{D} / \left(1 - \frac{H}{D} \right) = \frac{\phi}{d\theta/2\theta d\phi}, \quad (16)$$

where the values at the right-hand side must be taken at the dune top. This relationship is depicted in Figure 6a. The sediment transport is calculated by Engelund and Fredsøe (1976) and the data is the Fort Collins data amassed by Engelund and Fredsøe in 1974.

In the calculation of the dune form depicted in Figure 4, the value of H/D obtained from equation (16) has been used in Figure 4b

4. INFLUENCE OF SUSPENDED SEDIMENT ON DUNE HEIGHT AND LENGTH

a) Influence on dune height

At larger sediment transport rates, only a part of the sediment q_1 , carried past the crest, will settle on the dune front, while the remaining part q_2 will be carried in suspension further downstream towards the next dune. Even though some of the suspended sediment may be caught in the separation bubble in the lee of the dune front, it seems to be a good approximation to identify q_1 as the amount of bed load q_b at the dune top and hence take q_2 as the amount of suspended load q_s . This hypothesis has been tested by the author, Fredsøe

Figure 6

- a. Variation in dune height with bed shear stress in case of pure bed load. The data are from Guy et. al. (1966).
 b. Variation in dune height with bed shear stress for different grain sizes.

(1980), by calculating the migration velocity of dunes and comparing it with that which is actually measured. Now, a perturbation analysis like that carried out in the previous section indicates that the dune height is determined by

$$\frac{H}{D} \left(1 - \frac{H}{D} \right) = \frac{\phi_b}{2\theta (d\phi_b/d\theta + d\phi_s/d\theta)} \quad (17)$$

where index b denotes bed load and s, suspended load.

In Figure 6b, equation (17) is sketched for different grain sizes, according to those used in the Fort Collins experiments. Equation (17) has been calculated by use of the sediment formulas presented by Engelund and Fredsøe (1976), by which it is possible to calculate the bed load and suspended load separately. It is seen from the figure that at low values of θ , the height is determined by a single curve where all the sediment is carried as bed load, while this curve at higher θ -values splits up into a family of curves, depending on the value of the grain size. This is due to the fact that the amount of suspended load not only depends on θ' but also on other parameters, of which w/U_f (w = fall velocity, U_f = shear velocity) is the most important, and d/D large values of θ' also has some influence.

b) Influence on dune length

Also at higher values of θ it is observed that the height-length ratio increases to be larger than about 15-18, where the point of maximum bed shear stress is located, see Section 2.

This can be explained as follows: as pointed out by Kennedy (1963) in connection with stability analysis, a phase shift exists between local sediment transport, local flow velocity and local bed height. In case of fully developed dunes, there is no phase shift between local bed height and local sediment transport rate, but because of the final fall velocity of sediment, there exists a phase shift between the local rate of suspended sediment and the local bed shear stress, see for instance Engelund and Fredsøe (1974). Hence maximum transport of suspended sediment takes place at a certain distance δ downstream of the place of maximum bed shear stress.

This phase lag δ is approximately given by

$$\delta = \frac{\epsilon V_b}{w^2} \quad (18)$$

where ϵ is the eddy viscosity and U_b a typical near bed flow velocity. An evaluation of equation (3) is, for instance,

given in reference (3). Now, the distance of maximum sediment transport, and hence maximum bed level, is removed downstream to the point of maximum bed shear stress, namely from about $18 H$ to

$$L_1 = [18 H q_b + (18 H + \delta) q_s] / (q_b + q_s). \quad (19)$$

Still using the argument that the dune stops just downstream of the maximum bed level, it is possible to find the dune length at high sediment transport from equation (19), as sketched in Figure 7.

5. FLOW RESISTANCE

The total bed shear stress is normally split up into two parts: θ' , the mean value of the part acting as skin friction on the dune surface, and θ'' corresponds to a form drag, due to the expansion loss behind the crest of the dune. This latter may be calculated as a Carnot loss as done in Engelund (1978), to be

$$\theta'' = \frac{1}{2} \frac{V^2}{gD} \frac{\alpha}{(s-1)} \frac{H^2}{DL} \quad (20)$$

where α is a nondimensional coefficient (the velocity distribution coefficient) approaching unity as the angle of the dune front approaches the angle of repose (33°). In the following, α is taken to be unity. θ' is given by

$$\theta' = \frac{U_f'^2}{(s-1)gd} = \frac{V^2}{gd} \frac{1}{(s-1)} \left(\frac{U_f'^2}{V^2} \right) \quad (21)$$

where V/U_f' can be calculated from Einstein's approach. Hence, the relation between θ and θ' can be found by

$$\theta = \theta' + \theta'' = \theta' \left[1 + \frac{\alpha}{2} \frac{H}{D} \frac{H}{L} \left(\frac{V^2}{U_f'^2} \right) \right] \quad (22)$$

At low θ -values, we obtain a single resistance curve, while in the transition region, we have a family of curves. Transition occurs earlier in case of fine sediment, see Figure 8.

Figure 7
Variation in length-height ratio with bed shear stress for different grain sizes.

Figure 8
Relation between skin friction and total friction.

REFERENCES

- Bradshaw, P. and Wong, F. (1972) The Reattachment and Relaxation of a Turbulent Shear Layer. *J. Fluid Mech.*, 52, 113.
- Engelund, F. (1978) Hydraulic Resistance for Flow Over Dunes. Progress Report, No. 44, ISVA, Tech. Univ., Denmark.
- Fredsøe, J. (1980) Unsteady Flow in Straight Alluvial Streams. Part 2: Transition From Dunes to Plane Bed. *J. Fluid Mech.*, in press.
- Engelund, F. and Fredsøe, J. (1976) A Sediment Transport Model for Straight Alluvial Channels. *Nordic Hydrology* 7.
- Engelund, F. and Fredsøe, J. (1974) Transition From Dunes to Plane Bed in Alluvial Channels. Series Paper, No. 4, ISVA Tech. Univ., Denmark.
- Guy, H.P., Simons, D.B. and Richardson, E.V. (1966) Survey of Alluvial Channel Data From Flume Experiments, 195 6-61. U.S. Geological Survey Professional Paper 462-I.
- Kennedy, J.F. (1963) The Mechanics of Dunes and Antidunes in Erodible-Bed Channels. *J. Fluid Mech.*, 16, p. 521.

THE STABILITY AND PROPAGATION OF
SAND RIPPLES AND DUNES

By

I. A. Eltayeb

and

M. H. A. Hassan

ABSTRACT

The study of the development and movement of sand ripples and dunes has received considerable attention in recent years, largely due to its relevance to several meteorological, environmental and engineering problems. Several authors have developed various mathematical and physical models to study the deformation of sand beds consisting of non-cohesive particles under the influence of sheer flows. For a comprehensive review the reader is referred to the papers by Kennedy (1969) and Allen (1974). In general, two approaches have been followed in the theoretical treatment of the movement of sand. The first and older approach is to start directly from the equations of hydrodynamics, and adopts one of the numerous empirical laws for the transport of sediment. This approach was first started by Exner in 1925 using a hydraulic model and has subsequently been developed by several authors (Kennedy 1963, Reynolds 1965, Engelund 1970, Engelund and Fredsøe 1971, Reynolds 1976, Richards 1979). Although this approach has been followed for more than half a century, it is still not fully developed, due to the lack of sound knowledge of the mechanism of the transport of sediment. The second and more recent

approach adopts statistical methods to describe the formation and particle distribution of ripples and dunes.

In this treatment we shall follow the first line of approach. Most theoretical investigations using this method assume that the deformation of a sand bed is a manifestation of instability (Gradowczyk 1968, Fredsøe 1974). Among all theoretical models considered, the potential flow model used by Kennedy (1963) is the simplest. Nevertheless, it predicted results in agreement with experimental and observational data. Experiments and observations, however, further indicate that an erodible sand bed subjected to a uniform flow is deformed into ripples and then into dunes. These dunes evolve into larger dunes which move slowly with the flow, retaining their shape or evolving into yet larger dunes (Bagnold, 1954). Such behavior can not be described by the linear stability theory and so recourse must be made to non-linear analysis (Eltayeb and Hassan 1981).

SECTION 1: INTRODUCTION

In Section 2 we formulate the problem. In Section 3 the linear problem is investigated in the instance of a rigid top surface. This situation isolates the stability of the non-cohesive bed form, in the absence of the interaction of the bed form with surface gravity waves. Here, only one mode of propagation is present. When the surface is free, however, the analysis of Section 4 yields that there are three modes of propagation. The linear problems of both Sections 3 and 4 have already been studied by Kennedy (1963, 1964) for a potential flow perturbation. However, the analysis presented below deals with the general types of perturbations. This has made it possible to isolate the circumstances in which a potential flow is valid. Indeed, when the assumptions made by Kennedy are incorporated in the general results, it is found that the potential flow is the only possible solution. Moreover, the non-dimensionalization of the governing equations shows that the potential flow prevails whenever the time scale of sediment transport is much smaller than the time scale associated with the concomitant motion of the fluid. Instability is then associated with a growth rate of the order of the ratio of the former to the latter time scale. In relation to the free surface problem discussed in Section 4, the potential flow solution also filters out the two modes strongly influenced by surface undulations.

In Section 5, the nonlinear stability problem is studied. Three coupled nonlinear partial differential equations are obtained and some simple solutions relevant to some particular cases, are given. However, the general solution of the three evolutionary equations requires extensive numerical techniques and is not studied here although some general remarks about the nature of the solutions are made.

SECTION 2: BASIC EQUATIONS AND BOUNDARY CONDITIONS

We consider an incompressible inviscid flow of medium depth $\tilde{d}$, flowing freely with uniform velocity $\tilde{v}$, over an erodible bed consisting of non-cohesive particles. We choose a cartesian coordinate system $o(\tilde{x}, \tilde{y}, \tilde{z})$ in which $o\tilde{x}$ is along $\tilde{U}$ and (the vertical), $o\tilde{z}$ is perpendicular to the bed, and $o\tilde{y}$ completes the triad. The bed is assumed to be slightly perturbed from its horizontal position and its profile is given by

$$\tilde{z} = \tilde{\eta}(\tilde{x}, \tilde{y}, \tilde{t}) \quad (2.1)$$

Regarding the top surface, we shall deal separately with the two different cases: (i) a top rigid lid at $\tilde{z} = \tilde{d}$, and (ii) a free surface, given by

$$\tilde{z} = \tilde{d} = \tilde{\eta}(\tilde{x}, \tilde{y}, \tilde{t}) \quad (2.2)$$

We now suppose that the presence of the profile (2.1) causes small perturbations $\tilde{u} = (\tilde{u}, \tilde{v}, \tilde{w})$ and p in velocity and pressure. The subsequent flow will then be governed by the equations of motion and continuity

$$\frac{\partial \tilde{u}}{\partial \tilde{t}} + \tilde{u} \frac{\partial \tilde{u}}{\partial \tilde{x}} + \tilde{u} \cdot \nabla \tilde{u} = -\nabla \left(\frac{p}{\rho} = \tilde{g} \tilde{z} \right) \quad (2.3)$$

$$\nabla \cdot \tilde{u} = 0 \quad (2.4)$$

These equations must satisfy the following boundary conditions: (i) If the top is bounded by a rigid lid, the component of velocity, w , normal to the rigid lid $\tilde{z} = \tilde{d}$ vanishes, i.e.,

$$\tilde{w} = 0 \text{ at } \tilde{z} = \tilde{d} \quad (2.5a)$$

If the top is a free surface, we require that the component of velocity normal to the surface $\tilde{z} = \tilde{d} + \tilde{\xi}$ must vanish and (2.5a) is replaced by

$$\tilde{w} = \frac{\partial \tilde{\xi}}{\partial \tilde{t}} + \tilde{u} \frac{\partial \tilde{\xi}}{\partial \tilde{x}} + \tilde{u} \frac{\partial \tilde{\xi}}{\partial \tilde{x}} + \tilde{v} \frac{\partial \tilde{\xi}}{\partial \tilde{y}} \quad (2.5b)$$

$$\text{at } \tilde{z} = \tilde{\eta}$$

The conditions to be satisfied at the erodible bed $\tilde{z} = \tilde{\eta}$ are:
(a) The vanishing of the velocity component $\tilde{w}$, normal to the bed; this gives

$$w = \frac{\partial \tilde{\eta}}{\partial \tilde{t}} + \tilde{u} \frac{\partial \tilde{\eta}}{\partial \tilde{x}} + \tilde{u} \frac{\partial \tilde{\eta}}{\partial \tilde{x}} + \tilde{v} \frac{\partial \tilde{\eta}}{\partial \tilde{y}} \quad (2.6)$$

$$\text{at } \tilde{z} = \tilde{\eta}$$

(b) The continuity equation for the sediment motion; this can be written in the form

$$\frac{\partial \tilde{\eta}}{\partial \tilde{t}} + \nabla \cdot \tilde{\underline{G}} = 0 \quad (2.7a)$$

Where $\tilde{\underline{G}}$ is the flux of sediment per unit volume. Assuming that the sediment transport only takes place in the direction of the basic flow $\underline{U}$, (2.7a) reduces to

$$\frac{\partial \tilde{\eta}}{\partial \tilde{t}} + \frac{\partial \tilde{G}}{\partial \tilde{x}} = 0 \quad (2.7b)$$

To close the problem, we require an explicit expression for $\underline{G}$ in terms of $\underline{u}$ and $\tilde{\eta}$. The exact form of $\underline{G}$, however, is not known with adequate accuracy, even for the simplest of flows. Thus, it will be necessary to adopt one of the various empirical relations, based on experiments and observations. Following Kennedy (1964) we postulate the transport law:

$$\tilde{\underline{G}} = \tilde{m} [\tilde{U} - \tilde{U}_c + \tilde{u}(\tilde{x} - \tilde{\delta}, \tilde{\eta}, \tilde{t})]^n \quad (2.8)$$

where $\tilde{U}_c$ is the critical velocity for initiation of motion, $\tilde{m}$ is a dimensional coefficient and $\tilde{n}$ is a dimensionless exponent. The quantity δ defines the distance by which the local sediment transport lags behind the local velocity at the bed. This phase shift $\tilde{\delta}$ is introduced to account for a number of observed phenomena which are not fully taken care of by the relationship in (2.8). It should be emphasized, however, that the lag distance $\tilde{\delta}$ does not explain the full relationship between the local velocity and the rate of sediment transport, but it appears to give the correct cumulative effect of that relationship, and is a simple concept, consistent with the idealized models discussed here.

We shall find it convenient in the analysis below to convert the above basic equations into dimensionless form. This leads to the appearance of certain dimensionless parameters which provide a measure of the relative importance of various mechanisms in the physical problem.

We let $\tilde{U}$, $\tilde{d}$, $\tilde{d}/\tilde{U}$ and $\rho\tilde{U}^2$ be units of velocity, distance, time and pressure, respectively. The above equations then take the dimensionless form:

$$\frac{\partial \underline{u}}{\partial t} + \frac{\partial \underline{u}}{\partial x} + \underline{u} \cdot \nabla \underline{u} = -\nabla \left(p + \frac{z}{F^2} \right) \quad (2.9)$$

here $F = \tilde{U}/\sqrt{g\tilde{d}}$ is the Froude number

$$\nabla \cdot \underline{u} = 0 \quad (2.10)$$

$$w = 0 \text{ at } z = 1 \text{ (in case of a top rigid lid)} \quad (2.11a)$$

$$w = \frac{\partial \eta}{\partial t} + \frac{\partial \eta}{\partial x} + \frac{u \partial \eta}{\partial x} + \frac{v \partial \eta}{\partial y} \quad (2.11b)$$

at $z = 1 + \xi$

(in the case of a free surface at the top)

$$w + \frac{\partial \eta}{\partial t} + \frac{\partial \eta}{\partial x} + \frac{u \partial \eta}{\partial x} + \frac{v \partial \eta}{\partial y} \text{ at } z = \eta \quad (2.12)$$

and

$$\frac{\partial \eta}{\partial t} + \frac{H \partial \eta}{\partial x} = 0 \text{ at } z = \eta$$

where

$$H = \left(\frac{\partial \bar{u}^{n-1}}{\partial x} \right), \quad T = \left[1 - u_c \right. \\ \left. + u(x - \delta, \eta, t)^n \right] \quad (2.13)$$

SECTION 3: THE LINEAR STABILITY THEORY OF A HORIZONTAL LAYER WITH A TOP RIGID LID

We shall start with the simplest model in which the flow $\bar{u}$ is over a two-dimensional erodible bed at the bottom, and is bounded at the top by a rigid lid (Figure 1). The presence of a rigid lid at the top is assumed here for three reasons: First, the problem is simplified and it admits a simple analytical solution that gives some insight into the stability of the erodible bed. Second, the top rigid lid excludes the appearance on the top surface of gravity waves that are known to interact with the bottom undulations and hence complicate the problem. Third, the configuration of a rigid lid is experimentally possible. This model has been studied by Kennedy (1964) for potential flow perturbations. However, this restriction will not be assured here.

In this case the equations (2.4) through (2.12) reduce to

$$\frac{\partial \underline{u}}{\partial t} + \frac{\partial \underline{u}}{\partial x} + \underline{u} \cdot \nabla \underline{u} = -\nabla p \quad (3.1)$$

$$\nabla \cdot \underline{u} = 0 \quad (3.2)$$

$$w = 0 \text{ at } z = 1 \quad (3.3)$$

$$w = \frac{\partial \eta}{\partial t} + \frac{\partial \eta}{\partial x} + \frac{u \partial \eta}{\partial x} \text{ at } z = \eta \quad (3.4)$$

$$\frac{\partial \eta}{\partial t} + \frac{H \partial \eta}{\partial x} = 0 \text{ at } z = \eta \quad (3.5)$$

Figure 1
Horizontal layer with a top rigid lid.

Figure 2
Horizontal layer with a free surface.

Figure 3
The dependence of F^2 on k .

where $T = [1 - U_c + u(x - \delta, \eta, t)]^n$,

$$H = \frac{\partial \eta}{\partial x}^{n-1}$$

The above equations are linearized in the usual way by neglecting products and cross-products of u , η , and p . The linearized equations are:

$$\frac{\partial u}{\partial t} + \frac{\partial u}{\partial x} = -\nabla p \quad (3.6)$$

$$\nabla \cdot \underline{u} = 0 \quad (3.7)$$

$$w = 0 \text{ at } z = 1 \quad (3.8)$$

$$w = \frac{\partial \eta}{\partial t} + \frac{\partial \eta}{\partial x} \text{ at } z = 0 \quad (3.9)$$

$$\frac{\partial \eta}{\partial t} + \frac{H \partial T}{\partial x} = 0 \text{ at } z = 0 \quad (3.10)$$

$$T = (1 - U_c) + n(1 - U_c)^{n-1} u(x - \delta, 0, t) \quad (3.11)$$

we now assume that

$$\eta(x, t) = A(t) \exp [ik(x - U_b t)] \quad (3.12)$$

where U_b is the speed of the bed. It then follows that

$$[u, w, p] = [\hat{u}(z, t), \hat{w}(z, t), \hat{p}(z, t)] \exp [ik(x - U_b t)] \quad (3.13)$$

Substituting (3.13) into (3.6) and (3.7) and eliminating $\hat{u}$ from the equations we get the following equations for $\hat{w}$

$$\hat{L} \left(\frac{\partial^2}{\partial z^2} - k^2 \right) w = 0 \quad (3.14)$$

where

$$\hat{L} = \frac{\partial}{\partial t} + ik(1 - U_b) \quad (3.15)$$

equation (3.14) gives

$$\begin{aligned} \hat{L}\hat{w} &= \bar{E}(t) \cosh(kz) \\ &+ \bar{D}(t) \sinh(kz) \end{aligned} \quad (3.16)$$

and hence

$$\begin{aligned} \hat{w} &= \bar{\bar{E}}(t) \cosh(kz) \\ &+ \bar{\bar{D}}(t) \sinh(kz) \end{aligned} \quad (3.17)$$

in which

$$\begin{aligned} \bar{\bar{E}}(t) &= e^{-ik(1-U_b)t} \\ &\int^t \bar{E}(\xi) e^{ik(1-U)\xi} d\xi \\ \bar{\bar{D}}(t) &= e^{-ik(1-U_b)t} \\ &\int^t \bar{D}(\xi) e^{ik(1-U_b)\xi} d\xi \end{aligned} \quad (3.18)$$

The boundary conditions (3.9) and (3.8) give, respectively,

$$\hat{L}A = \bar{\bar{E}}(t) \quad (3.19)$$

$$0 = \bar{\bar{E}}(t) \cosh(k) + \bar{\bar{D}}(t) \sinh(k) \quad (3.20)$$

Inserting the expression (3.11) into (3.10) we have

$$\begin{aligned} & (\hat{\ell} - ik)A + \overline{H}n(1 - U_c)^{n-1} \frac{\partial}{\partial x} u(x - \delta, 0, t) \\ & = 0 \end{aligned}$$

and the further use of (3.7) gives

$$(\hat{\ell} - ik)A + \overline{H} e^{-ik\delta} \frac{(-\partial w)}{\partial z} \Big|_{z=0} = 0$$

Thus

$$(\hat{\ell} - ik)A = \overline{H} e^{-ik\delta} \overline{D}(t), \quad (3.21)$$

where

$$\overline{H} = n(H) (1 - U_c)^{n-1} \quad (3.22)$$

We now use (3.19) and (3.21) to eliminate $\overline{D}(t)$ and $\overline{E}(t)$ from (3.20). Thus, we find that the growth of the amplitude $A(t)$ of the bed form is governed by the equation

$$\hat{\delta}A + \frac{\tan h(k)}{k\overline{H}e^{-ik\delta}} (\hat{\ell} - ik)A = 0 \quad (3.23)$$

This is a first-order equation in d/dt and the solution is readily obtainable. Thus

$$A(t) = a_0 \exp(\gamma t), \quad (3.24)$$

where a_0 is a constant, and

$$\gamma = ikU_b - ik \left\{ 1 + \frac{\tan h(k)}{k\overline{H}e^{-ik\delta}} \right\}^{-1}$$

$$\begin{aligned}
= ikU_b - \frac{ik^2\bar{H} [\bar{k}\bar{H} + \tan h(k) \cdot \cos(k\delta)]}{k^2\bar{H}^2 + \tan h^2(k) + 2k\bar{H} \tan h(k) \cos k\delta} \\
- \frac{k^2\bar{H} \tan h(\bar{k}) \cdot \sin(k\delta)}{k^2\bar{H}^2 + \tan h^2(k) + 2k\bar{H} \tan h(k) \cos k\delta} \quad (3.25)
\end{aligned}$$

Since the oscillatory form of the solution is already specified in (3.12), then γ must be real. Consequently,

$$U_b = \frac{k\bar{H}[\bar{k}\bar{H} + \tan h(k) \cdot \cos(k\delta)]}{k^2\bar{H}^2 + \tan h^2(k) + 2k\bar{H} \tan h(k) \cos k\delta} \quad (3.26)$$

$$\gamma = - \frac{k^2\bar{H} \tan h(k) \cdot \sin(k\delta)}{k^2\bar{H}^2 + \tan h^2(k) + 2k\bar{H} \tan h(k) \cos k\delta} \quad (3.27)$$

Now the associated velocity and pressure perturbations can be found from (3.17), (3.19), (3.22), (3.25), and (3.7).

$$\hat{w} = a_0 [\gamma + ik(1 - U_b)] \sin h[k(1 - z)]/\sin h(k) \quad (3.28)$$

$$\hat{u} = ia_0 [\gamma + ik(1 - U_b)] \cos h[k(1 - z)]/\sin h(k) \quad (3.29)$$

$$\hat{p} = -a_0 [\gamma + ik(1 - U_b)] \cos h[k(1 - z)]/\sin h(k) \quad (3.30)$$

The perturbation pressure is non-vanishing on the top surface $z = 1$, a result that contrasts with the free surface case discussed in the next section.

Observations in experiments and natural systems indicate that bed forms of non-cohesive particles tend to possess a dominant wavelength. Within the context of stability theory the dominant wavelength is that which yields the maximum initial growth rate of the amplitude $A(t)$. Now the initial growth rate of amplitude is

$$dA/dt \Big|_{t=0} = a_0\gamma \quad (3.31)$$

The turning points of initial growth rate as a function of k are given by

$$\begin{aligned}
 0 = \frac{\partial \gamma}{\partial k} = -\bar{H} & \left| \frac{k^2 \tan h(k) + \sec h^2(k)}{k^2 \bar{H}^2 + \tan h^2(k) + 2k\bar{H} \tan h(k) \cos k\delta} \right. \\
 & - \frac{k^2 \tan h(k) \{2k\bar{H}^2 + 2 \tan h(k) \sec h^2(k)\}}{k^2 \bar{H}^2 + \tan h^2(k) + 2k\bar{H} \tan h(k) \cos k\delta} \\
 & + \frac{2\bar{H} \cos k [\tan h(k) = k \sec h^2(k)]}{2 \sin k\delta} \Big|_x \\
 & - k^2 \bar{H} \tan h(k) \left| \frac{\cos k}{k^2 \bar{H}^2 + \tan h^2(k) + 2k\bar{H} \tan h(k) \cos k\delta} \right. \\
 & + \frac{2k\bar{H} \tan h(k) \sin^2 k\delta}{\{k^2 \bar{H}^2 + \tan h^2(k) + 2k\bar{H} \tan h(k) \cos k\delta\}} \\
 & \frac{\partial}{\partial k(k\delta)} \quad (3.32)
 \end{aligned}$$

If we suppose that δ is independent of k , then (3.32) reduces to

$$\begin{aligned}
 k_c \tan h^2(k_c) \sec h^2(k_c) - k_c^3 \bar{H}^2 \sec h^2(k_c) - 2 \tan h^3(k_c) \\
 - 2k_c \bar{H} \tan h^2(k_c) [k_c^2 \bar{H}^2 + \tan h^2(k_c)] \cot(k_c \delta) \\
 2k_c \tan h(k_c) [k_c^2 \bar{H}^2 + \tan h^2(k_c)] \cot(k_c \delta) \\
 + 2k_c \bar{H} k_c \delta \tan h^2(k_c) / \sin k_c \delta. \quad (3.33)
 \end{aligned}$$

By the property of periodicity of $\cot(k_c \delta)$, equation (3.33) possesses an infinite number of zeros for k_c for each given pair (δ, H) . The zeros can only be calculated numerically.

Now the analysis presented here yields the expressions (3.28) and (3.29) for the velocity field which, in general, is not derivable from a potential, and consequently the perturbations are not irrotational. Kennedy (1964) studied the same problem subject to the conditions $|U_b| \ll 1$ and $|\gamma| \ll 1$ assuming a potential flow, and it is of interest to see if the solution to the general problem discussed here reduces to a potential flow when Kennedy's assumptions are applied to it. Indeed when $|\gamma| \ll 1$ and $|U_b| \ll 1$ the linearized equation of motion (3.6) takes the form

$$-\nabla \hat{p} = \ell \hat{u} \approx ik \hat{u} \quad (3.34)$$

so that

$$\hat{u} = \nabla \frac{(i\hat{p})}{k},$$

and the velocity is derivable from a potential $(i\hat{p}/k)$ which is proportional to the perturbation pressure. Thus when $|\gamma|, |U_b| \ll 1$, the only possible solution for the rigid lid model is a potential flow.

Now the condition $|\gamma| \ll 1$ implies that $|\bar{H}| \ll 1$ so that under the Kennedy assumptions,

$$\begin{aligned} U_b &= k\bar{H} \cot h(k) \cdot \cos(k\delta), \\ \gamma &= -k^2 \bar{H} \cot h(k) \cdot \sin(k\delta) \end{aligned} \quad (3.35)$$

and (3.33) reduces to

$$\frac{\sin h(2k_c)}{2k_c} = \frac{1}{2 + k_c \delta \cot(k_c \delta)} \quad (3.36)$$

The expressions (3.35) and (3.36) are exactly those obtained by Kennedy (1964) and have been studied by him. The fact that Kennedy's results can be obtained from the general expression (3.26), (3.27), (3.33) by making the single assumption

$$|\bar{H}| \ll 1 \quad (3.37)$$

suggests the usefulness of the clarification of the dimensionless number H defined by (3.5). In this respect, it is instructive to consider the transport equation (3.10) written in the form

$$\frac{\partial \eta}{\partial(Ht)} + \frac{\partial T}{\partial x} = 0 \quad (3.38)$$

We now see that the time scale associated with the transportation of the particles is Ht . Remembering that t is the time scale associated with the fluid motion we note that

$$H = \frac{\text{time scale of transportation of particles}}{\text{time scale of fluid motions}} \quad (3.39)$$

We may now conclude that a potential flow prevails when the particles are transported on a time scale much shorter than that of the fluid motions.

Returning to (3.36) we see that for a given δ , it has roots for $(\bar{n} + 1/2)\pi < k_c \delta < (\bar{n} + 1)\pi$, and for $(\bar{n} + 3/2)\pi < k_c \delta < (\bar{n} + 2)\pi$ where $n = 0, 1, 2, 3, \dots$. The zeros of the first are associated with $\sin(k_c \delta) = 0$, and according to (3.35) $\gamma < 0$, and the bed form decays to a flat bed. However, the second set of zeros corresponds to $\sin(k_c \delta) = 0$ and a positive growth occurs. The specific values of k_c as a function of δ have been calculated by Kennedy (1964, p. 1520) and will not be discussed in detail at this time. It may, however, be pointed out that a dominant wavenumber (and hence wavelength), according to the criterion used here, is indeed present for every value of δ .

SECTION 4: THE LINEAR STABILITY THEORY OF A HORIZONTAL LAYER WITH A FREE SURFACE

In this section, we consider the same model of the previous section; but in the absence of a top rigid lid (Figure 2). It will be shown that the presence of a free surface leads to surface undulations in the form of surface gravity waves which interact with the bottom undulations and strongly influence the behavior of the bed form.

Following the same procedure as in Section 3, we find that the relevant linearized equations and boundary conditions are:

$$\frac{\partial u}{\partial t} + \frac{\partial u}{\partial x} = -\nabla p \quad (4.1)$$

$$\nabla \cdot \underline{u} = 0 \quad (4.2)$$

$$w = \frac{\partial \xi}{\partial t} + \frac{\partial \xi}{\partial x} \quad \text{at } z = 1 \quad (4.3)$$

$$w = \frac{\partial \eta}{\partial t} + \frac{\partial \eta}{\partial x} \quad \text{at } z = 0 \quad (4.4)$$

$$\frac{\partial \eta}{\partial t} + H \frac{\partial T}{\partial x} = 0 \quad \text{at } z = 0 \quad (4.5)$$

where

$$T = (1 - U_c) + n(1 - U_c)^{n-1} u(x - \delta, 0, t) \quad (4.6)$$

The continuity of pressure gives the relation

$$p - \frac{\xi}{F^2} = 0 \quad \text{at } z = 1 \quad (4.7)$$

Assuming as before

$$(\eta, \xi, u, w, p) = \{A(t), B(t), u(z, t), w(z, t), p(z, t)\} \\ \exp \{ik(x - U_b t)\} \quad (4.8)$$

we find that (4.1) - (4.7) reduce to

$$\hat{\ell} \hat{w} = - \frac{\partial \hat{p}}{\partial z} \quad (4.9)$$

$$\hat{\ell} \hat{u} = -ik \hat{p} \quad (4.10)$$

$$ik \hat{u} = \frac{\partial \hat{w}}{\partial z} \quad (4.11)$$

$$\hat{w} = \ell B \quad \text{at } z = 1 \quad (4.12)$$

$$\hat{w} = \ell A \quad \text{at } z = 0 \quad (4.13)$$

$$(\hat{\ell} - ik)A + ik\overline{HT} = 0 \quad \text{at } z = 0 \quad (4.14)$$

where

$$\hat{T} = e^{-ik\delta} \hat{u}(x, 0, t), \quad \bar{H} = n(1 - U_c)^{n-1} H \quad (4.15)$$

and

$$\hat{p} - \frac{B}{-2} = 0 \quad \text{at } z = 1 \quad (4.16)$$

If we differentiate (4.10) with respect to z , multiply by ik , add the resulting two equations and use (4.11) we get

$$\frac{\partial^2 \hat{p}}{\partial z^2} - k^2 \hat{p} = 0 \quad (4.17)$$

Thus $\hat{p}$ is a harmonic function with time t as a parameter. Hence,

$$\hat{p} = D(t) \cos h(kz) + E(t) \sin h(kz) \quad (4.18)$$

The boundary condition (4.16), demands that

$$\frac{B}{F^2} = D(t) \cos h(k) + E(t) \sin h(k) \quad (4.19)$$

From equations (4.9) and (4.12) we have

$$\hat{\ell}^2 \hat{2}_B = \hat{\ell}^2 \hat{w} = - \left. \frac{\partial \hat{p}}{\partial z} \right|_{z=1} \quad (4.20)$$

$$= -k[D(t) \sin h(k) + E(t) \cos h(k)]$$

Now (4.19) and (4.20) can be solved for D and E to give

$$-k E(t) = \cos h(k) \cdot \hat{\ell}^2 \hat{B} + \frac{kB}{F^2} \sin h(k)$$

$$kD(t) = \sin h(k) \cdot \hat{\ell}_B + \frac{kB}{F^2} \cos h(k) \quad (4.21)$$

Also (4.9) and (4.13) give

$$\hat{\ell}_A = \left. \frac{\partial p}{\partial z} \right|_{z=0} = -kE(t) \quad (4.22)$$

while (4.14) and (4.11) yield

$$(\hat{\ell} - ik)A + \bar{H}e^{-ik\delta} \left. - \left(\frac{\partial w}{\partial z} \right) \right|_{\substack{z=0 \\ x=x-\sigma}} = 0$$

The application of $\hat{\ell}$ to this last equation and the use of (4.9) at $z = 0$ lead to

$$(\hat{\ell} - ik) \hat{\ell}_A = -\bar{H}e^{-ik\delta} k^2 D(t) \quad (4.23)$$

By using (4.22) and (4.23) in (4.19) we obtain

$$\frac{-kB}{F^2} - \frac{e^{ik\delta} \cos h(k)}{k\bar{H}} (\hat{\ell} - ik) A + \sin h(k) \cdot \hat{\ell}^2 A \quad (4.24)$$

Equation (4.22) together with (4.24) and the first of (4.21) yields

$$\hat{\ell}^2 A = \cos h(k) \left[\frac{-F^2 e^{ik\delta}}{k\bar{H}} \cos h(k) \cdot (\hat{\ell} - ik) \ell^3 A - \frac{F^2}{k} \sin h(k) \cdot \ell^4 A \right]$$

$$- \sin h(k) \cdot \left[\frac{e^{ik\delta} \cos h(k)}{k\bar{H}} (\hat{\lambda} - ik)\hat{\lambda}A + \sin h(k)\hat{\lambda}A \right]$$

Thus the amplitude A of the bedform is governed by the equation

$$\left[e^{ik\delta} (\hat{\lambda} - ik)\hat{\lambda}^2 + k\bar{H} \tan h(k)\hat{\lambda}^3 + \frac{k^2\bar{H}}{F^2} \sec h^2(k)\hat{\lambda} + \frac{ke^{ik\delta}}{F^2} \tan h(k) \cdot (\hat{\lambda} - ik) + \frac{k^2\bar{H}}{F^2} \tan h^2(k) \right] \hat{\lambda}A = 0 \quad (4.25)$$

We note here that B is also governed by the same equation. The solution can be obtained by solving (4.25) and then evaluating B from (4.24), E from (4.22), D from (4.23), $\hat{p}$ from (4.18) and $\hat{u}$, $\hat{w}$ from (4.9) and 4.19).

Assuming solutions of the form

$$A(t) = a_0 \exp(\gamma t) \quad (4.26)$$

and disregarding the root $\gamma = -ik(1 - U_b)$, which represents a solution with $U_b = 1$, so that the bed form moves as part of the fluid, equation (4.25) yields a cubic for γ . Defining

$$\bar{\gamma} = \gamma - ik(1 - U_b), \quad (4.27)$$

The cubic is

$$a\bar{\gamma}^3 + ikb_1\bar{\gamma}^2 - k^2b_2\bar{\gamma} - ik^3b_3 = 0 \quad (4.28)$$

in which

$$a = 1 + k\bar{H}e^{-ik\delta} \tan h(k)$$

$$b_1 = 2 + 3k\bar{H}e^{-ik\delta} \tan h(k)$$

$$b_2 = 1 + 3k\bar{H}e^{-ik\delta} \tan h(k) - \frac{1}{kF^2} [\tan h(k) + k\bar{H}e^{-ik\delta}]$$

$$b_3 = k\bar{H}e^{-ik\delta} \left[\tan h(k) - \frac{1}{kF^2} \right] \quad (4.29)$$

The roots γ of (4.28) can only be calculated numerically, in general. Our main concern here is not to calculate the roots for various values of δ , k , F , and $\bar{H}$ but to illustrate certain points of general interest and find the roots in some simple cases relevant to some experimental data.

The fact that $\bar{\gamma}$ has, in general, three roots, means that the presence of the free surface on the problem is to introduce two more modes (as compared to the rigid lid case). Also the previous case of a rigid lid showed that Kennedy's solution can be obtained by setting $|\bar{H}| \ll 1$, which is the case here as well. For $|\bar{H}| \ll 1$, (4.28) can be solved by expanding the roots in terms of the small parameter $\bar{H}$ to find that, to leading order, the three roots of (4.28) are

$$\bar{\gamma}_1 = - \frac{ik\bar{H} \left[\tan h(k) - \frac{1}{kF^2} \right] e^{-ik\delta}}{1 - \tan h(k)/kF^2} \quad (4.30)$$

$$\bar{\gamma}_{2,3} = -ik \pm i[k \tan h(k)]^{1/2}/F \quad (4.31)$$

The mode $\bar{\gamma}_1$ is therefore associated with a growth rate γ_1 and bed form speed U_b given by

$$\gamma_1 = \frac{k\bar{H}[1 - kF^2 \tan h(k)]}{kF^2 - \tan h(k)} \sin(k\delta) \quad (4.32)$$

$$U_b = - \frac{k\bar{H} [1 - kF^2 \tan h(k)]}{kF^2 - \tan h(k)} \cos(k\delta) \quad (4.33)$$

and these are precisely the same as those obtained by Kennedy (1963). Indeed, when the assumptions

$$|H| \ll 1, \frac{dA}{dt} = O(HA) \quad (4.34)$$

are made in (4.9) and (4.10), we find that the flow (u, w) is obtainable from a potential. However, if the assumptions (4.34) are made from the offset, the modes (4.25) will be filtered out as in the analysis by Kennedy.

Now the mode (4.32), (4.33) is associated with both growing rate γ and speed U_b being of comparable magnitude (order $\bar{H}$). The modes (4.31), on the other hand, are associated with U_b being independent of $|\bar{H}|$ (to leading order) when $|\bar{H}|$ is small:

$$U_b = 1 \pm \frac{1}{F} \left[\frac{\tan h(k)}{k} \right]^{1/2} \quad (4.35)$$

The growth rates, however, calculated from the expressions for $\gamma_{2,3}$ up to order H are

$$\gamma_{2,3} = \frac{\bar{H} \{k + [k \tan h(k)]^{1/2}\}}{2F^2 \left[\frac{\tan h(k)}{kF^2} - 1 \right]} \sec h^2(k) \sin(k\delta), \quad (4.36)$$

and have a magnitude of order $\bar{H}$. Thus, their growth rates are much smaller than their speeds.

The dominant wave number for each mode can be calculated in the manner described for the rigid lid case and the details of the calculations are omitted here. The calculations are made more tedious because of the appearance of the Froude number. The dominant wave number k_e for the mode γ_1 can be found to be given by

$$F^2 = \frac{1 + k \tan h(k) + k\delta \cot(k\delta)}{k^2 + [2 + k\delta \cot(k\delta)] k \tan h(k)} \quad (4.37)$$

The solutions of (4.36) are given in Kennedy (1969).

Now the mode (4.32) represents instability of $\gamma_1 > 0$ and this depends on $\sin(k\delta)$ as well as on the signs of the expressions $1 - kF^2 \tan h(k)$ and $kF^2 \tan h(k)$. In other words,

for a given wave number k , the stability depends on the lag distance δ and on the Froude number F^2 . Figure 3 illustrates the dependence on F^2 on k for the two relations $1 - kF^2 \tan h(k) = 0$, and $kF^2 - \tan h(k) = 0$. The shaded region is governed by

$$\frac{\tan h(k)}{k} < F^2 < \frac{\cot h(k)}{k} \quad (4.38)$$

This region provides the only area of the (k, F^2) plane for which

$$\frac{1 - kF^2 \tan h(k)}{kF^2 - \tan h(k)} > 0 \quad (4.39)$$

Thus when the inequality (4.38) holds, then instability occurs when

$$\sigma < k\delta < \pi \quad (4.40)$$

Now in this range of $k\delta$, U_b can take both positive and negative values. If $0 < k\delta < \pi/2$, U_b is negative (for positive k) and the bed form moves upstream. The bed and free surface are in phase and the bed form is then of the anti-dune type. When $k\delta = \pi/2$, $U_b = 0$ and B and A are in phase so that instability occurs in the form of stationary anti-dunes. If, however $k\delta$ takes values between $\pi/2$ and π , then U_b is positive and the bed form moves downstream. A and B are again in phase and instability is manifested as anti-dunes. At $k\delta = \pi$ the bed form is neutrally stable.

For values of F^2 violating the inequality (4.38), instability is present if

$$\pi < k\delta < 2\pi$$

The sense of movement and type of bed form can be identified from (4.32), (4.33). A full list can also be found in Table 1 in Kennedy (1969).

Before we conclude this case, it is of some interest to observe that the bed form speed U_b associated with the mode γ_2 is greater than 1, so that the bed form moves faster than the fluid. This result is unlikely to occur and may well be

due to the inadequacy of the transport equation (2.7) at values of U_b comparable with U since then one would expect that the bed load transport may not be fully dominant.

It is also appropriate here to mention that more sophisticated mathematical models have been studied by Engelund, Hansen and Fredsoe. These models use the concept of eddy viscosity and take into account the bed shear stress in the analysis. They also include the influence of suspension load as well as bed load. The reader is referred to the lectures by Dr. Jorgen Fredsoe.

SECTION 5: THE NONLINEAR EVOLUTION OF SAND DUNES

We shall now deal with the nonlinear theory for the evolution of dunes. We shall consider the simplest model in which the top surface is a rigid lid. The analysis carried out in Section 3 showed that for most practical purposes, the dimensional parameter H is much less than one, in which case the perturbation flow is a potential one. We thus introduce a potential ϕ , by

$$\underline{u} = \nabla \phi \quad (5.1)$$

so that the continuity equation (2.10) gives

$$\nabla^2 \phi = 0. \quad (5.2)$$

We are then required to solve (5.2) subject to the boundary conditions (2.11)-(2.13) in Section 3 above. For a flow satisfying (5.1), these conditions take the following form

$$\frac{\partial \phi}{\partial z} = 0 \quad \text{at } z = 1 \quad (5.3)$$

$$\frac{\partial \phi}{\partial z} = \frac{\partial \eta}{\partial t} + \frac{\partial \eta}{\partial x} + \frac{\partial \phi}{\partial x} \cdot \frac{\partial \eta}{\partial x} + \frac{\partial \phi}{\partial y} \cdot \frac{\partial \eta}{\partial y} \quad \text{at } z = \eta \quad (5.4)$$

$$\frac{\partial \eta}{\partial t} + H \frac{\partial T}{\partial x} = 0 \quad \text{at } z = \eta \quad (5.5)$$

where

$$T = \left\{ 1 - U_c + \frac{\partial \phi}{\partial x} (x - \delta, y, \eta, t) \right\} n,$$

$$H = \frac{\sum_m \sum_n u^{n-1}}{\sum_d} \quad (5.6)$$

The system (5.2)-(5.6) is a deterministic one and the equation of motion (2.9), when $\underline{u}$ satisfies (5.1), can be integrated to give

$$p = -\frac{\partial \phi}{\partial x} - 1/2 \left[\left(\frac{\partial \phi}{\partial x} \right)^2 + \left(\frac{\partial \phi}{\partial y} \right)^2 + \left(\frac{\partial \phi}{\partial z} \right)^2 \right] \quad (5.7)$$

which determines the pressure distribution after the system (5.2)-(5.6) had been solved.

Now the analysis presented in Section 3 showed that maximum linear growth of the amplitude is attained when $\delta = 2/k$ in which case U_b and k are related by

$$U_b = \left\{ 1 + \frac{\tanh(k)}{\alpha_1 k H} \right\}^{-1} \quad (5.8)$$

when this maximum growth is attained, the amplitude will evolve on a much longer time scale τ , and nonlinear interactions will become potent.

The nonlinear treatment adopted here employs a multiple scale technique based on a small parameter ϵ which represents the magnitude of the ratio of perturbation to basic state quantities. This method has been used successfully in the study of various nonlinear systems (see, for instance, Davey and Stewartson 1974, Eltayeb 1977). The solution is obtained by expanding the dependent variables in terms of ϵ . Thus

$$\eta = \sum_{n=0}^{\infty} (\eta_n E^n + \bar{\eta}_n E^{-n}),$$

$$\phi = \sum_{n=0}^{\infty} (\phi_n E^n + \bar{\phi}_n E^{-n}), \quad (5.9)$$

where the overbar denotes the complex conjugate, and the coefficients η_n and ϕ_n are further expanded in terms of ϵ

$$\begin{aligned}\eta_n &= \sum_{j=n}^{\infty} \epsilon^j \eta_{nj} \\ \phi_n &= \sum_{j=n}^{\infty} \epsilon^j \phi_{nj}\end{aligned}\tag{5.10}$$

while E is defined by

$$E = \exp ik(x - U_b t).\tag{5.11}$$

The multiple scale method defines the new independent variables

$$\xi = \epsilon(x - U_g t), \quad \zeta = \epsilon y, \quad \tau = \epsilon^2 t\tag{5.12}$$

so that the amplitudes η_{nj} are functions of ξ, ζ and τ only while ϕ_{nj} depend on ξ, ζ, τ and z . Note that $\phi_{00} = \eta_{00} = 0$. In (5.12) the group velocity U_g is given by

$$U_g = \partial(kU_b) / \partial k = U_b + U_b^2 [\sigma - k(1 - \sigma^2)] / \sigma\lambda\tag{5.13}$$

where

$$\sigma = \tanh(k).\tag{5.14}$$

We now substitute the expansions (5.9) and (5.10), using (5.11) and (5.12) into (5.2)-(5.6) and equate the coefficients of $\epsilon^n \epsilon^j$ ($n = 0, 1, 2, \dots$; $j = 1, 2, 3, \dots$) to zero in every equation. As a result, we obtain a hierarchy of systems of equations, each system belonging to a pair (n, j) . These systems are solved in a series. For our purposes here, we find it sufficient to consider $n = 0, 1, 2$ and $j = 1, 2, 3$ only in order to determine the equations governing the nonlinear growth of the leading order terms in (5.9) and (5.10).

It is found convenient to consider the two equations (5.2) and (5.3) first and then proceed to consider the remaining

boundary conditions (5.4) and (5.5). Now the problems posed by (5.2) and (5.3) can be written in compact form as

$$\frac{\partial^2 \phi_{nj}}{\partial z^2} = \gamma_{nj} \quad \text{for } 0 \leq z \leq 1 \quad (5.15a)$$

$$\frac{\partial \phi_{nj}}{\partial z} = 0 \quad \text{at } z = 1 \quad (5.15b)$$

in which, for our purposes here,

$$\begin{aligned} \gamma_{01} &= \gamma_{02} = \gamma_{11} = \gamma_{22} = 0, \\ \gamma_{03} &= - \left(\frac{\partial \phi_{01}}{\partial \xi^2} + \frac{\partial \phi_{01}}{\partial \xi^2} \right), \quad \gamma_{12} = -2ik \frac{\partial \phi_{11}}{\partial \xi}, \\ \gamma_{13} &= -2ik \frac{\partial \phi_{12}}{\partial \xi} - \left(\frac{\partial^2 \phi_{11}}{\partial \xi^2} + \frac{\partial^2 \phi_{11}}{\partial \xi^2} \right). \end{aligned} \quad (5.15c)$$

The solutions of (5.15) are straightforward. We find

$$\phi_{01} = \phi_{01}(\xi, \zeta, \tau) \quad (5.16)$$

$$\phi_{02} = \phi_{02}(\xi, \zeta, \tau) \quad (5.17)$$

$$\phi_{03} = (1 - z) \frac{\partial^2 \phi_{01}}{\partial \xi^2} + \frac{\partial^2 \phi_{01}}{\partial \xi^2} \quad (5.18)$$

$$\phi_{11} = A \frac{\cos h[k(1 - z)]}{\cos h(k)} \quad (5.19)$$

$$\begin{aligned} \phi_{12} &= D \frac{\cos h[k(1 - z)]}{\cos h(k)} \\ &+ i \frac{\partial A}{\partial \xi} \frac{(z - 1) \sin h[k(1 - z)]}{\cos h(k)} \end{aligned} \quad (5.20)$$

$$\begin{aligned} \phi_{13} &= \tau \frac{\cos h[k(1 - z)]}{\cos h(k)} \\ &- \left(i \frac{\partial D}{\partial \xi} + \frac{1}{2k} \frac{\partial^2 A}{\partial \xi^2} \right) \frac{(1 - z) \sin h[k(1 - z)]}{\cos h(k)} \end{aligned}$$

$$-\frac{1}{2} \frac{\partial^2 A}{\partial \xi^2} \frac{(1-z)^2 \cosh[k(1-z)]}{\cosh(k)}, \quad (5.21)$$

$$\phi_{22} = F \frac{\cosh[2k(1-z)]}{\cosh(2k)} \quad (5.22)$$

Here A , D , γ and F are functions of ξ , ζ , τ only.

The next step is to consider the problems posed by the two boundary conditions (5.4) and (5.5) for $n = 0, 1, 2$ and $j = 1, 2, 3$. First we need to expand T as in (5.6), making use of (5.9).

$$T = (1 - U_c)^n + \alpha_1 \frac{\partial \phi}{\partial x} + \alpha_2 \left(\frac{\partial \phi}{\partial x} \right)^2 + \alpha_3 \left(\frac{\partial \phi}{\partial x} \right)^3 + 0 \left[\left(\frac{\partial \phi}{\partial x} \right)^4 \right] \quad (5.23)$$

in which ϕ is evaluated at $z = \eta$, $x = x - 2\pi/k$, and

$$\begin{aligned} \alpha_1 &= n(1 - U_c)^{n-1}, \quad \alpha_2 = \frac{1}{2} n(n-1)(1 - U_c)^{n-2}, \\ \alpha_3 &= \frac{1}{6} n(n-1)(n-2)(1 - U_c)^{n-3} \end{aligned} \quad (5.24)$$

We then expand $\partial \phi / \partial x$ here about $z = 0$ and substitute in (5.5). The various systems are best treated successively. We find that $n = 0$, $j = 1$ is automatically satisfied by (5.16).

$$n = 0, \quad j = 2$$

$$\frac{\partial \eta_{01}}{\partial \xi} = 0,$$

Thus

$$\eta_{01} = \eta_{01}(\zeta, \tau)$$

$$n = 1, \quad j = 1 \quad (5.25)$$

$$\left. \frac{\partial \phi_{11}}{\partial z} \right|_{z=0} + ikU_b \eta_{11} - ik\eta_{11} = 0,$$

$$\left. -ikU_b \eta_{11} - \alpha_1 k^2 \phi_{11} \right|_{z=0} = 0, \quad (5.26)$$

Thus two equations give (5.8) and

$$\eta_{11} = \frac{i_1 k}{U_b} A \quad (5.27)$$

$$n = 1, j = 2$$

Here

$$\left. -ikU_b \eta_{12} = U_g \frac{\partial \eta_{11}}{\partial \xi} + \left[2\alpha_1 H k^2 \eta_{01} \frac{\partial \phi_{11}}{\partial z} - 2ik\alpha_1 H \frac{\partial \phi_{11}}{\partial \xi} + \alpha_1 k^2 H \phi_{12} \right] \right|_{z=0} \quad (5.28)$$

$$\left. -ik(U_b - 1)\eta_{12} = U_g \frac{\partial \eta_{11}}{\partial \xi} + \left[2\eta_{01} \frac{\partial \eta_{11}}{\partial z^2} \frac{\partial \eta_{11}}{\partial \xi} + \frac{\partial \phi_{12}}{\partial z} \right] \right|_{z=0} \quad (5.29)$$

By using (5.19) and (5.20), these two equations reduce to

$$\eta_{01} = 0, U_b, \eta_{12} = ik\alpha_1 D = \frac{\alpha_1 kH}{\sigma + \alpha_1 kH}$$

$$(\alpha_1 H + 1 + \alpha_1 \sigma kH) \frac{\partial A}{\partial \xi}$$

$$n = 0, j = 3 \quad (5.30)$$

The two relevant equations here are

$$\begin{aligned}
 -2U_g \frac{\partial \eta_{02}}{\partial \xi} + \left\{ 2\alpha_{1H} \frac{\partial^2 \phi_{01}}{\partial \xi^2} + i\alpha_{1Hk} \left(\bar{\eta}_{11} \frac{\partial^2 \phi_{11}}{\partial z \partial \xi} \right. \right. \\
 \left. \left. - \eta_{11} \frac{\partial^2 \bar{\phi}_{11}}{\partial z \partial \xi} + \frac{\partial \phi_{11}}{\partial z} \frac{\partial \bar{\eta}_{11}}{\partial \xi} \right. \right. \\
 \left. \left. - \frac{\partial \bar{\phi}_{11}}{\partial z} \frac{\partial \eta_{11}}{\partial \xi} \right) + k^2 \alpha_{2H} \left(\bar{\phi}_{11} \frac{\partial \bar{\phi}_{11}}{\partial \xi} \right. \right. \\
 \left. \left. + \phi_{11} \frac{\partial \phi_{11}}{\partial \xi} \right) \right\} \Big|_{z=0} = 0 \quad (5.31)
 \end{aligned}$$

$$\begin{aligned}
 2(U_g - 1) \frac{\partial \eta_{02}}{\partial \xi} + \left[2 \frac{\partial \phi_{03}}{\partial z} - ik \left(-\eta_{11} \frac{\partial \bar{\phi}_{11}}{\partial \xi} \right. \right. \\
 \left. \left. + \bar{\eta}_{11} \frac{\partial \phi_{11}}{\partial \xi} + \phi_{11} \frac{\partial \bar{\eta}_{11}}{\partial \xi} \right. \right. \\
 \left. \left. - \bar{\phi}_{11} \frac{\partial \eta_{11}}{\partial \xi} \right) \right] \Big|_{z=0} = 0 \quad (5.32)
 \end{aligned}$$

The utilization of (5.18) and (5.19) reduces these to

$$\begin{aligned}
 -2U_g \frac{\partial \eta_{02}}{\partial \xi} + 2 \alpha_{1H} \frac{\partial^2 \phi_{01}}{\partial \xi^2} + Hk^2 \\
 \left(\alpha_2 - \frac{2\alpha_{1k}^2 \sigma}{U_b} \right) \frac{\partial}{\partial \xi} |A|^2 = 0 \quad (5.33)
 \end{aligned}$$

$$(U_g - 1) \frac{\partial \eta_{02}}{\partial \xi} + \frac{\partial^2 \phi_{01}}{\partial \xi^2} + \frac{\partial^2 \phi_{01}}{\partial \xi^2}$$

$$-\frac{\alpha_1 H k^2}{U_b} \frac{\partial}{\partial \xi} \left| A \right|^2 = 0$$

$$n = 1, j = 2 \quad (5.34)$$

Here we get

$$2ik(U_b - 1)\eta_{22} + \left[\frac{\partial \phi_{22}}{\partial z} + \eta_{11} \frac{\partial^2 \phi_{11}}{\partial z^2} + k^2 \eta_{11} \phi_{11} \right] \Big|_{z=0} = 0 \quad (5.35)$$

$$-2ikU_b \eta_{22} + \left[-4\alpha_1 H k^2 \phi_{22} - 2\alpha_1 H k^2 \eta_{11} \frac{\partial \phi_{11}}{\partial z} - ik^3 H \alpha_2 \phi_{11}^2 \right] \Big|_{z=0} = 0 \quad (5.36)$$

By substituting for ϕ_{11} from (5.19) and ϕ_{22} from (5.22) we obtain expressions for F and η_{22} .

$$F = \frac{ik^2 [H^2 \frac{1}{2} k \sigma (U_b - 1) + \alpha_1 U_b H - 1/2 H \alpha_2 U_b (U_b - 1)] A^2}{U_b [2k_1 H (U_b - 1) + \sigma_2 U_b]} \quad (5.37)$$

$$U_b \eta_{22} = 2ik\alpha_1 H F + k_2 \left[\frac{\alpha_1^2 k \sigma H^2}{U_b} - 1/2 \alpha_2 H A^2 \right] \quad (5.38)$$

where

$$\sigma_2 = \tanh(2k)$$

$$n = 1, j = 3 \quad (5.39)$$

The relevant equations here are

$$\begin{aligned}
 ik(U_b - 1)\eta_{13} + (U_g - 1) \frac{\partial \eta_{12}}{\partial \xi} - \frac{\partial \eta_{11}}{\partial \lambda} = - \left[\frac{\partial \phi_{13}}{\partial z} - 2ik\eta_{11} \frac{\partial \phi_{01}}{\partial \xi} \right. \\
 \left. + 2k^2 \eta_{02} \phi_{11} + k^2 (2\bar{\eta}_{11} \phi_{22} - \bar{\phi}_{11} \eta_{22}) \right. \\
 \left. + \frac{3k^2}{2} \eta_{11}^2 \frac{\partial \phi_{11}}{\partial z} + k^2 \left| \eta_{11} \right|^2 \frac{\partial \phi_{11}}{\partial z} \right]_{z=0} \quad (5.40)
 \end{aligned}$$

and

$$\begin{aligned}
 ikU_b \eta_{13} U_g \frac{\partial \eta_{12}}{\partial \xi} - \frac{\eta_{11}}{\partial \tau} = -H \left\{ -\alpha_1 k^2 \eta_{13} + \alpha_1 \left(\frac{\partial^2 \phi_{11}}{\partial \xi^2} \right. \right. \\
 \left. \left. + 2ik \frac{\partial \phi_{12}}{\partial \xi} \right) + 2\alpha_1 k^2 \left(\eta_{22} \frac{\partial \phi_{11}}{\partial z} + \bar{\eta}_{11} \frac{\partial \phi_{22}}{\partial z} \right. \right. \\
 \left. \left. - 2k^2 \alpha_1 \eta_{02} \frac{\partial \phi_{11}}{\partial z} - \alpha_1 k^2 \left(4\bar{\eta}_{11} \frac{\partial \phi_{22}}{\partial z} + \eta_{22} \frac{\partial \phi_{11}}{\partial z} \right) \right. \right. \\
 \left. \left. + 2\alpha_2 k^2 \left(-\phi_{11} \frac{\partial \phi_{01}}{\partial \xi} + ik\phi_{11}\phi_{22} \right. \right. \right. \\
 \left. \left. - \alpha_1 k^4 \left(-1/2 \eta_{11}^2 \phi_{11} + \left| \eta_{11} \right|^2 \phi_{11} \right. \right. \right. \\
 \left. \left. + ik^3 \alpha_2 \left(1/2 \eta_{11} \phi_{11} \frac{\partial \phi_{11}}{\partial z} \right) + \eta_{11} \bar{\phi}_{11} \frac{\partial \phi_{11}}{\partial z} \right. \right. \\
 \left. \left. - \eta_{11} \phi_{11} \frac{\partial \phi_{11}}{\partial z} - \alpha_3 k^4 \phi_{11} \bar{\phi}_{11} \right\} z = 0 \quad (5.41)
 \end{aligned}$$

We now write equation (5.40) and (5.41) in the form, after using (5.25) through (5.39),

$$\frac{i}{\sigma + \alpha_1 k H} \eta_{13} + \lambda = H_1 / \sigma k$$

$$\frac{1}{\sigma + \alpha_1 k H} \eta_{13} + \lambda = H_2 / \alpha_1 H k^2 \quad (5.42)$$

It is immediately clear that these equations are consistent only if

$$\alpha_1 H k^2 H_1 - \sigma k H_2 = 0 \quad (5.43)$$

where

$$\begin{aligned} H_1 = & i(\sigma + k) \frac{\partial D}{\partial \xi} + 1/2 (2 + \sigma k) \frac{\partial^2 A}{\partial \xi^2} + \frac{(U_g - 1)}{U_b} \alpha_1 H k \left\{ i \frac{\partial D}{\partial \xi} \right. \\ & + \frac{(\alpha_1 H + 1 + \alpha_1 H \sigma k)}{\sigma + \alpha_1 H k} \frac{\partial^2 A}{\partial \xi^2} \left. \right\} - \frac{i \alpha_1 H k}{U_b} \frac{\partial A}{\partial \tau} + \frac{2 \alpha_1 H k^2 A}{U_b} \frac{\phi_{01}}{\partial \xi} \\ & + 2 k^2 A \eta_{02} - \frac{2 i \alpha_1 H k^3}{U_b} \frac{F}{A^2} - k^2 \frac{(\eta_{22})}{A^2} \left| A \right| 2A \\ & + 1/2 \frac{\alpha_1^2 H^2 K^5 \sigma}{U_b^2} \left| A \right| 2A + \frac{1}{2K} [\sigma + k(1 - \sigma^2)] \frac{\partial^2 A}{\partial \xi^2} \\ H_2 = & i \alpha_1 H k^2 \frac{\partial D}{\partial \xi} + \frac{\alpha_1 H k^2}{2} \frac{\partial^2 A}{\partial \xi^2} - \frac{U_g \alpha}{U_b} 1 H k \left\{ i \frac{\partial D}{\partial \xi} \right. \\ & + \frac{(\alpha_1 H + 1 + 1 H \sigma k)}{\sigma + i \alpha_1 H k} \frac{\partial^2 A}{\partial \xi^2} \left. \right\} + \frac{i \alpha_1 H k}{U_b} \frac{\partial A}{\partial \tau} \\ & + 2 i k \alpha_1 H \left\{ \frac{\partial D}{\partial \xi} - i \sigma \frac{\partial^2 A}{\partial \xi^2} \right\} + \alpha_1 H \frac{\partial^2 A}{\partial \xi^2} \\ & - 1 \bar{H} \sigma k^3 \frac{(\eta_{22})}{A^2} \left| A \right| 2A + 2 k^3 \sigma \alpha_1 H \eta_{02} A \\ & - \frac{4 i \alpha_1^2 H^2 k^2 \sigma_2}{U_b} \left(\frac{F}{A^2} \right) \left| A \right| 2A + 2 i \alpha_2 H k^3 \left(\frac{F}{A^2} \right) \left| A \right| 2A \end{aligned}$$

$$\begin{aligned}
& - 2\alpha_2 H k^2 A \frac{\partial \phi_{01}}{\partial \xi} - \frac{3}{2} \frac{\alpha_1 H k}{U_b^2} \left| A \right|^2 A \\
& + \frac{5}{2} \frac{\alpha_1 H \sigma k^5 \alpha_2 H}{U_b} \left| A \right|^2 A - \alpha_3 H k^4 \left| A \right|^2 A
\end{aligned} \quad (5.44)$$

The next step is to substitute for H_1 and H_2 into (5.43) and, after some manipulations, we obtain

$$\begin{aligned}
& i a_1 \frac{\partial A}{\partial \tau} + a_2 \frac{\partial^2 A}{\partial \xi^2} + a_3 \frac{\partial^2 A}{\partial \xi^2} + a_4 A \frac{\partial \phi_{01}}{\partial \xi} \\
& + a_5 \eta_{02} A + a_6 A \left| A \right|^2 = 0
\end{aligned} \quad (5.45)$$

where

$$\begin{aligned}
a_1 &= -(\alpha_1 k H + \sigma)^2, \\
a_2 &= U_g (\alpha_1 H k - \sigma) (\alpha_1 H + 1 + \alpha_1 H \sigma k) \\
&\quad - \alpha_1^2 H k (1 + \sigma k) - \alpha_1 H \sigma (1 + 2\sigma k), \\
a_3 &= 1/2 \alpha_1 H [\sigma + k(1 - \sigma^2)], \\
a_4 &= 2k^2 \left(\alpha_2 H \sigma + \frac{\alpha_1 H k}{U_b} \right), \\
a_5 &= 2\alpha_1 H k^3 (1 - \sigma^2), \\
a_6 &= \frac{k^5 H}{\sigma^3} \left\{ - \frac{\alpha_1 \alpha_2 H}{U_b} (3\sigma^4 - \sigma^2 + 2) \right. \\
&\quad \left. - \frac{\alpha_1 H k (\sigma^4 - 5\sigma^2 + 2)}{U_b^2} \right\}
\end{aligned}$$

$$\left. - \frac{\alpha_2^2 \sigma^2 (1 + \sigma^2)}{2\alpha_1 k} + \frac{\sigma^4 \alpha_3}{k} \right\} \quad (5.46)$$

The problem is now reduced to the evaluation of the amplitudes A , η_{01} and η_{02} which are governed by the three coupled equations (5.33), (5.34), and (5.45). The only known solution for these equations, in their entirety, is the uniform Stokes train in which

$$A = A_0 \exp(i p \tau) \quad (5.47)$$

where the frequency p is given by

$$p = - \left[a_6 + \frac{5K^2 H}{2U_g} \left(\alpha_2 - \frac{2\alpha_1^2 k^2}{U_b} \right) \right] A_0^2 - a_5 Q_0 \quad (5.48)$$

in which Q_0 is the constant of integration in (5.33). This simple solution, however, does not seem to be explicable in terms of existing experimental and observational data. Since complete solutions of partial differential equations are strongly dependent on boundary conditions, it is anticipated that other solutions may exist. It is evident that other solutions to the three coupled equations can only be obtained by considerable numerical computations.

If we impose the boundary conditions

$$\left| A \right| \rightarrow 0, \quad \frac{\partial \phi_{01}}{\partial \xi}, \quad \frac{\partial \phi_{01}}{\partial \xi} \quad 0 \text{ as } \xi^2 + \xi^2 \rightarrow \infty \quad (5.49)$$

relevant to localized solutions, we can integrate (5.33) to get

$$\eta_{02} = \frac{\alpha_1 H}{U_g} \frac{\partial \phi_{01}}{\partial \xi} + \frac{H k^2}{2U_g} \left(\alpha_2 \right.$$

$$- \frac{2\alpha_1 k^2 \sigma}{U_b} \Big| A \Big|^2 \quad (5.50)$$

The substitution for η_{02} into (5.34) and (5.45) leads to the two coupled equations

$$\begin{aligned} & \left[1 + \frac{\alpha_1 H}{U_g} (U_g - 1) \frac{\partial^2 \phi_{01}}{\partial \xi^2} + \frac{\partial^2 \phi_{01}}{2} \right. \\ & \quad + H k^2 \left[\frac{U_g - 1}{U_g} \left(\frac{\alpha_2}{2} - \frac{2_1 k^2 \sigma}{U_b} \right) \right. \\ & \quad \left. \left. - \frac{\alpha_1}{U_b} \right] \frac{\partial |A|^2}{\partial \xi} = 0 \right] \quad (5.51) \end{aligned}$$

$$\begin{aligned} & i a_1 \frac{\partial A}{\partial \tau} + a_2 \frac{\partial^2 A}{\partial \xi^2} + a_3 \frac{\partial^2 A}{\partial \xi^2} \\ & \quad + \left(a_4 + \frac{\alpha_1 H}{U_g} a_5 \right) \frac{A \partial \phi_{01}}{\partial \xi} \\ & \quad + \left[a_6 \frac{H k^2}{2 U_g} \left(\alpha_2 \frac{2_1 k^2 \sigma}{U_b} \right) \right] A |A|^2 \\ & = 0 \quad (5.52) \end{aligned}$$

It should be noted that the simple solution (5.47) does not comply with the boundary condition (4.49).

Furthermore, if we restrict our attention to the two-dimensional case, in which A , ϕ_{01} and η_{02} are independent of ζ , equation (5.51) can be integrated to give $\partial \phi_{01} / \partial \xi$ in terms of $|A|^2$. The substitution of the result in (5.52) leads to the equation

$$i \frac{\partial A}{\partial \tau} + \frac{\partial^2 A}{\partial \xi^2} + \nu |A|^2 A = 0 \quad (5.53)$$

where

$$\mu = \frac{a_2}{a_1}$$

$$\nu = a_1^{-1} \left[a_6 + \frac{Hk^2}{2U_g} \left(\alpha_2 - \frac{2\alpha_2 k^2 \sigma}{U_b} \right) - \alpha_7 \left(a_4 + \frac{\alpha_1 H}{U_g} a_5 \right) \right]$$

and

$$a_7 = Hk^2 \left[\frac{U_g - 1}{U_g} \left(\frac{\alpha_2}{2} - \frac{a_1^2 k^2 \sigma}{U_b} \right) - \frac{\alpha_1}{U_b} \right] \quad (5.54)$$

The two coupled equations (5.51) and (5.52) and the nonlinear Schrodinger equation (5.53) have the same form as those derived for surface gravity waves by Davey and Stewartson (1974) and Hasimoto and Ono (1972), respectively. The coefficients of the present equations however, are very different from those derived for surface waves in that they are dependent on the characteristics of sand grains which enter the equations through the transport law (5.6). It should be emphasized that if a different transport law is adopted, the coefficients will be different, and it remains to be seen how sensitive the present results are to the variety of transport laws that exist in the literature.

The solutions of the equation (5.53) have been investigated in some detail by Hasimoto and Ono (1972). It is known that the equation possesses a plane wave solution.

$$A = A_0 \exp [i(\beta\tau - k\xi)] \quad (5.55)$$

where A_0 is constant and $\beta = -\mu k^2 + \nu |A_0|^2$

When $k = 0$, (5.55) reproduces the uniform Stokes train

$$A = A_0 e^{i\beta\tau}$$

In addition to this simple solution, the equation (5.53) has equilibrium solution

$$A = A(\xi) e^{i\alpha\tau} \quad (5.57)$$

where α is constant and

$$A(\xi) = (a/\gamma)^{1/2} \tanh [(-\alpha/2 \mu)^{1/2} \xi] \quad (5.58)$$

if $\mu \nu < 0$

$$A(\xi) = (2\alpha/\nu)^{1/2} \operatorname{sech} \left\{ \frac{(\alpha)^{1/2}}{\mu} \xi \right\} \quad (5.59)$$

if $\partial \nu > 0$

It was shown by Hasimoto and Ono that if $\mu \nu < 0$, the Stokes train given by (5.56) is neutrally stable against modulational disturbances and if $\partial \nu < 0$ it is unstable, they conjectured that in the region $\partial \nu > 0$, the unstable Stokes wave eventually degenerates into a solitary modulational wave with amplitude $A(\xi)$ given by (5.59) and width $(\mu/\alpha\epsilon^2)^{1/2}$. Since a soliton is believed to be stable we may conclude that in the present problem ripples or small dunes of the Stokes train type may eventually degenerate into a single dune which moves as a soliton with $(\mu/\alpha\epsilon^2)^{1/2}$.

SECTION 6: CONCLUDING REMARKS

Eltayeb and Hassan (1981) showed that the non-linear theory of the movement of sand dunes yields results in agreement with observations, when long time cumulative effects are considered. This agreement seems to suggest that the transport equation used here is adequate enough for the description of sediment transport. However, the detailed solutions of the three-dimensional problem are necessary for precise evaluation of the above transport equation, since the

correct transport law must account for the detailed evolutionary process of sand dune formation.

Another remark concerns the phase shift δ . In the linear theory adumbrated by Kennedy and extended above, the presence of δ is crucial for the presence of instability and also in determining the dominant wavelength for which growth is maximized. When δ vanishes, the linear theory predicts neutral stability. Now the equations for the evolution of the amplitude in the nonlinear theory predict dune development even if $\delta = 0$, determined by the particular cases considered. Whether $\delta = 0$ will rule out nonlinear growth in the general three-dimensional case or not, can only be determined by finding the solutions to these equations, but it seems reasonable to anticipate nonlinear growth.

REFERENCES

- Allen, J. R. L. (1974) *Earth Sci. Rev.*, 10, 262.
Bagnold, R. A. (1954) *The Physics of Blown Sand and Desert Dunes*. Methuen, London.
Davey, A. and Stewartson, K. (1974) *Proc. Roy. Soc.* 338A, 101.
Eltayeb, I. A. (1977) *Proc. Roy. Soc.*, 356A, 161.
Eltayeb, I. A. and Hassan, M. H. A. (1981) *Geophys. J.R. Astr. Soc.*, 65, 31.
Engelund, F. (1970) *J. Fluid Mech.*, 42, 225.
Engelund, F. and Fredsoe, J. (1971) *Nordic Hydrology*, 2, 93.
Fredsoe, J. (1974) *J. Fluid Mech.*, 64, 1.
Gradowczyk, M. H. (1968) *J. Fluid Mech.*, 33, 93.
Hasimoto, H. and Ono, H. (1972) *J. Phys. Soc. Japan*, 33, 805.
Kennedy, J. F. (1963) *J. Fluid Mech.*, 16, 521.
Kennedy, J. F. (1964) *J. Geophys. Res.*, 69, 1517.
Kennedy, J. F. (1969) *Ann. Rev. Fluid. Mech.*, 1, 147.
Reynolds, A. J. (1965) *J. Fluid Mech.*, 22, 113.
Reynolds, A. J. (1976) *Nordic Hydrology*, 7, 161.
Richards, K. J. (1979) *J. Fluid Mech.*, 99, 597.

TURBULENT FLOW OVER TOPOGRAPHY WITH APPLICATIONS
TO SAND-WAVE DEVELOPMENT

By

K. J. Richards

ABSTRACT

The theory of turbulent flow over topography is reviewed. By considering the relevant timescales of the problem, it is shown where particular turbulence theories are appropriate in determining the structure of the flow, and rule-of-thumb estimates are given for the change in flow velocity and turbulence intensity over the topography. A particular turbulence model is applied to the study of the development of sand-waves. Both the initial growth and the later development are considered. It is shown that a proper specification of the turbulent flow is required in order to model the growth of sand-waves qualitatively as well as quantitatively.

1. INTRODUCTION

The occurrence of periodic transverse bed forms or sand-waves on a surface composed of a mobile sediment, at a number of different length scales, is common and remarkably similar in both aeolian and aqueous environments. Their often striking two-dimensional and periodic appearance is very attractive to the mathematician, allowing a good deal of simplification to the equations put forward to explain their development.

To explain the transport of sediment over the wave we require a knowledge of the flow field above the wave. The flow

is turbulent, which has important implications on the development of the wave. As we shall show later, only with a model for the turbulent flow can we account for the occurrence of both ripples and dunes.

The presentation is by no means a comprehensive review but is very much my own point of view on the subject. Section 2 very briefly introduces certain concepts relating to the atmospheric boundary layer. For a more detailed description, the reader is referred to the excellent book of Sutton (1953) or the more recent collection of papers in Haugen (1973). The effects of topography on the atmospheric boundary layer are discussed in Section 3. What I hope to show is that changes in the flow field can be adequately described by the simple theoretical argument of equilibrium shear layers and rapidly distorted turbulent flows. The results of a particular turbulence model are applied to studying the development of a sand-wave in Section 4. Both the initial growth and the later development are considered. My introduction to the subject has been through sand-waves formed by water flows. The examples given are relevant to that environment. Where appropriate, the applicability of the models to the desert environment is discussed.

2. THE ATMOSPHERIC BOUNDARY LAYER

An obvious feature of the atmospheric boundary layer is that the flow is turbulent. The flow is governed by many different length and time scales and by heating and cooling effects. We shall idealize the situation and consider a neutrally stable atmosphere, i.e., one in which buoyancy effects are negligible, and which, in a mean sense, is steady in time. Further, in this section we will assume the flow to be horizontally homogeneous.

Because the flow is chaotic, there is a need to describe its properties by statistical quantities rather than by its instantaneous values. We can define an Eulerian mean as a time average

$$\bar{u}_i = \lim_{T \rightarrow \infty} \frac{1}{T} \int_{t_0}^{t_0 + T} u_i(n, y, z, t) dt \quad (2.1)$$

(or an ensemble average if the flow is unsteady in a mean sense) at a fixed point in space for the velocity components u_i ($i=1,2,3$). The instantaneous velocity can then be divided into its mean and fluctuating components,

$$u_i = \bar{u}_i + u_i' \quad (2.2)$$

By definition $\overline{u_1} \equiv 0$. In practice we have to limit the averaging time because of the changing state of the atmosphere. Thus, T is usually taken to be approximately 12 minutes.

The correlation between the horizontal and vertical fluctuating components of velocity, $u_1'w_1'$, can be thought of as a stress. When this quantity is positive, the upward motions are transporting positive horizontal momentum upwards. This turbulent stress, or Reynolds stress, can be shown to be much larger than the viscous stress at a sufficiently high value of the Reynolds number (except at very short length scales). We define $\tau/\rho = -\overline{u_1'w_1'}$ where ρ is the density of the fluid (the minus sign is taken because close to a surface the momentum is transported downwards).

Sufficiently close to the ground and ignoring pressure gradients, the stress can be taken to be constant,

$$\tau/\rho = u_*^2 = \text{constant}, \quad (2.3)$$

where u_* is defined as the friction velocity. If we make the further assumption that the mean velocity gradient is a function of z , τ , ρ alone, i.e.

$$\frac{dU}{dz} = f(z, \tau, \rho)$$

where U is the mean horizontal velocity component and z is taken vertically upwards, then from dimensional arguments,

$$\frac{dU}{dz} = \frac{(\tau/\rho)^{1/2}}{Kz} \equiv \frac{u_*}{Kz}. \quad (2.4)$$

The constant K is called von Karman's constant and is assumed to be a universal constant of value 0.4. Recent measurements suggest that it may be slightly lower in the atmosphere (0.35) (Haugen, 1973).

Integrating 2.4 we obtain the logarithmic velocity profile

$$U = \frac{u_*}{K} (\ln z + B). \quad (2.5)$$

Sufficiently close to the boundary viscosity this will become important as the length scale of the turbulence decreases. This viscous sublayer will be of depth $\delta \sim \nu/u_*$. The constant B will therefore be a function of u_* , ν and k_s , a typical roughness height (Schlichting, 1968). For high Reynolds numbers $u_* k_s / \nu \sim 70$ the depth of the viscous sublayer becomes com-

parable with the height of the roughness elements. The stress at the surface is now transmitted by pressure forces in the wakes of the roughness elements, and viscosity is irrelevant. The flow is said to be hydraulically rough. The drag of the surface can be characterized by a roughness length z_0 defined such that

$$U = \frac{u_*}{K} \ln \frac{z}{z_0}.$$

Typically, $z_0 = k_s/30$. Note that for typical values in the atmosphere of $u_* = 1 \text{ ms}^{-1}$, $\nu = 0.15 \text{ cm}^2\text{s}^{-1}$ and for hydraulically rough flow, $k_s \gtrsim 0.1 \text{ cm}$. For smaller roughness elements we can define an effective roughness length, $z_0 \approx \nu/9u_*$ (Monin and Yaglom, 1971, p. 289).

When the surface is composed of mobile sand grains the moving sediment can change the value of z_0 . Owen (1964) relates z_0 to the thickness of the saltating layer and shows that $z_0 \propto u_*^2/2g$. From experiments the constant of proportionality is found to be 2.07×10^{-2} . Taking $u_* = 0.1 \text{ ms}^{-1}$ this gives $z_0 = 0.1 \text{ cm}$, an order of magnitude greater than the fixed bed value.

For laminar flow, the fluid viscosity is defined as

$$\nu = (\tau/\rho)/\frac{dU}{dz}.$$

For a well-behaved fluid at constant temperature, ν is a constant.

For turbulent flow we can define an effective or 'eddy viscosity' as

$$K_M = (\tau/\rho)/\frac{dU}{dz} \quad (2.6)$$

for the constant stress layer

$$K_M = K_U z. \quad (2.7)$$

The eddy viscosity is dependent on the velocity scale u_* and a length scale which increases linearly with height above the surface. This length scale is called a mixing length, considered to be the distance over which motions remain correlated, and can be defined by

$$l = (\tau/\rho)^{1/2}/\frac{dU}{dz} \quad (2.8)$$

$$= Kz \text{ (constant stress).}$$

We would not expect this to hold true throughout the whole depth of the boundary layer. Indeed, wind tunnel studies suggest that the mixing length becomes constant after $1/5$ of the boundary layer height.

The constant stress layer model is sufficient for the first few tens of meters of the atmospheric boundary layer. For greater heights, the rotation of the earth becomes important. Far above the surface the horizontal pressure gradient is balanced by the Coriolis force

$$fU_g = -\frac{1}{\rho} \frac{\partial p}{\partial y},$$

where f is the Coriolis parameter ($= 1/2$ the vertical component of the rotation rate). The flow is said to be in geostrophic balance with the flow velocity, $(U_g, 0, 0)$ the geostrophic wind. As the ground is approached, the turbulent stresses become important and the flow direction changes with height until, at the ground, it is between $15-30^\circ$ to the left of the geostrophic wind. This turning of the wind with height is called the Ekman spiral. The height of the atmospheric boundary is considered to be approximately

$$\delta = 0.4 u_* / f.$$

Various mixing lengths have been put forward for the atmospheric boundary layer. Perhaps the most widely used is that proposed by Blackadar (1962)

$$l = Kz / (1 + Kz/\lambda)$$

with $\lambda = 2.7 \times 10^{-4} U_g / f$. Close to the surface, $l \approx Kz$ while tends to a maximum value of λ as $z \rightarrow \infty$.

For the length scales we are interested in [$0(1 \text{ km})$ or less]; the Coriolis force is unimportant dynamically and will be neglected. However, the turning of the wind with height may be important in determining flow directions.

3. TURBULENT FLOW OVER TOPOGRAPHY

How does the structure of the boundary layer change as it passes over topography. With regard to sediment transport, in particular to the bed load, we are interested in the distribution of the shear stress at the surface over the topography and how this affects the erosion/deposition of sediment. We shall be considering two-dimensional topography transverse to the flow, either in the form of an isolated hill or as a periodic train of waves of horizontal length scale L and height h . The flow will be either unbounded vertically, or have a free surface at a height $z=D$. In the atmosphere, the free surface corresponds to a strong inversion at that height.

Preliminary considerations

Dividing the velocity into its mean and fluctuating components, substituting into the Navier Stokes and continuity equations and taking the mean, we obtain the following mean horizontal and vertical momentum equations

$$U \frac{\partial U}{\partial x} + W \frac{\partial U}{\partial z} = \frac{-1}{\rho} \frac{\partial p}{\partial x} - \frac{\partial}{\partial x} \overline{u^2} - \frac{\partial}{\partial z} \overline{uw} \quad (3.1)$$

$$U \frac{\partial W}{\partial x} + W \frac{\partial W}{\partial z} = \frac{-1}{\rho} \frac{\partial p}{\partial z} - \frac{\partial}{\partial x} \overline{uw} - \frac{\partial}{\partial z} \overline{w^2} \quad (3.2)$$

$$\text{and the continuity equation } \frac{\partial U}{\partial x} + \frac{\partial W}{\partial z} = 0. \quad (3.3)$$

Note that primes have been dropped from the fluctuating quantities. This process has introduced the terms $\overline{uiuj}$ or Reynolds stresses. The viscous stresses have been ignored in comparison to these. We are faced with what is called the turbulence closure problem. In order to solve this set of equations we need to express the Reynolds stresses in terms of mean flow quantities. We can write down transport equations for each of the second-order correlations but these involve third-order correlations such as $\overline{uiujUk}$ and $\overline{pui}$. Various closure schemes, or turbulence models have been put forward to close this set of equations.

We can decide on which is the most appropriate scheme for our problem by considering how a typical eddy will change as it is advected over the topography (Figure 1). Consider an eddy starting upwind at a height z_1 . The changes in turbulence structure depend on whether the travel time T is large or small when compared with the Lagrangian time scale, or eddy turnover time scale, τ_L , as it is advected over the topography. Along those streamlines where $T \gg \tau_L$ the turbulence has had sufficient time to adjust to its surroundings and can be considered to be in local equilibrium. In this situation an eddy viscosity approach is appropriate, i.e. the Reynold stresses are related to the local velocity shear. When $T \ll \tau_L$ the turbulent kinetic energy of the eddies is dissipated relatively slowly while the turbulent eddies are being distorted by the mean flow, thus not allowing them to reach a local equilibrium. The history of the flow is important and we can apply rapid distortion theory. Taking $T = L/U(z)$ and $\tau_L =$

Figure 1
Typical eddy advected over the topography.

Figure 2
Variation of maximum and minimum values of the bed shear stress with wave slope for a sine wave predicted by the linear (—) and non-linear (---) models: $D/z_0 = 1600$, $L/D = 6.25$ (Richards, 1980).

$0.3z/U_*$ (Hunt and Weber, 1979), we find that these two time scales are approximately equal at a height $z = \ell_T$, where

$$\frac{0.3\ell_T}{u_*} = \frac{KL}{U_*\ell_n(\ell_T/z_0)}$$

or

$$\ell_T \ell_n(\ell_T/z_0) \simeq L; \quad (3.4)$$

below this height $T > \tau_L$, while above $\tau_L > T$.

Jackson and Hunt (1975) present an analytical linear theory for flow over a two-dimensional hill. They divide the flow into two regions, an inner region of height ℓ in which changes in Reynolds stresses are important for determining the mean flow, and an outer region where the pressure gradient set up by the hill is balanced by inertia forces and the perturbed flow is essentially inviscid. They find that ℓ is given by

$$\ell \ell_n(\ell/z_0) = 2K^2L, \quad (3.5)$$

note that this is of the same form as 3.4.

Combining these two results, we deduce that to calculate the changes in the mean flow or turbulent stresses close to the ground, we can apply an eddy viscosity approach. The mean flow in the outer region can be calculated using inviscid theory, while we can use rapid distortion theory to estimate the changes in turbulence structure.

Changes in flow structure

The analytical model of Jackson and Hunt (1975) already mentioned has proved very useful in providing estimates for the changes in the mean flow and shear stress close to the ground due to topography. They use a mixing length approach, i.e., the horizontal turbulent shear stress τ is related to the local velocity gradient by

$$\tau/\rho = K^2(\Delta z)^2 \frac{\partial U^2}{\partial \Delta z} \quad (3.6)$$

where Δz is the local height above the topography. They present results for the flow over the hill shape

$$h = \frac{h_0}{1+(x/L)^2} \quad (3.7)$$

Different hill shapes will result in a change in the numerical constant (see Hunt, 1980).

They find that the change in mean velocity, Δu , at the crest of the hill increases from zero at the surface to

$$\Delta u \sim \frac{h}{L} \frac{U^2(L)}{u_*^2 \ln(\ell/z_0)} \text{ at } \Delta z \sim \ell. \quad (3.8)$$

In terms of the fractional speedup ratio ΔS , defined as $\Delta u/U(\Delta z)$, which is found to be essentially constant with height, we have

$$\Delta S_{\max} \sim 2h/L. \quad (3.9)$$

The perturbation shear stress increases from zero at $\Delta z \sim \ell$, to the surface where

$$\Delta \tau \sim \rho u_*^2 \frac{(h/L)U(L)}{n(\ell/z_0)} \text{ at } \Delta z = 0. \quad (3.10)$$

Jackson and Hunt's theory is valid for $h/L \ll 1$ and within the limit $\ln(L/z_0) \gg \ln(\ell/z_0) \gg 1$. We shall be interested in cases where this is not strictly true, i.e. finite h/L and $\ln(\ell/z_0)$. We will therefore need to solve for the whole of the flow. This approach has been taken by Taylor, Gent and Keen (1976) and Deaves (1976), and applied to sand-waves by Richards and Taylor (1981). Richards and Taylor use closure assumptions based on an eddy viscosity proportional to the square root of the local value of the turbulent kinetic energy and a mixing length dependent upon the distance from the surface. Thus, the Reynolds stresses in 3.1 and 3.2 are determined by

$$-\overline{u_i u_j} + \frac{2}{3} E \delta_{ij} = K \left(\frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i} \right), \quad (3.11)$$

where

$$E = \frac{1}{2} \sum_{i=1}^3 \overline{u_i^2}$$

is the mean turbulent kinetic energy per unit mass and is determined from a separate transport equation.

The eddy viscosity K is defined by

$$K = (\lambda E)^{1/2} \ell, \quad (3.12)$$

where λ is the equilibrium value of E/u_*^2 , and ℓ is a mixing length which is a prescribed function of position. Close to the surface, ℓ is taken to be proportional to the normal distance from the surface.

The equations are solved numerically by finite difference methods (see Richards and Taylor, 1981, for details). Examples of the results for two particular waves are given in the next section.

To illustrate the effects of the inclusion of the non-linear inertial terms in the equations of motion, the maximum and minimum values of τ are plotted in Figure 2 as a function of ak (wave slope) for flow over a sine wave, $h = a \sin kx$, with a wavelength to flow depth ratio of 6.25. The results are compared with the linear theory of Richards (1980) which is essentially the same model linearized with respect to ak . For low values of ak , $\tau_{\max}$ and $\tau_{\min}$ tend to a linear function of ak as predicted by the linear model. For slopes greater than 0.1, the linear model overpredicts the effect of the wave by more than 20%. As the amplitude of the wave increases, the form drag of the wave itself becomes important; it being 25% of the total drag on the surface at $ak = 0.16$ and then increasing to 66% at $ak = 0.3$.

A recent study by Britter, Hunt and Richards (1981) has compared the linear theory of Jackson and Hunt (1975) and the non-linear numerical model described above with wind tunnel measurements. The wind tunnel measurements are taken over a rough hill of shape (3.7) with $L = 0.25m$, $h = 0.25m$, $h = 0.1m$ (max slope 0.26) in a turbulent boundary layer of depth $1m$. The increase in horizontal velocity at the crest as a function of the distance above the crest, Δz , is shown in Figure 3. Above $\Delta z/L = 1$, the two theories and experiments agree within 15%. This confirms that although this part of the flow is still within the boundary layer, the change in flow is essentially inviscid. A similar conclusion was drawn by Deaves (1980). Below this level, the two theories diverge. The inclusion of the non-linear inertial terms reduce the increase in velocity by 30%. Unfortunately, due to the relative height of the roughness elements, no flow measurements could be made in this region.

Changes in turbulence structure

From previous arguments we expect the changes in turbulence to be different close to and far above the surface of the topography. For $z \ll L$ (given by 3.4), the turbulence will be in local equilibrium and we expect all three turbulence intensity components $\sigma_i^2 = u_i^2/U^2$ to be increased in proportion to $\Delta\tau/\rho u_*^2$. The theory of Jackson and Hunt (1975) suggests that

$$\frac{\Delta\sigma_i^2}{(\sigma_i^2)} \sim 4 \frac{h}{L}, \quad (3.13)$$

where $(\sigma_i^2)_\infty$ is the upstream value.

Figure 3

Comparison of theory and wind tunnel measurements of the ratio of the mean velocity at the hill top to the undisturbed velocity $U_0(\Delta z)$ as a function of the displacement above the surface $\Delta z/l$ where l is the thickness of the inner region: $\circ$ measured values: ---- inner layer, — — — outer layer, (Jackson and Hunt); — — — numerical model, (Britter, Hunt and Richards, 1981).

Figure 4

Distortion of (a) vertical and (b) horizontal vortex elements by the mean flow.

For $z \gg \ell_T$ the effect of the change in the mean flow is to distort the eddies with there being no interaction between the distorted eddies and the mean flow or other eddies. This effect is best illustrated by considering individual vortices advected over the hill by the mean flow.

These vortices will be rotated and compressed/elongated. Over the crest of the hill, according to potential flow theory, there will be zero rotation (Figure 4). Thus, vertical vortices will be compressed only. Since they preserve angular momentum this implies a decrease in σu . Likewise, horizontal vortices will be stretched, implying an increase in w . Quantitatively, applying the results of Batchelor and Proudman (1954), we obtain

$$\frac{\Delta \sigma u^2}{(\sigma u^2)_\infty} = \frac{-4}{5} \frac{u(x, z)}{U_0(zs)} \quad (3.14)$$

$$\frac{\Delta \sigma w^2}{(\sigma w^2)_\infty} = \frac{-4}{5} \frac{\Delta u(x, z)}{U_0(zs)},$$

where the changes are considered along the streamline starting at a height z_s .

Brittar, Hunt and Richards (1981) measured the change in the streamwise turbulence intensity. Again, due to the relative heights of the roughness elements, no measurements were made for $z < \ell$. For $z > \ell$ they measured a decrease in σu^2 above the hill crest. This is compared with (3.14) using the measured Δu in Figure 5 showing good agreement. For comparison, the result of an eddy viscosity model is shown in Figure 5. This shows a different trend from the measurements. Bouwmeester (1978) has also measured a decrease in σu^2 over a bridge in a wind tunnel.

Field measurements were taken by Bradley (1980) on Black Mountain, Australia, which can be approximated by (3.7) with $h = 170$ m, $L = 250$ m (maximum slope 0.6). Close to the hill he measured an increase in all three components of i with $\Delta \sigma i^2 / \sigma^2 \sim 3.0$. This is compared with an estimate of 2.5 from (3.13). Away from the surface at $z = 87$ m, although σu was increased by $\Delta \sigma u^2 / \sigma u = 0.2$, the relative increase in σw , $\Delta \sigma w^2 / \sigma w^2 = 1.25$ was significantly greater, thus supporting the qualitative features of the theory.

4. APPLICATION TO SEDIMENT TRANSPORT

Having obtained an estimate for the flow field over a given topography we can calculate the sediment transported by the flow (assuming of course that the sediment is transported passively - a large suspended load will stratify the flow and introduce buoyancy forces). We will consider the effects of

Figure 5

Comparison of theory and wind tunnel measurements of the changes in streamwise turbulence velocity, at the hill top, normalized by the upwind values: 0 measured; — — rapid distortion theory; ---- eddy viscosity model (Britter, Hunt and Richards, 1981).

topography on the bed-load only, and in particular the development of ripples and dunes. As a definition, and as the following analysis will show it to be a natural one, we define a ripple as a transverse bed form whose wavelength is dependant solely on the local bed properties and a dune as one dependant on the flow depth. This classification is obvious for river and sea flows with a finite depth to the flow. Desert sand-waves are more difficult. In the atmosphere where there is no free surface, but where there is the possibility of a strong inversion, it is not clear whether the large-scale transverse features are 'large ripples' or dunes according to this nomenclature (see later).

Note on time scales

The models described in the last section for calculating the distribution of shear stress, etc. over the topography assume that the topography is stationary. In applying these to a deformable bed, we have to make the assumption that the time scale for the flow development is short compared to that for the growth of the bed wave. Taking the diffusive time scale of the flow and assuming the depth influenced by the topography as $O(L)$, then

$$T_f \sim \frac{UL}{(\tau/\rho)} . \quad (4.1)$$

Taking $U = 1 \text{ ms}^{-1}$, $(\tau/\rho)^{1/2} = .1U$ and $L = 20 \text{ cm}$, then $T_f \sim 20 \text{ s}$. From the linear model of Richards (1980) we can estimate the time scale for a ripple to be $T_b \sim 600 \text{ s}$, indicating that our assumption is valid. Richards (1980) shows this to be also true for dunes.

4.1 Stability of erodible beds

We shall consider a small perturbation to a plane bed

$$h = h_0 \exp i(kx - \sigma t) \quad (4.2)$$

i.e. a wave growing exponentially in time at a rate σ_i and propagating forward with a velocity σ_r/k . By finding estimates of σ_r and σ_i (which will be functions of k) we can decide on whether the wave will grow ($\sigma_i > 0$, unstable) or decay ($\sigma_i < 0$, stable) with time. We shall assume that the wave with the fastest growth rate will be most likely to form from a plane bed given a random perturbation. Other lectures at this meeting will discuss the stability problem. Here, I want to briefly describe the model of Richards (1980) which, with the inclusion of a turbulence model, differs significantly from

potential flow or constant eddy viscosity models. The reader is referred to the original paper for more detail.

We shall consider bed-load transport only and assume that it is proportional to some power of the local bed shear stress, i.e.,

$$q_b = C(\tau_b)^m, \quad (4.3)$$

where $m = 3/2$ in the theory of Bagnold (1956). It should be noted that the analysis is not dependant on the precise form of this equation.

If the bed load, q_b , is varying with x , its horizontal position along the bed, then the change in height of the bed can be found from the continuity equation

$$\frac{\partial q_b}{\partial n} = -(1-n) \frac{h}{t}, \quad (4.4)$$

where n is the porosity of the bed; i.e. if q_b is decreasing with x , deposition will take place, while erosion will occur if it is increasing. Assuming that the flow variables have the same form as the bed wave, then from (4.3) and (4.4) (noting that $\partial/\partial t \equiv -i\sigma$, $\partial/\partial x \equiv ik$),

$$\sigma = \frac{k^2 C}{(1-n)} m \frac{\tau^s}{h_0 k} \quad (4.5)$$

where $\tau_b = \tau^s \exp i(kx - t)$.

The bed is unstable when $\sigma i > 0$. This requires $\tau^s > 0$, i.e. the maximum in the shear stress and thus the transport rate must be upstream of the crest. In physical terms, the transport rate is decreasing over the crest and increasing over the trough.

We shall find later that to restrict the growth of waves to a finite wavenumber we need to consider the effect of the local bed slope on the sediment transport. This will have a stabilizing effect on the bed wave; less sediment is transported towards the crest and more away from it. Following Fredsøe (1974), we take the perturbed bed-load to be proportional to $(\tau^s - \beta \partial h / \partial x)^m$, then

$$\sigma = Ak^2 \left(\frac{\tau^s}{h_0 k} \right)^m - i\beta, \quad (4.6)$$

and we require $i/h_0 k > \beta$ for ripple growth.

Estimates for A and β can be found by applying a specific transport formula. Richards (1980) uses Bagnold's (1956) theory applied to a slope and obtains

$$A = \frac{8.5e_b}{\gamma \tan \phi} \frac{3\tau_0 - \tau_{cr}}{2\tau_0^{1/2}} \frac{1}{(1-n)} \quad (4.7)$$

$$E = \frac{1}{\tan \phi} \quad (4.8)$$

where e_b is an efficiency factor, ϕ the angle of frictional resistance, τ_{cr} the critical shear stress for sediment movement and $\gamma = (\beta s - \rho)/\rho$.

From experiments, Bagnold found that ϕ varied by

$$0.75 > \tan \phi > 0.32,$$

which gives the limits on β as

$$1.3 < \beta < 3.1. \quad (4.9)$$

Richards (1980) uses a linear model (linearized with respect to wave slope, ak) to predict the distribution of shear stress over the bed perturbation. Turbulence closure is made via (3.11) and equations (3.1) - (3.3) are solved using a shooting method.

To predict the growth rate of the bed-wave we require the imaginary part of the perturbed shear stress (equation 4.6). This is plotted against kz_0 in Figure 6 for two values of the ratio D/z_0 (D being the depth of flow). A positive value of τ_i implies that the maximum shear stress is upstream of the crest. We note that there are two maxima in each curve, the second of which is independent of the flow depth at $kz_0 = 7.0 \times 10^{-3}$. We shall treat the two maxima separately.

Ripple mode (independent of flow depth)

From (4.6) the growth rate of the bed wave is proportional to $\omega r = (kz_0)^2 / u_*^2 (\tau_i / h_0 k - \beta)$. The results for varying β are shown in Figure 7. For $\beta > 2.9$ the bed is always stable as $\omega r < 0$ for all kz_0 . For $\beta = 0$ the growth rate increases monotonically with kz_0 with no maximum value. Intermediate values give a maximum in the curve. For the range of β given by (4.9), the wavelength of the fastest growing mode is

$$50z_0 < \lambda r < 1000 z_0, \quad (4.10)$$

Figure 6
Variation of the imaginary part of the bed shear stress, τ_i with kz_0 ; — $D/z_0 = 10^5$; $D/z_0 = 10^4$, ---- (Richards, 1980).

Figure 7
Growth rate of ripple mode, ω_r , for various values of β (Richards, 1980).

with λr increasing for an increasing β . This mode is dependant on the local value of z_0 (the bed roughness) which we classify as being a ripple mode.

Typical observed values of the wavelength of ripples give

$$\lambda r = 1000d \quad (4.11)$$

Taking the fixed bed value of $z_0 \sim d/30$ gives the predicted λr , an order of magnitude too small. However, the moving sediment will increase the value of z_0 . Owen (1964) takes z_0 to be the thickness of the saltating layer which is typically $2d$. This puts the range (4.10) around the observed ripple wavelength.

Dune mode (depth dependant)

The second maximum is dependant on the flow depth. The growth rate $\omega_d = (kD)^2 / u_*^2 \tau_i / \rho k - \beta$ (now non-dimensionalized with respect to the flow depth) is shown in Figure 8 for $\beta = 2.8$. For the curves with $D/z_0 > 10^3$, the curve attains a second maximum (not shown) corresponding to the ripple mode. As the depth is decreased, the two modes move closer together with them coalescing for $D/z_0 = 10^2$. The results are compared with observations from flume, river, and sea environments in Figure 9. The two lines are

$$\lambda d = 2\pi D \quad (4.12)$$

$$\lambda d = 1.16D^{1.55} \quad (D \text{ in meters}) \quad (4.13)$$

from Allen (1970), fitted to the data for flume and sea dunes respectively. The increase in the wavelength to depth ratio exhibited by the data is also given by the model results.

We note that models employing potential flow theory (Kennedy, 1963) or a constant eddy viscosity (Engelund, 1970; Smith, 1970) for the calculation of the flow field do not exhibit this bimodal character. The single mode is dependant on the flow depth. Application of these theories to desert dunes is inappropriate unless the flow depth is bounded by an inversion.

We can estimate the growth rates, $1/\sigma_i$, and propagation speeds $C = \sigma r/k$ for the two types of waves from the model. These are typically

$$T_b = \frac{1}{\sigma_i} \sim 600s, \quad C \sim 3 \times 10^{-3} \text{ cm s}^{-1}$$

for ripple, $\lambda r = 12 \text{ cm}$.

$$T_b \sim 10 \text{ days}, \quad C = 1 \text{ m/day for dune } \lambda d = 60 \text{ m}.$$

Figure 8
Growth rate of dune mode, ωd , for various values of D/z_0 ; $\beta = 2.8$ (Richards, 1980).

Figure 9
Comparison of observations of the group mean wavelength of dunes from flume, river and marine environments (taken from Jackson, 1976) with the model's results: $\circ$ observations; o , $z_0 = 0.1$ cm; Δ , $z_0 = 0.25$ cm; $+$, $z_0 = 0.5$ cm; $\square$, $z_0 = 1.0$ cm (Richards, 1980).

Once the ripples have formed, the value of z_0 of the bed will increase. This is typically 0.5 cm for a rippled bed. The model of ripple formation still applies, so that a second ripple will form dependant upon the roughness of the first. The mid-range of 4.10 predicts the formation of a mega-ripple, which is comparable to mega-ripples observed in the sea. A hierarchy of ripple sizes can evolve until some limiting effect occurs, such as the depth of the layer.

We can apply these ideas to desert sand-waves (with some caution). The wavelength of small scale ripples is comparable to the saltation path length of the individual sand grains. Thus, our rather simple model of sediment transport may be inappropriate. However, the similarity of those features formed under air to those formed under water, where the path length of grains is much reduced, is very striking. It is therefore attractive to think that they are formed by the same mechanism. The origin of the larger scale sandwaves is more difficult to interpret. With the above theory there are two possible modes of development. Either they are formed by a succession of increasing ripple sizes and are therefore 'large ripples' or the flow depth, when they are formed, is limited by a strong inversion and their wavelength is dependant on the height of the inversion; thus there are dunes. The former would require evidence of intermediate ripple sizes (which may or may not have been eroded away) while the second requires the depth of the inversion to be relatively constant throughout the growth of the wave. The question remains open.

4.2 Flow over finite amplitude waves

The results presented in Figure 2 suggest that to model the flow over finite amplitude waves correctly, the nonlinear inertial terms should be included in the equation of motion. Richards and Taylor (1981) have applied a model similar to that of Richards (1980), but with the non-linear terms retained, to the problem of flow over sand-waves. This necessitates solving for the whole flow numerically. The bed load is then related to the local shear stress by

$$q_b = \frac{8.5eb}{\gamma g} \frac{\tau_0^{1/2}(\tau_0 - \tau_{cr})}{(\tan \phi + \tan \alpha) \cos \alpha}$$

(Bagnold, 1956, applied to a sloping boundary), where α is the local slope. We shall present their results for two waves. Figure 10 shows the model prediction for the surface shear stress, bed load rate and erosion rate for the flow depth D , over a sine wave, $z_b = a \cos 2\pi x/L$ with $L/D = 6.25$, $a/z_0 = 250$ (maximum slope 0.16) and $D/z_0 = 1600$. The maximum shear stress occurs 23° upstream of the crest. The effect of including the local bed slope, $\tan \alpha$, in the transport formula

Figure 10

Bed shear stress and bed load transport for flow over a sine wave; $D/z_0 = 1,600$, $L/D = 6.25$, $a/z_0 = 250$, $\tan \phi = 0.6$ (a) shear stress, (b) bed load transport rate, (c) erosion rate, (d) wave shape (Richards and Taylor, 1981).

is to decrease the phase lag of the maximum transport rate to 12° . This decrease in the phase lag is dependant on the value of $\tan \phi$ (here taken as $\tan \phi = 0.6$), it being greater for smaller values of $\tan \phi$. The erosion rate shows maximum erosion occurring approximately 34° upstream of the crest and maximum deposition about 22° downstream of the crest. This would lead to a forward translation of the wave with the wave developing a steeper lee slope and growing in amplitude.

The results for the sediment transport over a periodic asymmetric wave (more applicable to observed sand-waves) are shown in Figure 11. This is when $D/z_0 = 1600$, $L/D = 6.25$ and $a/z_0 = 200$ (maximum slope 0.16), a case where the flow only just remains attached. With $\tan \phi = 0.6$ the bed load maximum is 2° upstream of the crest. The resulting erosion rate curve shows deposition occurring at the crest and on the lee slope, with erosion on the stoss slope. The wave will become sharper crested and the lee slope steeper with the crest height increasing and moving downstream, with a consequent separation of the flow. The erosion rate curve with $\tan \phi = 0.3$ is also shown in Figure 11. The bed load is 5° downstream of the crest, implying that the crest of the wave will be eroded and the lee slope will become shallower. The wave shape will be stable for a value of $\tan \phi$, somewhere between these two values.

Richards and Taylor (1981) compare their model with flow measurements taken over sand-waves in the Columbia River reported by McLean (1976). They chose a non-separating and a separating case. The comparison between the predicted and the measured friction velocity for the non-separating case (maximum slope 0.12) is shown in Figure 12. Both the magnitude of the variation in $\tau^{1/2}$ and the phase relative to the bed wave are in good agreement.

For the wave to propagate forward without a change in slope with a constant velocity c , we require

$$z_b = z_b(x-ct).$$

From the sediment continuity equation (4.4), this requires the erosion rate to be proportional to the local gradient of the bed. Figure 13 shows a good correspondence between the phases and the ratio of the maxima and minima of σ and dz_b/dx . This implies that the wave is progressive (note that the flow is not separating) and the propagation speed is calculated to be 1-5 m/day. This compares with a reported 5m/day for the sand-wave field as a whole.

Finally, the comparison between the predicted and measured $\tau^{1/2}$ for the separating flow case (maximum slope 0.36) is shown in Figure 14. The negative value of $\tau^{1/2}$ indicates flow separation. Again the agreement is good with the position and length of the separation bubble being similar in both cases.

Figure 11

Bed-load transport over asymmetric wave; $D/z_0 = 1600$, $L/D = 6.25$, $a/z_0 = 200$, $\tan \phi = 0.6$

(a) Bed-load transport rate,

(b) Erosion rate (dashed line, $\tan \phi = 0.3$)

(c) Wave shape (Richards and Taylor, 1981)

Figure 12
Friction velocity $(\tau)^{1/2}$ for non-separating flow case (McLean, 1976): ---- measured, — model prediction (Richards and Taylor, 1981).

Figure 13
Erosion rate predicted by the model and bed wave slope: — erosion rate, ---- dz_b/dx (Richards and Taylor, 1981).

Figure 14
Friction velocity $(\tau)^{1/2}$ for separating flow case (McLean, 1976): --- measured, — model prediction (Richards and Taylor, 1981).

The lee slope is close to the angle of repose of the sand grains and therefore the sediment transport model used is inappropriate. However, the stress distribution suggests a slight flattening of the wave just above the lee slope and a forward movement of the wave.

The above presentation is aimed at showing how rather idealized models can be used to illuminate the dynamics of sand-waves, their development and propagation rates. Applying these results to the real world is difficult, particularly when the flow is strongly separating and the waves have avalanche lee slopes. However, the results are encouraging and it is hoped that they will prompt more detailed measurements of the flow over sand-waves in the future.

REFERENCES

- Allen, J.R.L. (1970) *Physical Processes of Sedimentation*. Elsevier.
- Bagnold, R.A. (1956) The Flow of Cohesionless Grains in Fluids. *Phil. Trans. Roy. Soc. (London) A*, 249, p. 235-297.
- Batchelor, G.K. and Proudman, I. (1954) The Effect of Rapid Distortion of a Fluid in Turbulent Motion. *Quart. J. Mech. Appl. Math.* 7, p. 83-103.
- Blackadar, A.K. (1962) The Vertical Distribution of Wind and Turbulent Exchange in a Neutral Atmosphere. *J. Geophys. Res.* 67, p. 3095-3102.
- Bouwmeester, R.J.B. (1978) *Wind Characteristics Over Ridges*. Ph.D. Dissertation, Colorado State University.
- Bradley, E.F. (1980) An Experimental Study of the Profiles of Wind Speed, Shearing Stress and Turbulence at the Crest of a Large Hill. *Quart. J.R. Met. Soc.* 106, p. 101-124.
- Britter, R.E., Hunt, J.C.R. and Richards, K.J. (1981) Air Flow Over a Two-Dimensional Hill: Studies of Velocity Speed-Up Roughness Effects and Turbulence. *Quart. J.R. Met. Soc.* 107, p. 91-110.
- Deaves, D.M. (1976) Wind Over Hills - Numerical Approach. *J. Ind. Aerodyn.* 1, p. 371-391.
- Deaves, D.M. (1980) Computations of Wind Flow Over Two-Dimensional Hills and Embankments. *J. Wind. Eng. and Ind. Aerodyn.* 61, p. 89-111.
- Engelund, F. (1970) Instability of Erodible Beds. *J. Fluid Mech.* 42, p. 225-244.
- Fredsoe, J. (1974) On the Development of Dunes in Erodible Channels. *J. Fluid Mech.* 64, p. 1-16.
- Haugen, D.A. (Ed.) (1973) *Workshop on Micrometeorology*. American Met. Soc. Boston.

- Hunt, J.C.R. (1980) Wind Over Hills. Survey paper for NOAA: NSF Workshop, Boulder, Colorado, August 1978 (to be published by Am. Met. Soc.).
- Hunt, J.C.R. and Weber, A.H. (1979) A Lagrangian Statistical Analysis of Diffusion from a Ground-Level Source in a Turbulent Boundary Layer. *Quart. J.R. Met. Soc.* 105, p. 423-443.
- Jackson, P.S. and Hunt, J.C.R. (1975) Turbulent Wind Flow Over a Low Hill. *Quart. J.R. Met. Soc.* 101, p. 929-955.
- Jackson, R.G. (1976) Sedimentological and fluid-dynamic implications of the turbulent bursting phenomenon in geophysical flows. *J. Fluid Mech.* 77, p. 531-560.
- Kennedy, J.F. (1963) The Mechanics of Dunes and Anti-Dunes in Erodible Bed Channels. *J. Fluid Mech.* 16, p. 521-544.
- Monin, A.S. and Yaglom, A.M. (1971) Statistical Fluid Mechanics, Vol. 1 (English translation) M.I.T. Press, Cambridge, Mass.
- McLean, S.R. (1976) Mechanics of the Turbulent Boundary Layer Over Sand Waves in the Columbia River. Ph.D. thesis, Univ. Washington.
- Owen, P.R. (1964) Saltation of Uniform Grains in Air. *J. Fluid Mech.* 20, p. 225-242.
- Richards, K.J. (1980) The Formation of Ripples and Dunes on an Erodible Bed. *J. Fluid Mech.* 99, p. 597-618.
- Richards, K.J. and Taylor, P.A. (1981) A Numerical Model of Flow Over Sand Waves in Water of Finite Depth. *Geophys. J.R. Astr. Soc.* 65, p. 103-128.
- Schlichting, H. (1968) Boundary-Layer Theory, 6th ed. McGraw Hill, New York, 747 pp.
- Smith, J.D. (1970) Stability of a Sand Bed Subjected to a Shear Flow of Low Froude Number. *J. Geophys. Res.* 75, p. 5928-5940.
- Sutton, O.G. (1953) Micrometeorology. McGraw-Hill, New York.
- Taylor, P.A., Gent, P.R. and Keen, J.M. (1976) Some Numerical Solutions for Turbulent Boundary-Layer Flow Above Fixed, Rough, Wavy Surfaces. *Geophys. J.R. Astr. Soc.* 44, p. p. 177-201.

INDEX

Acacia cyclops	125
Acacia senegal	56, 58
Aeolian activity	159
Aerosols	223
Africa	5, 9, 11, 74, 76, 198, 274
Agglutinated grains	313
Al-Hassa (Saudi Arabia)	113
Alluvial channels	331
Amboy lava field	290
Andes Mountains	26
Apollo-Soyuz	88, 193
Arabian Peninsula	110
Argentina	26, 88
Asia	15
Atacama Desert	9
Australia	18, 195
Automatic weather station	152
Bandiagara plateau	210
Bed shear	386
Biological productivity	52
Blowouts	162
Boundary layer	318

Brazil	27
Bristol Mountains	292
California	96, 290
Canada	22, 25
Chemical weathering	70
Circulation models	182
Clay coatings	205
Cliff retreat	76
Climatic change	54, 74, 155
Colorado River	24
Conservation	134
Creep	265
Cyprus conglomeratus	57
Dashte Kavir (Iran)	110
Dashe Lut (Iran)	110
Denmark	59
Depressions	
Kharga (Egypt)	82, 141, 148
Turpan (China)	88
Desert	
environment	74
varnish	225
winds	141
Desertification	1, 4, 5, 6, 27, 40, 42, 52, 230

Dongola	57
Dune	
afforestation	109
bundles	88
fields	74, 88
fixation	90, 119, 123
forms	70, 72
morphology	94
profile	94
sand	96, 191
slope	104
stabilization	90, 118, 119, 123
vegetation	124
Dunes	
antidunes	327, 329, 340
circular	86
coastal	109, 118, 239
complex	80
compound	80
crescentic	82
dome	82
Echo	86
fixed	231
free	70, 86
Lee	86
linear	80, 146

Dunes (Cont.)

longitudinal	241
pyramid	82
sand	29, 417
seif	80, 146
scarp	86
simple	80
transverse	249
u-shaped	86
whaleback	146
windward	86
Dust	251
aerosols	223
atmospheric	262
devil	344, 354
production	322, 361
storms	76, 171, 344
transport	35
Ebro River	31
Eco techniques	127, 132
Egypt	118, 207
Electron microscopy	192
Equilibrium profile	98
Erosion	
rate	375
soil	37, 62, 127, 261, 370

Erosion (Cont.)	
water	35
wind	35, 67, 185, 267, 274, 361
Eucalyptus	91, 125
Evapotranspiration	53
Fachi-Bilma	226
FAO	49, 262
Flagstaff, Arizona	152
Flow	
channel	383
resistance	393
subcritical	335
turbulent	433, 437
Fluvoglacial	59
Fodder grass	116
Friction speed	300
Gandolfi camera	199
GEMS	45, 50
Ghard Abu Muharik	82, 90, 121
Gilf Kebir Plateau	143
Glacial silt	78
GOES-1 satellite	142
Grain size	94
Gravity forces	337

Great Sand Sea	72, 119, 143, 199
Groundwater	112, 126
Hematite coating	191
Ignimbrites	171
Indus Valley	17
Inselbergs	143
Kaolinite	203, 206
Kufra Desert	113
Landsat images	72
Libyan Desert	57
Mali	274
Mariner 9	167
Mars	80, 160, 165, 344
Martian negolith	206
Mauritania	234
Mesopotamia	15
Meteorological data	142
Meteorological stations	141
Meteosat	88, 150, 226
Mexico	22, 25
Mineralogical composition	201
Mojave Desert	290
Mongolian Plateau	78
Munsell standard color	193
Neolithic civilization	54

New Valley (Egypt)	119
Niger River	228, 276
Nile	
Delta	198
River	143
Valley	14
NOAA	226
Nomadic pastoralists	13
Nomads	57
North Africa	76
North America	4, 25
North Sea coast	59
Nubian Sand Sheets	119
Oases	
Farafra	119, 148
Kharga	121
Orbital photographs	195
Overgrazing	13, 18, 24, 56
Particle	
diameter	371
flux	182, 257
spherical	366
Plant communities	111
Rajasthan Desert	91

Red Sea	191
Reddening of quartz	191
Reynolds number	372
Ripples	327, 340 396
wavelength	371
River beds	327
Rub Al Khali	167
Sahara	5, 9, 74, 198, 274
Saharan dust	39
Sahel	9, 74, 212
Saline seepage	20
Salinization	13, 20, 29
Saltation	78, 370
Sand	
drift	61, 67, 86, 146
movement	62, 76, 143
seas	70, 88
shadows	86
sheets	70
stabilization	90, 118, 119, 123
waves	433
Sandblasting	253
Saudi Arabia	72, 110, 113
Sediment inertia	337
Selima Sand Sheet	78

Semiarid	19
Shear stress	329
Shelter-belts	122
Simpson Desert	196
Soil	
conservation	54
moisture	132
stabilizer	124
Sorghum vulgare	58
Sorting	265
South America	26, 29
Spain	29
Strip cropping	268
Sudan	52
Suspension	265
Taklimakan Desert	9, 72
Tamarix articulata	125
Terminal speed	371
Threshold velocity	64, 104, 322
Titan	159, 177
Topographic maps	143
Topographic obstacles	78
Transport rate	387
UNDP	49

Physics of Desertification groups together the results of investigations of the process of desertification by twenty international specialists. The book represents a thorough report on the status of our knowledge of the cause and effect of degradation of arid lands worldwide.

Emphasis is placed on the regions that are chronically affected by drought in North Africa, particularly in the Sahel. It is a timely look at the scientific basis of our knowledge of how arid terrain changed in space and time. The seeds for the book were planted four years ago during an international gathering of scientists to discuss the topic under the auspices of the United Nations' Institute of Theoretical Physics, Trieste, Italy. The selected authors for this book have updated their views and integrated results of their studies of all deserts not only on Earth but also on Mars. Several of the reported research results were sponsored by the United Nations, the FAO, NASA and the Smithsonian Institution.

The book will be of primary interest to researchers dealing with the geology, geography, morphology, environment and development of arid lands.