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ADAM LINSENBARTH

A REMOTE SENSING APPROACH TO GEOMORPHOLOGICAL INVESTIGATIONS OF SAND DESERT AREAS

LIBYAN SAHARA CASE STUDY

WARSAW 1996

INSTITUTE OF GEODESY AND CARTOGRAPHY

To my wife Barbara and daughters Anna and Elizabeth.
In memory of a very pleasant and exotic episode of our life on the fringes of the Sahara

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FOREWORD

The world's deserts in general and the sand deserts in particular constitute that part of the globe which is the least understood. Deserts are generally isolated, inaccessible and inhospitable; hence, investigations of those areas have always tended to be infrequent and limited. If semi arid regions are included, desert comprise more than one - third of the Earth's land surface, but the term *desert* is often applied only to those hot and dry regions which cover one - fifth of the land area, i.e. some 31.6 x 10 sq km. In fact, only about 20% of arid land is covered by sand which has accumulated in the so called *sand seas*.

The location of desert areas is determined largely by the dynamics of atmospheric circulation; hence, the majority of desert areas are located along two belts of subtropical high pressure cells. Clearly, sand seas play a very important role in climatic and global environmental changes. Unfortunately, but perhaps not surprisingly, there are still some very large gaps in our knowledge of sand seas geomorphology and morphogenesis and their dynamics.

The Sahara is the largest climatic desert in the world; it is virtually coterminous with the whole of the E-W oriented northern part of Africa and forms the western end of the Afro - Asian desert zone. The Libyan Sahara, which constitutes the north - eastern part of the region, belongs to the super - arid desert and it occupies more than 90% of Libyan territory. The Libyan Sahara exhibits several climatic extremes: the world's highest aridity (insolation amounts to 3825 hrs per year); the mean annual precipitation of less than 20 mm; and the highest recorded temperature (58°C, at Al-Aziziyah).

The Libyan Sahara consists of stony tabelands, gravel plains and sand seas. The latter cover more than 300 000 sq km; they constitute the least known part of the entire Sahara and contain a large variety of characteristic geomorphic landforms. The very spectacular star dunes which appear in the Idhan Awbari rise more than 160 m above the general ground level, while the huge longitudinal sandridges in Idhan Murzuq extend over distances of 135 km.

During his several missions to Libya in the 1970s and 80s, the Author recognised that the morphology of sand seas is not adequately depicted on any of the available Libyan topographic or geological maps. In respect to the former, which are available only for the northern part of the country anyway, the aeolian bedforms are represented by contour lines but these only partially reflect the sand dune forms and patterns. On the latter, the sand seas and gravel plains are merely represented as 'areas covered by Quaternary deposits'. Reconnaisance studies performed by the Author in the Libyan sand seas indicated that the landforms which give rise to the sand desert topography result from both contemporary and ancient geomorphological processes; indeed, the sand seas must have formed by interactive processes of various kind over a long period back into the past. During that time, the deserts episodically expanded and contracted according to the ever - changing conditions. At the time of maximum aridity, which was probably between 18 000 and 12 000 B.P., the Sahara was broader by between 200 and 500 km along its E-W axis; later, however, these active ergs become stabilised and were partially vegetated. The southern edge of Sahara has again expanded, by as much as 650 000 sq km in only the last 50 years. As a consequence of this, the drought which began in the Sahelian countries in 1968 was responsible for the deaths of between 100 000 and 250 000 of the individuals and the collapse of the agricultural bases of five countries.

In desert areas, human activity is both regulated by and, to some extent, superimposed on the natural processes which tend to shrink or expand the deserts. For example in many localities in the Libyan Sahara (e.g. Al Kufrah, Jalu, Awbari) artificial irrigation systems are successful in converting sand areas into cultivated fields. However, human activity may stress the ecosystem beyond its tolerance limit, resulting in rapid desertification. This, in turn, often exacerbates a general trend toward increasing aridity and may initiate major, if only local climatic changes.

Clearly, desertification processes are strongly correlated with those involved in sand desert development. The problems resulting from anthropogenic interference in desert areas continue to multiply; for example, in China, agricultural mismanagement has already resulted in the sterilisation of over 50 000 sq km from formerly productive land and it is estimated that, in the near future, an additional 170 000 sq km will be turned into desert.

A knowledge of sand desert environment and its impact on both regional and global climatic and environmental changes is obviously very important but unfortunately, is nevertheless very limited. Given, that desert areas are both difficult of access and inhospitable, the application of remote sesnsing techniques may therefore provide and extremely useful tool for investigating sand deserts. Indeed, judging from his own expirience to date, the Author would go so far as to contend that morphological surveying of the huge desert areas can already quite adequately be carried out on the basis of remotely sensed data, using the modern remote sensing technologies.

The main objective of the investigations described herein on the study area of the Libyan Sahara was to define the contribution which remote sensing might make with regard to integreted morphological studies of sand desert areas. This, in turn, relates to the morphogenesis of particular dune forms and entire sand seas. The small - and medium - resolution satellite data and remote sensing techniques allow us to carry out geomorphical studies on both a global and a regional scale. Regional scale investigations reveal the desert landforms and processes in a wider geomorphic context and, by comparative analysis, demonstrate the interactions between the particular desert landforms. The high -resolution remotely sensed satellite data contribute to local scale studies relating to the particular dune forms.

Sand seas are excellent natural laboratories in which all aspects of aeolian geomorphology and the interaction of the wind with other elements of the environment may be studied. The large areas of aeolian sand are, of course, the most distinctive of desert landforms. The Author anticipates that an analysis of contemporary geomorphic processes and a knowledge of past changes might eventually form the basis for establishing the full cycle of sand seas development; from this, it might one day be possible to predict future changes and their impact, both on a regional and a global scale.

The main part of this monograph focusses on the evaluation and analysis of the remote sensing approach to sand desert investigations; on investigations of certain dune forms which appear in the Libyan Sahara sand seas and on the geomorphological analysis of Libyan sand seas using remote sensing data. From this, the Author proposes a working hypothesis of sand seas development. In the appendices to the monograph, additional information is given regarding the geomorphic evolution of the Libyan Sahara; climatic changes; the wind regime in the Libyan Sahara; the principles of sand movement by wind and sand characteristics. There is, of course, no reason why the same data should not be useful to other investigators for further studies in this field.

1. INTRODUCTION

1.1. DESERTS - THEIR LOCATION AND CHARACTERISTICS

1.1.1. Types of deserts

Deserts are areas in which vegetation is sparse or absent, where the land surface is left exposed to the atmosphere and the associated physical forces (Mabbutt, 1979). Deserts comprise more than one - third of the Earth's land surface, if the semi - arid regions are included. The deserts of the world are those areas which normally receive less than 25 mm of precipitation per year, thus including both the cold and hot deserts, though the term desert is more commonly applied to the hot and dry regions. The *hot and arid deserts* comprise one - fifth of the land area of the Earth, i.e. over 31,000,000 sq km. To the extremely arid areas belong: the central Sahara and Namib in Africa, the Rab al-Khali in Saudi Arabia, the Takla Makan in Asia, the Atakama Desert of Peru and Chile, parts of northern Mexico and south - western United States (Fig.1). Regions of the *cold desert*, covers an area of 23,000,000 sq km which constitutes one - sixth of the landmasses of the Earth.

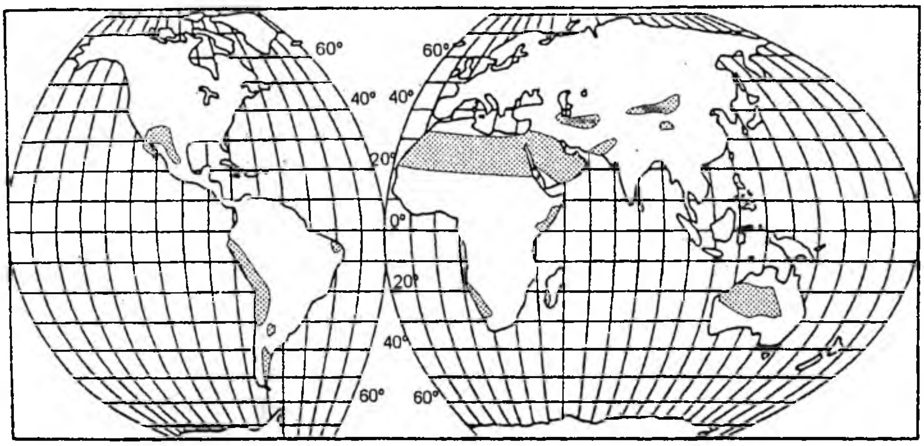


Fig.1 Global distribution of arid deserts.

1.1.2. Global distribution of deserts

The present location of desert areas results largely from the dynamics of atmospheric circulation. As is very well known, much more incident solar heat is received in the equatorial region than at both poles. This results in a poleward transfer of heat, and to a meridional atmospheric circulation pattern. The rotation of the Earth causes latitudinal wind zonation at the earth's surface giving rise to the trade winds in equatorial regions and the westerlies in the middle latitudes. The air which is heated at the Equator by incident solar radiation, rises, cools, condenses and releases its moisture content in the tropical zone and then subsides toward the Earth's surface near latitudes 30°N and 30°S. The two great belts of subtropical high pressure result from this mechanism. Most of the largest deserts are located beneath or near these high - pressure belts. The descending air is also the source of easterly winds which blow towards the equator.

Several of the desert areas can be characterised as high latitude types. They lie north or south of the principal arid belts like the North American Desert and the Patagonian Desert of Argentina. These arid areas can be attributed to an orographic effect, when air subsides in dry rain - shadows in the lee of mountain barriers. When the warm, moisture - laden winds must cross the mountain system, the air masses cool as they rise, and condensation and precipitation occurs over the mountains. The air which descends on the leeward side of the topographical barrier, is thus devoid of moisture and deserts result. Examples of such deserts are: North American Deserts located on the lee of the Sierra Nevada Mountains, and the Patagonian Desert of Argentina, the result of the orographic effect of the Andes mountains. The deserts of Central Asia, the Takla Makan and Gobi are also related to an orographic effect. Western coasts of the continents adjacent to the cold upwelling ocean currents may also be dry and deserts may form like the Namib Desert.

Generally the arid deserts are areas where the rainfall is small and irregular. The larger deserts of the world are those areas which receive less than 25 cm of precipitation per year. Arid climates can be defined as those under which potential moisture losses from evaporation and transpiration exceed incoming precipitation. The degree of aridity is defined by the indices of aridity or by the moisture indices.

1.1.3. Morpho - structural types of deserts

In respect of major geological structures and resulting gross relief, the desert areas can be divided into two morpho - structural types. There are *deserts of the shields and platforms*, and the *mountain - and - basin deserts* of tectonically more active zones (Mabbutt, 1979). Typical shield and platform deserts include broad plains on the granite rocks of the exposed shields, as in Western Australia, and tablelands and basin lowlands on subhorizontal platform strata, as in the Sahara. The plantation of the shields under savanna regime in the Early Cenozoic, was followed by stripping of weathered layers from higher parts and the extension of depositional surfaces on lower ground. Thus, gently sloping erosional plains were developed on the shields. The structural tablelands of the platforms are commonly reinforced by resistant weathering crusts. The lower parts are occupied by aeolian sands thus forming sandplains or dunefields. The shield and platform deserts correspond strongly with the extreme arid or arid climate and the homogeneity of their relief is emphasised by a stability and uniformity of the climate.

The mountain - and - basin deserts include both the mid - latitude deserts and low - latitude deserts. In general, the desert basins are essentially structural features, and were formed principally by the downward movement of large blocks of the Earth's crust between bounding uplifted blocks which are now mountain ranges. Thus, desert basins alternate with mountains ranges in such places as the western United States, Pakistan, Iran and elsewhere. Distinct from the down - warped basins such as Lake Eyre in central Australia, there are also faulted tectonic basins, such as Death Valley in California and the Northen Rift Valley in Tanzania. The mountain - and - basin deserts form areas of interior drainage and their lowlands are entirely depositional.

In addition to the morpho - structural classes, each desert region consists of a sequence of physiographic settings from upland to plain. These are made up from characteristic landform groupings and constitute areas of particular geomorphologic processes. According to Mabbutt (1979), the following sequence of physiographic settings occurs: *desert uplands, desert piedimonts and desert lowlands*.

The stony deserts which comprise the structural plateaux and stony plains are formed at a lower level. Desert rivers, in the form of ephemeral channels and floodplains, represent the riverine desert. Beyond the drainage terminals are the desert lake basins and finally, beyond the active drainage, are the sand deserts.

The desert plains are generally characterized by their vastness and extreme smoothness. Most authorities accept that the plains were formed by alluvial erosion in the distant past and have a very small gradient. A number of depressions in desert plains has been ascribed chiefly to deflation, i.e. erosion by wind action. The Qattara Depression in Egypt (- 134 m) and several basins in the Gobi, and in the Atacama are in this category. Generally, the deflation basins occur in flat areas, commonly in soft alluvium or in lacustrine sediments, but in the extreme arid areas, such as the eastern Sahara, they were found to be eroded in subhorizontal rocks (Mabbutt, 1984).

1.1.4. Sand deserts

Only about 20 percent of the arid land of the Earth is covered by sand, but there are large differences between continents. For example in Central Asia the sand deserts constitute over 45 percent of the desert surfaces; in Australia more than 30 percent; in the Sahara 10 percent and in North America, only 2 percent.

The sand seas occur in structural basins and/or in plains. In Central Asia, the largest sand seas occur in basins bounded by mountains. The Takla Makan Desert in China, located in the centre of the Tarim Basin, covers an area of 337,600 sq km (Zhenda, 1984). There are also plateau - type desert areas, like the Badain Jaran Desert (in Western Inner Mongolia), in which the compound sandridges surpass 420 m in height (Chao Sung-Chiao, 1984).

The Australian deserts mainly belong to the shield and platform morpho - structural type. The major sand deserts in Australia are: Simpson, Great Sandy and Great Victoria Deserts. The sandy deserts cover an area of 2 million sq km or nearly 40 percent of the arid zone (Mabbutt, 1984). Saudi Arabia is covered 30% by sand desert, the largest one being Rub Al' Khali, which has an area of 560,000 sq km. The deserts in Northern America are distributed over northern Mexico and western United States between 23°N and 45°N. All these deserts belong to the intermountain geomorphic region. Death Valley, within the Great Basin, constitutes the only hyper - arid zone in North America. The Great Basin is a geographic unit laying between the Rocky Mountains on the east and Sierra Nevada and Cascade Mountains on the west.

In arid regions, the main geomorphological processes, which effect landscape in general and individual landforms in particular in arid regions are: weathering, gravitational, fluvial and aeolian. In the case of the Sahara desert, fluvial processes are very limited at present due to the arid conditions. The gravitational processes are limited to the mountains areas. Hence, the most important are the weathering and aeolian processes, earlier responsible for the stony desert and later for sand desert forms and their development.

1.1.5. Desert pavement

Desert stone pavement, in which the stones are closely packed on flat or gently inclined surfaces, is most extensive in arid areas. The stony pavements range from rocky or boulder - strewn surfaces to smooth plains of the finest gravel. The rocky or bouldery surfaces are called *hamadas,* while smooth plains consisting of the finest gravel are named *regs* or *serirs*. Hamada is an Arabic name which means „unfruitful" and describes difficult bouldery terrain, and reg, also an Arabic name, means „becoming smaller" and is used for the trafficable finer pavements. The term serir is used in central and eastern Sahara. The dominant distinctions between hamada and reg are: the size of pavement material and the nature of the underlain surface. Hamada pavement consists of boulders transported only short distances, while reg pavement consists mainly of transported stones. Hamada boulders lay on the bedrock, while regs are commonly underlain by soil profiles. Hamadas are structural tablelands, associated with flat - bedded rocks or near horizontal, weathered, crusts; regs are generally alluvial plains. The term hamada is used for two types of surface, namely rock hamada, which constitutes rock outcrop, and for the boulder hamada. The hamada surface is largely determined by the mode of breakdown of the underlying rocks and their origin.

Regs, which have an alluvial origin, are characterized by sorted rounded gravels of mixed composition. Generally, regs appear near to or in the extension of ancient stream channels. Gravel concentrations in desert areas are sometimes called *lag gravels* in reference to the residue left by the removal of fine material. Such desert pavement is produced by the combined effects of wind and water. Evaporation of capillary water may cause the precipitation of calcium carbonate, gypsum and other salts, which cement the pebbles together to form a desert - conglomerate. The closely spaced pebbles act as a drag on the surface wind, thus, restricting the entraintment of finer matrials. Such pavements with large stone density and smoothness, are resistant to weathering, selective erosion and deflation. Reg pavement plays a protective role on the surface, which is reflected in the name *desert armour*. Reg surfaces belong to the most wind - stable of desert surfaces. The regs, due to the smooth character of their surface, are thereby suitable for caravan routes.

One pronounced effect of chemical weathering processes in arid regions, is *desert varnish* or *patina*. Desert varnish is a dark red to black mineral coating of iron and maganese oxides or silica, deposited on pebbles and rocks on the surface of arid regions. Oxides of iron produce red - brown coating, while maganese oxides produce dark brown to black hues. As dew and capillary water evaporate, their disolved minerals are deposited on the pebbles or rock surfaces. Wind abrasion removes the softer salts and polishes the patina to a glossy finish. The degree of darkening by varnish can be used as an indicator of stability and relative age of desert pavement, but Mabbutt (1984) claimed that comparative dating by the degree of varnishing is risky, in the view of variations between sites. Mabbutt also stated that the stones on the ground are more varnished than rock outcrops and that the process of varnishing is very slow on surfaces above the ground. The rates of patina formation vary, but it is generally considered to have required about 2,000 years for its formation.

1.1.6. Sahara

The name „Sahara" (As-Sahraal-Kubra) derives from the Arabic name *sahra*, meaning desert, and its plural *sahara*. It is also related to the adjective *ashar*, meaning desert - like and carrying a strong connotation of the reddish colour of the vegetationless plains.

The Sahara, with its axis along the Tropic of Cancer, is the largest hot desert in the world (ca 6,000,000 sq km). It fills almost all of the east - west oriented northern part of Africa and constitutes the western end of the Afro - Asian desert zone. Its main topographical features include plains at 180÷360 m above sea level, lowlands and depressions and two mountains chains: the Tibisti mountains, which rise to 3415 m in Emi Koussi, the highest point in the Sahara, and Hoggar (Ahaggar) mountains (3003 m).

The Sahara plains are covered with angular stone fragments of boulders (*hamadas*), gravel (known as *reg* in the west and *sarir* in the east) and sand (*erg*, *edeyin* or *ramlah*).

The Sahara exhibits climatic extremes: for example, it has the highest evaporation rates in the world (7 720 mm) per year at Bauroukou in Chad and the huge areas of perennial drought. Generally, the Sahara is a region where rain is extremely rare and drought is the norm. The average drought duration in the middle of the Sahara is five or six months in the west and two years in the east (Dubief, 1979). There are large differences between the northern and southern margins of the Sahara. In northern Sahara, like in the Libyan Sahara, the drought occurs mainly during the summer, while in the southern edges, where the thermic equator is situated, a long desert period alternates each year with a short rainy period. The evaporating power of the Saharan atmosphere is one of the most important factors creating a desert. The evaporating power depends mainly on the great saturation deficit of the Saharan air, and results in an exremely low relative humidity, which is about 10% in central Sahara and can be as litle as 2% in extreme cases.

The apparent migration of the sun in the Sahara region gives short winters and long summers. The Saharan ground receives great quantities of heat from the sun. The surface layer of the ground reaches the highest temperatures in the middle of the day and the lowest during the night. The heating of the ground leads during the day to a strong thermic turbulence, which disappears at the end of the day (Dubief, 1979).

The Sahara Desert contains several large sand seas which occur in two belts: one running along the northern flanks of the Hoggar-Tibisti mountains, and a second on the southern side of these mountains (Fig.2). The largest sand seas in the northern belt are: The Great Sand Sea, Ramlat Rabyanah, Idhan Awbari, Idhan Murzuq, the Great Erg Oriental and the Great Erg Occidental. Towards the west, these two belts of sand seas merge and hence the following sand seas are presented in the Western Sahara: Erg er Raoui, Erg Chech and Erg Iguidi. The southern belt constitutes of: Erg Qoz Kordofan, Erg Bilma, Erg Hausa, Erg Azouad, Erg of Ijafene, Erg of Mreyye, Erg Ouarane, Erg of Maqteir and on the most western side of the Sahara, Erg of Trarza. The largest of the Saharan sand seas is the Erg Chech in Algeria, which has an area of ca 320,000 sq km. The total area covered by Saharan sand seas amounts to ca 2,600 000 sq km.

1.1.7. Libyan Sahara

Libya is one of the largest African countries and occupies an area of 1,775,500 sq km. Libyan area is composed of basement rocks of Precambrian age (570,000,000÷460,000,000 B.P.) which are overlain by marine and continental deposits (App.1).

The Libyan Sahara constitutes the eastern part of the Sahara plateau and presents all forms of the Sahara desert landforms. Libya with 99% arid lands and Egypt, 90%, belong to the strictly Saharan states and truly desert lands.

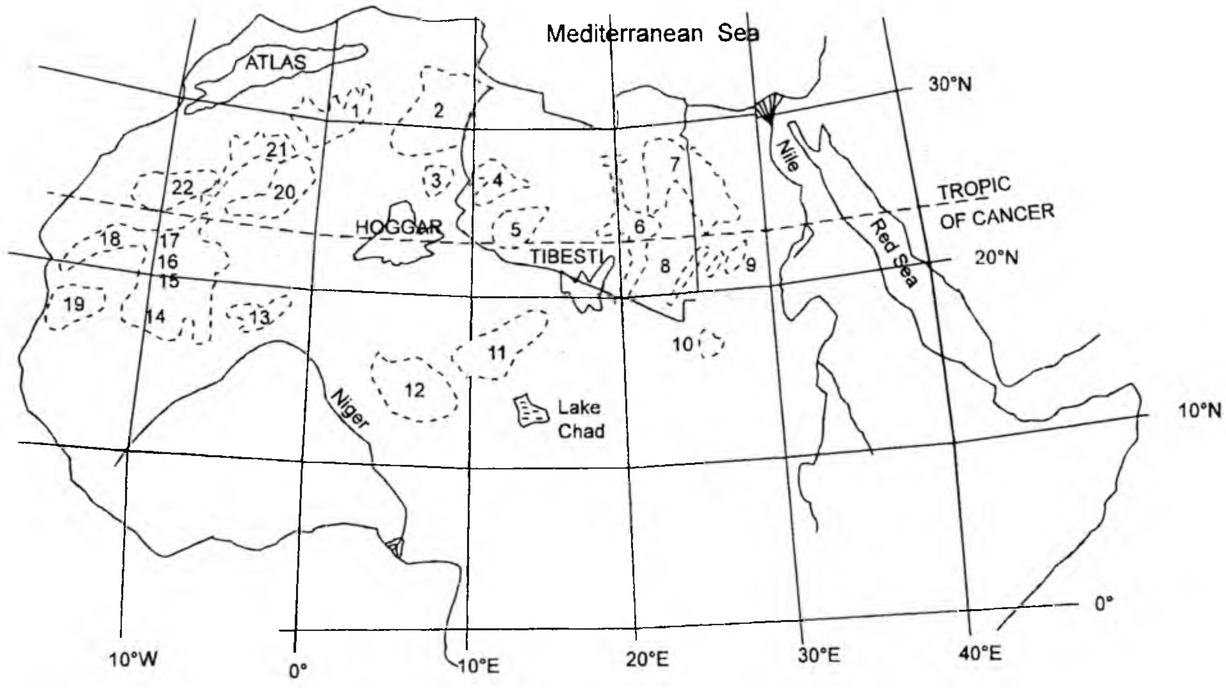


Fig.2 Sand seas in Sahara.
The northern belt : 1 - The Great Erg Occidental, 2 - The Great Erg Oriental, 3 - Erg Isauane N'Irarraren, 4 - Idhan Awbari, 5 - Idhan Murzuq, 6 - Ramlat Rabyanah, 7 - The Great Sand Sea, 8 - Irq al Idrisi, 9 - Erg of Selima; the southern belt : 10 - Erg Qoz Kordofan, 11 - Erg Bilma, 12 - Erg Hausa, 13 - Erg Azouard, 14 - Erg Aoukar, 15 - Erg Mreyye, 16 - Erg Ijafene, 17 - Erg Ouarane, 18 - Erg Makteir, 19 - Erg Trarza; North-western Sahara Ergs: 20 - Erg Chech, 21 - Erg Er Rowi, 22 - Erg Iguidi (Names of ergs outside

The sand - covered undulating plateau surface is broken by several physical features, including the al-Haruj al-Aswad and the Hamada al Hamra. Al-Haruj as-Aswad is a hilly basaltic plateau, which covers about 39,900 sq km of central Libya and rises to about 1200 m. The upper part of this plateau is crowned by volcanic peaks. The region surrounding al-Haruj as-Aswad is covered with angular stone fragments and boulders.

Al Hamadah al Hamra is a rocky plateau, which constitutes the eastern part of the Plateau of Tinrhert and lies behind the Jabal Nafusah. Al Hamadah al-Hamra contains bare rock outcrops and covers an area of about 49,000 sq km. In the southern part of Libya, the huge Tibisti massif culminates at 3376 m, the highest peak in the land.

The climate over most of the country is that of the hot, arid Sahara. The world's highest degree of aridity was found at Sabha, where isolation (the amount of solar radiation received at ground level) amounts to 3 825 hours per year. The annual potential evaporation for the Sabha region, between 2 250 mm and 2 500 mm, represents the highest in the world (Burdon and Gonfiantini, 1991). The highest world temperature of about 58°C was recorded in al-Aziziyah on the Gefara plain near Tripoli. By contrast, in the Tibisti mountains the temperature can fall to as low as −15°C. The mean annual precipitation in the Libyan Sahara is very low (Fig.3). All of the Libyan sand seas are located in the areas where the mean annual precipitation is less than 20 mm. The mean annual relative humidity for sand seas areas is below 40%, with the exception of the northern part of the Great Sand Sea where this value increases to 50%. For example: in Sabha, in the summer time, the mean relative humidity is ca 20%, while in the winter months this value rises to 50% (Fig.4). In Sabha, the average annual rainfall is 8.2 mm, the average temperature in July, 32°C and the maximum recorded temperature, 46.5° C.

The dry climate of Libya is accelerated by *ghibli*, a hot, arid wind which blows from the south several times per year. It is proceded by a lull in the prevailing winds, which is followed by the full force of ghibli. The wind carries large quantities of sand dust, which turn the sky a red colour and reduce visibility to less than 10÷20 m. The heat of the wind is increased by a rapid drop of the relative humidity, which can fall from 80% to 10% within hours. Similar saharan winds occur in other parts of the Sahara and are known under different names like *harmattan* in Sudan and Nigeria, *khamsin* in Egypt, and *sirocco* in Magreb.

The Libyan desert areas consist of stony tablelands called *hamadas* (eg. Al Hamadah al Hamra, Hamadat Murzuq), gravel plains called *sarir* (eg. Sarir Tibisti, Sarir al Qattusah and As Sarir) and sand seas called: *idhan*, *ramlat* or *irq* (eg. Idhan Awbari and Idhan Murzuq, Ramlat Rabyanah and Ramlat Zaltan, Irq al Idrisi).

The Libyan sands seas (Fig.5), analysed in this investigation occupy an area of 302.585 sq km. The particular sand seas cover the following areas listed in table 1.

Table 1

Sand sea	sq km
Idhan Awbari	55 190
Idhan Murzuq	70 685
Ramlat Zaltan	28 410
Great Sand Sea	80 150
Ramlat Rabyanah	25 620
Irg Southern Rabyanah	25 620
Irg Idrisi	26 040

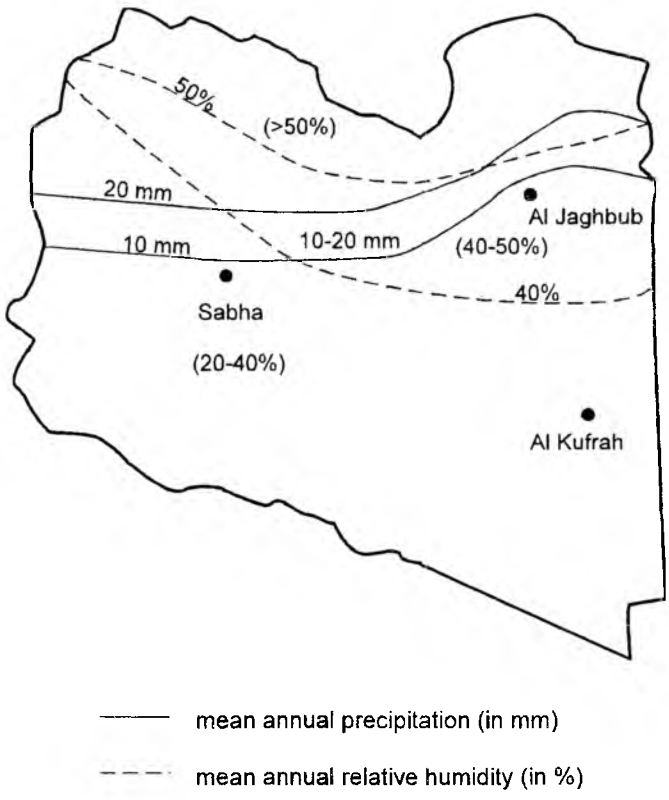


Fig.3 Precipitation and relative humidity in Libyan Sahara (extracted from the National Atlas of Libya, 1978).

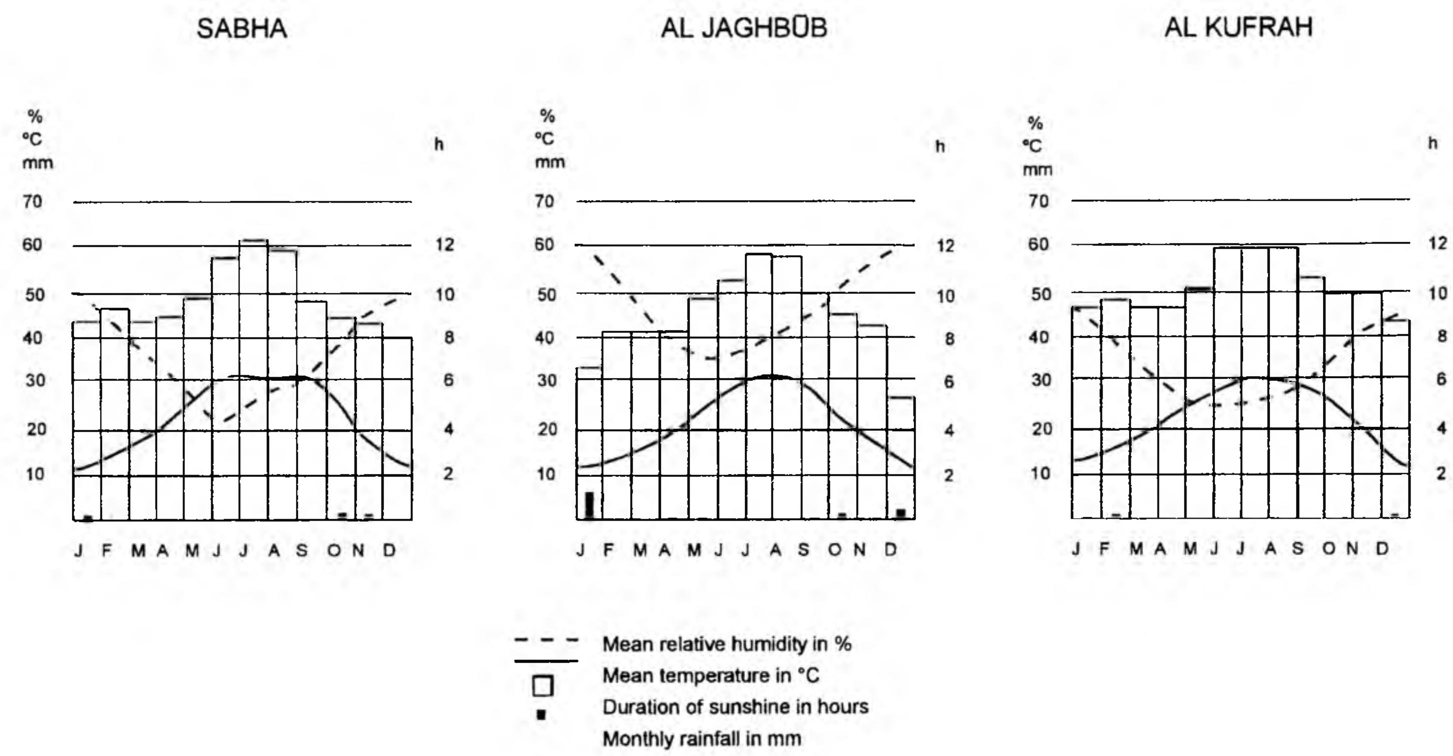


Fig.4 Climatic data for meteorological stations in the vicinity of sand seas (extracted from the National Atlas of Libya, 1978).

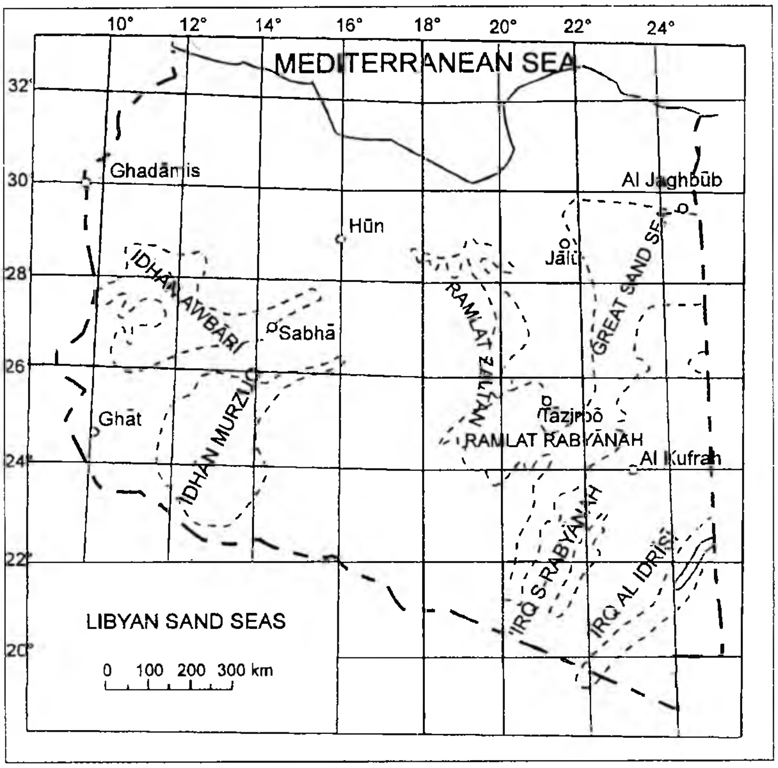


Fig.5 Sand seas in the Libyan Sahara
and meteorological stations in the vicinity of sand seas.

1.2. PREVIOUS INVESTIGATIONS OF SAND DESERTS

Geomorphological investigations of desert areas have been undertaken in the late - nineteenth and early twentieth centuries. As early as 1870, geomorphologists became interested in desert morphology, and, since then, the problems of dune formation have been a subject of considerable field observations and analysis. Rolfs, Medlicott, Blanford, Cornish, Udden, Carnegie and Sokolov were the preeminent pioneers in this field. In 1874, the german explorer Gerhard Rolfs explored the Western Desert in Egypt and described the dunes. He gave the name *the Great Sand Sea* to this part of sand desert. In 1876 Medlicott and Blanford described sand dunes of the Great Indian Desert. Cornish (1897) was involved in sand desert studies in the same region. In 1898, Udden published his monumental work on the grain size of aeolian deposits. In the same year, Carnegie presented the results of his explorations of the desert of the interior of Western Australia. Sokolov (1894) was involved in studies of aeolian sands. Huntington (1907) studied grain size distribution over dune flanks.

The first publications regarding the Sahara were presented by Beadnell (1910), Pawlowski (1926), Ball (1927) and Aufrere (1928). The early field investigations carried out during the desert expeditions were limited to certain routes, and the publications from these investigations have a rather descriptive character. In 1923 Ahmed Hassanein Bay, an Egyptian diplomat and explorer, had made a camel journey of 2000 miles, crossing for the first time the entire Egypto - Libyan Desert from north to south, from Sallum on the Mediterranean coast, via Al-Jaghbub and Al Kufrah, then, via Uweinat-Arkenu massif, to El Fasher far away in Sudan (Bagnold, 1982; McHugh, 1982). In early 1925, Prince Kemal El-Din Hussein, led an expedition of moto cars into the southern Libyan Desert which reached the Uweinat-Arkenu massif and Irq al Idrisi in the south - eastern part of Libya (McHugh, 1982)

The most significant investigations regarding sand desert morphology, have been performed by Colonel Ralph Bagnold, a British military engineer. Commissioned in 1915, he spent much of his army life in the Corps of Signals, among elsewhere, in Egypt (1926 ÷ 1929 and 1939 ÷ 1944) and India (1929÷1931). During his mission to Egypt, he explored both the Libyan and Sinai deserts. In late 1932, Bagnold led an expedition to Gilf-Uweinat area and explored far to the south and west of Uweinat (Bagnold, 1933). In Hong-Kong Bagnold (1935) wrote his first book :„Libyan Sands". In 1941 his second book „The Physics of Blown Sand and Desert Dunes" was published. This book represents the seminal textbook in this subject and most of the workers involved the sand desert investigations repeatedly refer to his findings and theories.

In 1940 Bagnold was posted to Cairo, where he formed the „Long Range Desert Group" operating anywhere in the uninhabited interior of Libya. His group was able to navigate deep into the territories behind the enemy lines due to detailed knowledge of desert areas and their characteristics.

Apart from his field investigations, Bagnold performed a number of fundamental experimental studies of sand movement by wind (Bagnold, 1936a, 1936b, 1938 and 1985).

The earliest post - World War II expedition to Gebel Archenu and south-eastern Libya was organized in 1962 by the Royal Military Academy, Sandhurst, England (Williams and Hall, 1965).

Later investigations and publications dealing with sand deserts focused on selected problems regarding sand dunes and their behaviour. Many of the investigations were based on laboratory wind-tunnel experiments, which allowed the principles of sand movement and dune formation to be detremined (Sharp, 1963; Belly, 1964; Warren and Knott, 1983; Wasson and Hyde, 1983a; Iversen, 1985; Jensen and Zeman, 1985; Napalnis, 1985; Rasmussen *et al,* 1985; Reid, 1985; Willets and Rice, 1985a and 1985b; Pye, 1985; Nicking, 1988; Rubin and Ikeda, 1990; Sherman and Hotta, 1990).

Many workers focussed their efforts on the study of sand grain distribution and relationships within the dune fields, and within particular dune bodies

(Bagnold, 1938; Besler, 1982; Livingstone, 1987; Lancaster, 1981, 1986; McArthur, 1987).

In the last two decades, a significant growth of interest in sand desert investigations has been observed. Several detailed field investigations, connected with local meteorological measurements on particular dunes, has permitted to determine the relationship between the characteristics of wind and particular dune forms to be determined (Howard *et al*, 1978; Fryberger, 1979; El-Baz, 1982; El Baz and Prestel, 1982; Hyde and Wasson, 1983; Howard, 1985; Howard and Walmsey, 1985; Kolm, 1985). A few studies were oriented towards investigations of the internal structure of particular dune forms (McKee and Tibbitts, 1964; McKee, 1971 and 1979). A further group of investigations was directed towards studies of coastal dunes which are developed in environmental conditions quite different to those of inland dunes (Miszalski, 1973; Borowka, 1980, 1990a, 1990b; Pye and Tsoar, 1990; Rotnicki, 1995).

Field investigations and studies relating to the behaviour of the sand desert areas represented by the huge sand seas were rather limited. Such works were restricted to the Australian sand desert areas (Brookfield, 1970; Mabbutt, 1984); the Algerian sand seas (Wilson, 1973); the North American Deserts (Dregne, 1984); the China deserts (Chao Sung-Chiao, 1984a and 1984b; Zhenda, 1984) and to the Sahara desert (Mainguet and Chemin, 1983; Mainguet, 1984b).

Dust transport in suspension and the problems of desert loess formation were the subjects of several investigations (Morales, 1979; Tsoar and Pye, 1987a and 1987b). Desertification problems were studied by the investigators such as Mainguet and Chemin (1983), El Baz and Hassan (1986); and Mainguet (1986).

In the last two decades, several investigations have focussed on studies of dune forms recognized in the stratigraphic record (Clemmensen and Abrahamsen, 1983a and 1983b; Rubin and Hunter, 1985; Collinson, 1986). Studies of aeolian stratification types, facies and sequences have provided much new information on the development of different dunes and interdune environments in ancient sediments.

The application of aerial photographs and satellite images and photographs created many new possibilities for studies of the sand desert geomorphology at the meso - scale. Verstappen (1970) performed aerial photographic studies of the southern part of Thar Desert. In 1970 Verstappen and Zuidan applied Gemini and Appollo photographs in geomorphological investigations of the Central Sahara. Wilson (1972, 1973) applied aerial photographs for morphological studies of the Algerian ergs. Later, aerial photographs and satellite data were applied by Mainguet (1982, 1984b) in the elucidation of transsaharan and transsahelian aeolian currents and aeolian dynamics. El-Baz (1982, 1984) has applied Landsat data for sand desert investigations in Egypt, and for establishing analogies between the features of terrestrial deserts and the Martian surface.

Maxwell (1987), Maxwell and Jacobber (1986) and Maxwell and Hayness (1989) applied Landsat and Spot data in investigations of desert landforms in Egypt. Miszalski (1973, 1974) applied multi - temporal aerial photographs for investigation of the contemporary aeolian processes on the Baltic coast. On the basis of aerial photographs in the Misurata region of northern Libya Aparta and Szczypek (1987) analysed dune forms.

In 1979, Mc Kee edited the USGS professional publication: „A study of global sand seas". This publication was based on the interpretation of Landsat MSS images covering selected world sand seas (Breed and Grow, 1979; Breed *et al.*, 1979). In this publication only a small part of the western Libyan Sahara has briefly been described.

In 1979 Mabbutt published his book *Desert Landforms* which is devoted to desert morphology problems. A review of the world's deserts was presented in the book *Desert and arid lands* (El-Baz, 1984).

Aeolian geomorphology constitutes a very significant part of the many text - books devoted to geomorphology or remote sensing (Chorley and Schun, 1984; Ciolkosz and Kesik, 1989; Ciolkosz *et al.*, 1978; Goudie, 1983; Klimaszewski, 1978; Lillesand and Kiffer, 1979; Ostaficzuk, 1978; Strahler, 1975; Sunnerfield, 1991; Warren, 1979).

In Poland, the stabilised inland dunes which occur in several parts of the country, were investigated by several geomorphologists (Dylikowa, 1969; Kobendzowie, 1958; Kozarski, 1962; Nowaczyk, 1986; Pernarowski, 1958; Szczypek 1977; Urbaniak, 1969; Urbaniak - Biernacka, 1973; Wojtanowicz, 1965).

There are very few investigations relating to the Libyan Sahara. Capot - Rey, best known for his explorations of the Idhan Murzuq and Grand Ergs Occidental and Oriental, travelled through Idhan Murzuq in 1947. He recognised several types of dune in this area and noted the bimodal character of the Idhan Awbari sand grains (Capot - Rey, 1947). During their field expedition in 1961 to Idhan Awbari, McKee and Tibbitts (1964) studied the primary structures of seif dunes present in the vicinity of Sebha. The application of space photographs for desert studies, as acquired over the south - eastern part of Libya, was reported by Pesce (1971). The phenomenon of dune reddening in the western part of the Libyan Sahara was studied by Walker (1979). All these investigations have been carried out on a rather limited scale and were focussed on selected problems.

In recent years, the sand desert areas have been briefly described in the series of explanatory booklets, of the Geological Map of Libya, at the scale of 1:250,000, edited by the Industrial Research Centre in Tripoli. Aeolian deposits have briefly been analysed in the chapters dealing with the stratigraphy of the Pleistocene - Holocene period.

During his mission to Libya in the 1970s, the Author was astonished by the variety of dune forms and patterns appearing in the Libyan sand seas, which were not marked on any of the geological and other thematic maps. The analysis of aerial photographs and, later, of satellite images of the vast areas of the Libyan Sahara, stimulated the Author to undertake more comprehensive investigations of sand desert landforms, recognising that the Libyan Sahara has been omitted from the geomorphologic investigations conducted on the regional scale. The first results of the research into sand desert landforms, based on the interpretation of the Landsat Photomaps of Libya, were presented to the 3rd Symposium on the Geology of Libya, held in Tripoli in September 1987. Since then, the author has applied Metric Camera photographs to geomorphological analysis of Ramlat Zaltan. The results of these investigations were presented to the international symposium „Mapping of the Earth Surface" held in Leipzig in 1987. The remote sensing approach to sand sea mapping was presented to the Willi Nordberg Symposium held in Graz in 1987 (Linsenbarth, 1987b). The results of futher studies on the geomorphology of Libyan sand seas were demonstrated at the Second International Conference on Geomorphology held in Frankfurt in 1989 (Linsenbarth, 1989) and at the XVII Congress of the International Society of Photogrammetry and Remote Sensing in Washington in 1992 (Linsenbarth, 1992). The results of the application of MOMS-02 data for the analysis of dune development were presented during the MOMS symposium held in Cologne in 1995 (Linsenbarth, 1995).

The preliminary results of the investigations undertaken by this Author on the Libyan Sahara, indicated beyond much doubt that remote sensing data, especially high resolution satellite images or photographs, can contribute to a large extent to integrated geomorphological studies of sand desert areas. The Author belives that remotely sensed data can be used as spatial reference data, which are of the highest importance in investigations related to the sand seas morphology, their morphogenesis and morphochronology. The Libyan Sahara, which contains a large variety of dune forms and patterns, was chosen as the study area. The objectives and scope of the investigations are explained in Chapter 1.3.

1.3. OBJECTIVES AND SCOPE OF THE INVESTIGATIONS

The main goal of the investigations undertaken was to develop a methodological approach to the integration of the remote sensing data and techniques with the geomorphological investigations of sand desert areas.

The Libyan Sahara sand seas were selected as the test study because in the Author's opinion, they represent the largest variety of dune forms and patterns in the world. The phenomenal variety is the product of both the present and past environmental conditions prevailing in this part of the Sahara (App.1 and 2) and also the terrain topography.

The methodological approach was based on the integration of satellite remote sensing data, with other data concerning desert geomorphology. It was assumed that, if the satellite remote sensing data reflect the present day morphology of the sand desert area hence, they must also reflect the influence of the different factors governing the formation of sand seas, and their development. If the present forms of sand desert also reflect the changes which occurred within the process of sand seas formation and development, the remote sensing satellite data can thereby be used, not only for morphological studies but can also provide an input to the morphogenesis and morphochronology of sand desert. The satellite images present a true image of desert areas, without any selective extraction of information and/or generalization, as is the case for topographic or thematic cartographic materials. Also the satellite data were selected for these integrated studies, due to the fact that they are the only homogeneous medium which covers an area as large as the Libyan Sahara; moreover these materials can be used for investigations undertaken at both a regional and local scale.

In the adopted methodological approach, it is assumed that remote sensing materials will be treated as the core data in the investigations undertaken, because they reflect the interaction of the different factors governing the sand seas formation and development. Of course, the remote sensing data has to be supported by the other available data necessary in the geomorphologic studies of sand desert areas. These supplementary data refer to the general characteristics and development of the Libyan Sahara, in terms of geology, topography and climatic conditions and with special reference to the prevailing wind regime. This concept is graphically presented in Fig.6.

The investigations carried out were divided into several stages, which can broadly be divided into *preliminary* and *basic stages. The preliminary stages* are related to the studies referring to environmental conditions, governing the Libyan Sahara and to review of up - to - date literature on the theory of sand movement and sand desert characteristics. The *basic stages* of investigations relate to remote sensing data and their applications in the geomorphological investigations of the Libyan sand seas.

Some general information regarding the world deserts and their characteristics is given in Chapter 1.1, while more detailed information concerning the environmental conditions of the Libyan Sahara is presented in Appendices 1 and 2. The results of the detailed studies of wind regime in Libya, which plays the principal role in sand seas formation and development, are presented in Appendices 3 and 4.

Preliminary studies performed by this Author in the Libyan Sahara (Linsenbarth, 1987a, 1987b) indicated that sand seas formation and development is influenced not only by the present invironmnetal conditions, but also, undoubtedly by past environmental conditions. This was a reason why, in this investigation special attention is paid to studies of the geomorphologic evolution of the Libyan Sahara (App.1)and to climatic changes in the Sahara (App.2).

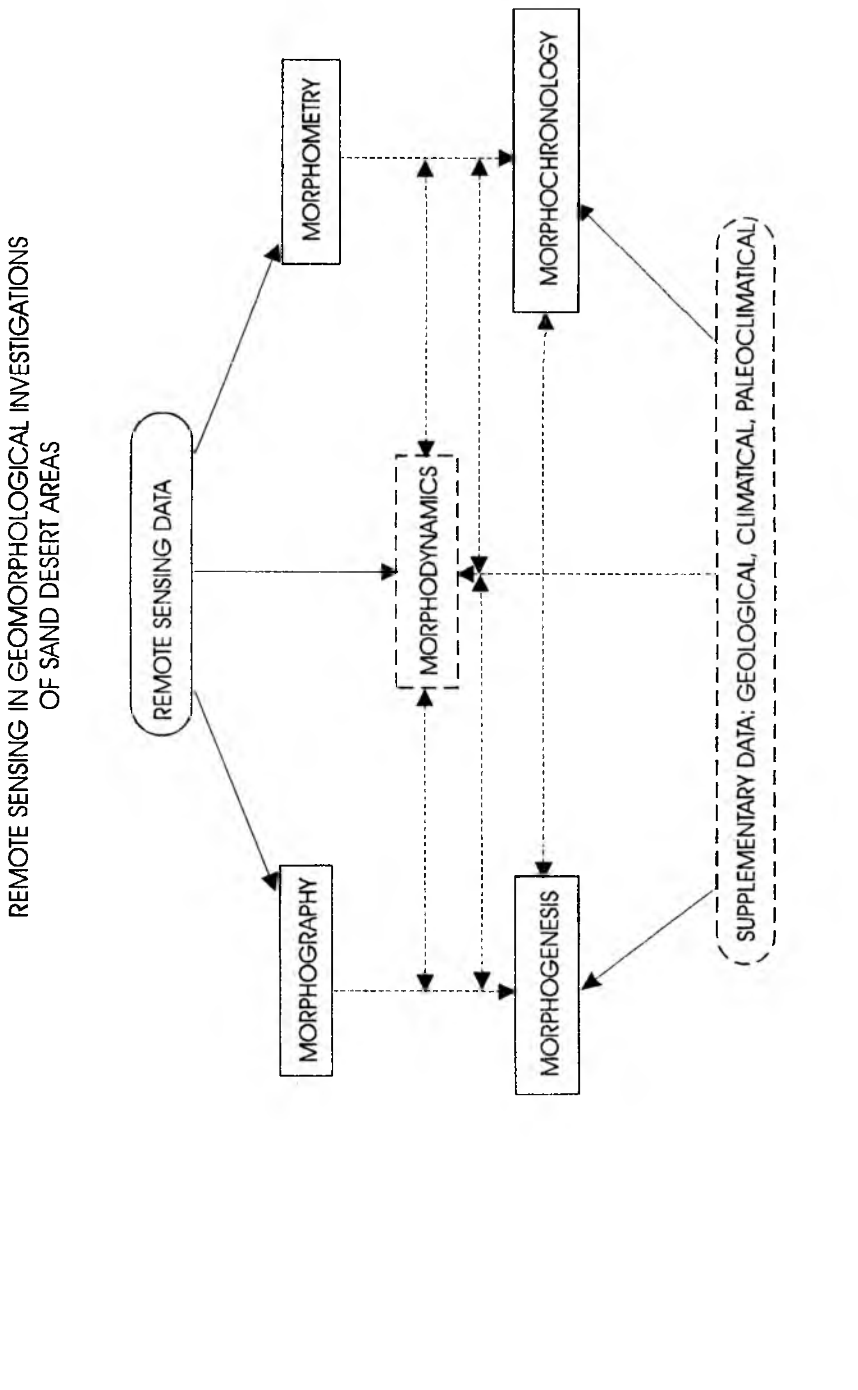


Fig.6 Remote sensing in geomorphological investigations

The remote sensing approach to geomorphological investigations of the Libyan Sahara sand seas, is preceded by a review of the theory of the physical principles governing sand movement (App. 5). The intention of the Author is to present this information in a systematic and synthetic way and to focus attention on those principles and factors governing sand movement and dune formation which are not very well known and which have been presented only in very narrowly oriented research publications. The results of this synthetic review were applied later, at the basic stage of investigations, i.e. in the evaluation of satellite remote sensing data for sand desert studies.

As the basic material for these investigations, different satellite remote sensing data were applied. The intention of the Author was to verify the different remote sensing data and techniques and to evaluate their usefulness in geomorphologic studies. For these purposes, the following satellite remote sensing data were used: those from the operational systems Landsat MSS and SPOT XS, and those from experimental remote sensing missions on the Space Shuttle: Metric Camera, MOMS-01, MOMS-02 and SIR-A. As basic material for regional scale investigations, the Landsat Photomap of Libya at scale 1:250,000 was applied. For local investigations, the SPOT data and those from the experimental remote sensing missions, carried out on board of the Space Shuttle were used. The main aim of this part of the investigation, was to determine what kind of information regarding the sand desert geomorphology could be derived from different satellite remote sensing data and what should be the methodological approach for intergrating the remote sensing data within the process of geomorphologic investigations. Multitemporal images were used for determining the contemporary dynamics of a sand desert. In the Idhan Awbarii, the dynamics of the sand desert was determined on the basis of SPOT data from 1988 and Metric Camera photographs from 1983 and photomosaics produced in 1955. For dynamics analysis in Ramlat Zaltan the MOMS-02 data from 1993 were referred to Metric Camera photographs taken in 1983. For remote sensing analysis, different techniques of data processing were applied, special attention being paid to digital methods. In Chapter 2, the results of investigations related to remote sensing data in geomorphologic investigations of sand desert are presented.

Based on satellite remote sensing data, the proposed methodological approach of intergrated investigations was applied on the study area, referring to the Libyan sand seas. This part of the investigation was carried out in three consecutive stages. At first, the morphology of particular dune forms was studied (Chapter 3). Secondly the morphology of sand seas was considered (Chapter 4) and thirdly comparative geomorphologic analysis of the Libyan sand seas has been carried out (Chapter 5).

The investigations of the particular dune forms appearing in the Libyan Sahara were preceded by the classification of dune forms and patterns adopted by the Author. The methodological approach was based on an assumption that it is necessary to define the particular dune forms, in terms of their location and the environmental conditions governing their formation and development. For these purposes, the Landsat MSS data, were accepted as the basis and for selected areas, other data from the experimental missions were also used. For each single dune form, its locality, and main characteristics and morphometric parameters were determined and, finally its hypothetical morphogenesis interpreted (Chapter 3).

The second stage of the investigations undertaken relates to the geomorphology of the Libyan sand seas (Chapter 4). These investigations have been based mainly on Landsat Photomaps of Libya at a scale 1:250,000 (Linsenbarth, 1987a, 1991); these were used for the classification of sand seas into those fields showing homogeneous dune forms and patterns and also for the interpretation and geomorphologic description of the particular dune fields (Linsenbarth, 1991). The results of this part of the investigation are presented on the Libyan sand seas maps compiled at a scale 1:1,000,000. The geomorphologic analysis of particular sand seas is preceded by a general description of location, terrain topography and sand sources. For each sand sea or for chosen parts of particular sand seas the hypothetical morphogenesis and morphochronogy has been assessed.

The third stage in the investigation was devoted to a comparative geomorphologic analysis of the Libyan sand seas (Chapter 5). This stage concentrated on defining the differences between the main characteristics of the individual sand seas, such as their shape and extent, sand sources, dune forms and patterns and their appearance and sequence of dune forms within particular sand seas. All these characteristics were referred to the factors which govern sand seas formation and development.

As a result of this integrated approach to remote sensing in sand desert investigations, a hypothetical scenario of sand seas development is proposed by the Author (Chapter 6). General conclusions, based on the adopted methodological approach and the investigations performed, are presented in the last chapter (Chapter 7).

2. REMOTE SENSING IN SAND DESERT INVESTIGATIONS

2.1. THE ROLE OF REMOTE SENSING DATA IN SAND DESERT STUDIES

Remote sensing data represents a very usefull tool in investigations of sand desert geomorphology. As sand deserts have a very limited map coverage, remote sensing data can be used both for topographical mapping of such areas and for geomorphologic studies in terms of the recognition of sand desert landforms and the determination of their properties (Linsenbarth, 1987b)

The main objective of the investigations undertaken by this Author was to evaluate the different remote sensing data available for geomorphologic studies of sand desert areas, and to establish the most convenient methodological approach, in fulfilment of the geomorphologic requirements.

The investigations performed were based on all available satellite remote sensing data taken over the Libyan Sahara. The following data, acquired by different sensors and from different space platforms has been used in investigations: Landsat MSS, Metric Camera, MOMS-1, SIR-A, SPOT-XS and MOMS-2.

The analysis of the processes involved in sand desert geomorphology, as described in previous chapters and in Appendixes 1÷6, resulted in the definition of problems which can be directly or indirectly solved on the basis of remote sensing data. Direct input can be given by satellite remote sensing data to morphological mapping and to morphometric and morphologic studies. The data derived from these materials can be used indirectly for morphochronologic and morphogenetic studies (Fig.6).

For morphometric and morphologic studies, the remote sensing data can be used for the mapping of sand desert forms and for determining the morphometric parameters of particular dune fields and dune forms. The main factors limiting the usefulness of satellite remote sensing data for such studies are: spectral characteristics of sensors, geometric resolution, and stereoscopic capability.

The analysis of data acquired by sensors working in different sectors of the visible and near infrared part of electromagnetic spectrum indicated, that, for sand desert areas, the best - quality images are recorded in the red and near - infrared part of the spectrum. Here, the highest differentiation between sand and other terrain surfaces occurs. The multispectral data can indicate the sand colour and colour changes both within singles dune and within the entire dune field. It is possible to discriminate between sand covering dunes and interdune areas, and to define the differences between sand covering the windward and lee - side of dunes and even the differences in sand along the windward side, lee - side and dune crest. These differences are a direct function of sand grain size characteristics and sorting. The sand colour may be used as an indicator of the sand sources while the differences in sand colour may be used for delineation of different sand sources and sand supply in specific dune fields.

The geometric resolution limits the detection of smaller dune forms. Certainly, it is not possible to detect ripples on satellite images or photographs, but the question is open as to what are the smallest dune forms which can be recognized on satellite images or photographs. Individual dunes should be defined by the dune outline and by the crest line, and by indicating peaks and saddles of dune ridges (Linsenbarth, 1988). Stereoscopic data are very useful in the mapping of dune forms. In the case of compound dunes, besides the primary forms, the detection of secondary forms is very important, because they can be used as indicators in the analysis of contemporary wind regime, and in morphogenetic studies.

In interdune areas, information concerning the sand cover and appearance of a rocky basement is of first importance. The detection of smaller dune forms (or zibars) which may appear in interdune areas is very important for the analysis of dune formation and their development.

As well as being able to map dune fields and particular dunes on the basis of satellite remote sensing data, the multitemporal data can provide a tool for studies of the morphodynamics of sand deserts. In this case, the resolution of the applied satellite data should be higher than the expected changes of sand dunes or their migrations in a given period of time.

All remote sensing satellite data, used in this investigations, have been analysed and their usefulness for geomorphologic studies determined. Finally, the methodological approach relating to the application of remote sensing satellite data for geomorphologic studies is described.

2.2. ANALYSIS AND EVALUATION OF REMOTE SENSING DATA AND TECHNIQUES

2.2.1. Landsat data

2.2.1.1. Landsat Photomap of Libya

The Landsat Photomap of Libya at a scale of 1:250,000 was employed as a basic material for regional scale investigations. The first National Landsat Photomap of Libya consists of 127 map sheets (1.5°WE×1°NS) which cover the total territory of Libya. The Landsat Photomap was prepared in 1980 by: Pacific Aero Survey Co. Ltd-Japan, ESRI International and the Earth Satellite Corporation, USA.

For the Landsat Photomap, 100 Landsat scenes acquired in the period 1975 ÷1978, were selected for processing. As a rule, winter scenes (October - February) were used for photomap compilation due to the lesser incidence of sandstorms in this period. All Landsat scenes were registered by the Landsat MSS system.

The Landsat Photomaps were produced in several technological stages (Linsenbarth, Gammudi, 1984). All Landsat scenes were digitally processed by EarthSat, using GEOPIC software for geometric and radiometric corrections, edge enhancement, gray scale adjustment and scanline suppression. Digital processing was performed on an individual scene - by - scene basis, because due to serious reduction of information contents, the results of the developed „national Landsat histogram for Libya" were not acceptable. Due to this fact, the photomaps do not present the homogeneous colour image of desert surface.

The corrected and enhanced digital data for spectral bands 4, 5 and 7 were separately recorded at a scale of 1:1,140,000, onto black and white film using the Optronics P-1500 Photowrite System. At the next stage of photomap production, the colour composite negative images were produced by additive filtration. These colour composite negatives were used for production of photographic enlargements at the scale of 1:500,000. For scaling the existing cartographic materials were used. These enlargements were applied for GEOPIC mosaicing which was performed in three separate sections. The geographic grid overlay 1.5°×1° was used for defining sheet corners, and then the photomap sheets were cut from the original mosaic section. Next, the photomap sheets were copied onto 8"×10" colour negatives at a scale of 1:750,000. As the final product, the 1:250,000 Landsat Photomaps were printed as both colour and black - and - white copies (Fig.7).

A verification of Landsat Photomaps of Libya (Linsenbarth, Gammudi, 1984) indicated that the scale of photomaps varies between 1:249,000 ÷ 1:253,000. The mismatching errors of mosaicing are between 0.25 and 1 mm (0.1 ÷ 0.4 km), but, in several cases they are up to 24 mm (6 km). The sheet - to - sheet fitting is generally better than between scenes within the particular photomap sheets. The largest differences occur between sheets on the boundary between sections of the original mosaicing at a scale of 1:250,000.

Generally, the image quality of photomaps is good, but several sheets covering sand areas have a rather poor contrast. Futhermore; the season differences of scene acquisition caused changes of the image contrast and changes in image colour. The sand seas are presented in various shades of yellow, light brown and white, and have a strong contrast with surrounding areas.

2.2.1.2. Application of Landsat Photomaps for sand desert investigations

The Landsat Photomap of Libya represents the only homogeneous cartographic base map of Libya, and also the only, full coverage of the available remote sensing satellite data over Libyan territory, available at that time. Hence, in this investigation, these photomaps were used as the basis for geomorphologic mapping, the determination of morphometric parameters of sand dunes, and for geomorphologic investigations.

As a result of photointerpretation, the outlines of particular sands seas were defined, and, within these sand seas, the various dune fields, characterised by

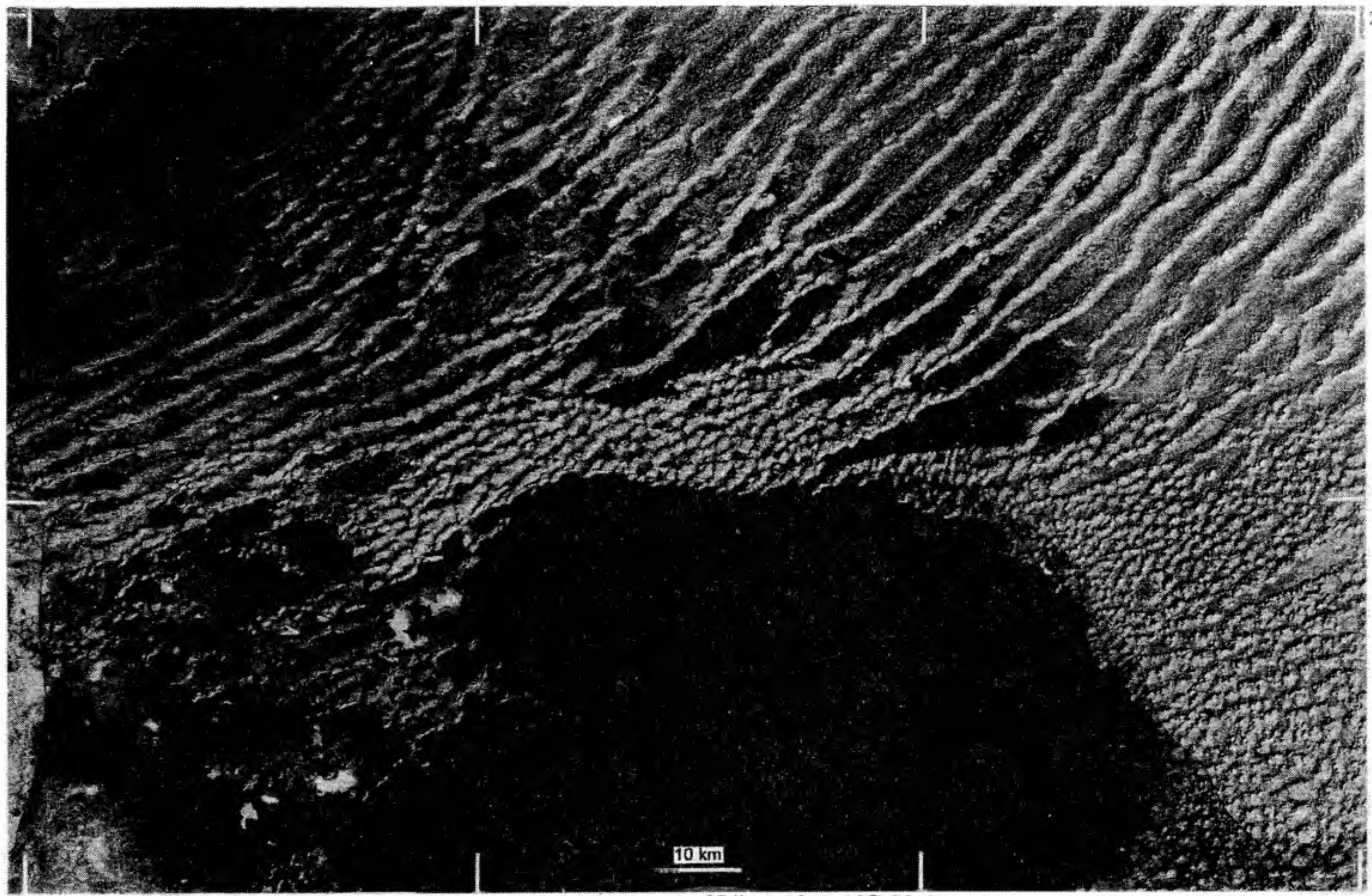


Fig.7 Landsat Photomap of Libya (sheet NG-32-4).

homogeneous dune forms were differentiated. The results of the interpretation were presented on sand seas maps compiled at a scale of 1:1,000,000 for: Idhan Awbari, Idhan Murzug, The Great Sand Sea, Ramlat Zaltan, Ramlat Rabyanah, Irq Southern Rabyanah and Irq al Idrisi (Linsenbarth, 1986a, b, c, d, e, f, g). Sketchmaps of all these sand seas are presented on figures 47 ÷ 53. Within particular sand seas, the dune fields occupied by the same dune types and trends were defined. In case of sandridges, all ridges were plotted (usually only the crest lines of the sandridges were plotted), also, if visible, the trends of secondary dunes superimposed on sandridges, were marked. In the case of very densely spaced linear or crescentic dunes, due to the scale of mapping, only dune trends were shown.

Photomaps have also been used for determining the main morphometric parameters of dune fields (length, width and spacing of dune or dune ridges) and for elaboration of a more detailed geomorphologic description of dune fields and particular dune forms (Linsenbarth, 1986a, b, c, d, e, f, g). The reliminary results of this investigation were presented to the 3rd Symposium on Libyan Geology, held in Tripoli in 1987 (Linsenbarth, 1991).

In the investigations reported in this paper, the sand sea maps created from the Landsat Photomaps of Libya, have been used for analysis of various dune forms which are present in the Libyan Sahara (Chap. 3), for geomorphologic analysis of particular sand seas (Chap. 4) and for comparative analysis of the Libyan sand seas (Chap. 5).

The Landsat Photomaps of Libya have been used for the determination of sand colours in the Libyan sand seas. The sand seas depicted on the Landsat Photomap of Libya (colour version) are presented in various yellowish - red to yellowish - brown colours. The western sand seas: Idhan Murzug and Idhan Awbari are strikingly redder than those in the east: the Great Sand Sea and Ramlat Rabyanah. Idhan Awbari is generally depicted as redish - brown (2.5YR6/8 in Munsell Colour System) to light brown (7.5YR7/10) while the Great Sand sea is presented in light brown (7.5YR6/10) to yellowish - brown (10YR7/8). The sands in Ramlat Zaltan are different in colour compared with other Libyan sand seas. The north - western part has a dark reddish - brown colour (10R6/10); the norh - eastern part is depicted in yellowish - brown (7.5YR6/8) wheras in the middle and southern part the sand is reddish - yellow (2.5YR6/8). The sands in Ramlat Rabyanah are reflected in large varieties of yellowish and yellowish - brown colours (10YR7/8-7.5YR6/6). The southern wing of Ramlat Rabyanah generally radiates yellowish - brown hues (7.7YR7/8). The sands of Irq al Idrisi are lighter than the sands in the other sand seas and reflect light yellowish - brown hues (10YR8/6).

There are distinct changes of sand colour within the sand seas. For example, in the northern part of the Great Sand Sea there is a significant change of sand colour in an west - east direction. In the western part, sands are depicted in light yellowish - brown colours, in the middle part, in yellowish light brown hue, wheras the eastern part of sand sea is depicted in light yellow hue. Towards south, in the narrowest part of the Great Sand Sea, the sand hue changes into reddish - brown and on the boundary with Ramlat Rabyanah the sand has golden - red hue. In the Ramlat Southern Rabyanah the sand colour changes from golden hues in the northern part, through yellowish - gold in the central part, into reddish - brown towards Chad border.

In Irq al Idrisi, the sand incorporated in the linear dunes has very light yellow hue in the north - eastern part while the interdune areas reflect reddish - brown hues. Towards south - west direction the dunes are presented in darker hues and the interdune areas are also depicted in darker values. In Idhan Murzuq a very distinct trend of sand colour changes can be traced from NE to SW. The sand hue changes from light brown in NE, through dark brown, into yellowish - red on the SW margin of sand sea. In Idhan Awbari there are several changes in sand colour. The central part of the sand sea is generally much redder than the other parts. The north - western part has a much brighter hues, which was later confirmed on SPOT images (see Chapter 2.2.5).

At several places within the sand seas, generally on the borders of sand sea, the glaucous colours appear. In Idhan Murzuq, such a colour of sand appears in the north - eastern - most part as well as in the north - western part of Idhan Awbari.

The differences in sand colour in respect of the individual sand seas can be accounted for by assuming that there were different sand sources. The differences in sand colour within the particular sand seas may be attributed to the different sand sources in several parts of sand sea but also to the changes of sand colour along the direction of sand transport. The local differences of sand colour along the boundaries of sand seas, relflect a local supply of sand from the adjacent areas.

In all investigated sand seas, the crest of dunes are brighter than dune flanks, and dunes are depicted on satellite images in much brighter values of colour than the interdune areas (with the exception of the south - western wing of Irq al Idrisi). The hammadas and sarir areas have generally very dark reddish - brown hue (10R4/5) thus indicating the boundary between sand seas and sarir surface. It is necessary to emphasise, that Landsat Photomaps of Libya have no homogeneous colour presentation; thus the colour analysis can give only general information of sand colour and their changes.

2.2.1.3. Digitally processed Landsat data

Landsat scene No. 202/42 acquired on 10 June 1978, registered by Landsat MSS system, was used for digital processing. The digital processing was performed on the ERDAS PC system Version 7.5. Within the process of digital image processing, several enahcement techniques (Mather, 1987) were applied for improving the visual interpretability of the sand desert forms.

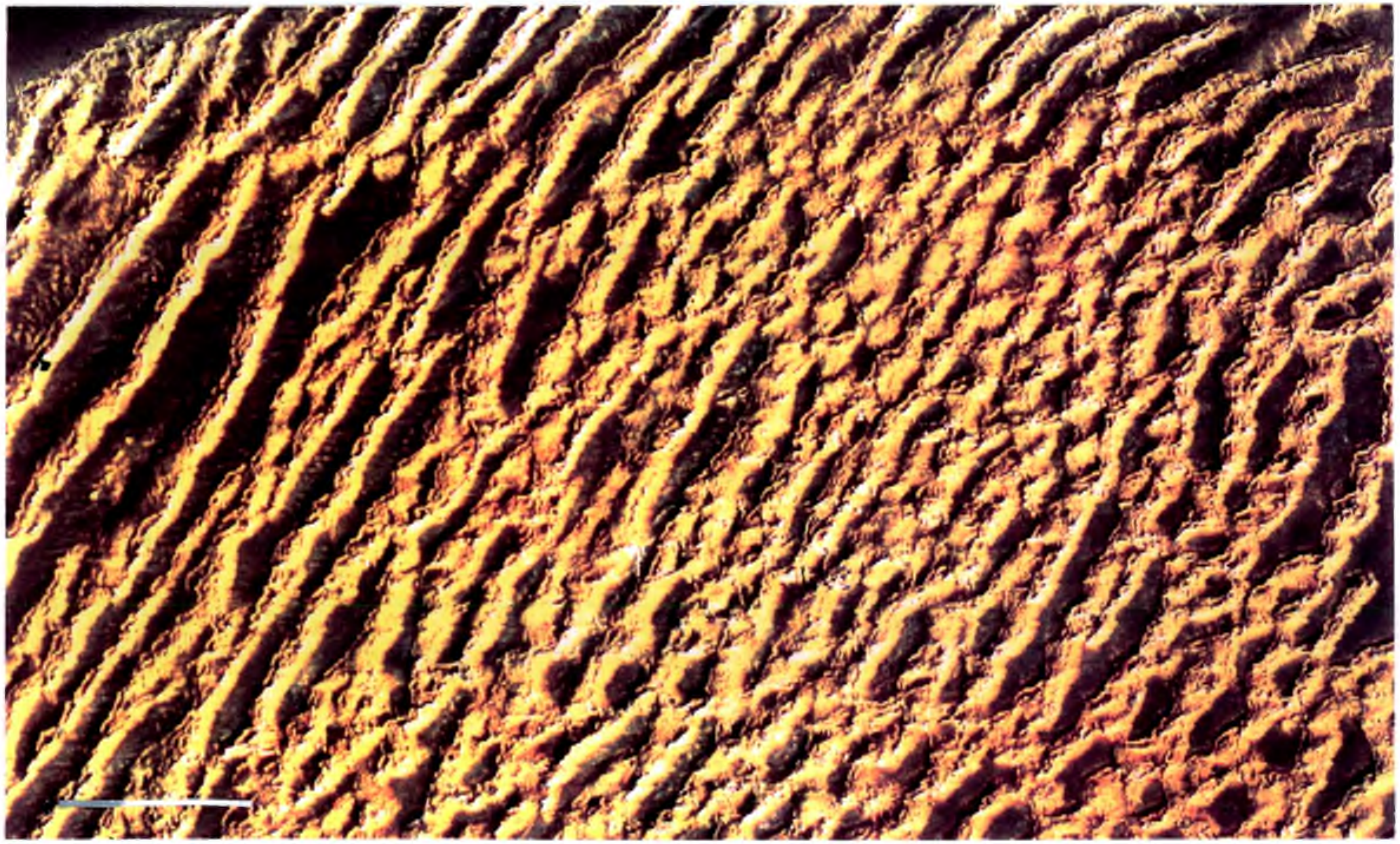


Fig.8 Landsat colour composite - northwestern part of Idhan Murzuq. (White bar represents 10 kilometres)

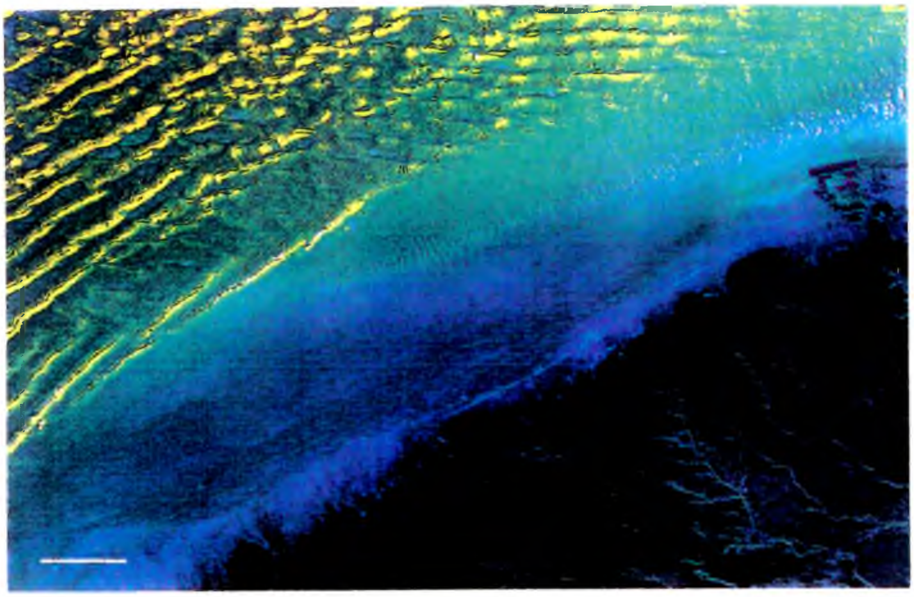


Fig.9 Landsat colour composite -
southern margin of the central part of Idhan Awbari.
(White bar represents 10 kilometres)

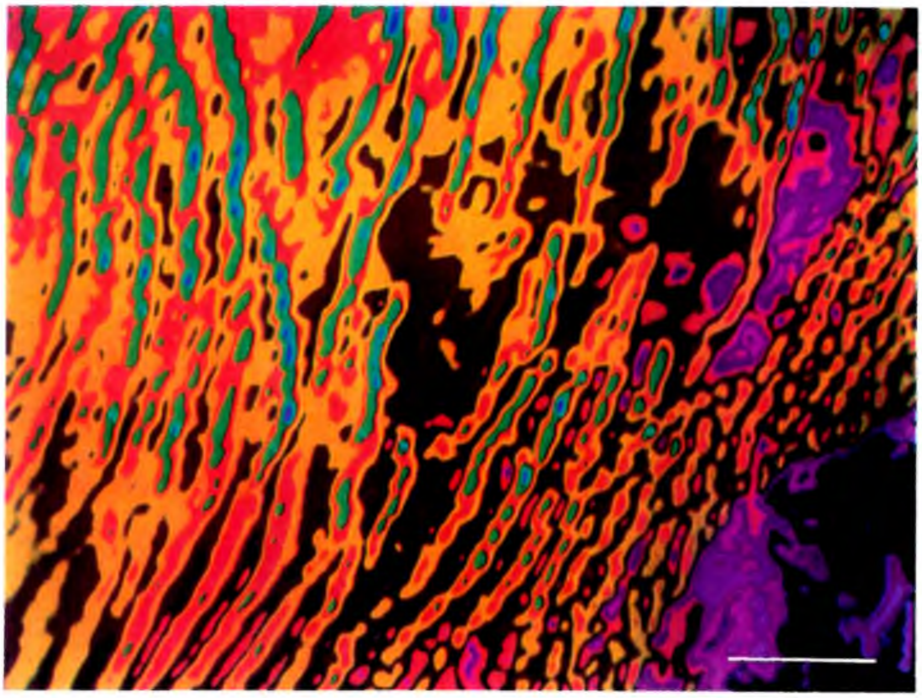


Fig.10 Density slicing of a Metric Camera space photograph -
western part of Idhan Awbari. (White bar represents 10 kilometres)

For standard colour composites the following MSS channels were used: channel 7 - red, channel 5 - green and channel 4 - blue. Simultaneously, a linear contrast - stretching technique was applied. On the standard colour composite, the sand sea area is presented in various intensities of yellow and light brown colour (Fig.8 and 9). Generally, dune crest and partial dune flanks are depicted in light yellow, wheras the interdune areas are presented in dark yellow. Very distinct differences in the sand colour can be distinguished between particular parts of sand seas. There is a narrow belt of dunes in the vicinity of Ramlat Murzuq which is expressed in very light yellow colour. Another wide belt of the lighter colour sands, stretching SW-NE across the image, can be also delineated. The colour differences indicate the different spectral characteristics of sand, which are in turn a function of sand origin and granulometric characteristics of sand particles.

The evaluation of colour composites using a histogram equalization procedure, indicated that a redistribution of pixel values across the 0÷255 scale did not improve the image interpretability. On the processed image the dune pattern is less distinct than in the case of standard composites. Also, the delineation of sand colours is more complicated. The transition from one sand colour to another is less gradual. Generally, it may be concluded that the histogram equalization method reduces contrasts of sand sea image.

The applied five - level density slicing of the colour composite did not improve the interpretability of the dunes forms in the sand seas area, but greatly improved the differentiation of the deposits in the wadi areas adjacent to the sand sea. Also, in the interdune areas, the differentiation of various surfaces is more distinct.

The modification of selected bands of the colour composite resulted in a change of image colours. For example modification of red and green bands resulted in presentation of the interdune areas in green, without change of the dune light brown. The modification of red, green and blue bands introduced a pink colour to the dune surfaces and blue and green - blue to the interdune areas. This procedure led to discrimination of various surfaces in the interdune areas.

The application of the technique based on inversion colours highly increased the visual perception of dune patterns. Shifting the colour composite with to the threshold value of 50 units, suggested some advantages of this technique. More distinct differences in the sand colour of sandridges and additional information regarding interdune areas result from this. On the northwestern flanks of sandridges, the depicted white dots can be correlated with different sand characteristics, which have result from selective sand movement.

Clearly, the interpretability of desert sand areas can be improved when other enhancemnet techniques are combined with standard colour composite.

2.2.2. Metric Camera

During the first Spacelab Mission in December 1983, a Metric Camera was part of the earth observation payload of the Space Shuttle STS-9. The objective of the mission was principally to test the mapping capabilities of the high resolution space photographs. The Metric Camera, a modified Aerial Survey Camera Zeiss RMK A30/23, which uses standard aerial film of 23×23cm, was applied for the first time to obtain high - quality photographs of the Earth's surface.

2.2.2.1. Characteristics of the Metric Camera

The Metric Camera was built by the German Aerospace Research Establishment (DFVLR) and German Industry on the basis of a proposal by prof. G. Konecny from Hannover. The Metric Camera consists of the following components:

* camera body with optics and exposure meter,
* film magazines containing aerial films of 24 cm width,
* remote control unit,
* camera suspension mount.

The camera objective has a Topar A-1 lens with 7 lens elements. The calibrated focal length is 305.128 mm. The maximum distortion is 6 μm and the resolution 40 lp/mm over the whole image size on Aviphot PAN 30 film. The horizontal field of view is 41.2° and diagonal 56°. The Aerotop rotating disc shutter (between lens shutter) has a speed of 1/250 ÷ 1/1000 sec. in 31 steps. The apertures stops are 1:5.6, 1:8 and 1:11 and the shortest cycling time (exposure frequency) is 4 ÷ 6 sec.

The Metric Camera has two film magazines, each with a film length of 150m which is equivalent to 550 image frames. Each magazine may contain a different film type.

2.2.2.2. Spacelab Mission parameters

During the Spacelab Mission, the Metric Camera was in operation on 2nd and 5th December 1983. In use, the camera was assembled on its suspension mount, fitted to an optically flat high quality window, and operated via a remote control unit installed in an experimental unit of the Space Shuttle. During the operation of the Metric Camera, the Space Shuttle must be at such an attitude that the optical axis of the camera is exactly radial to the Earth. The camera is oriented to an accuracy of 0.5 degree.

The space photographs have been taken from tan orbit altitude of ca 250km, hence the scale of photographs was ca 1:820,000. Each photograph covers an area of 188×188 km, which is roughly equivalent to the Landsat scene. During the first Spacelab Mission, 36 hours were planned for earth observation. The timing of the photography was also limited by illumination conditions. The film is only properly exposed when the sun elevation is greater than 15°. There are two strips of satellite photographs taken over Libyan territory: 24 and 4 (Fig.11).

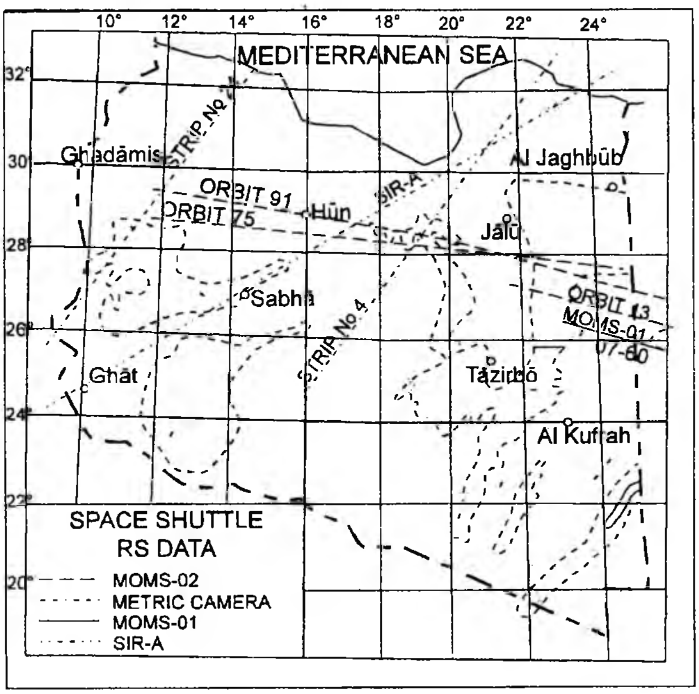


Fig.11 Space Shuttle missions over Libyan Sahara.

The space photographs of strip 4, taken on 2nd December 1983, were exposed on Kodak Aerochrome Infrared Film 2442, with shutter exposure time 1/500 sec, and aperture of 6.8 with 60% forward overlap. With a relative forward velocity between the spacecraft and the earth of approximately 7.55 km/sec, the image motion is equal to 18 μm (14.8 m on the ground). Photographs over Libyan territory were acquired between $7^h58'13''$ and $8^h02'32''$ GMT, with ca 10″ time interval between successive photographs, therefore; the sun angle was ca 28° and sun azimuth ca 140°.

The space photographs of strip 24, taken on 5 December 1983, were exposed on the Kodak Double-X Aerographic Film 2405 (B/W), with exposure time 1/625 sec and aperture of 5.7. The influence of the image motion is thus 16 μm (11.1 m on the ground). The photographs taken over Libyan territory were exposed between $7^h25'21''$and $7^h27'00''$ GMT, hence, the sun azimuth is ca 127° and sun elevation angle ca 17°. The base to height ratio, photographs being taken with 60% overlap, is 1 : 3,3, which is rather inconvenient for elevation determination (Linsenbarth, 1987a).

2.2.2.3. Evaluation of Metric Camera data for sand desert investigations

The Metric Camera space photographs from strip No.4 were used for geomorphologic mapping and analysis of Ramlat Zaltan, and the results presented to the International Symposium: Mapping of the Earth's Surface, held in Leipzig in 1987 (Linsenbarth, 1987a). The space photographs 157 and 158 from strip 4 were used for photogrammetric plotting performed on a Wild A-10 stereoplotter. The map was compiled at a scale of 1:100,000 with emphasis on detailed mapping of all visible dune forms and patterns. As well as the very spectacular star dunes appearing in Ramlat Zaltan, smaller dune forms namely linear dunes, crescentic dunes and crescentic chains were also recognized and their geomorphology analysed. Owing to a larger scale and higher resolution, these data, have given much more detailed information of the sand desert forms, their sequence and relation to terrain topography when compared with the Landsat MSS data.

Further to the photogrammetric mapping, density slicing of space photograph 158 from strip 4 was carried out on the NAC MCDS (*Multi Colour Data System*) (Sitek, 1991). The technique of density slicing converts a black - and - white image into pseudocolour image. This method is an effective way of emphasising the different but internally homogeneous areas within an image, at the expense of loss of detail (Mather,1987). The number of discrete grey levels of the Metric Camera photograph was reduced to 8 or 10 levels which were then presented by different colours. Several tests allowed the adjustment of the slice boundaries in such a way that the differences of sand characteristics were greatly enhanced. The density slicing performed on the Metric Camera images taken over Idhan Awbari permitted the delineation of the different zones of sand covering the slopes of sandridges (Fig.10). In the case of Ramlat Zaqqut, the application of density slicing technique enabled the determination of the boundaries between active aeolian sand, i.e. that taking part in the development of dunes, and the coarser sand incorporated in sand sheets or sarir areas.

The space photographs Nos. 157, 158 and 159 of strip 4 were converted into digital form by scanning on PhotoScan P1. The images obtained were digitally processed on an Intergraph Microstation to enhance the sand dune forms. The results of the enhancement are presented in Fig.12 on which the star dunes, as well as linear dunes and crescentic chains, are very well depicted.

The black and white photographs Nos.708 and 709 from strip 24, covering the eastern part of Idhan Awbari, were used for detailed investigations of the sand desert forms which appear in this area. These have a much higher resolution than the Landsat MSS images, and their stereoscopic capability coupled with a low sun angle, permitted the esay resolution of the dune forms and their trends. The secondary forms superimposed on sandridges which give indirect information about the present wind regime in this region are very obvious. Also, it is sometimes possible to detect the giant ripples which occur in the interdune

Fig.12 Digitally enhanced Metric Camera photograph.
Eastern part of Ramlat Zaqqut.
(White bar represents one kilometre on all photographs if not stated otherwise)

Fig.13 MOMS-01 space image, taken on 22 June 1983 over the Libyan-Egyptian border (orbit 07-60, band 853/900 nm). Example of the classical linear dunes on the margin of the Egyptian part of the Great Sand Sea. (Courtesy of DFVLR-Germany).
(White bar represents ten kilometres)

areas. These forms are closely correlated with strong winds and their direction. The effect of shadows gives a better impression of dune heights and can thus be used for their height determination, at least in relative term.

The stereomodel 708/709 was used for stereoplotting performed on an analytical stereoplotter Planicomp P1. On a selected part of this model, spot heights were measured and a digital terrain model created; later contour lines were generated with 50 m contour intervals. The contour lines at such intervals (this is the contour interval which can be accepted due to a very incovenient base to flight height ratio 1 : 3,3) give a general overview of the dune forms but only partly relate to their geomorphology. Linsenbarth, (1988) discussed this problem and concluded that the dune forms, if presented solely by contour lines, should either be depicted by very densely spaced contour lines or combined with planimetric outlines of dune elements such as dune crests, dune base outline etc., and supported by the plotting of secondary forms.

The digital approach to the Metric Camera data resulted in a much more detailed interpretation and analysis of the image contents. The space photographs were converted to digital form by scanning on a photogrammetric scanner PhotoScan P1 with resolution of 50 μm.

The analysis of different approaches to digital processing performed on the Intergraph Microstation, indicated that, in the case of Metric Camera data, after image enhancement, the optimum scale for interpretation, is an image at the scale of ca 1:47,000 on the system monitor. The scale of images on the monitor screen is ca 19-times larger than the scale of the original space photographs. Due to a rather low sun angle (17°) the sandridges appear as 3-D forms. The windward sides and downwind sides of sandridges are very obvious. The windward gentle slopes are covered by secondary dunes, and the downwind, steeper, slopes are without secondary dune forms. These data led to the identification of the trends of secondary dune formation on the windward side of primary dunes, their preferred orientation and changes along the slope. In some parts, the giant ripples, the so - called zibars, can be distinguished. The processed segments of the Metric Camera photographs taken over Idhan Awbari have been applied as examples of different dune forms such as longitudinal sandridges (Fig.44), barchanoidal sandridges (Fig.45) and domical sandridges. When converted to the digital form, the Metric Camera photographs were used as reference data in a comparative analysis of SPOT and Metric Camera data (see Chapter 2.2.7).

2.2.3. MOMS-01

2.2.3.1. Characteristics of MOMS-01

The Modular Optoelectronic Multispectral Scanner (MOMS) was used for the first time during the experimental missions of the Space Shuttle flight STS-7 in June 1983 and, later, on the STS-11 in February 1984. During these two missions, an area of ca 3.5 milion square km was covered by space images.

MOMS-1 was the first German Remote Sensing System to operate in space. For the first time, MOMS-1 made use of the CCD technology by means of push - broom principle; thereby no mechanical moving parts were used for scanning. MOMS-01 constitutes a space - adapted version of two two - channel airborne scanner MOMS-EM, which was produced in 1980 and tested for thematic mapping. MOMS-1 was designed as a multispectral sensor, which consists of two bands, but which in the future, can be extended to include a larger number of spectral bands. The modular concept is characterized by individual modules, each representing one spectral band. Each module consists of filters, dual lens optics, 4 CCD sensors and preamplifier electronics. Channel one operates in the bandwidth between 575 ÷ 625 μm and channel two in the region of 825 ÷ 975 μm.

The pixel size is 16×16 μm, representing a ground pixel of 20×20 m for an orbit of 300 km. The focal length of the optical module is 237,2 mm and the instantaneous field of view 67.5 μrad.

The MOMS-1/STS-7 mission started on 18 June 1983. During a 6 day mission, MOMS-1 was in operation for 24 hours with a total recording time of 27 min. The space shuttle orbit inclination was 28.5° and the orbit altitude 292km. The ground swath width of MOMS imagery was 138 km. During the STS-7 mission, images were taken over parts of Africa, East Asia, India and South America, on a total area of 1,644,000 sq km.

The same system MOMS-1 was used on the STS-11 mission which started on 3 February 1984 and, during 4 days of MOMS operation, the images covered 1,872,000 sq km of the Earth's surface. The data from both missions, as derived from the original HDT, were available to users as CCT copies and negatives and B/W hard copies at a scale of 1:800,000.

The Libyan Sahara is covered by images taken from orbit 07 ÷ 60 on the descending path. The images started from 23°E towards ESE and pass over Egypt and Saudi Arabia (Fig.11). The images, were taken on 22 June 1983, from an altitude of 291 km over the Libyan Sahara, the sun azimuth being 249° and sun elevation, 11°.

2.2.3.2. Evaluation of MOMS-1 data for sand desert investigations

Unfortunately the MOMS-1 data over Libyan territory cover only the eastern part of the Great Sand Sea, but in the east, the MOMS-1 data present a very spectacular view of the Egyptian part of the Great Sand Sea. Comparison of data from the two channels indicated that the data images registered in the near infrared (in channel 2), are much better than the images registered in channel 1. In the Libyan part, sandridges and star dunes are both detected. On the Egyptian side of the sand sea, there are classical examples of longitudinal sandridges, which overlie the rocky basement (Fig.13). The interdune areas are free from dune forms and are covered by only the thin blanket of sand. Composite longitudinal sandridges which, in this location, consist of several subparallel linear dunes arranged along the main dune crests may also be identified. These sandridges represent typical examples of dunes formed under the influence of an unidirectional or bidirectional wind regime.

2.2.4. MOMS-02

2.2.4.1. Characteristics of MOMS-02

Since the mid - seventies, an Earth Observation Programme called MOMS has been under development in Germany. The basis of the MOMS instrument development was a concept of a three - line - scanning method for mapping the Earth's topography and for improved thematic interpretation using 3-D information. The laboratory version, EOS, was built in 1978 to prove the CCD technology for remote sensing sensors. MOMS-2 constitutes the second generation of the MOMS system, which was sucessfully flown on board the second German Spacelab Mission (D2) from 26 April to 6 May 1993. MOMS-02 combines a high resolution three - channel stereo - module with a four - channel multispectral unit (MOMS-02-D2, 1994). MOMS-02 provides along - track stereo capability, to be operated in a panchromatic mode, alone or in various combinations with spectral channels. The system was designed both for thematic mapping on the basis of mutispectral data (channels 1, 2, 3 and 4) and for photogrammetric mapping on the basis of the stereo images registered by this system (channels 5, 6 and 7).

MOMS-02 multispectral system (M/S) consists of four sensors working in the following spectral bands: channel 1: 449÷511 nm, channel 2: 532÷576 nm, channel 3: 645÷677 nm and channel 4: 772÷815 nm. The wavelength centres of the seven channels are distributed over a range between 480 nm and 793 nm. The corresponding bandwidths of the multispectral channels vary between 35nm and 65 nm. The ground resolution of multispectral data acquired from the altitude of 300 km is ca 13.5 m and the ground swath of imagery is ca 78 km.

The stereo - system works along the orbit track. In the stereo mode, one image (channel 5) is registered at nadir orientation, with ground resolution of 4.5 m, while the two others (channels 6 and 7) are taken at an angle of ± 21.4° with a resolution of 13.5 m. The swath width of high resolution imagery is 37km. The stereo - channels cover the panchromatic region; a bandwidth of 240 nm with a center wavelength of 640 nm was selected for those channels. Channel 5 operates in the bandwidth 512÷765 nm, while channels 6 and 7 operate within the bandwidth 524÷763 nm.

The MOMS-02 imaging geometry is depicted on Fig.14. Because of the viewing angle of 21.4° between the two off - nadir stereo channels (6 and 7), the linear imagery of those channels is separated by ca 120 km from the swath of the nadir channels (1,2,3,4,5 and 6). The three - fold stereo images are recorded with a time lag of about 20 seconds.

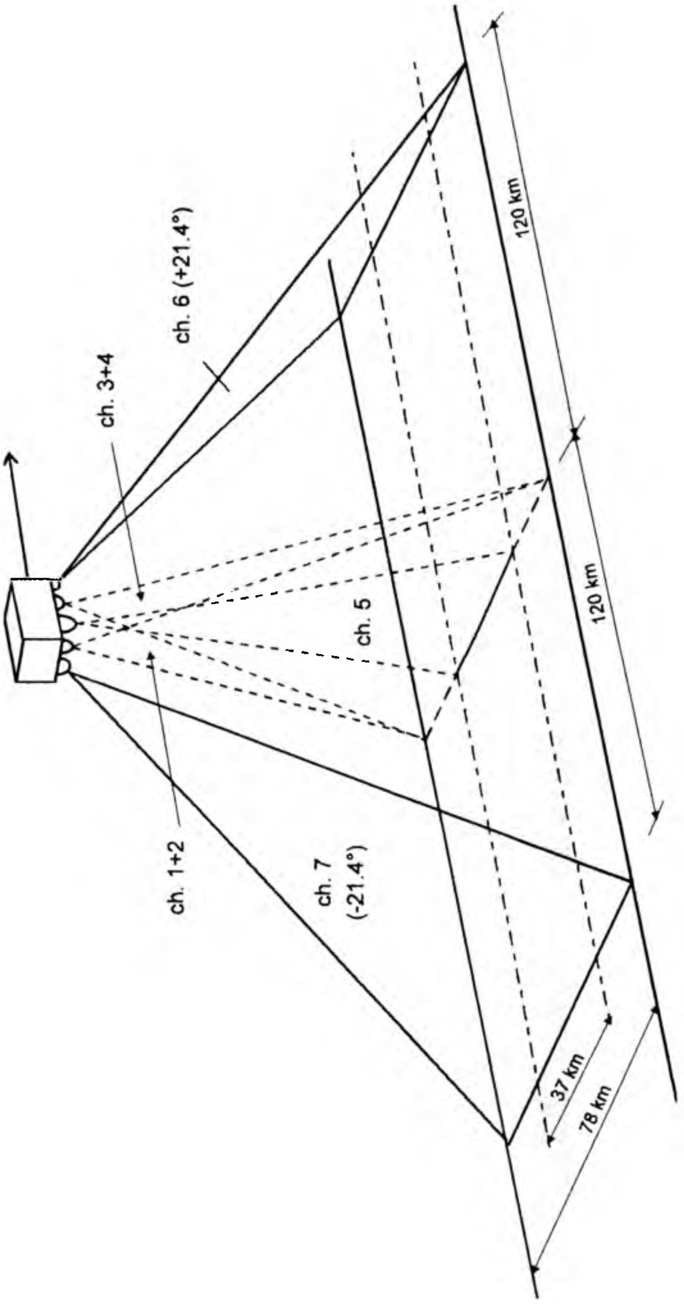


Fig.14 MOMS-02 imaging geometry (modified after MOMS-02-D2 data catalogue).

The MOMS-02 optical system consists of five lenses: three are designed for stereoscopic images, whereas the other two are for the acquisition of multispectral data. The high resolution imagery is taken by the central lens, which has a focal length of 660 mm. The two other stereo images are taken by two stereo lenses, each with a focal length of 237.2 mm, tilted at an angle of +21.4° and –21.4°, respectively, relative to the flight direction. The focal length of these lenses was selected in such a way that the ratio of the ground pixel size of high resolution and tilted stero images will be 1:3. The instantaneous field of view (IFOV) for high resolution images (channel 5) is 15.15 μrad, while, for off - nadir stero images (channels 6 and 7) it is 42.16 μrad. The CCD arrays of the stero module for channels 6 and 7 consist of 5800 active sensor elements, whereas the nadir - looking CCD array comprises 2 arrays with 6000 sensor elements each, which are optically combined to one array with 8304 active sensor elements The stereo images are acquired quasi - simultaneously. The three CCD-lines are imaging different terrains at the same time, whereas the three lines image the same terrain at different times, as the space shuttle moves.

The remaing two lenses are designed for acquisition of multispectral images. There are two sensors in the focal plane of each multispectral camera, together with their corresponding filters. To achieve the requirement of identical pixel size relative to the tilted stereo channels, the selected focal length is 220 mm and IFOV is equal to 45.45 μrad.

Table 2

Channels		1	2	3	4	5	6	7
Orientation		Nadir					+21.4°	-21.4°
Band width (nm)		449 -511	532 -576	645 -677	772 -815	512 -765	524 -763	524 -763
IFOV [μrad]		45.45	45.45	45.45	45.45	15.15	42.16	42.16
Mode 1	Number of pixels					8304	2976	2976
Mode 2		5800	5800	5800	5800			
Mode 3				5800	5800		5800	5800
Mode 4		5800		5800	5800		5800	
Mode 5		5800		5800	5800			5800
Mode 6			3220	3220	3220	6000		
Mode 7		3220		3220	3220	6000		

For the MOMS-02 system, seven operation modes were defined, which combine the different channels selected for various thematic applications. In table 2 the different modes are listed together with the bandwidth of particular channels, IFOV and number of pixels registered in each channel. The maximum data recording rate of the on - board magnetic tape recorder is only 100 Mbit/sec, thus there was no possibility of recording all the channels simultaneously.

During the Space Shuttle STS-55 mission in 1993, the data were recorded only in modes 1, 2, 3 and 6. Mode 1 is a full stereo mode with the high resolution nadir channel 5 and two tilted channels, 6 and 7.

During the MOMS-02 mission mode 1 was activated in two shuttle orientations: nadir and with a roll angle of 30°. Mode 2, can be classified as a full multispectral mode which consists of four channels 1, 2, 3 and 4 and can be applied for thematic applications. Mode 3 combines two multispectral channels 3 and 4, and two stereo - channels 6 and 7, thus providing thematic information in connection with three - dimentional data. Mode 6 combines three multispectral bands 2, 3 and 4 with a high resolution band, thereby increasing the resolution from 13.5 m to 4.5 m. During the MOMS-02 mission, the multispectral channels were recorded without compression, with the full radiometric resolution of 8 bits, while the data in the high resolution channel were compressed from 8 to 6 bits.

2.2.4.2. Evaluation of MOMS-02 data for sand desert investigations

During the MOMS-02/D-2 mission, images were acquired only over the northern part of Libya. The images were recorded from orbits 13, 60, 75 and 91 (Fig.11). There are four images, recorded from orbit 13, in multispectral mode 2, which cover the middle part of the Great Sand Sea but unfortunately these images are partialy cloud covered. Twelve scenes from orbit 75 were recorded in stereo mode 1, and scenes 14 and 15 were registered over the Great Sand Sea. The images taken from orbits 60 and 75 were acquired in mode 1/30°. The images from these orbits have been taken over the norhern part of Ramlat Zaltan and the middle part of the Great Sand Sea.

For the evaluation of MOMS-02 for sand desert investigations the images recorded from orbits 75 and 91 were selected. Scene 14, acquired from orbit 75 in mode 1, covers the central part of the Great Sand Sea, whereas scenes 22 and 23 recorded from orbit 91 in mode 1/30° were acquired over the northern part of Ramlat Zaltan. All the scenes were registered from the Space Shuttle orbit altitude of 303 km. For scene 14, the sun elevation was 32° and sun azimuth 89°, and for scenes 22 and 23, 28° and 88° respectively.

The main aim of these investigations was to evaluate the MOMS 02 data for geomorphologic studies and to determine the advantges of these data when compared with other remote sensing data. The selected scenes have been digitally processed on the Intergraph MicroStation and the processed images were analysed on the monitor screen. In the case of scene 14, recorded in mode 1, the main objective was to evaluate and to compare the data recorded in the high resolution channel 5 (4.5 m) with images taken in channels 6 and 7 which have a resolution of 13.5 m. The images from high resolution channel 5 and one of the off - nadir channels 6 or 7, were simultaneously projected on the monitor screen at the same scale (Fig.15). The results of the processing and enhancement indicated that the maximum scale on the monitor, with good visibility and interpretability of sand desert forms is 1:8,000 for the images recorded in channel 5 and 1:16,000 for the images acquired in channels 6 and 7. The images presented on the screen at the smaller scale permit the analysis of the general dune forms and trends, as in the case of the Landsat data. On the images monitored at larger scales, small dune forms as well as zibars can be detected in the interdune areas. Owing to the high resolution, the possibility of detection of the small simple or composite dune forms exists. In this respect the MOMS-02 data are more advantageous than other sources of data.

Scene 14 overlays a terrain swath ca 37 km wide (NS) and ca 110 km long (WE). The images recorded in channels 6 and 7 cover the same terrain. The images registered in channel 5 consist of 3 sub - scenes, described as *a*, *b* and *c*. After geometric matching, these three sub - scenes correspond with the same part of terrain surface as the images from channels 6 and 7. The longer axis of image strip extends almost W-E direction, from the boundary between As Sarir and the Great Sand Sea up to the eastern boundary of sand sea. The different sand dune forms appearing in this part of sand sea were easely discerned in this scene. On the western border, linear dunes and their components represented by very open crescentic dunes are clearly visible. That part of image which covers the central part of sand sea, where two trends of dunes can be distinguished, is very instructive. On the image presented on the screen at a smaller scale, only the larger dune forms representing subduded sandridges, are visible, but after enlargement, the contemporary dune trends, represented by linear dunes, can also be detected. This is a very spectacular part of the Great Sand Sea, in which the transition process, from one dune form to another, can be observed. Zibars covering interdune areas can also be detected. The radiometric quality of the image permits the detection and delineation of the differences in sand spectral responses which are a function of sand characteristics related, among others, to the grain parameters and sand sources.

In the northern part of Ramlat Zaltan and particularly in Ramlat Zaqqut, the MOMS-02 images recorded on scenes 22 and 23 were investigated. These focussed on the evaluation of the interpretability of the different dune forms which appear in this area (mainly linear dunes, crescentic chains and star dunes), in comparison with the Metric Camera photographs and on the evaluation of the stereoscopic capability of the stereo mode of MOMS-02 system.

The processing of MOMS-02 images was performed on the Intergraph MicroStation.

The MOMS-02 data were referred to the Metric Camera space photograph No.158 which was converted to the digital form on the photogrammetric scanner Photoscan PS1. The data registered in the high resolution channel 5 led to the determine of the small linear dunes and crescentic chains as well as very well displayed zibars. The analysis of these data led to the establishment of the relationship between dune forms and zibars and their relationship to the wind regime in this region. The application of stero images led to the establishment of the relationship of the linear dunes to terrain terrain topography and their preferencial development along small terrain depressions. The trend of linear dunes thus recognized is evidently, determined by terrain topography; dune trends are obviously tailored and adjusted to the terrain topography (Fig.16). Such trends can also be related to the secondary surface winds affected by terrain topography. Composite linear dunes, formed by connected, very open, crescentic dunes were also detected from these high resolution data (Fig.40).

The most spectacular dune forms which are present in the northern part of Ramlat Zaltan are the so - called star dunes which are located in several parts of this area (Linsenbarth,1987a and 1995). Owing to the high resolution and steroscopic capability of the MOMS images, it was possible to study the different types of star dunes: single star dunes, twin - star dunes and chains of star dunes, their connection with other dune forms and their relationship to the terrain topography (Fig.16). The opportunity to observe the same terrain feature from three positions (so - called three - fold images), which highly improves the interpretation capability is made possible by the application of the stereo mode (Fig.17).

The comparison of the MOMS-02 images with the Metric Camera photographs indicated that many smaller dune forms, which were not depicted on Metric Camera photographs, are easily visible on MOMS-02 images. The giant ripples, called zibars, are perfectly recognisable on MOMS-02 images, wheras on Metric Camera photographs, they are not visible. Zibars are very important in morpogenetic studies, because they reflect the influence of the strongest winds responsible for their direction and development.

2.2.5. SPOT data

The SPOT-XS multispectral images recorded on 17 April 1987 over Idhan Awbari, were considered in this investigation. Scene number 72 ÷ 294 was registered by an HRV-2 instrument in three spectral bands, XS1 (490÷590 nm), XS2 (610÷710 nm) and XS3 (800÷910 nm). At the time of scene acquisition (10h 03m 26s GMT local time) the sun elevation was 65.6° and sun azimuth, 133.5°. The scene was processed at level 1B by the Spot Image. Histograms of the spectral bands are presented on Fig.18. As can be seen, all bands occupy a very limited part of 254 DN. The peak of spectral band No1 is at 97 DN, and extends between 80 and 115 DN. In spectral band 2, the image is registered between 140 and 190 DN, with a peak at 162 DN and, in the spectral band No3

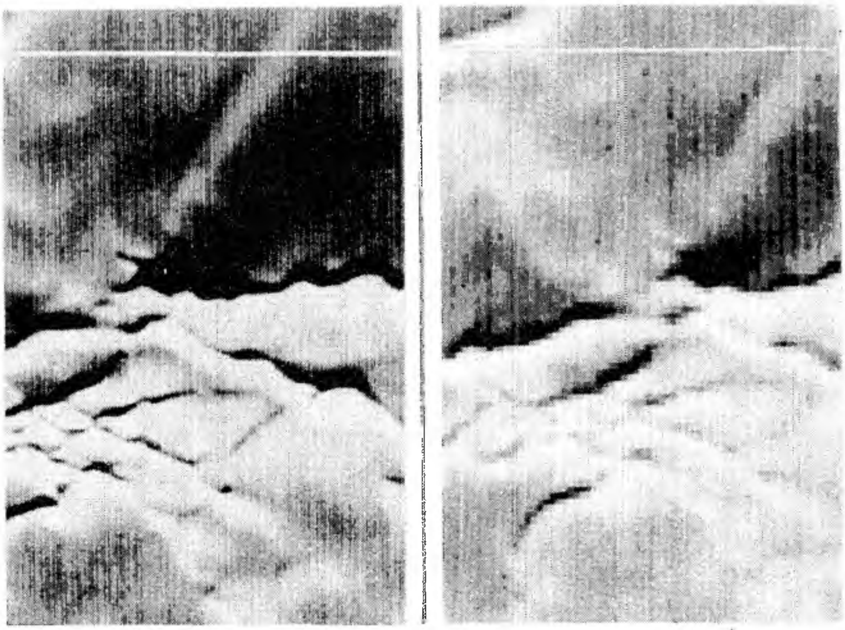


Fig.15 A comparison of maximum enlargement of MOMS-02
images recorded in mode 1, channel 5 (left picture) and channel 7 (right picture).

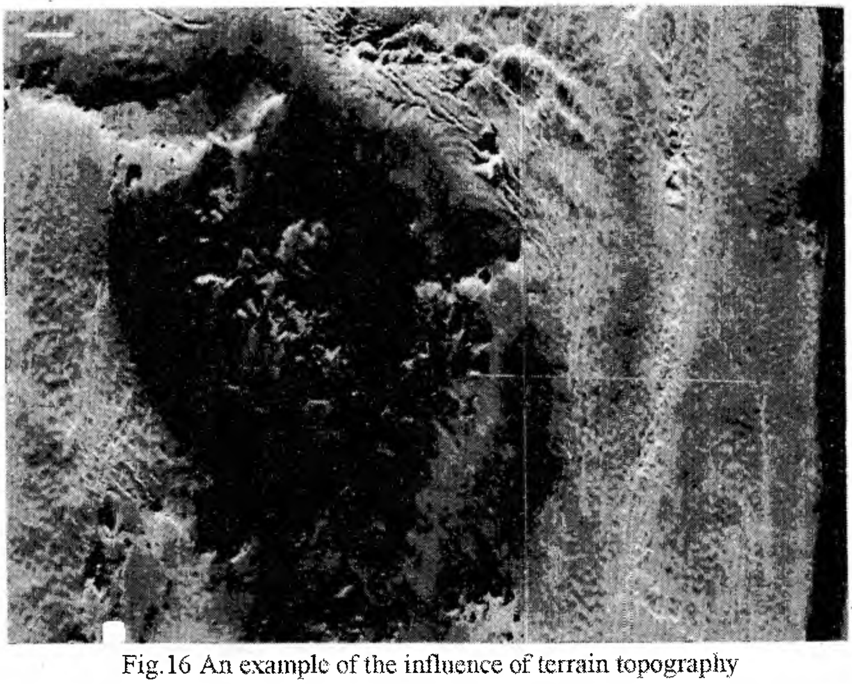


Fig.16 An example of the influence of terrain topography
on the dune forms in Ramlat Zaqqut. (MOMS-02 , orbit 91, scene 23, channel 7).

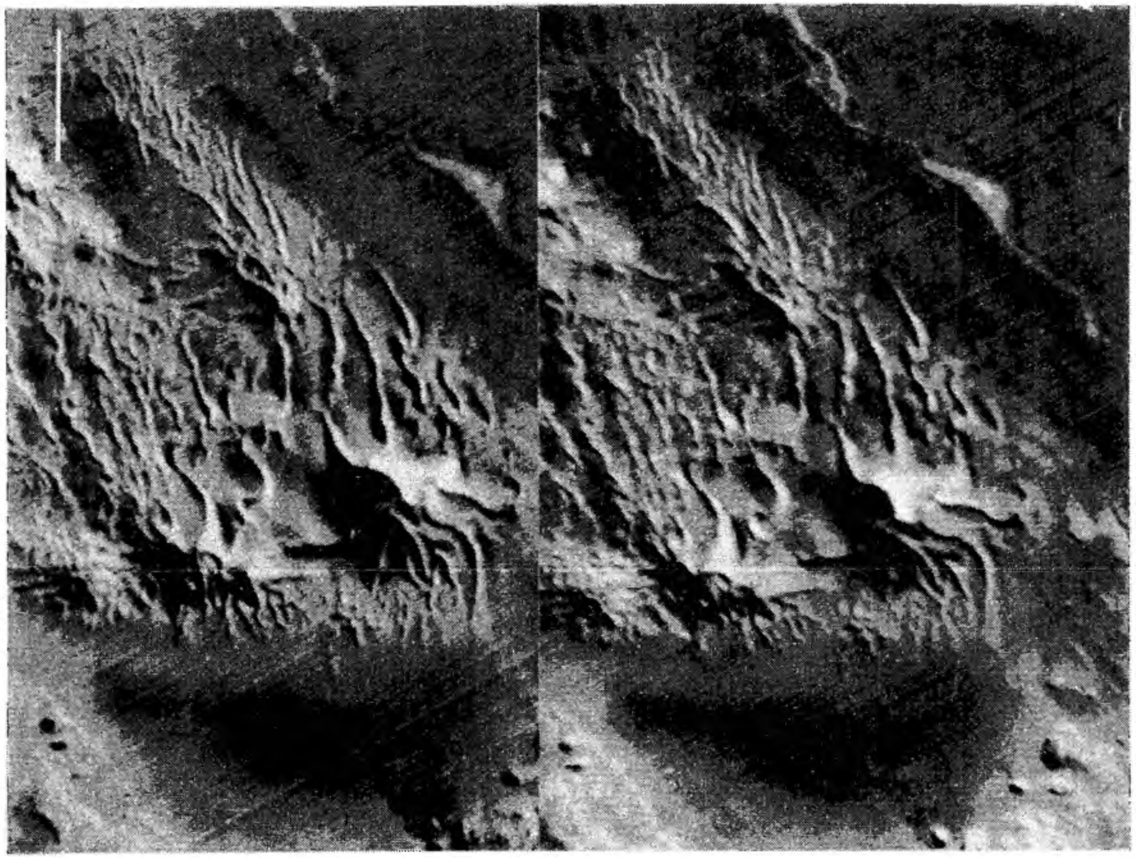


Fig.17 Stero pair of star dunes in Ramlat Zaqqut. MOMS-02, orbit 91, scene 23, channel 6 (left picture) and channel 7 (right picture).

the image occupies an area between 130 and 180 DN with a peak at 152 DN. The histogram data indicate that the spectral responses from the terrain covered by sand are very limited in DN when comparised with other areas with a wide range of land cover types. The range of values cover some 50 counts for each band. The similarity between the spectral bands 2 and 3 is very obvious. The peak of DN for band 1 has the lowest value.

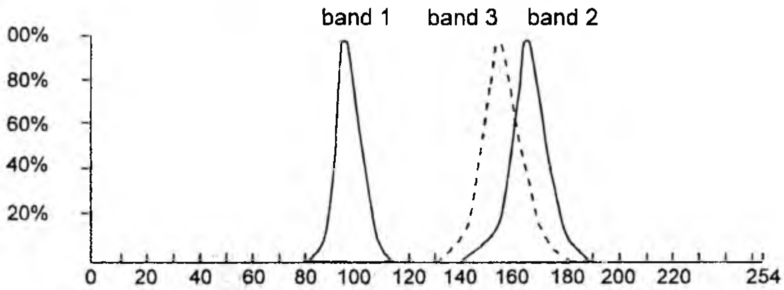


Fig.18 Histogram of spectral bands of SPOT-XS
scene 72-294 taken on 17 April 1987 over Idhan Awbari.

A comparison of images recorded in the different SPOT channels is shown in Fig.19a, 19b and 19c. The analysis of image contents and their validity for sand desert investigations indicated that the amount of information registered by channel 3 is the most appropriate for such studies. In this channel, both the dune and interdune areas are very well depicted. For comparison, on Fig.19d, the Metric Camera image of the same part of sand sea is included.,

The digital processing of SPOT data was performed on an Intergraph MicroStation-ISI-2. The analysis of six combinations of spectral bands indicated that the interpretability can be best achieved when the multispectral image is composed from the channels 1, 2 and 3, presented respectively as B,G and R.

The full scene, digitally processed and presented at a scale of ca 1:250,000 on the monitor screen, gives a general view of sandridges, their spacing and the interdune areas (Fig.21 - right photo). The full SPOT image provides the best information concerning the sand colour and sand reflectivity. It is quite obvious that there are large differences in the spectral responses of the sand cover, reflecting the different sand sources, different mineralogical composition and different grain size characteristics. There is a very significant belt of much brighter spectral responses from sand in the north - western part of the image, and, in the furthest north - eastern part, in the comparison with the other parts of the image. The sand covered interdune areas and those sand - free interdune areas where the bedrock is exposed could easely be distinguished. In the interdune areas, brackish areas, salt lakes and other types of surface water bodies can easily be detected.

For detailed analysis, six subscenes, each related to the particular parts of sand sea and types of sand dunes, were selected. The scale of these subscenes on the system monitor was about 1:66,000. On each subscene smaller representative areas were chosen and presented on the screen at a scale of ca 1:33,000.

The main aim of this analysis was to evaluate the improvement of geomorphological interpretation of all the elements comprising sand desert morphology (i.e. the interrelationship between sandridges, secondary forms, interdune areas and zibars). As the result of digital processing and enhancement of SPOT data it was possible to detect all these forms, their relaive abundance as well as sand characteristics.

Typical longitudinal sandridges are depicted on Fig.22. The very significant differences between western and eastern flanks of these sandridges are clearly visible. The eastern, downwind slopes are steeper and display lighter tones. The line which delineate the dune base and interdune area is very easily defined. The western, more gentle slopes are covered by the secondary dunes formed in crescentic chains. In the uppermost part of sandridges, the crescentic chains are very densely spaced, wheras towards the base of sandridge, the spacing is wider. The line dividing the dune flank and interdune area is not very well defined, because the crescentic chains extent towards the interdune area. In the uppermost part of sandridges, the crescentic chains are oriented nearly parallel to the dune crest. Towards lower parts of the western flanks, the trend of crescentic chains is more oblique and the elementary crescents of the crescentic chains are larger and more curved. The crescentic chains are enhanced by shadow effects which thereby give the information concerning the heights of the secondary dunes. The uppermost parts of sandridges appear in lighter tone than the dune flanks, thus indicating the finer grain size of sand incorporated into this part of dune.

Zibars, which occur in the interdune areas are easily detected on the SPOT images. As they are indicators of strong winds, they constitute very important element of the sand desert morphology. On the southern most part of the image, where the interdune areas are sand - free, the rocky basement is very well exposed. The different colours reflect the characteristic of the component rocks.

The investigations performed on the basis of SPOT images of Idhan Awbari indicate a sequence of changes in the sandridge forms and in their relationship to secondary forms and interdune areas. In the southern part of the SPOT scene (Fig.22), the sandridges are fully developed and are separated by the interdune areas in which the bedrock is exposed. Towards the north, the interdune areas are covered by sand on which zibars can be detected. The longitudinal sandridges appear as segmented ridges with well defined peaks located on the south - western parts of these segments, and saddles dividing the sandridges into segments. The peaks of the segmented sandridges are enhanced by the white

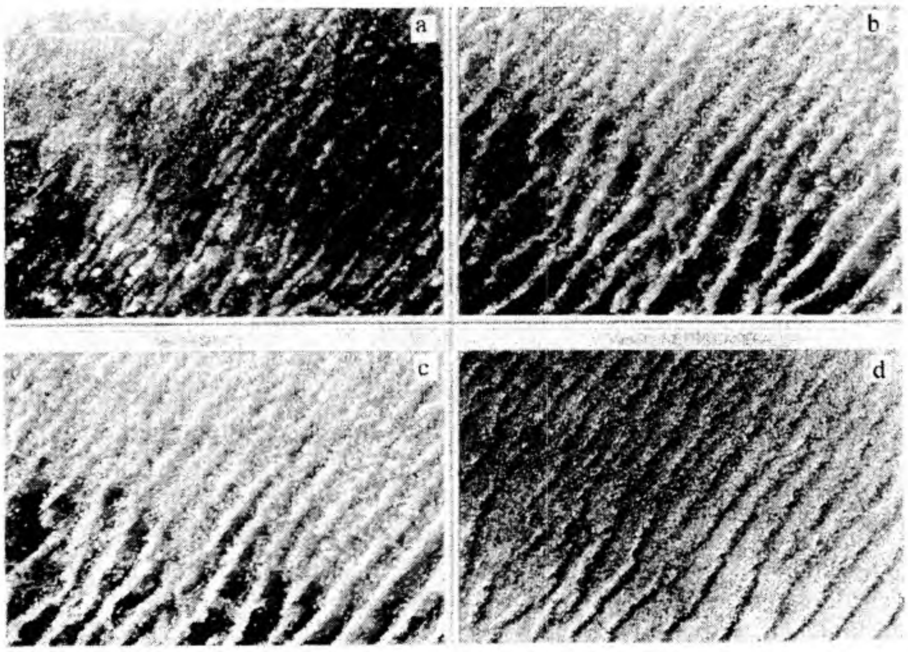


Fig.19 A comparison of images recorded by
SPOT (channels 1, 2 and 3) and Metric Camera.
(White bar represents ten kilometres)

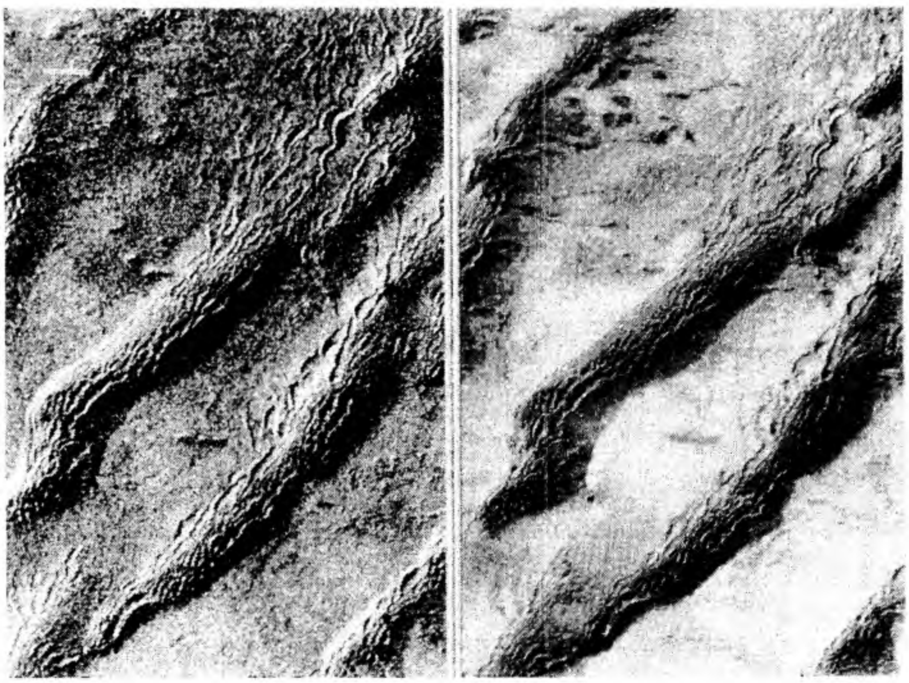


Fig.20 A comparison of Metric Camera image (left) and SPOT image (right).

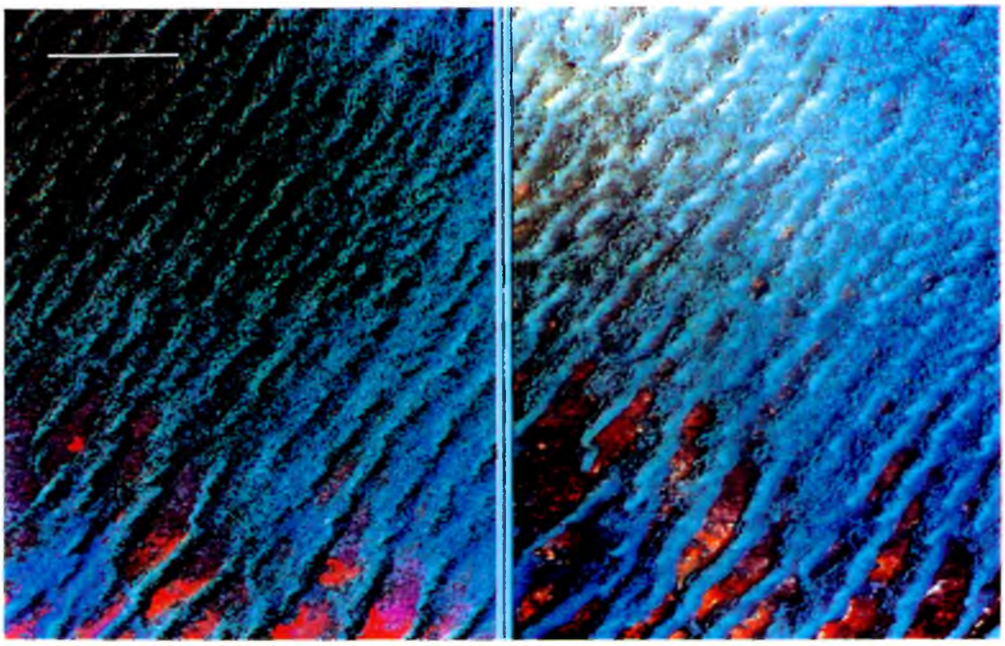


Fig.21 Right image -SPOT colour composite of Idhan Awbari
(scene 72-294 taken on 17 April 1987),
left image: superposition of SPOT and Metric Camera image.
(White bar represents 10 kilometres)

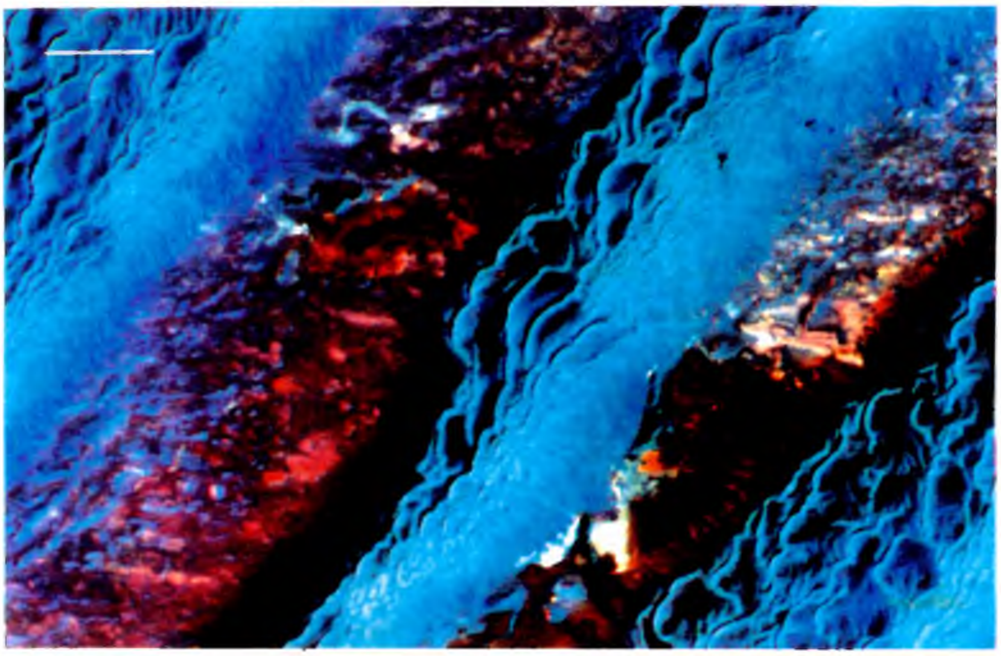


Fig.22 SPOT colour composite image - longitudinal sandridges in Idhan Awbari.

tone of the image, thus indicating the sand composition and grain characteristics of that part of the dune. Together with the changes in primary forms, the secondary forms also change their character and appearance. The secondary forms, represented by crescentic chains, are located both on the flanks of sandridges and in the interdune areas. Towards the north the segmented sandridges are transformed into more domical forms on which the secondary forms are very densely spaced. The summits of these forms are very bright.

2.2.6. Shuttle Imaging Radar-SIR-A

2.2.6.1. System characteristics

The Shuttle Imaging Radar - SIR A, formed part of the scientific payload of the Space Shuttle launched in November 1981. The main objectives of this mission were:

a) to explore areas which are difficult to study using visible and thermal sensors,

b) to carry out geoscientific investigations which require a radar system.

During 2.5 days mission of Columbia STS-2, which started on 12 November 1981, SIR-A system was in operation for 8 hours and collected data from 10 million sq km of the Earth's surface. The Space Shuttle orbit had an inclination of 38°÷40° and the altitude of 245 km. The radar imaging was acquired between 41°N and 36°S and covered both the ocean and land. Strip 29/30 covers a track across the Africa from Tobruk in Libya in SW direction over Libya, Algeria, Mali and Guinea (Fig.11).

SIR-A radar is a synthetic aperture radar operating in the L-band at 1.278 GHz corresponding to a 23 cm wavelength. The antenna of 9.5 m length was assembled relative to the shuttle at an angle of 47° to the left of the subnadir track corresponding to an incidence angle of 50° at the centre of the swath. The imagery had a resolution of the order of 38 m. Horizontal polarization was used on both transit and receive (HH). The SIR-A data were optically recorded onto film carried by cassette and then, after processing, optically converted onto 13 cm film. The scale of the radar image is ca 1:500,000, thus the imagery covers a swath of ca 65 km. In SIR-A the new higher incidence angle of 50° (in Seasat the incidence angle was 22°) reduced the effect of slopes and topography and increased the backscatter of surface texture, thereby improving the interpretation.

2.2.6.2. Backscatter in sand desert areas

The radar images are a function of the amount of energy backscattered from the terrain surface. The factors which control the radar backscatter depend on system parameters and terrain surface characteristics. The system parameters which influenced the backscatter are: *look direction, look angle and system wavelength*. In the case of the spaceborne radar, the look direction is determined by the orbit parameters and is pre - determined for complete global coverage. The look direction is generally not important in the observation of targets and surfaces without strong directional elements, but in the case of the sand desert areas, with sand fields covered by linear dunes or longitudinal sandridges, the look direction effects can be very significant. The linear dunes or longitudinal sandridges can be recognized on space radar images, providing they are oriented favourably to the radar illumination. Field studies performed by Blom and Elachi (1981) indicated that linear features, which are more than 60° from perpendicular to the illumination direction, will not be visible unless they have considerable topographic variability.

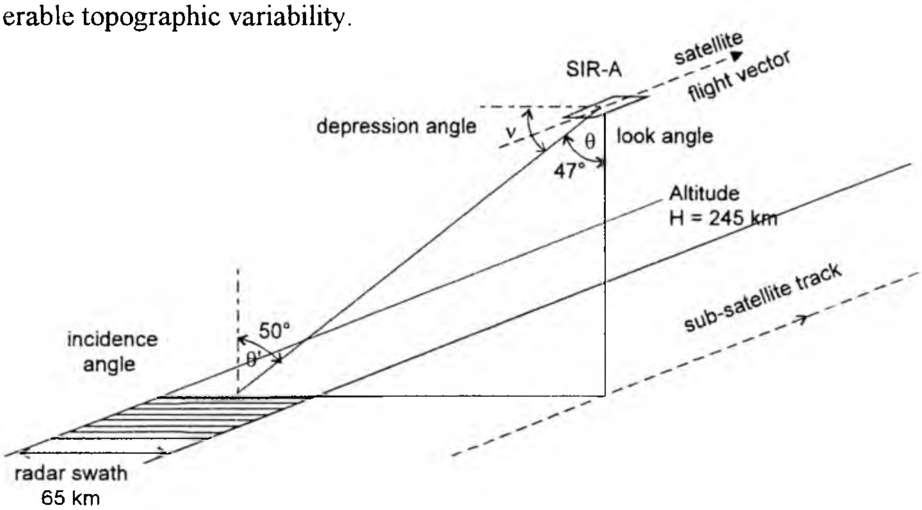


Fig.23 SIR-A geometry.

The system look angle θ is connected with the *incidence angle* ϕ (Fig.23). For a spaceborne radar, the incidence angle is slightly larger than the look angle, owing to earth curvature. The variation of the look angle along the image is minimal due to the great distance to the earth surface. The *radar reflectivity* of the surface is a *function of the look angle*. The backscatter curves for different terrain surfaces (Blom, 1988) indicated that the backscatter differences have the largest values for low look angle (~25°), but stay nearly constant at the larger angles. The backscatter constrast is easier to detect at smaller look angles.

The system wavelength should be analysed in connection with terrain characteristics. The terrain characteristics which control radar backscatter are: *surface roughness, surface slope and dielectric constant* (primary moisture content). From these factors only surface roughness response is strongly correlated with wavelength. Surface roughness is a relative concept depending upon wavelength and incidence angle. The surface roughness is expressed by the Rayleigh roughness or flatness criterion (Trevet, 1986). In accordance with the incoherent scattering roughness criterion, a surface would look smooth for radar sensors if:

$$h \leq \lambda(8\cos\phi)^{-1};$$

or rough if:

$$h > \lambda(8\cos\phi)^{-1};$$

where:
h - mean height of surface variations,
λ - radar wavelength,
ϕ - local incidence angle.

For SIR-A with $\lambda = 23.5$ cm and $\phi = 50°$ (for flat surface), h = 43 mm, and for $\phi = 17°$ (referring to local dune slope of 33°), h = 31 mm.

This criterion is very important in the radar imagery of sand areas, especially of sand sheets and interdune areas. The sand sheets form the widespread blankets of aeolian sediment, and, as a whole, have biomodal size distribution (Maxwell, 1982). The sand represents poorly - to moderately - sorted, very fine - to fine - sand and very coarse sand - with grain size not exceeding 2 mm. The values of the mean height of surface variation, for various satellite radar systems (Table 3.), indicate that the *sand sheets* fulfil the roughness criterion, and *can be regarded as smooth surface* for radar sensors.

Table 3

Radar system	Frequency (GHz) (band)	Wavelength in cm	Look angle θ	Surface roughness $h_c < \lambda\ (8\cos\phi)^{-1}$ For flat surface (for $\phi = \theta$) in mm	For slope surface (for $\phi = 0$) in mm
SIR-A	1.28 (L)	23.5	50°	43	-
SIR-B	1.28 (L)	23.5	15°- 60°	30 - 59	29
Seasat	1.28 (L)	23.5	22°	32	29
Almaz	3.0 (S)	8.6	30° - 60°	12 - 22	11
ERS-1	5.3 (C)	5.7	23°	8	7
ERS-2	5.3 (C)	5.7	23°	8	7
JERS-1	1.3 (L)	23.1	35°	35	29
Radarsat	5.3 (C)	5.7	20° - 45°	8 - 10	7

Generally, smooth surfaces will reflect radar energy away from sensors in a specular manner; hence, resulting in a dark image tone. The backscattering of the sand surface is generally specular, because the sand surface irregulations are the inadequate to cause incoherent scattering. The sand sheets and interdune areas appear in dark tone on radar images and contrast with rough surfaces which are strong, diffuse scatterers and are presented in very bright tones on radar images. Inland water bodies, which tend to be relatively smooth, generally appear in dark tone on radar images.

Another radar - scattering mechanism is the so called Bragg scattering (Blom and Elachi, 1981), related to surface undulations. Bragg scattering will take place if the surface ripples have a wavelength Λ equal to:

$$\Lambda = \lambda(2\sin\phi)^{-1};$$

In the case of SIR-A, λ = 23.5 cm and the local incidence angle varies from 50° for flat desert surface to 17° for dune local slope, corresponding to the repose angle. The equivalent values for Λ are 153 mm and 402 mm. Bragg scattering will occur only if the sand ripples have a wavelength larger than 153mm or 402mm, but the measurements performed by Sharp (1963) indicated that the wavelengths of sand ripples are generally between 7 and 15 cm; hence, Bragg scattering from sand ripples will not take place.

The reflective ability of any target is related to its moisture content or the magnitude of the dielectric constant. The relative dielectric constant of a target material consists of a real part often referred as relative permittivity and an imaginary part referred to as the loss factor (Trevett 1986). Both parts are highly dependent on the moisture content of the material considered. In the microwave region, most materials have a dielectric constant around 3 to 8 in dry conditions (Radar Imagery, 1993). When the moisture content is high, the radar reflectivity is larger.

The moisture content of sand sheets and dunes may influence the image tone, but this factor is relatively small. Generally, dune moisture never exceeds 6% at a depth of 10 cm, and in the Libyan Sahara it is much lower. The moisture variation, if any, has a small effect towards image tone variation. On the contrary, in the Idhan Awbari the desert lakes and sabkhas appear very distinctly. For water, the dielectric constant is equal to 81 and is at least 10 times higher than that for dry sand. As a result, all sabhas and desert lakes are depicted very well in very bright tone and can be very easy detected as the dark tones of sand sheets and interdune areas.

The backscattering from the dune fields depends on dune forms and their orientation to radar illumination. The radar energy is backscattered in a specular manner from the sand dunes. The effect of backscattering depends on the orientation of the slip faces and their angle. In the case of sand dunes the maximum angle of the slip face is equal to the angle of repose, which is about 33°, therefore the radar backscatter from the dune face is possible only at the look angle - less than 33° (Fig.24c). The highest backscattering effect will be in the case when the dune slope is normal to the radar illumination (Fig.24a). At a small look angle, dune slopes oriented towards radar illumination have strong reflection. In such a case, the foreslopes are highlighted, while the backslopes are darkerned.

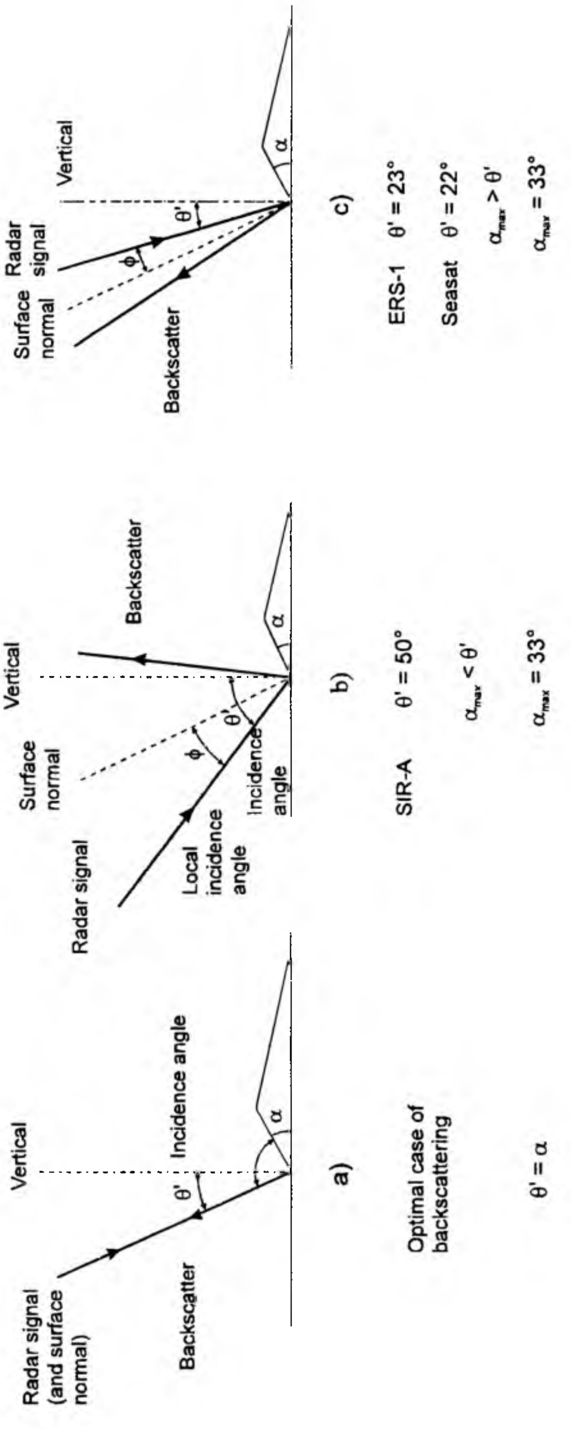


Fig.24 Backscattering effect (α - dune slope, θ' - incidence angle, ϕ - local incidence angle).

In the case of small look angle radar systems (e.g. Seasat), the sub - pixel terrain features can be detected if they produce strong echos. In the case of SIR-A, with a higher look angle, the strong radar echo will never occur, because the backscatter is not oriented towards the radar receiving antena (Fig.24b). In such a case, dunes with smooth faces will not be depicted on radar images. Nevertheless, if the dune faces have secondary dunes or large ripples superimposed over them, then the surface of dune faces will not be smooth and a backscattering effect will occur. Dune irregularity, as expressed in peaks and saddles, will also cause the backscattering effect.

The detection of sand forms is strongly dependent on the resolution of radar system. The results of research performed by Blom and Elachi (1981) indicated that the linear dunes or longitudinal sandridges will be recognized when the spacing between them is more than about 3 to 4 resolution elements. In the case of star dunes, domical dunes or barchans, the size of such dune forms should be about 10 or more resolution elements in each direction. On the SIR-A images, the minimal recognized spacing for linear dunes will be between 114 m and 152 m, and for other dune forms their dimensions should be larger than 380 m.

2.2.6.3. Radar subsurface imaging

The ability of the microwave to penetrate certain cover targets and return signals from targets below the surface is one of the advantages of radar remote sensing. The *depth of penetration is proportional to the wavelength of the radar signal.* The subsurface imaging will occur only when the topographic surface is radar smooth and the subsurface is rough. Further, the penetrated material must be fine grained, very dry and not too thick. All these conditions must be met simultaneously (Blom *et al.*, 1984). They are fulfilled in the case of sand sheets, which consist of fine grain material and create a smooth surface. Usually, the thickness of sand sheets is very limited (up to 2÷3 m), but the experiments indicated that penetration of sand cover occurs when the thickness is below 2 m (Blom *et al.*, 1984). The sand to be penetrated should be homogeneous and extremely dry, i.e. with less than 1 percent of moisture.

The losses from the scattering of radar signals are small when the grain sizes are less than one - tenth of a wavelength, but are considerable for a grain size larger than one - fifth of the wavelength (Roth and Elachi, 1984). For L-band radar systems, the grain size of cover material should be less than 1.5 cm (medium pebbles or smaller) and, particularly for SIR-A, the grain size should be smaller than 2.3 cm.

The electromagnetic wave penetration into an object is an inverse function of water content. The moisture content of sand is correlated with the thickness of sand cover and the power skin depth. The power skin depth (the depth at which power is attenuated to 37% of the value at the surface) ranges from 2 to 0.5 m where the moisture varies from 0.25% to 1% (Blom *et al.*, 1984). If the thickness of the cover material is less than the skin depth, the radar echo from the subsurface layer is substantially increased and can yield a stronger echo than if it was exposed on the surface.

Both the Seasat (Linsenbarth, 1984) and SIR-A L-band radar systems (Linsenbarth 1978), can detect subsurface features. The Seasat radar images of Means Valley in the Mojave Desert in California revealed igneous dikes, buried beneath a 2 m thick alluvium cover (Blom *et al.*, 1984) and SIR-A has revealed the paleodrainages channels of the Eastern Sahara covered by sand sheets (McCauley *et al.*, 1986). The results of the interpretation of SIR-A data taken over the Eastern Sahara have changed the previous geomorphic models of this region, revealing an erosion surface cut by Tertiary rivers long before the onset of the general aridity of the Sahara.

2.2.6.4. Evaluation of SIR-A images over the Libyan Sahara

The radar imagery registered by SIR-A in November 1981 over the northern part of Libyan territory, partially covered the eastern part of Idhan Awbari, Ramlat Az Zallaf and Ar Ramlah As Saghirah (a ca 60 km swath between 13°E and 15°E). The sand sea area is generally depicted in a very dark tone which strongly contrasts to the surrounding areas which are represented in a brighter tone (Fig.25a and b). The trends of large sandridges, which constitute the main dune forms in this area, are clearly visible. The huge sandridges which occur between 13°30'E and 14°20'E and which are oriented parallel with the orbit track are easily discerned. The sandridge faces, oriented SE towards the illumination of the radar beam, are depicted as bright areas and contrst strongly to the dark back slopes and interdune areas. Towards the east, the dune trend changes (here, dune axes are oriented ca 30° to the orbit track), but the sandridges are also well depicted. In both cases, the sandridge spacing is ca 3 km and the bright depicted slopes are 0.5÷0.7 km wide. The intensity of the bright tone along these slopes is not homogeneous, which probably reflects the local changes of dune slope and irregularities of the sandridges, expressed by peaks and saddles or by the secondary dune forms superimposed on the sandridges.

The image of sandridges located between 13°30'E and 14°20'E is different, when compared with the sandridges in the western part of the radar image. The main dune trend here is NE and the spacing varies between 2.5 and 3.5 km. The dune axes are oriented at ca 40° to the orbit track. The dune faces oriented towards the radar beam are depicted as trains of bright points, appearing with changing density along the dune slopes. The change of dune slope expression is probably caused by the change of relative orientation of the dunes to the radar signal. The bright points result from the backscatter effect caused by dune irregularities and not the dune slope itself. In the northern part, the second trend of sandridges is very easily traced, and the SW slopes are expressed as trains of bright points. In the northern part of Ramlat Az Zallaf in the vicinity of Wadi ash Shati the very narrow subparallel bright lines, spaced of ca 1 to 1.2 km intervals and trending SE can be detected. These bright lines could be interpreted as linear dunes but, due to the fact that linear dunes were not recognized in this area on Landsat images, they are probably better interpreted as ventifacts eroded on the surface.

Generally, sandridges are better depicted in the northern part of Ramlat Az Zallaf than in the southern part. Towards the southern boundary of the sand sea, the image is darker, and the dune forms depicted on Landsat images cannot be detected on SIR-A imagery. For example: the linear dunes described by McKee and Tibbitts (1964) which appear NW of Sabha are not visible on radar imagery. The dune rim pattern, very well presented on Landsat images (Fig.47 area 13), is only partly visible on radar imagery. In the eastern part of Ramlat Az Zallaf and in Ar Ramlah Ar Saghiran, the sand sea appears in a very dark tone, indicating the sand sheet. The smaller linear dunes which were detected in this area on Landsat images cannot be detected on SIR-A imagery. However, on these generally dark images of sand sheets relatively lighter areas can be distinquished, and may be interpreted as sarir areas, covered by coarser sand or gravel, thereby giving stronger backscatter signals than the homogeneous sand cover of sand sheets.

Within the sand sea area, the desert lakes and sabhas appearing in NW direction from Sabha are very obvious on SIR-A imagery, forming very bright spots. Also, the roads from Sabha towards the N and NE are very well depicted and represented by lines composed of bright points. Irrigated fields appearing in the eastern part of Ar Ramlah Ar Saghirah are very vivid. These circular shaped fields (the field diameter is ca 0.6 km) are expressed in bright tone. The two phenomena connected with radar scatter of such surfaces were recognized. The bright circles of the irrigated fields are partly superimposed, thus confirming that the signal reflected from water or very moist areas can give a signal response greater than its normal target signature. The other observed phenomenon is that, on the radar image of the circular fields, the very bright lines passing through the middle of the circle and oriented parallel to the radar beam are clearly visible. These bright lines are a little wider at both ends than in the middle of these fields.

The southern boundary of the sand sea is easily delineated. In this part of the image, the escarpment of Msak Mastafat is perfectly represented as a very bright ca 2 mm - wide line. Further to the south, wadies trending towards Idhan Murzuq and depicted in very dark tone, with a white bordering line, are also easily recognisable. Along the boundary of the sand sea, the agricultural areas with watering systems, as well as all villages and towns (eg. Sabha which is expressed in very bright tone, caused by the corner reflections from buildings and other constructions) is instantly recognisable.

The northern boundary is not as well defined as the southern boundary. The eastern part has a rather distinct boundary between the dark sand areas and rather bright sarir and sandstone areas to the north, while in the middle part of the image, approximately north to Sabha (between 14°20'E and 14°35'E), there

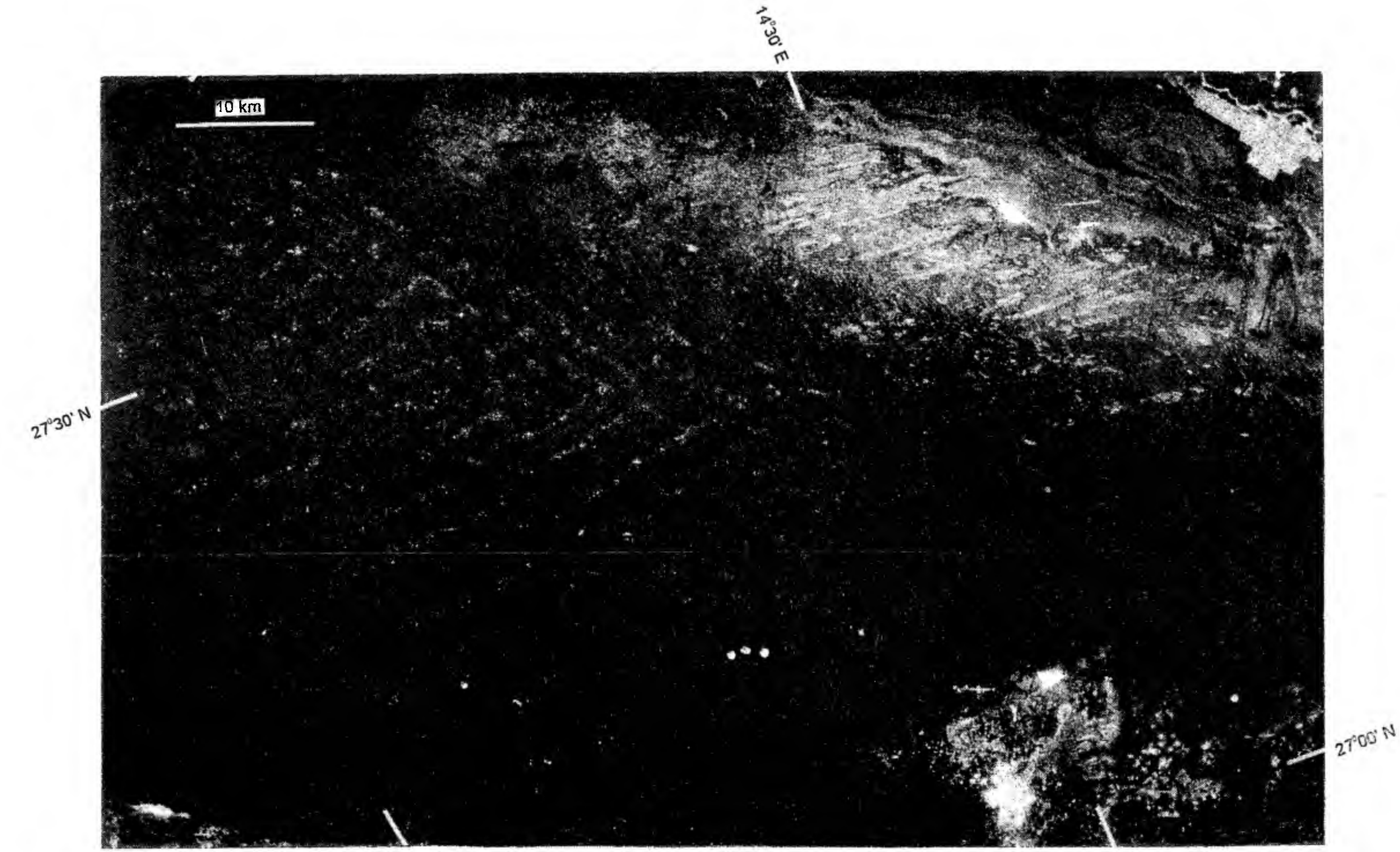


Fig.25a SIR-A image of the eastern part of Idhan Awbari - Ramlat Az Zallaf.
Bottom right - Sabha expressed in very bright tone; right top - artificial irrigated fields in Brak. (cont.of Fig.25b).

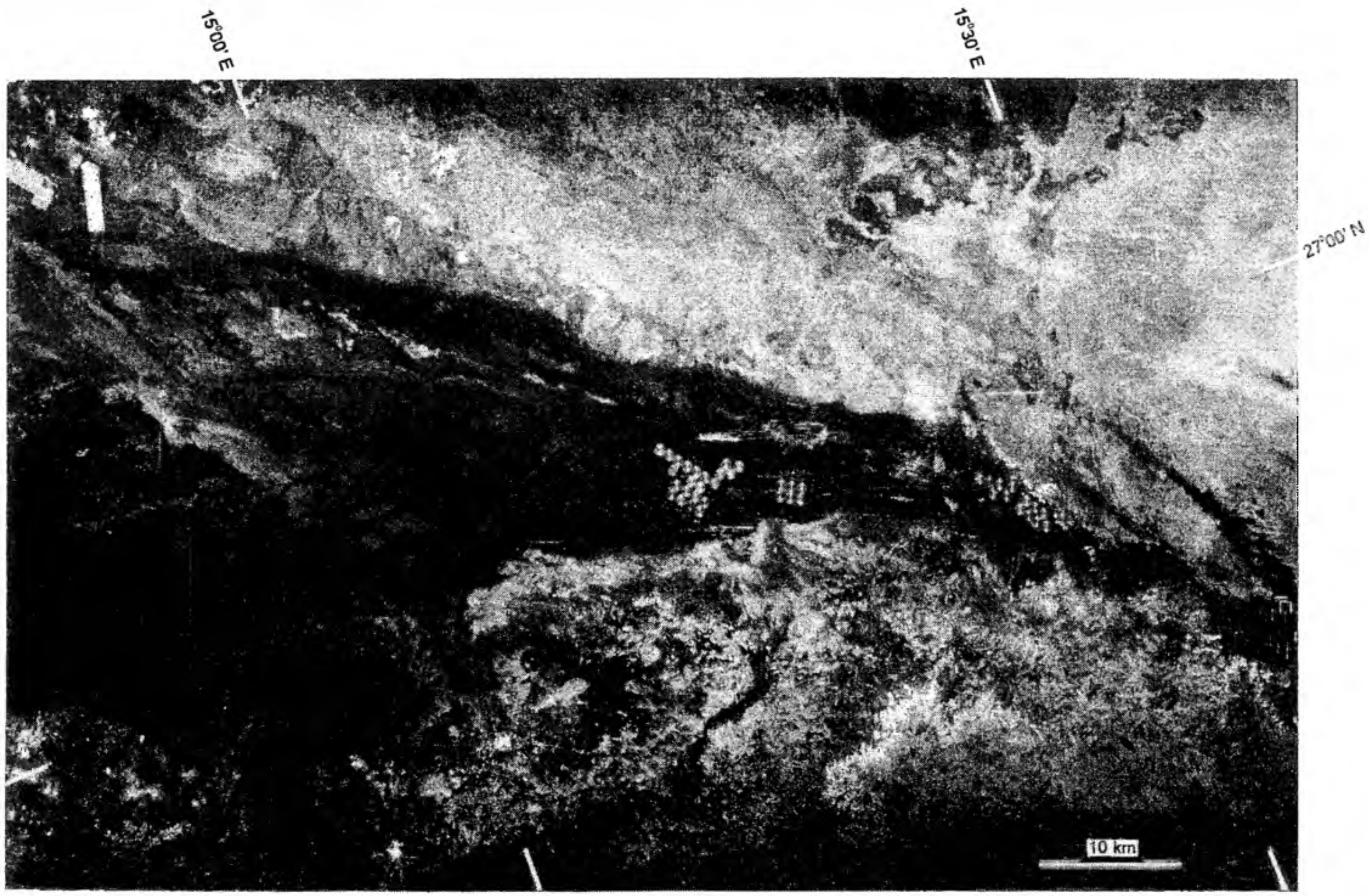


Fig.25b SIR-A image of the eastern part of Idhan Awbari - Ar Ramlat Az Saghirah. Left side - road between Sabha and Brak depicted by line of bright spots. Middle part - circular irrigated fields in Ar Ramlah ar Saghirah. (cont.of Fig.25a).

is a transition area between sand sea and its surrounding area. Here, subparallel very long ventifacts caused by wind erosion in the bedrock material are visible. The trend of these ventifacts is almost the same as the trend of sandridges in the southern part of the sand sea. The same strong winds were probably responsible for the creation of these ventifacts and the longitudinal sandridges. The ventifacts, distinquished on radar images, are only partly shown on the existing geological maps. In this part of thc image, as well as the agricultural areas, wadies are also presented in dark tones, and sabhas appear in very bright tone.

2.2.7. Comparison and superposition of SPOT and Metric Camera data

SPOT and Metric Camera data were used both for the comparative evaluation of both sources of data and for the detection of temporal changes (see chapter 2.2.8). The merging of the SPOT XS image with Metric Camera photograph, converted to digital form, was carried out by application of RGB to IHS transformation (Fig.26).

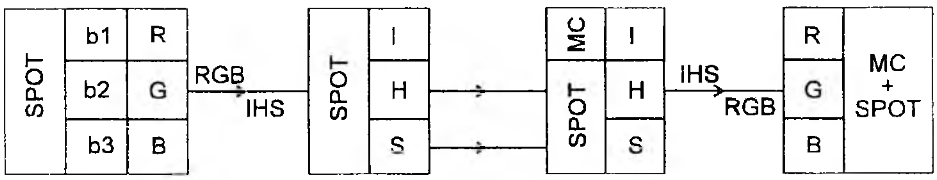


Fig.26 Merging of SPOT and Metric Camera images - RGB to IHS transformation.

In the IHS (Intensity, Hue, Saturation) representation, all colour information is condensed into two bands, while, in the third band, the intensity information is concentrated. By definition, the intensity band is almost the same as a panchromatic band, bacause the *Intensity* is the total intensity of the colours as obtained by adding up all grey levels of the composing bands (RGB) and representing it in the 0÷255 scale. The *Hue* indicates the colour range as occuring in the rainbow, and *Saturation* is the percentage of colour. In the applied transformation, SPOT XS bands defined in RGB colour system were transformed to the IHS colour system. Then, the intensity band was exchanged with the panchromatic image taken by the Metric Camera. Finally, the new IHS image was retransformed into the normal RGB image, because the screen of the Intergraph MicroStation is made for RGB displays.

The comparison of the Metric Camera and the SPOT images (Fig.20 and 27) indicated that, on the Metric Camera image, the primary forms represented by sandridges are perfectly depicted, whereas the secondary forms and zibars are missing. By contrast, on the SPOT image the general forms of sandridges are not enhanced but the secondary forms and zibars are clearly visible. On Fig.28 three images are presented: Metric Camera, SPOT-channel 3 and reversed image of SPOT-channel 3. As can be seen, on the Metric Camera image the general forms of the sandridges are depicted perfectly, while, on the SPOT image, the 3-D impression is missing. The inversed image of SPOT channel 3 greatly improves such an impression. Generally, the analysis indicated that Metric Camera images can be used for geomorphic investigations of sand deserts on the regional level while SPOT images are particularly useful for more detailed studies carried out on a local level.

The Metric Camera stereo images (708 and 709) were used for the measurements of spot heights on an analytical plotter Planicomp P-1 for the area corresponding with the selected subscene of the SPOT image. The spot heights were limited to the dune crest and interdune areas. The determined spot heights were then superimposed onto the SPOT image. In this part of Idhan Awbari, the crests of sandridges are elevated ca 120 m over the interdune areas. The heights of interdune areas oscillate between 470 and 500 m a.s.l.

2.2.8. Investigation of sand desert dynamics

Sand desert dynamics can be investigated on the basis of multitemporal remote sensing data or on the basis of new remote sensing data relating to existing photogrammetric or cartographic materials. The changes in dune morphology can be detected only in the case where these changes are greater than the resolution of remote sensing data. Change detection analysis was carried out on the basis of three data sets:

- Metric Camera data from 1983 versus photomosaics from 1957,
- SPOT data from 1987 versus Metric Camera data from 1983,
- MOMS-02 data from 1993 versus Metric Camera data from 1983.

2.2.8.1. Comparison of Metric Camera data from 1983 with photomosaics from 1957

For the evaluation of the changes in sand dune morphology, the space photograph No.708 taken by the Metric Camera on 5 December 1983 over the western part of Idhan Awbari and photomosaics at scale 1:50,000 compiled in 1958 from aerial photographs taken in 1957 were examined. A detailed comparative analysis was performed on four sheets of photomosaics partly covering the space photograph. The change detection was carried out on a Zeiss Jena Kartoflex, on which the space photograph image can be superimposed onto the photomosaic. That part of Idhan Awbarii being investigated is characterized by huge longitudinal sandridges which are covered, on their western slopes, by

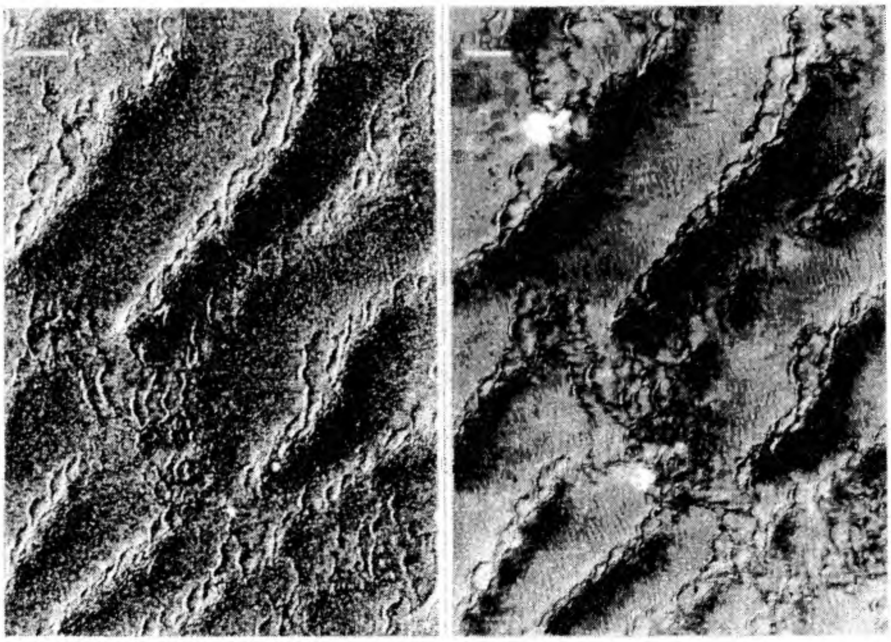


Fig.27 Comparison of SPOT image (right) with Metric Camera photograph (left).

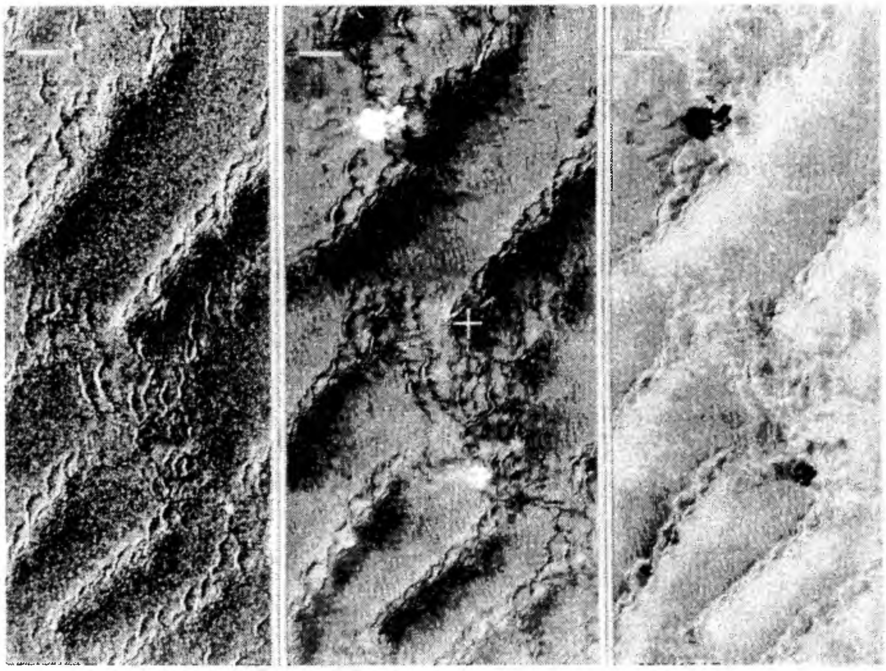


Fig.28 Comparison of Metric Camera photograph (left), inversed image of SPOT-channel 3 (middle) and SPOT-channel 3(right).

secondary forms represented by crescentic chains and linear dunes. The detailed comparative analysis indicated only very limited and very small changes in the secondary forms, which occurred mainly in the forms present in lower parts of sandridges and in the interdune areas. Generally, no significant changes in secondary forms were detected over this 30 year period.

2.2.8.2. Comparison of SPOT images from 1987 with Metric Camera photographs from 1983

The SPOT scene 72 ÷ 294, acquired in multispectral mode on 17 April 1987 over the western part of Idhan Awbari was used for comparison with Metric Camera space photograph No 708 taken over the same territory on 5 December 1983. The Metric Camera photograph was converted into digital form on a photogrammetric scanner PhotoScan P-1. For the digital processing of both sets of data an Intergraph MicroStation was employed. The Metric Camera data were used as references. After the transformation process, the comparative analysis was performed using two methods. In the first, the SPOT image was superimposed onto the Metric Camera image, and in the second method, both images were presented on the left and right part of the monitor screen respectively. Generally, no changes in the period of 3 years were detected in the case of both in the secondary and in primary dune forms. The only significant change was detected in the northwestern part of the SPOT image (Fig.27), where a small desert lake was recognised which was not discernible on the Metric Camera photograph. In the same place, on the SPOT image where the desert lake is located, a semi - crescent linear dune is clearly visible on the Metric Camera photograph.

2.2.8.3. Comparison of MOMS-02 images from 1993 with Metric Camera photographs from 1983

The scene 23, acquired by MOMS-02 from the orbit 91 in mode 1/30 in May 1993, and space photograph No 158, taken by the Metric Camera on 2 December 1983, were used for detection of time changes in the dune morphology in the northern part of Ramlat Zaltan. As in the case of previous examples, the Metric Camera photograph was converted into digital form, and then both sets of data were processed on an Intergraph MicroStation.

The detailed comparative analysis indicated several changes in dune forms. On the lefthand side of Fig.29a a new trend of linear dunes can easily be recognised. The new dune trend is presumably related to the changes in the wind regime in this region. Significant changes were also detected in the star dunes surrounding the horseshoe - shaped depression in Ramlat Zaqqut (Fig.29b). The changes in the star dune arms are quite obvious.

The results of these investigations have indicated that there are significant differences in the desert dynamics between Idhan Awbari and Ramlat Zaltan. In Idhan Awbari, the detected changes are few and very small, whereas, in Ramlat

Zaltan, they are quite significant. Idhan Awbari belongs to a very old sand sea on which huge sandridges were developed through several cycles of dune development. The small changes (detected only in the interdune areas and and in the lower parts of sandridges) may be related to the secondary winds generated by the appearance of the huge sandridges. On the contrary, the northern part of Ramlat Zaltan, and particularly Ramlat Zaqqut, belong to the modern sand seas in which contemporary processes of dune development occur and the primary surface winds play the most important role in dune development.

2.2.9. Conclusions and recommendations regarding remote sensing in sand desert investigations

The investigations carried out in the Libyan Sahara case study using various remote sensing data and techniques demonstrate their usefulness in geomorphological studies of sand desert areas. A comparison of various remote sensing data and techniques applied to the case study leads to several conclusions and recommendations regarding the role of remote sensing in sand desert investigations.

The remote sensing systems of data acquisition applied in these investigations can be classified into the *passive* remote sensing systems operating in visible and infrared portion of electromagnetic spectrum and the *active* systems operating in the microwave region. Due to the different characteristics of these two systems and image properties, the two systems will be discussed separately.

The remote sensing data acquired by passive remote sensing systems are based on various techniques of data acquistion such as: multispectral scanners (Landsat MSS and Landsat TM), CCD remote sensing systems (SPOT, MOMS-01 and MOMS-02), and satellite photogrammetric cameras (Metric Camera on the board of Space Shuttle). Data acquired by Landsat MSS and TM, SPOT and MOMS-01 and MOMS-02 systems are recorded in a digital form suitable for digital processing. Metric Camera photographs can either be elaborated in analogue or analytical method or can be converted into digital form by a scanning procedure.

For the geomorphic investigations of sand desert areas, the most important characteristics of remote sensing sensors are: *spatial resolution*, *spectral resolution* and *radiometric resolution* or *radiometric sensitivity* (Mather,1987). The spatial resolution, called also geometric resolution, is the main factor which allows the detection of the various dune forms and the relationship between compound, composite and simple dune forms as well as the morphology of particular dunes. The comparative analysis of different remote sensing data clearly confirmed that medium - resolution satellite data, such as Landsat MSS permits the delineation of the boundaries of sand seas and the identification of the homogeneous dune fields and the mapping of the dune trends and patterns. The higher resolution data (SPOT-XS and Metric Camera) permits the identification of the smaller dune forms, secondary dunes superimposed on sandridges and the

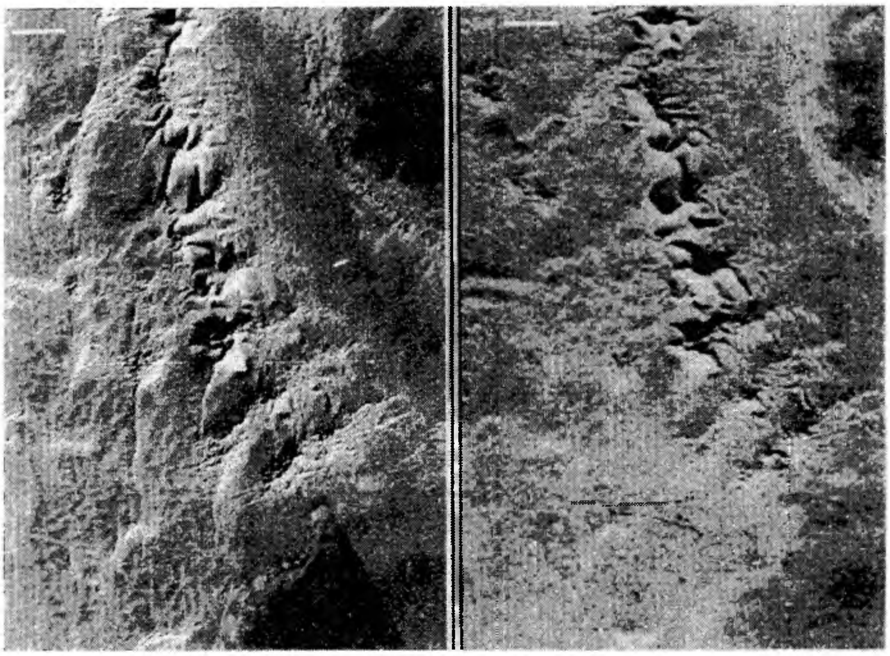


Fig.29a Changes in dune forms in Ramlat Zaqqut.
Left image: MOMS-02, orbit 91, scene 23a, ch.5. May 1993.
Right image: Metric Camera, photo 158, strip 4, December 1983.

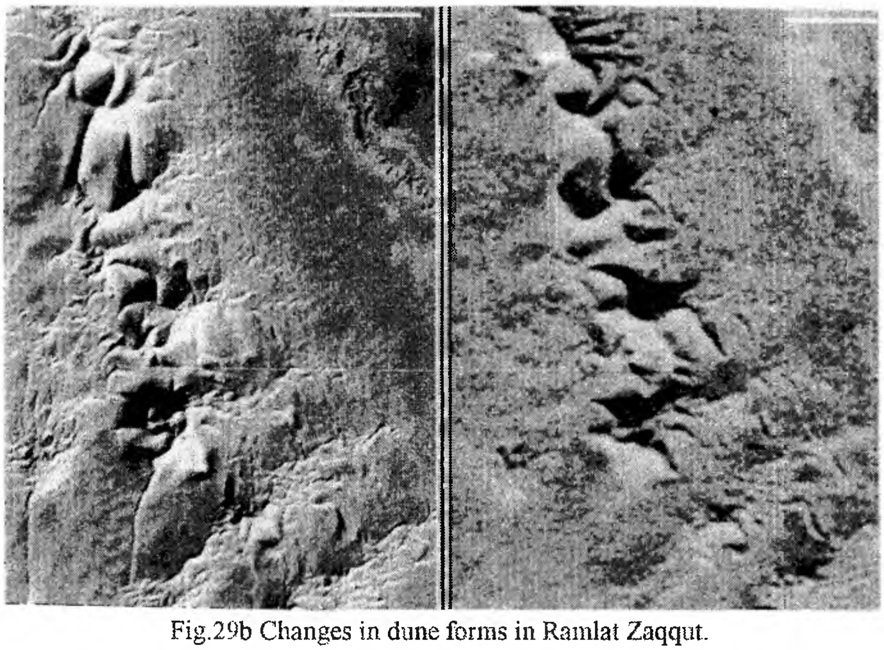


Fig.29b Changes in dune forms in Ramlat Zaqqut.
Left image: MOMS-02, orbit 91, scene 23a, ch.5 May 1993.
Right image: Metric Camera, photo 158, strip 4, December 1983.

zibars in the interdune areas. The MOMS-02 data registered in channel 5 with 4.5 resolution present the highest resolution images. These are very suitable for sand desert investigations.

The term spectral resolution refers to the position of spectral bands in the spectrum and to the width of spectral bands. The digital images collected by the satellite sensors can be multispectral or multiband; i.e. the images are separately recorded in the discrete spectral bands. The number of spectral bands and parameters of the selected spectral bands can greatly improve the dicriminatory power of the acquisition system. The spectral resolution of the sensor must match, as closely as possible, the spectral reflectance curve of the particular targets. The different targets have various reflectance characteristics expressed by the spectral reflectance curves; hence, the applied system should be capable of distinquishing different targets. The spectral response of the sand surface is a function of sand charcteristics such as granulometric parameters, sand parent material and sand colour. In the case of sand surface areas, the most important is the discrimination of dry and wet sand and the delineation of sand desert areas from other targets on the earth surface.

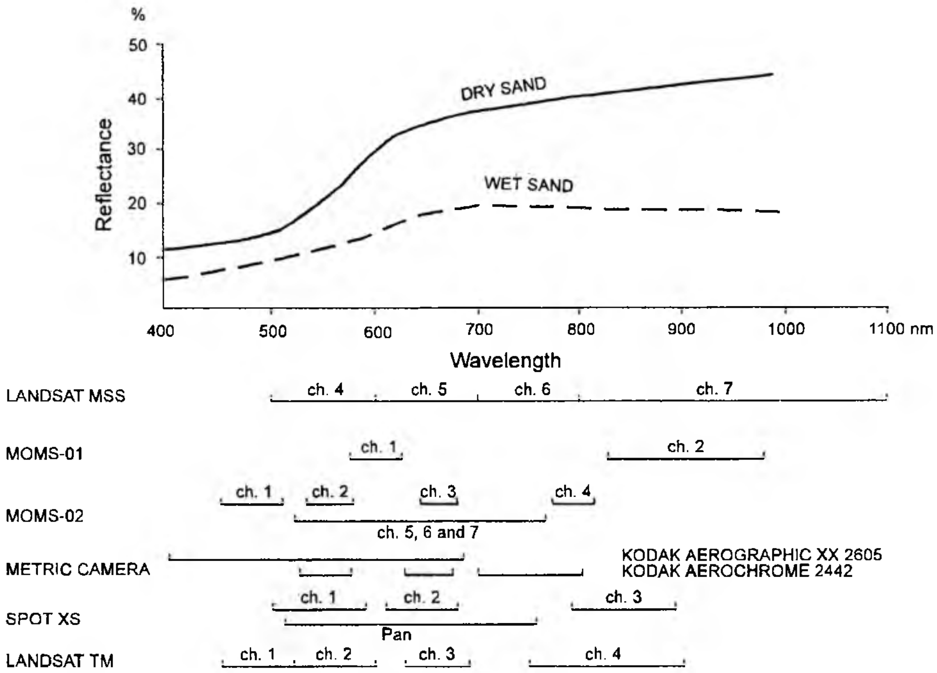


Fig.30 Spectral resolution of various remote sensing sensors and reflectance of dry and wet sand.

As can be seen from Fig.30, both the highest reflectivity and differentiation between dry and wet sand is registered in the red portion of the visible region of the spectrum (around 700 nm) and in the near - infra red region. These investigations, have confirmed that the images registered in this part of electromegnetic spectrum present the best quality and interpretability of sand surface elements (Landsat MSS: channels 6 and 7, SPOT: channel 3, MOMS-01: channel 2). The panchromatic images (MOMS-02:channels 5, 6 and 7) or space photographs (Metric Camera on b/w film) are taken in a very wide region of the spectrum; hence, the differentiation of appropriate elements is limited.

The radiometric resolution, or radiometric sensitivity, refers to the number of digital levels used to express the data collected by sensor (Mather, 1987). Generally, the greater number of digital levels increases the recorded information. The satellite remote sensing systems have 256 or 64-level imagery. The radiometric resolution of Landsat MSS images was 6 bits (64 levels), while, for SPOT multispectral mode, the resolution is 8 bits (256 levels) and 6 bits (64 levels) in panchromatic mode. In the MOMS-02 system, the multispectral channels were recorded with the full radiometric resolution of 8 bits (256 levels), while, the high resolution channel was compressed from the original 8 to 6 bits (64 levels).

The multispectral images (Landsat MSS, SPOT-XS) offer better interpretability of sand surfaces, thus permitting the determination of sand colour, thereby reflecting sand parent material and sand transport. The colour version of the Landsat Photomap of Libya at the scale of 1:250,000 led to the definition of the differences in sand colour between the Libyan sand seas and the recognition of trends in sand colour changes within particular sand seas (Chapter 2.2.1.2). The Landsat multispectral images of the north - western part of Idhan Murzuq permitted the delineation of the regions of different sand supply (Fig.8 and 9). Also, on the north - western part of Idhan Awbari, as recorded by SPOT-XS, a the zone of different sand colour was determined.

The stereo data enables 3-dimensional observations to be made, measurements and analysis of the dune forms and their relationship to terrain relief. The remote sensing systems produce stereo - images both in across - track mode (SPOT) or along - track mode. The along - track mode enables images to be recorded from the same orbit in a very short time difference, while the images registered in across - track mode are recorded from the different orbits at different times. The second mode, due to the elapse of time in data acquistion, can caused problems in the stereoscopic perception; there are due to changes in terrain coverage and to different weather conditions. In the stereoscopic measurements, the so - called *base to height ratio,* is very important. In the case of Metric Camera was this 0.3, while in SPOT images, it varies between 0.08 and 0.45. The highest accuracy of stereoscopic measurements is offered by the MOMS-02 images recorded in mode 1. In the case of the stereomodel composed of images taken in channels 6 and 7, the base - to - height ratio is 0.4 whereas in the case of the stereomodel created from images taken in channels 5 and 6 or 5 and 7, the base - to - height ratio is 0.8. The comparison of stereo images taken by the Metric Camera over Idhan Awbari and Ramlat Zaltan and stereo images registered by MOMS-02 over Ramlat Zaltan and Great Sand Sea, confirmed that the MOMS-02 images produced the more useful stereoscopic models.

To sand desert investigations the microwave remote sensing systems are evidently of little application, due to the fact that the meteorological conditions in desert areas allow the registration of good quality imges in thes regions. The observations reported here and analysis of existing radar systems indicated that several conditions relating to microwave systems should be fulfilled. The detection of sand desert forms is strongly correlated with the resolution of the radar system, wavelength and look angle. The higher resolution radar systems allow the detection of smaller dune forms; hence, the high resolution radar systems such as JERS-1, ERS-1 and ERS-2 and Radarsat are the most appropriate for desert investigations. The wavelength is correlated both with the roughness criterion and with backscattering from the dune slopes. In the case of flat sand surfaces, such as sand sheets or interdune areas, the smooth surface can have larger rms height for a longer wavelength of radar system. Radar systems operating in band L or S, such as JERS-1 or Almaz, permit on increase in rms height (Table 3). The roughness criterion is also dependent on the look angle. For the larger look angle, the rms height can be higher. Thus the systems with the larger wavelength and larger look angles can increase the criterion of smooth surface.

In the case of sand desert areas covered by dunes, both look angle and the look direction play very important roles in the backscattering effect. The roughness criterion is related to the local incidence angle, which is a function of incidence angle and local slope. The local slope will reduce the local incidence angle; hence, lowering the value of rms height. The backscattering from the dune slope will be highest, when the radar illumination is normal to the dune axis. This condition can be fulfilled only for radar systems where the incidence angle is smaller than the angle of repose i.e. smaller than 33°. This requirement is fulfilled by ERS-1 and ERS-2 and by Radarsat operating in the mode of low incidence angle.

Regardless of the characteristics and quality of the remotely sensed data, the applied techniques of data processing can improve the final results of image interpretation and analysis. Both traditional analogue methods and digital techniques were applied in the research programme. It is out beyond doubt that digital techniques, which permit the image to be enhanced, greately improved the interpretability of sand desert areas. Merging of the data from different sensors (Metric Camera and SPOT data) has resulted in the upgrading of the image value. Temporal changes and the investigations of sand desert dynamics have derived from the multitemporal satellite data.

The Libyan Sahara case study and analysis of existing and developed remote sensing systems, has led the Author to propose the following recommendations:

a) For continental or regional scale geomorphological investigations of sand desert areas medium resolution data, acquired by operational remote sensing systems such as Landsat TM or SPOT can be applied successfully.
b) For more detailed studies, conducted on a local scale, high resolution satellite data, acquired both in multispectral mode and stero mode should be used (e.g.MOMS-02 - Priroda).
c) For mapping and morphological analysis of sand desert forms, the digital technology of data processing offers the most effective approach. Digital photogrammetry methods can be used for compilation of digital orthoimages or digital orthophotomaps, which can used for geomorphological interpretation and for deriving the necessary morphometric information. A digital terrain model can greatly improve the spatial analysis of dune forms, their interrealtionship and relation to the terrain topography.
d) For the studies of sand desert dynamics, the multi - temporal stereo data, taken by the high resolution sensors, should be applied. It is highly recommended that the same system should be used and images should be registered in the same meteorological conditions. For analysing of long - term changes (tens of years) the consecutive data should be taken in the same period of the year. For the analysis of short time changes, the images should be taken before and after the period of expected changes (for example, in Libyan Sahara, before and after spring time when the winds are most effective).
e) For precise and detailed studies of dune forms and their dynamics large scale photographs should be applied. The creation of sequential DTM will allow the dynamics of dune forms to be assessed.
f) Terrestrial photogrammetry methods (Linsenbarth, 1974; Ostaficzuk, 1962, 1964; Muller and Ostaficzuk, 1971; Sitek, 1991) can be applied for precise measurements of dune development and analysis of sand transport.

3. GEOMORPHOLOGY OF SAND DUNES IN THE LIBYAN SAHARA

3.1. CLASSIFICATION OF SAND DUNES IN THE LIBYAN SAHARA

3.1.1. Review of sand dune classification

The preliminary studies of sand forms performed by the author in 1986÷87 (Linsenbarth, 1986a-g, 1987b) indicated the necessity of establishing the classification of sand dunes, with respect to forms appearing in the Libyan Sand Seas. The analysis of existing systems of dune classification showed that no one system can usefully be applied for Libyan case studies where such a large variety of dune forms appears.

Aufrère (1933) proposed a dynamic classification, based on the division of winds into: winds of conjunction, winds in opposition, multidirectional winds and incident winds, and specified the dune forms appearing in these wind categories. Mainguct (1984a) proposed that dune classification should be based on aeolian dynamics and sand budget. She distinguished:

a) active depositional dunes, including barchanic edifices and sandridges, linear dunes and star dunes,
b) erosional forms including parabolic edifices and sandridges.

Hunter *et al.* (1983) proposed a morphodynamic classification of dunes. In their classification, the orientation of an aeolian bedform with respect to the average wind regime is used as a discriminator. They assumed that the dunes have unique behaviours depending on whether they are oriented perpendicular, oblique or parallel to the long - term sand drift direction. In accordance with their definition a longitudinal dune is that where the long axis of dune is less than 15° from the resultant sand drift direction, whereas for the transverse dune it is within 15° of the normal to the sand drift direction. Others, where the orientation to wind direction ranges between longitudinal and transverse types, are classified as oblique dunes.

Wilson (1972) introduced a hierarchy of aeolian bedforms. According to Wilson, aeolian bedforms should be divided into: two types of ripples (aerodynamic and impact ripples), dunes and draas. He pointed out that there is a relationship between dune spacing and sand grain size and that there is no transition between these forms. Unfortunately the most recent investigation in this field has not confirmed Wilson's hypothesis of granulometric control (Wasson and Hyde, 1983b).

McKee (1979) divided sand dunes into: simple, compound and complex. Simple dunes were defined as individual dune forms which occur separately and have no connection with adjacent dunes. All dune forms which consist of two or more dunes of the same type, both coalesced or superimposed, were classified as compound dunes. Complex dunes, i.e. dune forms which consist of two or more different types of simple dunes, coalesced or superimposed, were the third category.

3.1.2. Proposed classification of sand dunes in the Libyan Sahara

The classification proposed by the Author was based on the assumption that, with respect to the application of remote sensing approach to the geomorphology of sand forms, any dune classification should be based on the geometric dune shapes and their relative location, both in the horizontal and vertical plane (Linsenbarth, 1987a, 1987b). Such a classification of dune forms allows us distinguish the different dune forms on satellite or airborne images or photographs, regardless of their morphogenesis. In accordance with such an assumption, the dune forms can then be divided into: *simple or elementary forms, composite forms and compound forms.*

The *simple forms*, representing elementary dune forms, are separated one from another. Dunes in *simple* or *elementary form* appear only as *barchans (crescent dunes), linear dunes* or *star dunes.* In accordance with the Author's definition, composite dunes forms are composed of simple dunes, of the same or different types, but appearing on the same level of terrain surface. Compound forms are created *by two sets of sand forms being superimposed, one onto the other.*

In the proposed classification, a distinction is drawn between the forms appearing on the terrain surface in one level (simple dunes and composite dunes) and the sand dunes consisting of *two levels of sand bodies*, here referred to as *primary* forms and *secondary* forms. The primary forms constitute a lower level on which the secondary dunes forming the upper level are superimposed.

The proposed classification of dune forms is presented graphically on Fig.31. In the middle part of this graph, the simple forms: barchans (crescentic dunes), linear dunes and star dunes are depicted. In the upper row, the different types of composite forms are presented: *hooked dunes, crescentic chains, composite linear dunes and star dune chains.* The bottom part expresses the compound forms, which comprise primary forms usually called sandridges, megadunes or draas. The basic forms of primary dune forms recognized in the Libyan Sahara are: *barchanoid sandridges, longitudinal sandridges, domical sandridges* and *reticulate sandridges.* The secondary forms represented by simple or composite dune forms can be superimposed on the primary forms.

In the proposed classification, the distinction is made between dunes and sandridges forming the compound forms. The term *sandridge* is equivalent to the term draa (used in Sahara for a large sand hill) as proposed by Wilson (1972) in his hierarchy of aeolian bedforms.

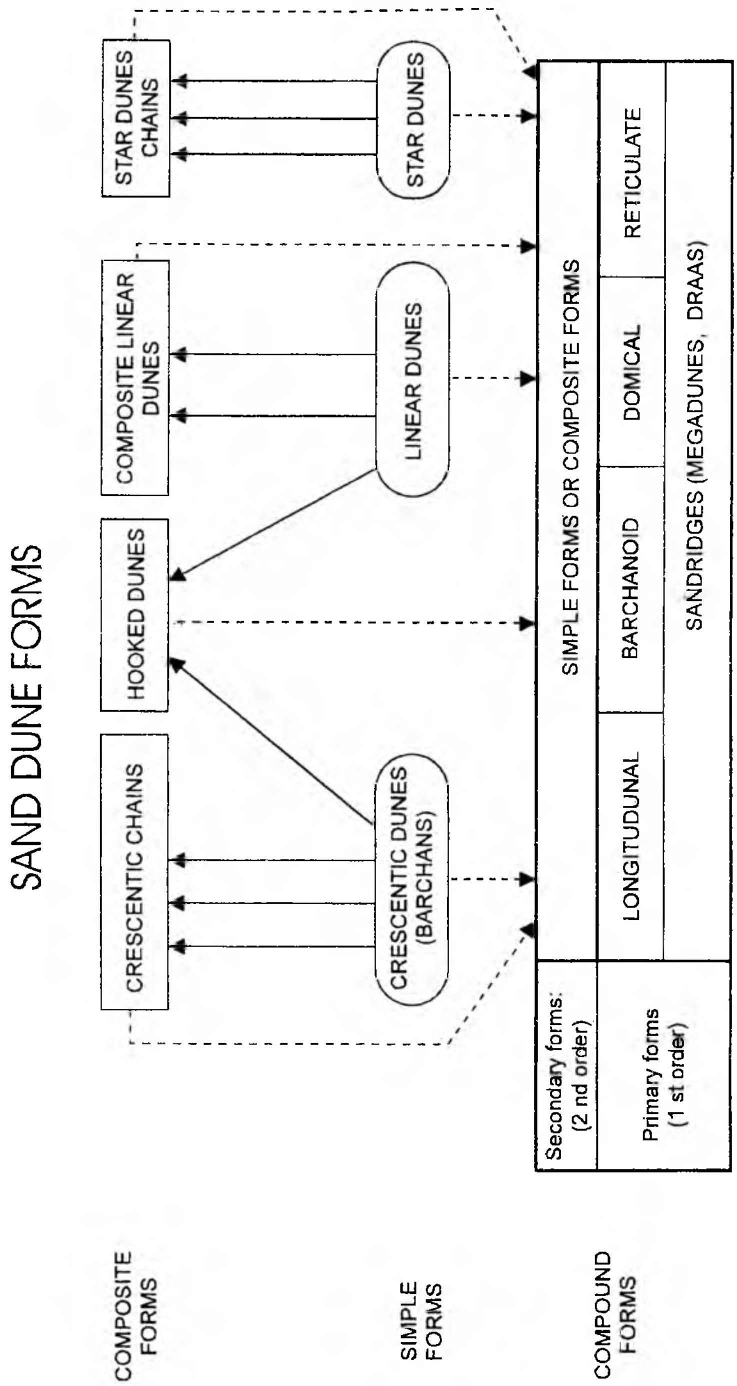


Fig.31 Schematic classification of sand dunes.

It should be emphasised that, in 1988, Havholm and Kocurek, proposed a similar approach to the sand dune classification, in which the term *draa* is used for any aeolian bedform with smaller superimposed dunes.

In the proposed classification of aeolian bedforms, the sand ripples are omitted, because, due to their small dimensions, they can not be detected on satellite or even aerial photographs. The geomorphology of particular dune forms in the Libyan Sahara will be discussed in detail in Chap. 3.2.

3.2. GEOMORPHOLOGY OF PARTICULAR SAND DUNE FORMS

3.2.1. Simple forms

3.2.1.1. Barchans

Localities and characteristics

Barchans are the most typical forms of active dunes which are entirely depositional. In the Libyan Sahara, barchans constitute minority of sand bodies and generally represent the marginal forms of sand seas or occur outside sand seas. Due to the limited resolution of Landsat images, only the larger forms of barchans can be detected, but, on the areas where Metric Camera space photographs or space images taken by SPOT or MOMS-02 were used for analysis, both barchans and crescentic dunes can be very easily recognised.

In the Libyan Sahara, barchans occur as simple forms or as secondary forms on the compound dunes. The barchans detected in Ramlat Zaqqut represent a more crescentic shape, opposed to the typical barchans occuring in southern Libya. These barchans are located within the central part of an elongated depression trending NWW to SEE through Ramlat Zaqqut. The spacing of the dune horns varies between 100 and 200 m.

In the western part of Ramlat Zaqqut large barchans are present (Fig.32). The spacing of dune horns ranges from 300 to 500 m. The adjacent barchans are connected and form „barchan - like" composite dune forms.

The analysis of barchans in the Libyan Sahara indicated that these forms are associated with sand sheets or areas having a thin sand cover. Their axes correspond with the resultant wind direction, a conclusion which is generally accepted by all desert authorities. In the Author's opinion, the slightly crescentic forms, with poorly developed horns, result from the wind pattern. Regardless of the WE resultant wind direction, there, are additionally two opposite wind directions, N and S, which extend the dune arms in these directions, thereby preventing the crescentic dune from changing into a typical barchan

The limited number of barchans were detected in the southern Libyan irgs. They appear both in the Irg al Idrisi (Fig.53, area 5) and on the westernmost flank of Ramlat Rabyanah. In both cases, these forms correspond with resultant

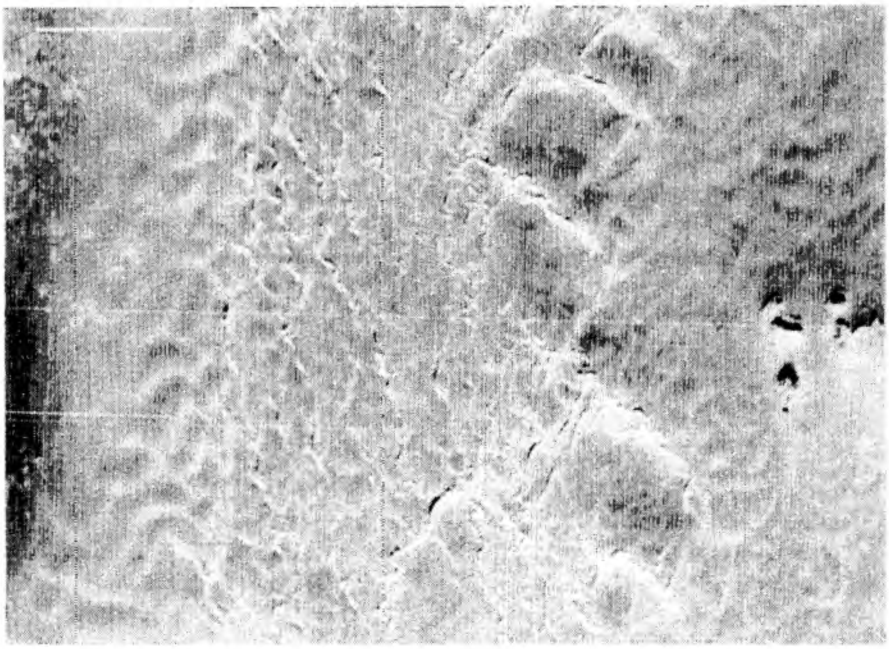


Fig.32 Barchans in Ramlat Zaqqut (MOS-02 image, orbit 91, scene 23, ch.5c).

Fig.33 Star dunes in Ramlat Zaqqut around horse-shoe depression (MOMS-02, orbit 91, scene 23, ch 7).

wind directions of a unimodal regime and their occurrence is associated with a diminishing sand cover.

Large, barchan - like dune forms appear on the north - central part of the Great Sand Sea. These forms can be regarded as asymmetrical barchans with elongated western arms. In the nortwestern part of area 6 (Fig.49) the dune width varies between 2 and 4 km and they are spaced at an average of 2 km apart. Towards the southeast, the dune width decreases and they are very infrequent.

Barchans appearing as secondary forms were recognized both in Idhan Awbari and Idhan Murzuq. Generally, they possess a crescentic shape and are usually arranged in chains, thus forming so - called crescentic chains.

Hypothetical morphogenesis

The appearance of barchans or crescentic dunes in the Libyan desert, confirms the generally accepted theory, that such forms occur in areas with a limited sand supply and in a monodirectional wind regime (Bagnold, 1941; McKee, 1966; Howard *et al.*, 1978; Mainguet, 1984a). Barchans are developed generally under the influence of strong winds blowing from a nearly constant direction. The shape of barchan dunes is a function of grain size, velocity, degree of saturation of the oncoming winds and the variability in the direction of the oncoming winds (Howard *et al.*, 1978).

In projection, barchans have a cresentic form, convex to the wind, with two faces: *a windward face*, called very often *the stoss side*, and *a lee - ward face*, which generally forms the slipface. The windward face angles range between 6° and 12°, while the lee - ward slipface has a steeper slope between 20° and 34°. The mechanism of barchan development creates a stable, sharpely curved crestal brinkline. The lee - ward side is either convex, convex - straight or straight in form. Convex and convex - straight slopes prevail on small dunes, wheras straight slopes dominate on the lee - ward side of large barchans (Embabi, 1982).

Barchans are the only dune forms which preserve their form and size, while migrating downwind. Self - preservation of form (i.e. equilibrium), requires that the rate of translation in the direction of oncoming wind should be constant on all parts of the dune (Howard, *et al.*, 1978). Fulfillment of these requirements is connected with the speed - up factor of wind velocity on the windward slope, which depends on the dune profile. The dune development and transition is a continuous process of wind erosion and deposition. If the dune has to be in equilibrium, the rate of sand transport per unit area should be the same on each part of the windward slope. The rate of sand transport must increase, so that in a very unit area, the wind should be able to carry all the sand eroded before the unit area, and to continue eroding more sand (Tsoar, 1986a). According to Tsoar, the steady state profile, which guarantees the dune equilibrium, should be convex. The barchan shape can be approximated to a chordal section of an elipsoid of revolution with the downwind portion scooped out (Howard *et al*, 1978). The stoss side geometry of a barchan can be characterized by two ratios: H/W and W/L, where H is the crest high, W is the wing - to - wing width, and L is the distance from the tip of the windward edge to the crest. The increase of wind speed or decrease in grain size, causes a moderate increase in H/W and rapid increase of W/L. Thus, the *smaller grain size or higher wind speed produces a steeper and blunter stoss - side.*

The transition of a barchan also requires the migration of the brinkline and slip face. The position of the brinkline must be in balance between the interception of sand from upwind and sand avalanching down the slipface as well as between interception of sand from upwind, and loss from the dune by streamers from the barchan wing tips (Bagnold, 1941).

The shape of a barchan dune is also a function of a degree of saturation of the oncoming flow. The sand which develops the barchan, is generally transported over a sand sheet or desert pavement or other immobile surface. It seems that such sand may be unsaturated, relative to the loose sand surface, but it was observed (Bagnold, 1941) that, due to the high rebound of sand grains on a hard surface, the flow could be oversaturated relative to a loose sand surface. Howard *et al.* (1978) stated that, if the equilibrium transport is assumed, then the degree of upwind saturation affects the rates of erosion and deposition only at the first area element of the barchan upwind edge. But, on the other hand, the degree of saturation also affects the dune shape. The lower degree of saturation implies a greater potential for erosion along the upwind edge, resulting in a steeper dune and increased divergence at the upwind edge, causing a large W/L ratio. As a result, there is a shorter distance between the brink and the upwind edge, and the barchan becomes more open, forming a crescent - moon - shaped dune. Such open barchans, named in this analysis *crescent dunes*, occur in the Maradah area and can be treated as indirect confirmation of low - saturated sandflow.

In the case of high saturation of the oncoming flow, the barchans develop a whaleback form with a small slip face.

The barchan size, in the case of the Libyan desert, varies between few metres in width to several hundred, and even, rarely, 2 km. The factors controlling their size may originate from the sand transport processes or from the natural scales of atmospheric turbulence. The transport factors can be related to the saltation load and to the upwind roughness length. The distance lag between change of shear stress and response of the saltation load, governs the minimum size of a stable dune. The minimum size of a dune is probably proportional to the average path length of grains in saltation, and hence the minimum wavelength would increase with larger grain size and stronger wind shear (Bagnold, 1941). The smaller sand bodies would not have the transport rates necessary to cause the pattern of erosion and deposition necessary for equilibrium. The upwind roughness length seems to have a very important influence on the dune scale. Howard *et al.* (1978) suggested that the dune size would be proportional to the upwind roughness, and that the relative dune steepness H/L, is independent of the dune scale. Embabi's (1982) results, from field investigations in Egypt, did not conform with this suggestion, and he argued that slope angles of the windward side tend to increase as the dunes grow in size. If the upwind roughness is controlled by the fixed roughness elements such as the rock cover of the desert floor, the dune size will be larger with the greater upwind roughness.

In the case when the sandflow is near saturation, and if the natural roughness elements are small, the roughness is controlled by the saltation process. Howard *et al.* (1978) concluded that if the dune size is proportional to the saltation roughness, grain size and dune scale would be inversely correlated. But, on the other hand, they concluded that the dominant shear stress for dune formation increases with the grain size, due to the greater threshold velocities; hence, depending upon the natural variation with grain size of the ratio of dominant stress to threshold shear, the saltation roughness and the dune size might either remain constant or increase as grain size becomes larger. This conclusion agrees with the direct proportionality between grain size and dune size observed by Wilson (1971).

The dune size is also a function of time (Embabi, 1982). The elapsed time since the initiation of the dune may be an important dune scale factor. The large barchans occuring in the Great Sand Sea, seem to be governed by both the terrain roughness and larger grain size due to selective transport of smaller grains towards the south.

Due to the natural fluctuation of the wind direction, and also to non - unidirectional wind regime, the mechanism of barchans development can be modified, and intermediate forms between the barchans and crescentic dunes can occur. In the case of winds blowing in some range, the centreline is much more eroded, because this part plays a role of crestline on which a greater relative shear stress occurs.

The rate of barchans movement varies between 20 and 100 m per year, and exhibits a strong correlation with dune size, length of the windward side and the slope angle of the windward side (Embabi, 1982). The smaller barchans migrate faster than the larger. In the same dune fields, there can be barchans of quite different size. Due to the inverse relationship between migration rate and dune height, the smaller barchans can merge with the larger, and frequently create a single barchan dune of larger size. Very often, small barchans originate on the streamers from large barchans. Generally, barchans belong to the best known sand bedforms, and the theory of their development is quite advanced.

3.2.1.2. Star dunes

The star dunes represent the most conspicuous form of sand dunes. In Libyan sand seas they appear both as simple forms and as secondary forms in compound dunes. As simple forms they are developed in the Ramlat Zaqqut and, as secondary forms, are superimposed on the sandridges in the Idhan Awbari and Idhan Murzuq.

Simple star dunes.

Localities and characteristics

The star dunes developed in Ramlat Zamlat, which are very well depicted on the Metric Camera space photographs, were described in detail by Linsenbarth (1987a). Star dunes are present in three different topographical locations:

a) star dunes developed around huge depressions,

b) star dunes developed in the vicinity of topographical barriers,

c) star dunes located on plateau.

a) Star dunes located around depressions

The largest star dunes are located around the horse - shoe - shaped depression between 28°08'E and 20°20'E and 28°52'N and 28°57'N (Fig.33). The horse - shoe depression, the base of which lies at between 55 and 70 m a.s.l., surrounds rocky hill ca 180 m high. On the northern flank of this depression there are 7 star dunes; the largest on the eastern flank has a diameter of 1.8 km. To the west, the dunes reflect a diminishing tendency; the smallest dune having a diameter of 0.8 km. The dunes are spaced between 1 and 1.8 km apart. The peak of the highest star dune has an elevation of 218 m, and the relative height between the bottom of the depression and the dune top is 164 m. The star dunes have three or four arms oriented radially in all directions. The arms are slightly curved and form large semi - barchans. The steeper arms are oriented towards the rim depression.

The largest star dunes surround the most north - eastern part of the depression with sabkha located in the lowest part. The biggest star dune has a diameter of 2.2 km and is connected with an other smaller star dune, forming a *twin - star dune*. The southern arms are steep and short, while the northern arms are gentle and long, and extend into several linear dunes trending NNE-SSW.

b) The star dunes at the front of topographical barriers

Simple and isolated star dunes were detected on satellite photographs in the central part of Ramlat Zaqqut (Fig.34). The largest dunes have a diameter of up to 2 km. The location of these dunes is related with the appearance of residual flat - topped hills forming topographical barriers. In all cases, the dunes are located at distance of 1.5 to 2 km to the topographical barriers. These dunes are located along the southern flank of Ramlat Zaqqut in the vicinity of a conspicuous escarpment of Jabal al Zaltan. The dune heights range about 100m above the dune base.

The star dunes which appear at the front of topographical barriers were also detected in the westernmost part of Ramalat Zaltan (Fig.51 area 1). They

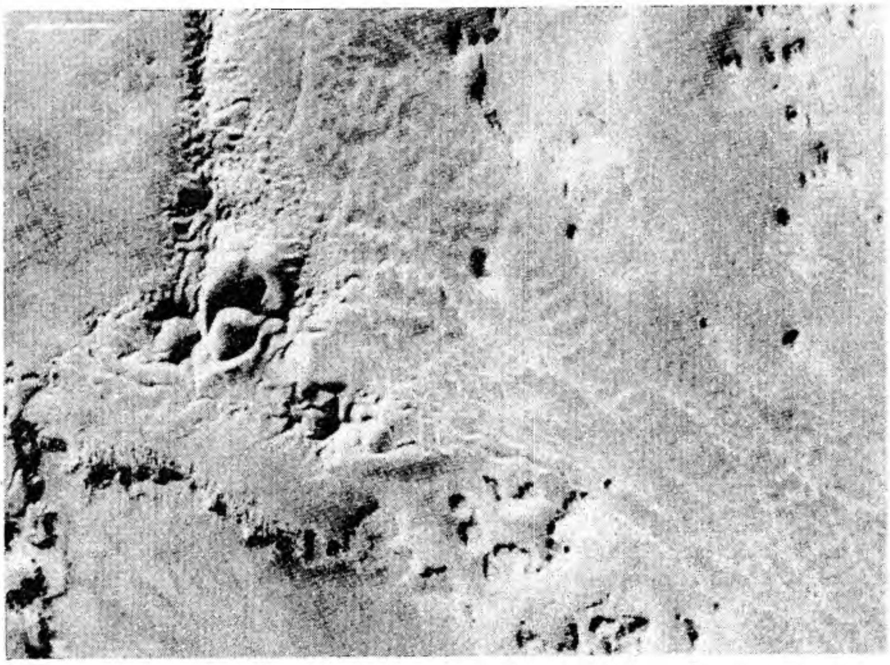


Fig.34 Star dunes in front of a topographical barrier
(MOMS-02, orbit 91, scene 23, ch 5,c).

Fig.35 Star dunes as secondary forms on longitudinal sandridges
in the northwestern part of Idhan Awbarii (Metric Camera, strip 24, photo 709).

are ca 1 km indiameter and are spaced 1.2 ÷ 2.0 km apart. The dunes are arranged in chains.

In the Great Sand Sea there is only one small area (Fig.49 area 20) of star dunes located at the southern margin of the sand sea, at the front of the escarpment formed by Jabal al Qardabah. The dune diameter varies between 0.8 and 1.0 km, and the dunes are spaced 1.2 km apart.

c) Star dunes on plateaux

Star dunes formed on plateau surfaces appear in two distinct parts of Ramlat Zaqqut. One field of star dunes is located north of the horse - shoe - shaped depression (Fig.33). The dunes are developed on a thin sand sheet. The field extends ca 17.5 km in a W-E direction and ca 5 km in a N-S. The star dunes cover the most easterly flank of a plateau dipping gently towards the NE. The dune diameter varies between 0.4 and 1.0 km (average ca 0.7 km) and dune spacing is between 0.6 and 1.2 km. In the most easterly flank of this field, the star dunes present a random pattern, whereas, toward the west, they are arranged in chains.

Another field of star dunes is located on the nortwesternmost flank of Ramlat Zaqqut, called Ramlat al Qazzun. The field is located on the plateau which rises gently toward the ESE. The dunes are arranged in chains which trend NW-SE. The spacing between these chains varies between 1 to 2 km and the distance between dune peaks varies from 0.6 to 2.0 km. The average dune diameter is 0.5 km; the maximum 1 km. The dunes have 3 to 5 asymmetrical arms diverging radially. Generally, the dunes have two longer arms oriented NE-SW. Because the chains of star dunes form very low sandridges, these dunes can be regarded as transitional forms between the simple forms and the star dunes formed on the large sandridges.

Hypothetical morphogenesis of simple star dunes

The location of the star dunes in the Libyan Sand Seas confirms that such forms are strongly associated with very high annual drift potential, which, for Jalo, is 414 VU. The sand rose for Jalo (Fig.60) indicates a complex and multidirectional wind regime, with the highest DP from north, west and south, and with the resultant drift potential towards the SE. The ratio of RDP to DP, an index of directional variability of wind, is very low (0.30) in comparison with other Libyan meteorological stations.

Very little sandy sediment was generated from the escarpments surrounding Ramlat Zaqqut depression (Damaci, 1985). The analysis of star dunes appearance indicates that, as well as relating to a strong multidirectional wind regime, the dune fields are located on very well exposed plateaux or in the vicinity of the topographical barriers. In the case of the star dunes developed on open plateau areas, the form of dune can be related to sand transported from different directions during various seasons of the year (Cooke and Warren, 1973; McKee and Tibbitts, 1964). It is possible that star dune development, started from the small barchan forms, which under the seasonal changes of wind direction, created additional horns or arms. When the dunes start to grow, an interaction took place between airflow and dune morphology (Lancaster, 1989). As in the case of linear dunes, when firstly the airflow approaches the windward side of the dune arm and then passes the crest, the sand is transported along the lee - side of the arm into the peak direction. When the wind direction changes, sand transport is directed along another arm. This process, due to the changing wind direction can be repeated on other arms, as well as along the same arms, but from different sides. Hence, the sand deposition becomes concentrated on the central parts of the dune.

In the case of the large star dunes which are present around the topographical depressions, another hypothetical model of dune developmet may be proposed. The major and longest arms of dunes trend N-S which corresponds with the strongest winds recorded in March and April which blow from both N and S. In May, the northern winds are stronger than the southern. The northern winds, with reduced sand wind potential, also blow during the summer months from June to August. In late winter and in spring, between January and March, winds blow mostly from the W and NW. The dune arms which point to the north are longer and subparallel to the principal flow direction. These arms extend into linear dunes. From the southern side, the arms of star dunes are shorter and steeper. These sides are effected both by the southern winds which can be modified and enhanced by the influence of the depression, and by the northern winds, when the sand flow passing the top of the dune is separated and when strong helical eddies on the lee - side move the sand towards the dune peak.

The star dunes developed at the front of topographical barriers (isolated hills or huge escarpments) seem to be related to the deflation of wind from these barriers, which create the conditions similar to a bi- or multidirectional wind regime. The sand supply and sand deposition is also higher at that location, thus allowing the growth of large sand bodies. As discussed previously, the sand dunes have not developed close to the topographical barriers, but at some distance, usually ca 0.5 to 1.5 km, thereby creating a sand free corridor between the dunes and these barriers.

Star dunes as secondary forms

Localities and characteristics

In the Libyan sand seas, most of the star dunes have developed as secondary forms on the large sandridges (Fig.35). In the Idhan Awbari, the star dunes are present both on longitudinal sandridges and on domical sandridges. On the longitudinal sandridges, they are formed in the western part of area 2 and in area 18 (Fig.47). The star dunes are located at the tops of sandridges or on the cross - section of sandridges arranged in reticulate pattern. The peaks of sandridges are elevated up to 120 m above the interdune areas. The star dunes are composed of 3 to 5 arms of different lengths.

In many cases, the star dunes form the peaks of the huge domical dunes (areas 4, 7 and 15 - Fig.47) which are up to 160 m above the basement. Sometimes it is very difficult to make a clear distinction between a domical dune representing a primary form and a star dune which has developed as a secondary form.

Hypothetical morphogenesis of secondary star dunes

The morphogenesis of the star dunes developed as secondary forms on the top of primary forms is strongly correlated with the supposed morphogenesis of the primary forms, both in character and morphology. Star dunes which appear as secondary forms in compound dunes, represent the classical example of active accumulation dunes, developed under contemporary wind activity. The primary forms can be treated mainly as normal topographical forms which influence the local wind pattern.

As secondary forms, the star dunes are located in the areas governed by the multidirectional wind regime; they are characterised by a very high drift potential (e.g. Ghadamis DP = 588 VU, Sabha DP = 655) but a rather low index of the directional variability of wind RDP/DP, which are 0.23 and 0.33 respectively. These values correspond closely to the range of directional variability given by Wasson and Hyde (1983b). Both at Ghadamis and at Sabha there are 3 main wind directions, and winds from opposing directions blow with high frequency changes. In the Author's opinion, *the frequency of wind direction variability* is as important in star dune formation as the low index of directional variability.

In uppermost parts of primary forms, the wind velocity can be up to two times higher than that of the interdune areas; thus, the secondary forms are influenced by very strong winds, reworking and remodeling the uppermost parts of star dunes. The flanks of primary forms seem to be the only area of sand transport; i.e the sand in transit between the upper parts of successive primary dunes.

3.2.1.3. Linear dunes

The linear dunes, referred to by many authors as the longitudinal dunes (which term has a genetic connotation), represent the most widespread type of dunes in sand desert areas. They are dominant in the Sahara desert (Bagnold (1933), McKee and Tibbitts (1964), Stengel and Busche (1989), Warren (1970, 1972)), Linsenbarth (1987a), in Saudi Arabia (Besler (1982), Holm (1960), in the Sinai Desert (Tsoar (1986b) in Australia (Mabbutt (1968), (Mabbutt *et all* 1968), Madigan (1936), in South Africa (King, 1960; Lancaster, 1981) and in Asia (Fedorovich, 1970). Longitudinal dunes are the most controversial form of sand bodies, but in the Author's opinion, this controversy has its background in different approaches to the classification of these forms. The basic mistake has been the assumption that linear dunes (or longitudinal dunes), belong to the same category (or group) as longitudinal sandridges called draas (Draa is the arabic name for large sand dunes; it was introduced by Wilson 1972).

The linear dune, in simple or composite form, is developed in completely different conditions, for those pertaining to longitudinal sandridges. Both types, in several variants, appear in the Libyan Sahara and the differences of those forms were first detected, interpreted and analysed by the present Author. The morphogenesis of „pure" linear dunes is based on the wind deposition theory (Bagnold 1941, Tsoar 1986b), while the morphogenesis of longitudinal sandridges is explained by the wind drift theory (the erosional theory) (Madigan 1936, King 1960, Hanna 1969, Folk 1971a). The sand deposition theory seems to be accepted by a majority of scientists, but there are two basic approaches to explaining the development of linear dune (Verstappen,1960, 1970, 1972). According to the first approach (McKee and Tibbitts, 1964), the linear dunes develop as the result of bi - directional winds; in the second approach, oblique winds (Carson and MacLean, 1986) thought to be responsible for the formation of linear dunes.

The linear dunes form elongated sand bodies, which are generally rectlinear or sinuous in detail. The profile of a linear dune is symmetrical and is formed by two steep slip faces which meet at an angular crest. The crests show considerable variation, but generally appear as a sharp edged crest. In the Libyan sand seas, sharp crests are usually very well depicted on satellite images due to a very high spectral response.

So called *seif dunes* can be classified as linear dunes. The seif dunes *sensu stricto* are characterised in horizontal section by elongation for a few to a few tens of kilometres, a sinusoity of the crest line and, in vertical section, by a sharp crestal profile with alternating peaks and saddles along the dune, forming a *a tear drop pattern*. On satellite images, seif dunes can be recognised by the sinusoity of crest lines.

Localities and characteristics

The analysis of different localities of linear dunes which appear in the Libyan Sahara suggest that the following types should be recognised

- linear dunes as main sand bodies in ergs and sand seas,
- linear dunes on the margins of sand seas,
- linear dunes in interdune corridors,
- linear dunes in the depressions of sand fields,
- linear dunes as secondary forms on the windward side and crest of huge sandridges.

Linear dunes as the main sand bodies in ergs and sand seas

The linear dunes are the dominant form of dunes in the southern ergs of Libya, namely Irq al Idrisi, erg Sayf as Saniyah, Irq al Mayar and Irq Southern Rabyanah. In Irq al Idrisi, a majority of linear dunes appear as single linear dunes, and, to lesser extent, as composite linear dunes. The linear dunes constitute the main form of dunes which appear both in the NE part of this erg and the SW part (Fig.53). In the narrow NE part of the erg, the linear dunes are aligned obliquely to the axis of elongated sand fields, passing between the topographical barries created by Jabal al Bahrt, Jabal Arkunu and Jabal al Awaynat (1852 m). Within the main sand corridor of the erg, the dunes are arranged obliquely to the outer flanks, forming an angle of 30°÷35°. The length and spacing between dunes, show wide variations. The dune width varies between 0.2 and 0.5 km and the dune length between 5 and 80 km, and the spacing ranges from 1 to 7 km. Interpretation of satellite images indicated that the majority of linear dunes started on small topographical barriers and some of them also terminated on topographical barriers.

In the NE part of the erg, where large topographical barriers occur, the dune trend mirror the form of the topographical barrier, i.e. instead of straight forms, more arcuate forms of linear dunes appear. The linear dunes, arranged obliquely to the defined erg boundaries, terminate at some distance before the erg boundary. The boundary zone on the NW flank of the erg is free from sand dunes. The linear dunes are staggered downwind. As depicted on satellite images, the dunes seem to appear on the sand sheets without any other associated forms or on sand surfaces on which zibars appear. In the preliminary studies of sand forms in Libya (Linsenbarth, 1987b) the Author has used the term *giant sand ripples*, which he now regrets because this term can cause confusion with pure ripples. The analysis of wind regime, supported by the analysis of terrain topography, indicates that the sand - moving winds have bidirectional trend, with the dominant trend following the axis of the terrain depression in which the erg is located; hence, the linear dunes are oriented obliquely to the wind direction. Apart from meteorological observations, the wind direction can also be determined on the basis of *ventifacts* (corrosion tracks detectable on satellite images). The direction of ventifact traces is also parallel to the axis of the erg. The wind direction is also confirmed by the crescentic chains (Fig.53 area 8) the axes of which are at right angles to wind direction. The single linear dunes are also located outside the main field of the erg, and their trend is similar to linear dunes within the erg; in several cases these linear dunes are much longer.

In the erg called Sayf as Saniyah (Fig.52) the linear dunes constitute the only form of dunes appearing in this sand body. The orientation of the dunes is oblique to the erg border, and the morphometric parameters are generally the same as in Irq al Idrisi.

The classical linear dunes - very rectilinear - appear in Ramlat Az Zullaf in Idhan Awbarii, ca 30 km northwest of Sabha (Fig.47). The linear dunes form lines which are consistently parallel and straight for long distances. These dunes are around 15 m in high, ca 80 m wide and are spaced between 400 and 800 m. Thus, the *dune average frequency* is 1.5 (number of dunes per 1 km), and the *interdune index* varies between 5 and 10 (the ratio of interdune width to the width of dune base). The dunes are located on the margin of the main body of the sand sea and have developed on the sand sheet surface. The sand sources are to the NE and sand is transported from this direction.

Linear dunes on the margins of sand seas

Linear dunes and composite linear dunes constitute the marginal zone of the Great Sand Sea. The MOMS-02 data has permitted the interpretation of these forms which represent the boundary between the Sarir from west and the main sand bodies of sand seas (Fig.36). In the middle part of the western boundary of the Great Sand Sea (between 27° and 28° N), the linear dunes are arranged in a downwind staggered pattern. Their length varies between 5 and 40 km, their width, on average, is 8 km. Generally the linear dunes are very straight but are divided into several segments ca 5 km long. The crest lines are very clearly depicted and their shape indicates that the windward sides are on the western flanks of dunes.

Linear dunes in interdune corridors

The linear dunes in interdune corridors generally occur in those sand seas where large sandridges are developed (Idhan Awbari, Idhan Murzuq) or in areas of subdued longitudinal sandridges, where the linear dunes pass through both the ridges and the interdune corridors (Ramlat Rabyanah).

Linear dunes in the depressions of sand fields.

Ramlat Zaqqut, which constitutes the northern part of Ramlat Zaltan, can be used as a typical example of linear dunes developed along the terrain depression within the sand fields. As can be seen in Fig.51 the linear dunes are arranged along the axis of the depression. These are probably the smallest linear dunes in the Libyan Sahara. They are only detectable owing to the high resolution of both the Metric Camera photographs and the MOMS-02. The dunes are from 1 to 6 km long; their width ranges between 0.2 and 0.3 km and the average spacing is 0.45 km. Between the linear dunes the zibar pattern can be recognized, thus indicating the relationship between development of zibars and linear dunes.

Linear dunes as secondary forms on the sandridges

As seconadry forms, the linear dunes appear mainly on the windward sides of sandridges or on the crests of these forms. Generally, linear dunes are arranged obliquely to the main axis of the longitudinal sandridges (e.g. Idhan Awbari). In many cases, the linear dunes are developed along the crest of longitudinal sandridges (the southern part of the Great Sand Sea - area 18 on Fig.48). In some cases, two or three linear dunes, spaced very closely form braided patterns along the crest lines (e.g. Ramlat Rabyanah, Idhan Murzuq).

Hypothetical morphogenesis of linear dunes

In the Libyan Sahara, linear dunes detected on satellite images constitute a rather minor part of sand dune forms appearing in sand seas. As major dune forms, they occur only in Irq al Idrisi and in Irq Southern Rabyanah, and, in

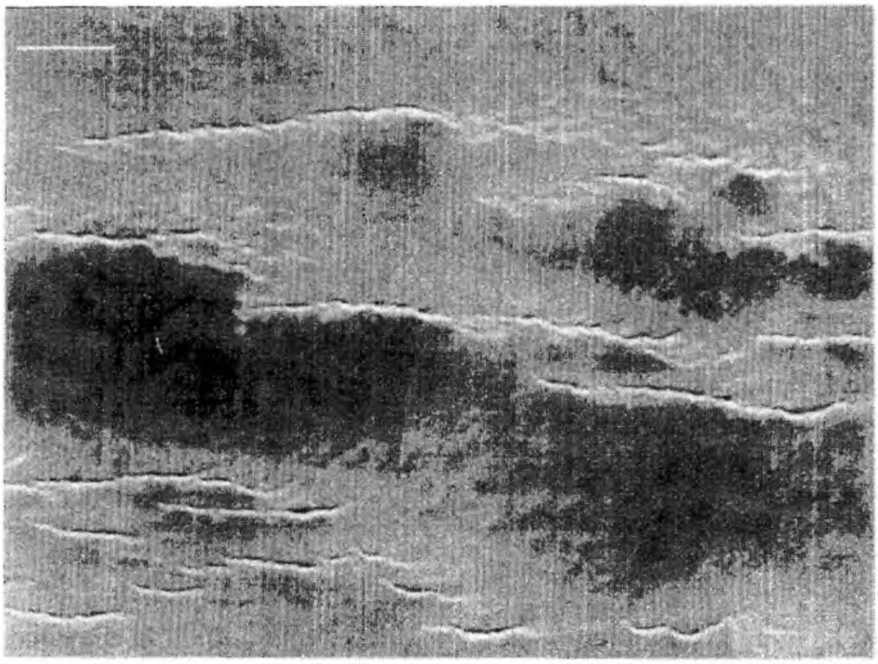


Fig.36 Linear dunes on the margin of the Great Sand Sea
(MOMS-02, orbit 75, scene 14, ch.6).

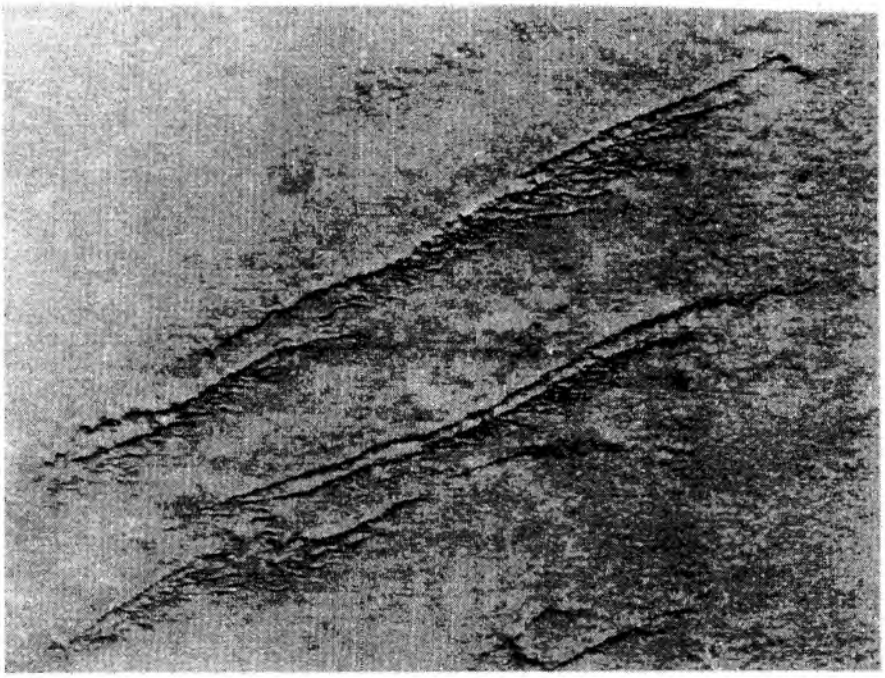


Fig.37 Composite linear dunes in the central part of Ramlat Zaltan
(Metric Camera, strip 4, photo 158).

other sand seas, they appear only as marginal forms or as secondary forms superimposed on huge primary sandridges.

The localities of the linear dunes are characterised by having only a small amount of sand supply. In the case of Irq al Idrisi and Irq Southern Rabyanah, they appear on the thin sand sheet with allochthonous sand, imported from the sand seas located to the NE (the Egyptian Sand Seas) and from Ramlat Rabyanah.

The irgs Irq al Idrisi and Irq Southern Rabyanah, are located in elongated depressions which are surrounded by topographical barriers. In such topographical configurations, the winds are funnelled along the axis of the depressions. This wind direction is confirmed by very striking ventifacts which occur on both sides of these irgs. An additional confirmation of wind direction is provided by the orientation of crescentic (transverse) dune chains (Fig.32, areas 5 and 8 in Irq al Idrisis). Futher indirect confirmation provided by the resulting drift direction, as calculated for the meteorological stations in Kufra and Tazirbu and by the general wind circulation, which is governed in this region by anticyclonic circulation. The wind regime in Kufrah can be classified as wide unimodal or bi - directional (between 350° and 18°), the strongest winds coming from the N and NE (Fig.65). The dune axes are oriented at an angle of 30°÷34° to the resultant wind direction, thereby confirming the hypothesis that oblique winds create the linear dunes (Carson and MacLean, 1986).

The initiation of linear dunes in Irq al Idrisi and Irq Southern Rabyanah is promoted by small topographical barriers. Wind flow, encountering the obstacle, deposits sand on the downwind side, which, later on, is transported and reworked in the process of dune formation.

The small linear dunes, which appear in Ramlat Zaqqut are oriented along the axes of depressions; hence, they can be treated as indicators of modern winds funnelled along terrain depressions. In this case, the linear dunes are developed under very acute bi - directional winds created within the depression.

The linear dunes in the vicinity of Sebha were in investigated in detail by McKee and Tibbitts (1964, 1971). The results of their investigations confirmed the hypothesis that the trend of linear dunes is controlled by the bi - directional wind regime. Their investigations indicated that the area of linear dunes is influenced by strong winds - in the morning from the south - east while, in the afternoon, dominantly north - east winds occur. The general dune trend corresponds with the resultant wind direction computed by the Author of this paper on the basis of data from the meteorological station in Sabha for the period 1967 ÷ 1984. The sand rose for Sabha (Fig.63) confirms a wide, unimodal wind regime. The very high resultant drift potential RDP = 277 VU (the highest RDP from all meteorological stations used in this investigations) is very significant.

The examples of linear dunes in the Libyan Sahara, their occurrence and relationship to the sand supply and wind regime and the results of investigations performed by other authorities in different locations (Tsoar, 1986b; Stengel and

Busche, 1989), has provided the basis for a generalised hypothesis of linear dune development. The location of linear dunes in Libya confirms the theory that linear dunes are developed in a bi - modal wind regime. In such conditions, the dune axes are located between two wind directions, and wind striking the dune obliquely is responsible for erosion and deposition along the lee - flank and for net transport and dune elongation.

When the wind encounters the crest obliquely (Tsoar, 1986b) a three - dimensional flow separation occurs which is caused by the perpendicular component of the incident wind. At the line of reattachment, this component has zero velocity and the second component is presented by a line parallel to the crest line. The magnitude of this wind component depends on the cosine of the angle of incident wind (Tsoar, 1986b). When the angle is less than 40°, the velocity of deflected wind is higher than that on the crest line. Sand eroded from the wind-flank is transported along the lee - flank. When the incident wind drops below the threshold velocity, deposition took place.

In the second phase of linear dune development, when the wind encounteres the crest line from the opposite direction, the windward flank and lee - ward flank are reversed and the erosion and deposition process take place on the opposite side. When, in the sucessive phases of linear dune development, the wind velocities and directions are uniform from both sides, a straight linear dune is developed and its elongation occurs. The Author of these investigations proposed the hypothesis that dune straightness depends on the frequency of the consecutive phases of linear dune development and the symmetry of dunes depends on the drift potential in each phase. In the case of linear dunes in the vicinity of Sabha which were described by McKee and Tibbistti (1964), there were two consecutive phases each day: the morning and afternoon phases, but the wind potential was not equal in each phase and this resulted dune asymmetry.

Detailed field investigations of seif dunes were performed by Tsoar (1983) in the Sinai Desert. His field measurements and observations led him to a definition of the mechanism of seif dune development. This model is based on the assumption that, due to the dune sinusoity, the wind, which blows obliquely, approaches the dune flanks at a different angle. Hence, in some parts, wind approaches the dune at a acute angle, while, in the other parts, at nearly right angles. The different angles of the wind, which approach to dune flanks, are responsible for the different rates of erosion on the windward flank and different rates of sand transport and deposition on the lee - side. Similar processes occur when the wind blows from the opposite direction. Due to the consecutive and repetitive changes of wind approaching dune flanks, the dune peaks and saddles are redeveloped and a foreward movement of the dune occurs.

The movement (elongation) of seif dunes was emperically confirmed by Tsoar (1983). His field investigation of a seif dune 1500 m long, located on a flat sand strip, with peaks between 12 and 14 m and saddles between 5 and 8 m, was carried out over a period of 32 months. It indicated that, during that time, the dune was elongated 39 m, and that the mean displacement of peaks and saddles was 22 m (average 1.2 m and 0.7 m per month accordingly).

The mechanism of seif dune development nearly explains the further development of existing seif dunes but fails to expain the initiation and formation of the primary form of seif dune.

The field investigations of Stengel and Busche (1989) in Tenere in Niger, based on observations and measurements of ripple patterns allowed them to define the present morphodynamics of seif dunes.

A completely different hypothesis of longitudinal (seif) dunes formation was proposed by Verstappen (1960, 1970, 1972). As the results of his analysis of aerial photographs of the dune areas of the Thar Desert, Verstappen (1960) showed that parabolic dunes are only the initial stage of seif dunes formation. In the initial phases of their development, seif dunes were denudation/deflation forms. Verstappen pointed out that the early phases of seif dune formation were characterized by wind erosion and the later phases by deposition. In opinion of the present Author, the dune forms described by Verstappen belong rather to compound forms which have been built in several cycles of their development and cannot be classified as pure linear dunes (which are are considered to be typical depositional forms). Also, earlier workers hinted at wind - made blow holes and horse shoe - shaped pits as the primary stage of seif dune formation. Aufrere (1928) believed that seif dunes ridges are essentially a residual relief and that depressions between them, so - called *gassi*, are due to aeolian erosion. King (1960) has reported that linear dunes were formed due to the separation of the trailing arms of the parabolic dunes.

Another hypothesis of linear or seif dune development is based on the develoment of such dunes from barchans (Bagnold, 1941, Tsoar, 1984, Linsenbarth, 1987b). In the nomenclature used by the present Author, such forms can be treated as composite forms and were named *hooked dunes.* The localities of hooked dunes in Libyan Sahara and their hypothetical development are described in Chapter 3.2.2.3.

3.2.2. Composite forms

3.2.2.1. Composite linear dunes

Localities and characteristics

The composite linear dunes represent dune forms which are unique to the Libyan Sahara. They have not been recognised in the other sand desert areas except, perhaps, as the composite linear dunes in th Erg of Fuchi Bilma described by Mainguet, (1984a).

Composite linear dunes detected in the Libyan sand seas may be divided into: the main forms of sand seas, the marginal forms and the solitary dunes developed outside sand seas. Composite linear dunes represent the main dune forms in Ramlat Zaltan and marginal forms in the Great Sand Sea.

The southern part of Ramlat Zaltan which extends over 360 km from 28°45'N to 25°15'N, is mainly covered by large composite linear dunes. These are arranged into two separate belts: the western and the eastern (areas 10 and 11 on Fig.51). In both cases the composite linear dunes possess the same trend. The composite dune consists of so - called *principal dune* and *subsidiary dunes* arranged obliquely to the principal dune (Fig.38). The principal dunes are located on the western side of composite forms. The length of the dunes varies between 5 and 30 km, and the width reaches up to 1.6 km at the feathered end. The dunes are staggered downwind and are spaced between 2.5 km and 4 km apart. The interdune areas are covered by zibars. The similar composite linear dunes, which form arrow - shaped figures appear in area 11. Their length varies between 15 km and 50 km, width between 1.0 km and 1.5 km and spacing from 2 km to 3 km. The interdune areas are also covered by zibars. The composite linear dunes which appear in the northern part of ramlat Zaltan are presented on Fig.38.

Another form of composite linear dunes can be recognised on the western flank of Ramlat Rabyanah (Fig.50 area 5). The dunes are composed of several long arms organised in an acute fan with the head pointed to the west. The length of these dunes varies between 15 km and 35 km and spacing between 2 km and 3 km. The tails of adjacent dunes merge at their eastern ends.

Composite linear dunes are very well developed on the western margin of the Great Sand Sea (Fig.49 area 16). These composite linear dunes appear in a asymmetrical feathered form, with the principal, most rectilinear dune from the west, and with the shorter subsidiary linear dunes attached obliquely to the principal dune on its eastern side. The dune heads point towards the south and the feathered parts are oriented to the north. The dune length varies between 5 and 40 km, width between 0.6 to 1.0 km and spacing from 5 to 10 km. The dunes are up to 40 m high. The crests of the principal dunes are composed of 2 or 3 linear dunes forming a braided pattern. The more detailed analysis performed on the topographical maps at a scale of 1:50,000 indicated that the western flanks are steeper than the eastern flanks, thus confirming that the formative winds blew from a NE direction. Another type of composite linear dune appears in the eastern margin of reat Sand Sea (Fig.39). These dunes are much wider than these developed on the western margin of the Great Sand Sea.

Hypothetical morphogenesis

The mechanism responsible for the formation of such composite dunes is not known. On the basis of a knowledge of the distribution of such dunes in the Libyan Sahara and they relationship to the wind regime, the following hypothesis may be erected. Generally, composite linear dunes are located at the margins of sand seas. In the case of Ramlat Zaltan and the Great Sand Sea, they are formed at the western margin, which is bordered by sarir areas; hence, the sand

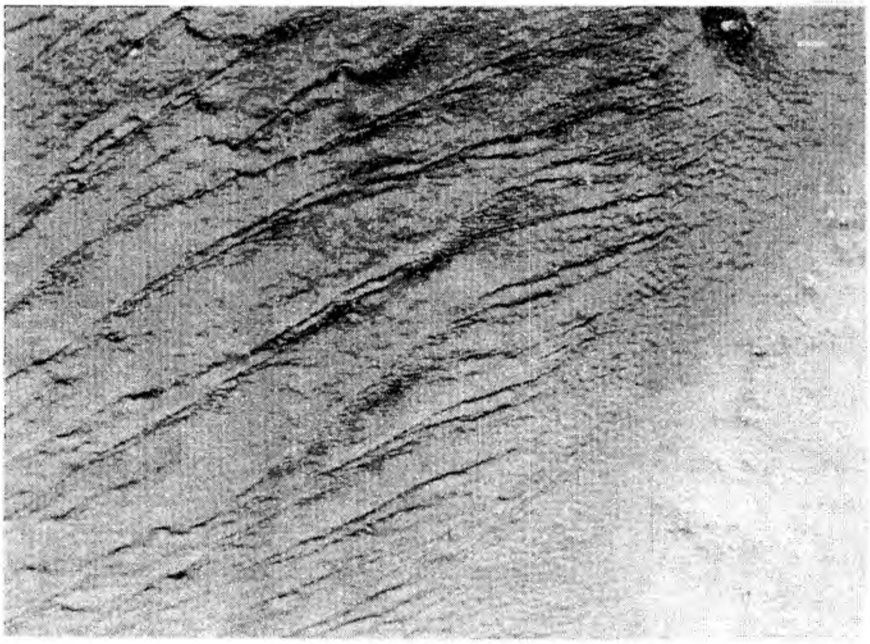


Fig.38 Composite linear dunes in the central part of Ramlat Zaltan (metric Camera, strip 4, photo 158).

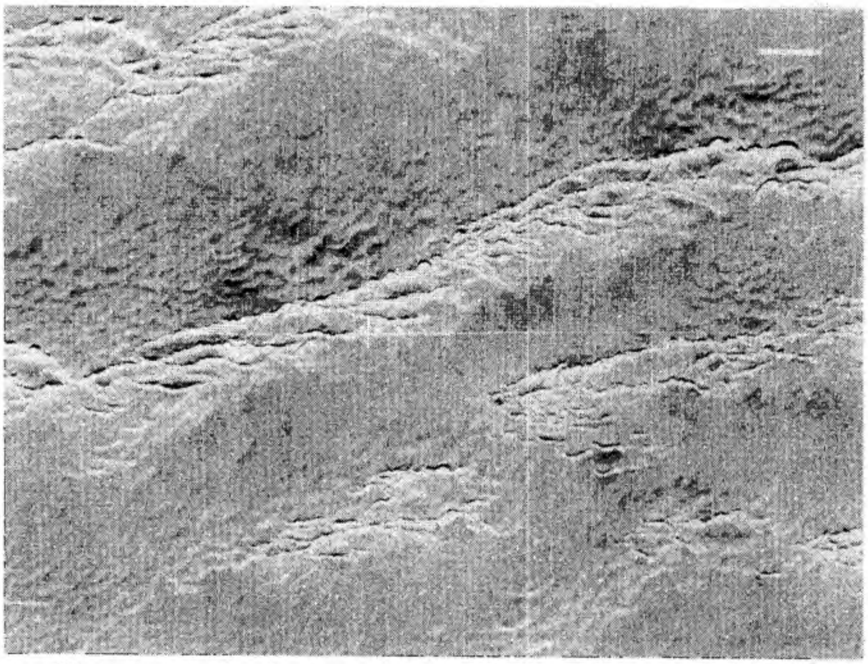


Fig.39 Composite linear dunes on the eastern flank of the central part of the Great Sand Sea (MOMS-02, orbit 75, scene 14, ch.6).

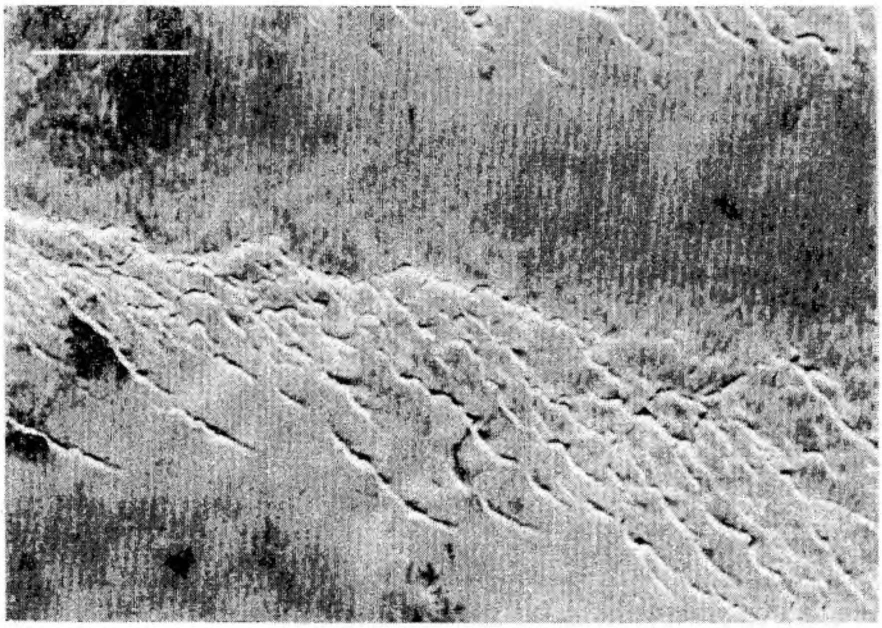


Fig.40 Crescentic chains in Ramlat Zaltan (MOMS-02, orbit 91, scene 23, ch.7).

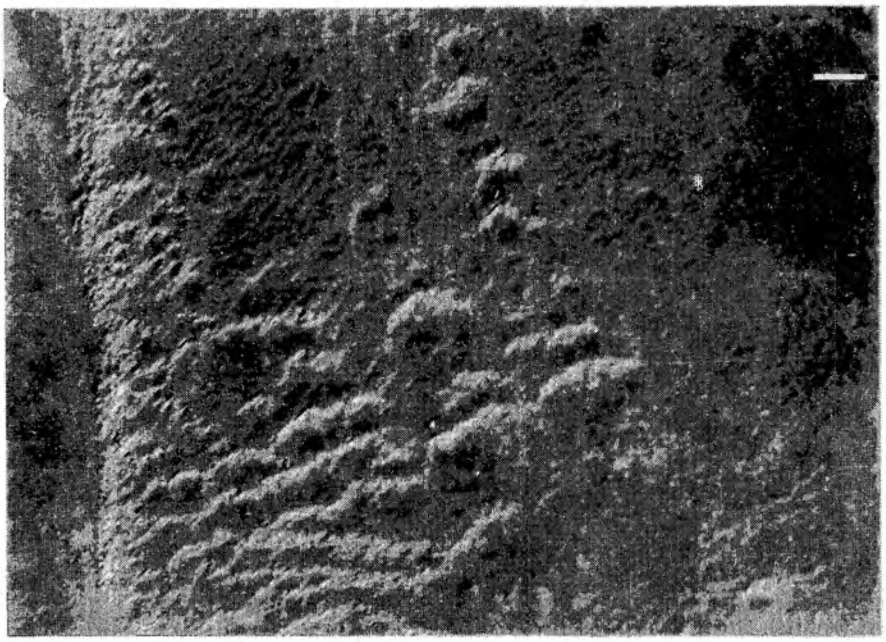


Fig.41 Crescentic ridges composed of crescentic chains - northwestern part of Ramlat Zaqqut (MOMS-02, orbit 91, scene 23, ch.7).

supply is mainly from the eastern side, i.e. from the inner side of the erg. The sand supply, if any, from the sarir area is restricted to the particular, bi - modal population of sand grains represented by a large grain size and finer grain size particles. The larger sand grains when move in creep mode, are transported and deposited on the western flanks forming a fixed layer. The resultant winds together with the erosional lineaments, indicate that the winds blew from the NW and were caused by moving Mediterranean depressions. Those strong winds are responsible for the development of the principal dunes into these composite forms. The principal dunes probably form a topographic barrier for winds and are responsible for changes of wind direction; hence, the subsidiary dunes are arranged at angle acute to the principal dune. The subsidiary forms can be attributed to the gentle winds from the NNW and NNE. The seasonal changes of winds can be explained by the unstable position of the anticyclonic centre which governs the wind regime in this part of Libya. The appearance of composite linear dunes confirms the Author's hypothesis that such dunes can be treated as *boundary forms* between sarir and sand sea areas.

3.2.2.2. Crescentic chains

Localities and characteristics

Crescentic chains represent lines of linked crescentic dunes oriented at right angles to the dominant wind direction. In the Libyan Sahara these forms appear in most of the sand seas, but constitute a marginal and rather minor part of the sand seas. The crescentic chains are clearly exhibited on Metric Camera photographs and also on MOMS-02 images. Owing to the higher resolution of the Metric Camera photographs, the crescentic chains were investigated in detail in the Ramlat Zaltan (Linsenbarth, 1987a). The crescentic chains are located in several places within the area studied. The chains are composed of coalescing crescentic dunes forming coalescent crest lines (Fig.40). The coalescing chains are spaced from 150 to 300 m apart. In some cases, the crescentic chains resemble reversed forms with different orientations of crescent crests. The crescentic changes are separated by interdune areas with only a thin sand cover.

A somewhat different pattern results from the composition of several very densely spaced crescentic chains. Such forms occur on northeastern part of the dune field in Ramlat Zaltan (Fig.51, area 7). These composite forms present small ridges, ca 1.0 to 2.5 km long and spaced from 0.6 to 1.0 km apart. Besides the crescentic form of the modular elements in chains, the general shape of these composite chains is also crescentic (Fig.41).

The fields of crescentic chains can also be discerned in Ramlat Rabyanah (Fig.50, areas: 1, 12, 13, 14 and 17). The fields of crescentic chains are located on the upwind margin of the sand sea at the boundary with sarir areas, and between the western and eastern parts of Ramlat Rabyanah.

In the Great Sand Sea, the crescentic chains occur in several fields (Fig.49, areas 4, 17, 19, 21 and 23) located at the western flank of the sand sea. The crescentic chains are spaced 0.6 to 0.8 km apart. Different types of crescentic chains cover area 11. They are large crescentic chains spaced at 1.5 km intervals.

A very limited number of crescentic chains can be detected in Idhan Awbari and in Idhan Murzuq. In Idhan Awbari only a small field of crescentic chains is located in the vicinity of Sabha (Fig.48, area 20). In Idhan Murzuq, crescentic chains are developed only in the eastern part. The very densely spaced crescentic chains (6 crest/km) appear in the eastern part of Ramlat Zuwalyah (Fig.48, area 4 and 8).

The crescentic chains are also developed as secondary dune forms which are superimposed on large sandridges, both in Idhan Awbari and in Idhan Murzuq.

Hypothetical morphogenesis

The occurrence of crescentic chains in Libyan sand seas indicates that they are developed on the marginal upwind parts of sand seas. Generally, the fields with crescentic chains occur in the vicinity of sarir areas. In all cases, the dominant wind direction is at right angles to the lines of crescentic chains; this confirms a generally accepted theory that both barchans and crescentic chains (also called transverse dunes), have the axis oriented parallel to the dominant wind direction. In the case of the crescentic chains in Ramlat Zaltan, their shape is more crescentic than barchanic. Development of a such a crescentic form may be explained as the result of a particular wind regime: the strong E wind responsible for the orientation of dune axes parallel to its direction and two other winds blowing perpendicularly to the first and from two opposite directions - i.e. from N and S.

In Ramlat Rabyanah (Fig.50), the orientation of crescentic chains corresponds with the windrose for Tazirbo (Fig.64), on which the north and south components are the strongest.

The source of sand and its granulometric nature are evidently controls governing the appearance of the crescentic chains. In all cases analysed, the sand supply is from the sarir areas, which suggests a very limited range of grain size characteristics. The medium - grain particles were probably moved earlier by saltation to the interior parts of the sand sea and only the coarser sand particles, moving in a creep mode, supply sand for the development of crescentic chains. For the larger - size particle, the threshold velocity is higher, thus only the stronger winds can move them. This may explain why the crescentic chains have a different orientation: when the coarser sand sources are located of the upwind side to the dune field, then only the strongest winds govern the development of the crescentic chains and their orientation to the direction of the strongest winds. In the case where the sand source is less coarser, hence, such particles can be moved by more gentle winds. The evidence for such a hypothesis is very well documented in Ramlat Rabyanah.

The location of areas covered by crescentic chains can also be directly related to the terrain topography. In the Idhan Murzuq area 4 (Fig.48), which is covered by crescentic chains, their development clearly correlates with the elongated terrain depression which represents the old drainage system. Here, there are probally two factors controlling dune development: a sand source characterised by coarser grains and, if any, wetter conditions causing changes in threshold velocities.

Lancaster (1985) indicated that as the result of the velocity acceleration over the transverse dune, the dune chains become broader and more rounded when wind velocity is increased. When the wind velocities are lower, such a wind regime develops steeper profiles. Crest - to - crest spacing is also controlled by wind velocity. In the areas of high wind velocities, the dunes will be spaced further apart and, in areas with low wind velocities, the dunes will be relatively closely spaced. The Libyan examples of cresentic chains fully support Lancaster's (1985) observation.

3.2.2.3 Hooked dunes

Localities and characteristics

Hooked dunes have a form of barchans, with linear dunes formed by the extention of one horn. The hooked dunes can be classified as transition forms between barchans and linear dunes and represent an excellent example of the mechanism of linear dune formation.

Hooked dunes in different stages of development prevail in the northern part of the Great Sand Sea (Fig.49, area 5). In the case of a hooked dune, the barchan like form appears on the northern side and the linear part on the western side. The width of barchans varies from 0.5 to 1.2 km, whereas the length of linear dunes ranges from 3 to 20 km and width from 0.4 km to 1.2 km. The hooked dunes are spaced at distances from 1.5 km to 5 km. The hooked dunes are developed on the sand sheet without other smaller forms like barchans or zibars.

Very spectacular hooked dunes are present in the northern part of Ramlat Zaltan. These forms are very well depicted on the Metric Camera photographs . The northern parts of hooked dunes constitute barchans in crescent form (Fig.37). The western wings of barchans are developed into very long linear dunes.

Hypothetical morphogenesis

The development of hooked dunes is controlled by two directional winds blowing from the north and the west. The resultant wind direction is from NW as it is evidenced by the meteorological stations both in Jalu and in Al Jaghbub (Fig.60 and 61). The resultant drift direction (RDD) corresponds with the axes of the barchan parts of hooked dunes, whereas the winds the from north and northwest govern the development of linear forms.

The development of such forms was explained by Bagnold (1941) and Tsoar (1984) as one version of seif dune development. According to Bagnold, the primary gentle winds form barchans, while strong winds from a secondary direction elongate one horn, that on the side from which the winds blow. Tsoar (1984) proposed another version of such a model. In this, the strong winds form the barchan part, while the gentle winds from a secondary direction elongate the horns on the opposite side. Tsoar pointed out that, when a barchan starts to develop one arm, this part elongates faster that the barchan itself.

The analysis of hooked dune development in the Great Sand Sea leads to a different hypothesis which, lies somewhere between models proposed by Bagnold and Tsoar. The analysis of winds recorded at the meteorological stations in Jalu and Al Jaghbub indicated that the strongest winds (April - May) blow from N and that these winds are responsible for elongation of the western horns in a southerly direction. Such an assumption partly agrees with Tsoar's model in which the horn opposite to the wind direction is elonagated into linear form. But, according Tsoar, this elongation is caused by gentle winds. Libyan observations confirmed Bagnold's suggestions that it is, in fact, the gentle winds which are responsible for barchan formation and dune advance; however Bagnold indicated that stronger winds develop the extended horns opposite to the wind direction.

3.2.2.4 Zibars

In various sand seas of the Libyan Sahara, giant ripples, called *zibars*, have been detected on satellite images and analysed. On Landsat images, zibars are only partially depicted, but on the high resolution images and photographs (SPOT, MOMS-02, Metric Camera) zibars are plainly visible. Generally, zibars are low amplitude and long wavelength sand bedforms (in vertical plane). Zibars appear mainly on sand sheets and in interdune areas (for more details see Appendix 6, p.3.1.).

In Idhan Awbari, zibars are clearly depicted both on the Metric Camera space photographs and on SPOT images. They appear in the interdune areas between large sandridges and are oriented obliquely to their axes. They possess very regular trend and spacing, but in the plan view they present more of a striped than a wavy pattern. The zibar spacing ranges between 100 and 150 m, and their width between 200 and 300 m. In the western part of Idhan Awbari, a significant pattern of zibars occurs between sandridges. Here the zibars are oriented ca 40 ÷ 45° to the sandridges. They present a striped pattern characterized by spacing of between 150 ÷ 250 m and widths ranging from 150 ÷ 300 m.

In Ramlat Zaqqut, the MOMS-02 images permitted the detection and analysis of zibars which are developed along sand sheets. Their spacing varies between 150 ÷ 200 m in the western part and rises to 300 in the eastern part. Their wavelength (width) ranges from 300 m to 600 m accordingly. In comparison with zibars in Idhan Awbari, the zibars pattern in Ramlat Zaqqut is more

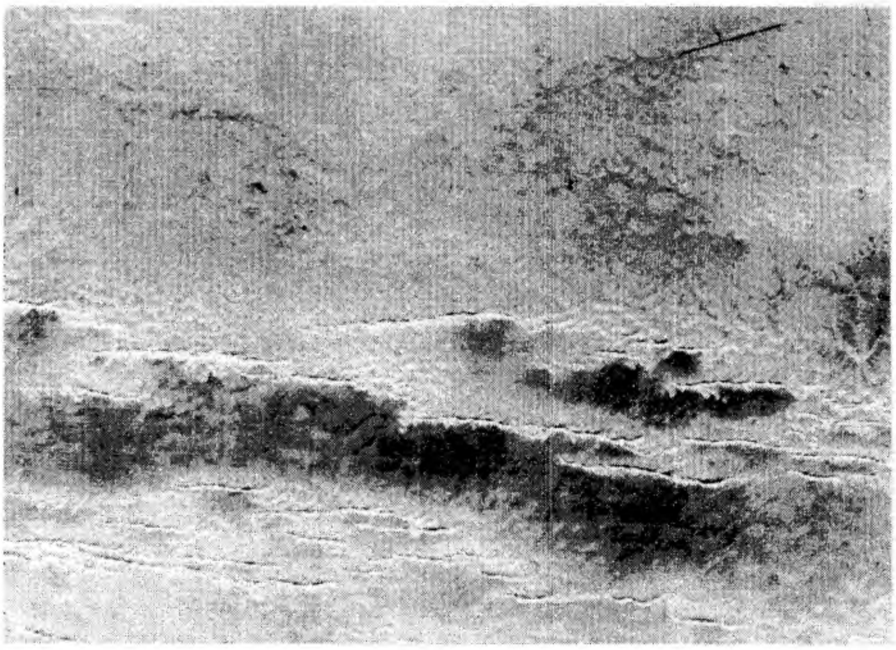


Fig.42 Zibars in Ramlat Zaqqut (MOMS-02, orbit 91, scene 23, ch.7).

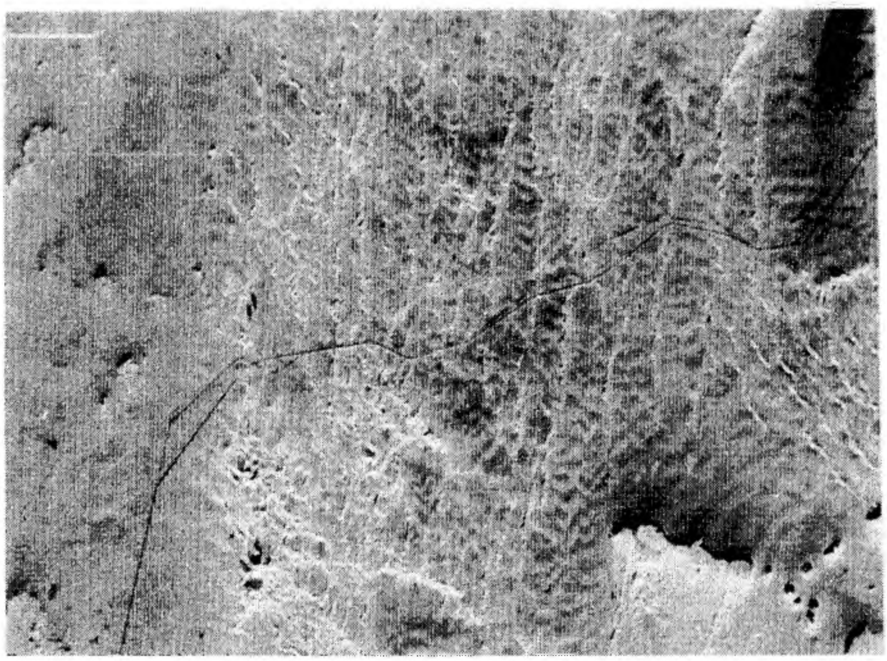


Fig.43 Zibars on the western boundary of the Great Sand Sea
(MOMS-02, orbit 75, scene 14, ch.7).

wave - like. The amplitude of this wave pattern ranges between 100 and 400 m. In the western part of Ramlat Zaqqut (Fig.42), where zibars occur on the sand sheet, linear dunes, oriented at a right angle to the zibars pattern, also occur. The segmented pattern of these linear dunes, caused by zibars, is very distinct.

On the boundary between the Great Sand Sea and As Sarir, the zibar pattern is clearly visible on the MOMS-02 images. The zibar pattern appears both on the sarir area as well as between the linear dunes bordering the Great Sand Sea (Fig.43). The spacing of the zibars ranges between 200 and 250 m and wavelength varies between 200 and 600 m. Towards the west, the amplitude of the wave pattern decreases to 100 m. In the vicinity of the liear dunes which define the Great Sand Sea boundary, the wave pattern has higher amplitude. Between the linear dunes, both the spacing and the amplitude of the zibars is larger and varies between 350 and 400 m.

The linear dunes, developed on the margin of the Great Sand Sea, are arranged parallel to the axis of zibar pattern. The relationship between linear dunes and zibars is somewhat different when compared with Ramlat Zaqqut, where the linear dunes are arranged at right angles to zibar pattern. This difference may be attributed to the different sand characteristics. Sand in the As Sarir area is much coarser than that the in the sand sheets of Ramlat Zaqqut.

As well as being detected on the Metric Camera photographs, SPOT images and MOMS-02 images, zibars were also partially detected on Landsat images in the central and southern part of the Great Sand Sea and in Irq al Idrisi.

As detected on the satellite images of Libyan Sahara, zibars represent low relief sand bedforms, which can be regarded as indicators of high - energy winds. Both in Ramlat Zaqqut and in the vicinity of the Great Sand Sea, the orientation of zibars strongly correlates with the direction of the high energy winds recorded in March and April at the meteorological stations in Hun and Jalu. The pattern of zibars in Idhan Awbari is very similar to the so - called festooned liquid zibar patterns in the Tenere desert (Warren, 1971). The linear dunes developed on the zibars appearing in Ramlat Zaqqut are also similar to the seif dunes in central Tenere. Warren (1971), concluded that zibars are developed in bi - modal sands. The zibar spacing in the Libyan Sahara is similar to that in Sudan (400 m) and in the Selima sand sheet in Egypt. The Libyan Sahara zibars are rather smaller than the large - scale low - amplitude bedforms (*chevrons*) observed in the Selima sand sheet by Maxwell and Hayness (1989).

3.2.3 Compound forms

Compound forms of sand dunes constitute the largest dunes in the Libyan Sand Seas. The analysis of satellite images has enabled the differentiation of particular compound forms such as: *longitudinal sandridges, barchanoid sandridges, domical sandridges, reticulate sandridges and deflation rims.* The sandridges representing the *primary forms* are partially covered by simple or composite dunes, here referred as *secondary forms* (see Fig.31)

3.2.3.1. Longitudinal sandridges

Localities and characteristics

The typical longitudinal sandridges occur in Idhan Awbari (Fig.47) and Idhan Murzuq (Fig.48). The longitudinal sandridges represent large sand bodies arranged in parallel (Fig.7 and 44). Such forms occupy areas 2, 11, 17 and 19 in Idhan Awbari (Fig.47) and areas 1, 9, 11 and 16 in Idhan Murzuq (Fig.48). In area 2 of Idhan Awbari (Fig.47), the ridge trend is NE-SW, the length of ridges varies between 60 and 85 km, their width between 2 and 3 km and spacing range from 2 to 3 km. The dune heights are up to 120 m. The sandridges are separated by interdune corridors some 1 to 2 km wide constituting bedrock with a thin mantle of sand. The cross - section of sandridges is asymmetrical, the northwestern slopes being gentle, whereas the southeastern steeper. The sandridge tops are rounded and without a distinct crest line. Brink lines show well on satellite images.

The gentle north - western slopes are partially covered by secondary forms, represented by linear dunes and/or crescentic chains. The trend of the linear dunes is oblique to the main trend of the sandridges. In several places along of the most nortwestern flank of this field, star dunes are superimposed on the top of longitudinal sandridges.

Similar sandridges, with the same orientation, occur in area 11. In the eastern part of this field, the sandridges are arranged in a very regular pattern of parallel straight ridges, whereas, on the western flanks, the sandridges swing slightly towards the south.

Very long sandridges appear in area 17. This area contains 11 sandridges trending ENEE-WSW and ranging in length between 35 and 80 km. The dune spacing varies between 3 and 4.5 km. The interdune areas comprise long, sand - free corridors without dune forms. The northern gently sloping flanks are partially covered by oblique linear dunes some 4 to 5 km long. In the middle of this field, the top part of the sandridges consists of several closely - spaced linear dunes forming a braided pattern. The longitudinal ridges in this field are the highest in the Libyan sand seas and reach up to 140 m over the dune base.

Hypothetical morphogenesis

The morphogenesis of compound forms has to be discussed separately in respect of the primary forms and to the secondary forms. The Libyan Sahara examples of compound forms indicate that the primary forms are much older than the secondary forms, and that they were developed under climatic conditions different to that of the present day.

The examples of the longitudinal sandridges which appear both in the Idhan Awbari and in the Idhan Murzuq, confirm the *theory of the winddrift origin of sandridges*. Both sand seas are located in depositional basins in which huge amounts of sand were deposited during the Pleistocene. In the more arid periods, wind channelled these fluvial deposits to form the windrift dunes. Such an

Fig.44 Longitudinal sandridges in the western part of Idhan Awbari
(Metric Camera, strip 24, photo 708).

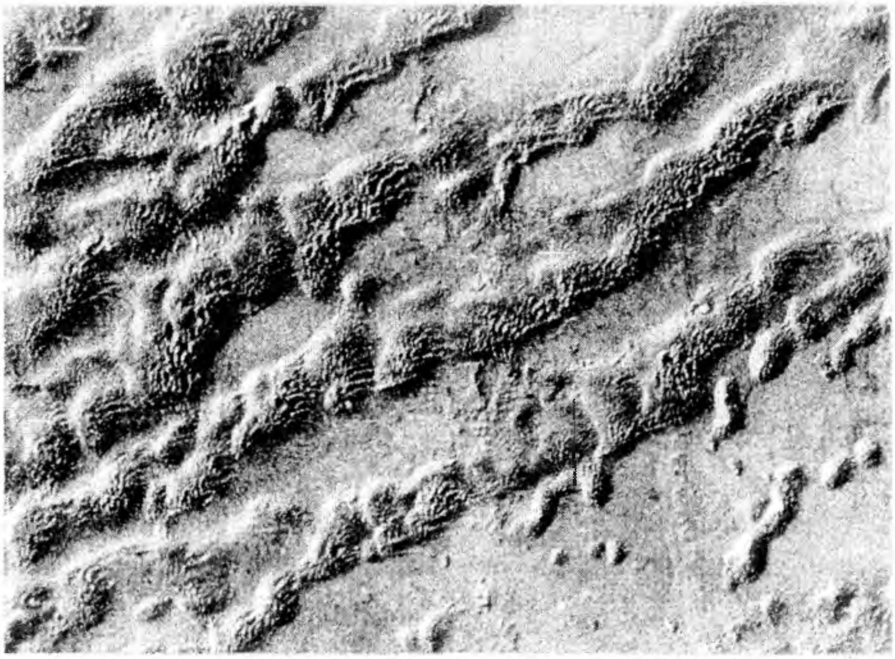


Fig.45 Barchanoid sandridges in the western part of Idhan Awbari
(Metric Camera, strip 24, photo 708).

idea was postulated to explain the origin of the longitudinal dunes of the Simpson Desert (King, 1960; Mabbutt, 1965; Folk, 1971a). However, Verstappen (1960) hypothesised that the seif dunes (which in this work are referred as longitudinal sandridges), represent denudation/deflation forms in the earlier phase of their development. He believed that, under favourable conditions, the ridges tend to become aerodynamically stabilized, due to the differences in wind velocity over the ridges and depressions and particularly because of the edddy current so caused. The mechanism of dune development can be explained by assuming that the flow of the wind is helical (Bagnold, 1953; Hanna, 1969) (Fig.46).

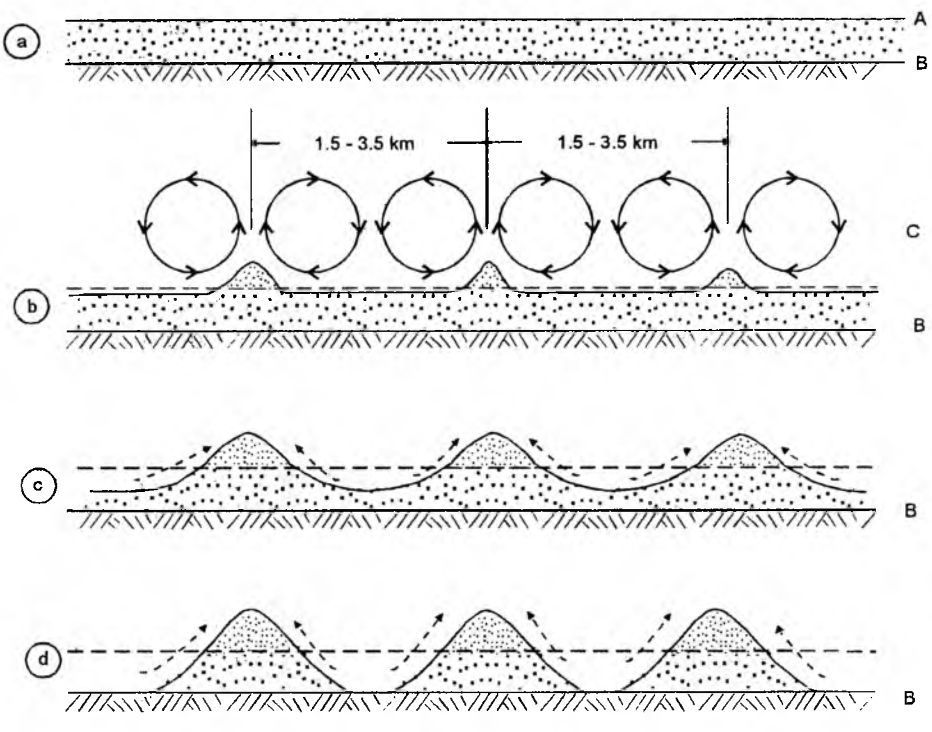


Fig.46 Mechanism of longitudinal sandridges development (modified after Bagnold,1953 and Hanna,1969). a/ alluvial plain before formation of sandridges, b/ first stage of sandridges development, c/ further development of sandridges, d/ mature stage of sandridges. A/ upper level of alluvial plain, B/ upper level of rocky basement, C/ paired roller vortices.

When still air is heated, convection cells with vertical axes are formed, but when the wind is moving, heat causes the formation of paired roller - vortices with horizontal axes which are parallel the wind direction. The poorly - sorted alluvial plains are thus reworked by the paired roller - vortices which cause the sand to follow an elongated helical path (Fig.46). The sand which is eroded from the alluvium along such helical paths, is transported up the dune flanks. The sand is then deposited on the ascending parts of the adjoining roller - vortices. Hence, in this hypothesis the dunes are sites of deposition caused by helicoidal (Langmuir) circulation, in which winds meet and dropp their saltatory load. The interdune areas are the sites of descending winds which bite deeply into the sand surface, causing deflationary erosion there. A final result of above described mechanism is a production of a regular system of parallel depositional dunes and erosional interdune areas would. Such compound dunes comprise of two parts: a lower, which originated from the alluvial plain, and an upper part produced from depositional sand. Thus, such dunes (sandridges) are *partly erosional and partly depositional*. The system described is self - propagating because the dunes are hotter than the interdune areas and so air trends to rise over them (Folk, 1971a).

Bagnold (1953) has emphasised that, whether or not the sand of dune ridges becomes hotter than the bare ground on either side, far more intense saltation of sand driving over the dune ridge will cause greater heating of the air immediately above. Bagnold believed that the vortices due to this differential heating tend to maintain their initial orientation, irrespective of wind direction. Thus, once a dune ridge or even a flat tongue of sand has formed, it will tend to perpetuate its intial orientation irrespectively of wide seasonal changes of wind direction.

There are, however different hypotheses relating to the spacing of sandridges and to the mechanisms which are responsible for this spacing. Bagnold (1953) suggested that the spacing of ridges is a function of the instability of air over a great desert plain. In his estimation, during hot weather, the unstable atmospheric layer has a height from 300 to 1000 m. It was found that the distance between the axes of vortices is about three times the height of atmospheric instability; hence, the dune ridges are separated by distances equal to the width of a vortex - pair, which is about six times the height of the unstable atmospheric layer. Thus, the spacing between sandridges should be between 2 and 6 km.

More evidence regarding the pattern of wind velocities has been given by Thornthwaite, cited by Bagnold (1953). On the basis of the microclimatological network in the mid - western area of the United States, consisting of more than two hundred weather stations, it was found that there were belts of high and low velocity which were always always parallel to the wind direction. These belts were located at a distance of about 2.5 km.

The orientation of the huge longitudinal sandridges in Idhan Awbari and Idhan Murzuq indicates that the sandridges were formed under the influence of strong winds which mantained a constant direction. Such winds, with pronounced unidirectional orientation and much stronger than present winds, were pravailed in the late Pleistocene and Holocene (Nicholson and Flohn, 1980).

The sandridges, formed by this suggested mechanism, can be in different stages of development, depending on the level of the eroded alluvial cover. Full maturity will be reached only when the alluvial mantle is eroded to basement level and the interdune corridors will show underlying rocks with a very thin cover of sand. Thus, the height of the sandridges seems to depend on the thickness of the original alluvial plain and duration of dune development.

Author hypothesises is that the primary forms of sandridges could have been developed in either one cycle of arid conditions or during several cycles. It is possible that in both Idhan Awbari and Idhan Murzuq the longitudinal sandridges were formed in one cycle, during which they reach maturity or semi - maturity, or in several cycles, under the same wind regime. In both the process of excavating the troughs between the dunes and heaping up the eroded sand into dunes progressed. During the humid periods, the dune crests became more rounded, and then on the dune surfaces, the fossil developed, thereby fixing and immobilising the sandridges.

When the final stage of sandridge development is reached, futher development then depends on sand supply or recycling of existing forms. Analysis of the Idhan Awbarii and Idhan Murzuq areas revealed that the sand supply from external sources is very limited and the active sand cover is only present on the uppermost parts of the sandridges. The active sands accumulated in secondary forms, passing over the sandridges. Their form depended on the present day wind regime which is characterized by seasonal directional variability.

The secondary forms, developed on the primary forms, reflect the present wind regime. *The sandridges have created the present terrain topography, which, in turn, influences the local wind regime*. In the majority of sandridges, the secondary forms are developed only on their windward flanks in the form of linear dunes or crescentic chains, oriented obliquely to the main axis of the sandridges. These forms can be regarded as sand transport bodies or vehicles which transport the sand from the interdune corridors or from the lower parts of sandridges to the upper parts. A detailed analysis of satellite images indicates that contemporary dune activity is restricted mainly to the dunes uppermost parts of the dunes, which are influenced by winds stronger than these in the lower parts or in the interdune areas. In some parts the Idhan Awbari, on the top of sandridges, the star dunes developments give evidence of a multidirectional wind regime in these regions. At the top of the dune, the wind velocity is accelerated and has much greater energy than on the flanks or in the interdune areas. In alternate seasons, the winds create reversed dunes and/or the star forms which appear on the tops of the sandridges. In the case of the segmented sandridges, the star dunes are developed on the top of the ends of segments, thus forming the highest points of the segmented sandridges.

The longitudinal sandridges in Idhan Murzuq have a somewhat different morphogenesis. The geological records reveal a very thick layer of unconsolidated sand, hence the interdune corridors are not fully eroded here, and the rocky basement is not exposed. The model of sandridge development differs a little when compared with supposed sandridge development in Idhan Awbari. Assuming that the origin of sandridges dates to the same geohistorical period and that both areas were influenced by the same wind regime, they should have much the same morphometric parameters and, more or less the same morphogenesis. In the case of the sandridges in Idhan Murzuq, the sand from the troughs was eroded to the same level (or deeper) as in Idhan Awbari, but, due to the deeper sources of the underlying alluvial sand, the interdune areas are floored by unconsolidated sand alluvium instead of a rocky basement. As the result of consecutive cycles of arid and humid climate episodes, the process of sandridge development was repeated. The sand eroded from the corridors was heaped up to the top, to a stage where the dunes reached a state of equilibrium (constant ratio of dune height to dune basis). The surplus of sand heaped up was then transported over the top and deposited on the other flank or transported along the dune.

3.2.3.2. Barchanoidal sandridges

The barchan - like sandridges are composed of the large mega - barchans arranged in chains and covered by secondary dunes. In the Libyan Sahara there were only two fields of such forms which detected on the satellite images: one in the Idhan Awbari (Fig.47, area 4 and 14) and second in Idhan Murzuq (Fig.48, area 23). In Idhan Awbari (area 14), the sandridges are composed of short barchan - like segments of 2 to 5 km length, arranged in chains spaced 1.8 to 2.5 km apart. Their flanks are covered by linear dunes or crescentic chains, which appear along the windward slopes. Fig.44 shows the barchanoidal sandridges which are present in easternmost flank of area 4 (Fig.47) in Idhan Awbari.

The morphogenesis of barchan - like sandridges mainly relates to the wind regime in different cycles of sand sea development. In the case of barchan - like sandridges, both the primary forms and seceondary forms are similar, which confirms that the direction of the winds during the initiation of dune formation and during the consecutive phases of dune development was much the same.

3.2.3.3. Domical sandridges

Localities and characteristics

In the Libyan Sahara, domical sandridges appear only in Idhan Awbari. These sandridges have a domed form. The vertical cross - section is in the form of a symmetrical triangle, whereas a horizontal cross - section approximates to a circle. These forms can be further divided, according to their pattern, into the dunes arranged in a random pattern and dunes arranged in chains (Fig.7). The domical sandridges in random pattern occupy the largest part of area 7 and the southern part of area 4 (Fig.47). The diameter of domical sandridges varies between 1 and 2.5 km, and their spacing between 1 and 3 km. The dune heights range between 50 and 160 m over the interdune areas. Generally, the upper parts are flat - topped. In several cases star dunes are superimposed on the domical sandridges and these forms are the highest. Such compound forms looks like large star dunes, but the arms of star dunes are only restricted to the upper part of the dune, and never extend to the dune base. The flanks of the dunes are generally covered by secondary dune forms, such as crescentic chains, the axis of which is oriented towards the dune summit.

In several places, the domical sandridges are arranged in a distinct pattern of chains (Fig.47 - the easternmost part of area 7, area 3 and area 4). The dune chains are spaced at 1.5 to 2.5 km intervals and the average distance between dunes in chains is ca 1.8 km.

Hypothetical morphogenesis of domical sandridges

The location of domical sandridges in the central part of Idhan Awbarii indicates that their occurance is related to an abundance of sand. The random pattern of sandridges is probably generated by the multidirectional wind regime, highest drift potential being recorded for Sebha (DP = 655 VU) and Ghadamis (DP = 588). The chain - like pattern of domical sandridges may be attributed to the pre - existing longitudinal sandridges, redeveloped in time into domical sandridges. Indirect evidence of such a hypothesis lies in the location of domical chains: in the eastern part of area 7, the chains of domical sandridges continue the trend of longitudinal sandridges in area 11. The chains of domical sandridges in area 3 constitute an extention of longitudinal sandridges of area 1 and partly the extention of sandridges in area 2. In the case of chains of domical sandridges in the eastern part of area 1, these forms may be explained as the result of the wind regime. However, the abrupt change of dune trend, as in area 7, cannot result solely from a change of wind regime. The existance of such forms can be only related to *a different sand origin*. It is possible that the sands forming this dune field were deposited later during a pluvial phase. The shape of this field seems to correlate with the depression area, and it is possible that this depression was reworked later during a pluvial phase, when the existing longitudinal sandridges were lowered and partly destroyed and a new wind regime, in the consecutive arid phase, created such a pattern of domical sandridges.

3.2.3.4. Pyramidal sandridges

Localities and characteristics

Typical pyramidal sandridges (Fig.8) occur only in area 13 of Idhan Murzuq (Fig.48). These pyramidal dunes are based on a triangular plane view; thus they have three faces. The pyramidal dunes are spaced between 2 and 3 km, and the length of their base ranges from 1 to 2 km. On the western flank of this field, they are arranged in a random pattern, while, on the eastern flank, they form chains of linked pyramids. The flanks of the dunes are covered by crescentic chains oriented towards the dune peak.

Hypothetical morphogenesis

The location of the dunefield with pyramidal sandridges is the most significant evidence in the interpretation of such dune genesis. The dune field is located in the vicinity of the Msak Mastfat mountains complex and in the vicinity of Wadi Barjuj. Assuming that the dune forms have a complex history, it is possible that this part of Idhan Murzuq was strongly influenced by a pluvial phase, both in terms of water activity and sand transport as well as the destruction of previous dune forms. On the eastern part of area 13, the pyramidal dunes resemble the chain pattern and are similar to the domical sandridges. This chain - like pattern may be related to the partial destruction of the longitudinal sandridges occuring in area 9. The new wind pattern in the consecutive arid phase, with different wind directions, tailored the final pyramidal forms.

3.2.3.5. Reticulate sandridges

Localities and characteristics

The reticulate sandridges constitute very distinct forms and patterns in Idhan Murzuq and Idhan Awbari. In these patterns, the two dune trends cross nearly at right angles and form *closed interdune areas*, thus resembling *a reticulate pattern*. The reticulate sandridges in Idhan Murzuq are arranged in a very regular pattern over area 14 (Fig.48). The reticulate sandridges are very clearly depicted on the Landsat colour composite image (Fig.8), where the differentiation in sand colour between dune crests dune flanks and interdune areas is enhanced. This pattern is characterized by two trends of sandridges, NE-SW and NW-SE. The more significant and continuous is the NE-SW trend, and the NW-SE trend is less distinct. These two trends of sandridges create *elliptical hollows*. The width of sandridges is up to 1.5 km, while in a typical hollow the longer axis varies between 3 and 5 km (NE-SW), and shorter axis between 1.5 and 2.5 km (NW-SE). In several cases, the deflation hollows are exceptionally large, up to 3×15 km. A somewhat different pattern of reticulate sandridges is developed in area 15. The main trend is composed of sandridges 3 to 10 km long, while the lateral ridges are 3 to 5 km long. The modular ridges are between 1.0 to 1.4 km wide and their flanks are covered by secondary dunes arranged in lines running parallel to the main ridges.

The reticulate sandridges also appear in the southwest part of Idhan Awbari (Fig.47 area 16) and in the middle part of Idhan Awbari (Fig.47 area 18). In area 16, the two trends of sandridges cross at right angles forming the closed interdune areas, which resemble the reticulated pattern. The more distinct and dominant is the E-W trend, in which the modular sandridges vary in length between 6 and 15 km. The sandridges in the second, N-S, trend are shorter - between 1 and 3 km. The dune tops are up to 70 m in height.

Another type of reticulate sandridge occurs on the southern part of area 15 in the Idhan Murzuq. In this pattern, the lateral ridges are not completely associated with the dominant trend of sandridges; thus, they form a *backbone - like* *pattern* with short arms on both sides of the main sandridges. The interdune areas are not completely closed but are linked together.

Hypothetical morphogenesis

The hypothetical morphogenesis of reticulate sandridgeis is based on the supposed multicyclic development of such pattern. Analysis of the reticulate pattern of sandridges in Idhan Murzuq indicates two trends of dune development. The primary trend, the more significant, represents the longitudinal sandridges formed within the earlier stages of sand sea development. The second trend, which is less significant, was developed in the later stages, when different winds prevailed. During these stages, the shorter dune ridges, those perpendicular to the main sandridges, were developed, thus forming the reticulate pattern. The different stages of reticulate pattern development can be observed in Idhan Murzuq. In area 15 (Fig.48) the shorter arms extend from the eastern towards western sandridges, but both main trends of the primary sandridges are disassociated. In the area 14, the secondary trend extends between the main sandridges, hence, forming the elliptical hollows. The first example is essentially an immature stage of the reticulate pattern, whereas the second example depicts a mature stage. The development of reticulate patterns reflects new orographic conditions and hence, a new wind regime. The interdune hollows create cellular circulation of winds, which is responsible for further deflation of the interdune areas and transport of sand towards crest of sandridges. It should be noted that, in the interdune areas, both megaripples and simple dune forms are not observed.

3.2.3.6. Rim pattern of sandridges

Localities and characteristics

Sandridges arranged in *rim pattern* appear only in Idhan Awbari in areas 5 and 13 (Fig.47). These forms surround large deflation areas. In area 5, the longer semirims are oriented NE-SW and shorter ones ESE-WNW. Their respective lengths range from 5 to 15 km and from 3 to 4 km. In area 13, the rim pattern reflects a larger scale. In the western part, the elliptical depression areas surrounded by these rims are up 12 km long and 10 km wide which diminish to 6 and 3 km respectively towards the northeast. The flanks of these sandridges are covered by secondary forms, mainly crescentic chains and linear reversed dunes which are located on the upper parts of sandrdges.

Hypothetical morphogenesis

Both fields covered with rim patterns of sandridges are located at the boundaries of different dune forms. Area 5, is surrounded by longitudinal sandridges in the west; in the north, by chains of domical sandridges and in the east and south by domical sandridges arranged in random pattern. A very similar situation is presented in the area 13, where, in the north, this field is bounded by longitudinal sandridges trending NE-SW, in west by longitudinal sandridges trending WNW-ESE. The location of those fields at the intersection of different dune trends, indicates both the sand origin and wind regime. The earlier dune forms which existed in the past were reworked and, due to the strong wind action, deflationary hollows were created. Presently, the elliptical depressions are covered by a rather thin sand layer without visible dune forms. The secondary dunes located on the flanks of sandridges indicate the mutidirectional wind regime. There are very sharp crests, on the upper parts of dunes, indicating reversed flow connected with seasonal wind changes.

In the depression hollows in the Ramlat and Dawadah, there are saline lakes and oases, such as the famous oasis of Mandarah which is situated along the desert lake (Burdon and Confiantini, 1991). The village is situated at the base of the sandridge, thus indicating that the sand is not invading the oasis. The movement of active sand is probably limited to the upper parts of these sandridges. All salt lakes, as well as the depressions, are very clearly shown on the radar satellite images from the Space Shuttle Mission. The appearance of salt lakes and sabkhas is clearly related to higher water levels and, thus, with higher moisture of deflationary areas.

4. GEOMORPHOLOGY AND MORPHOGENESIS OF LIBYAN SAHARA SAND SEAS

4.1. IDHAN AWBARI

4.1.1. Location, topography and sand sources

Idhan Awbari (Fig.47) belongs to the sand seas which are located in structural basins. Such locations strongly influence the sand origin and transport. The geomorphological development of this region (see Appendix 1) indicates that the sand was deposited in this basin during a single pluvial phase or during several pluvial phases. The sand was derived from the surrounding mountains. Rivers transported sand into the basin, thereby forming a thick alluvial plain which constitutes the source material for aeolian reworking of these deposits. The sands incorporated in Idhan Awbari are thus to be regarded as being *autochthonous origin*.

Idhan Awbari constitutes a semi - closed sand sea basin, which extends from the border with Algeria (10°E) through Ramlat az Zallaf and terminates at Ar Ramlah As Saghirah (15°30'E) (Fig.47). Idhan Awbari covers an area of 55,190 sq kms. Hamadat Tanghirit, a moderately undulating plain, and Qararat al Marar, delineate the sand sea in the north. The northeastern flank of Idhan Awbari follows the 400 m contour line, running N-S along the foreland of Qararat Al Sifar and Jabal As Duwaysah. At latitude 27°20' N, the sand sea boundary turns to the SEE and runs through a discontinuous belt of depressions with sabkhas.

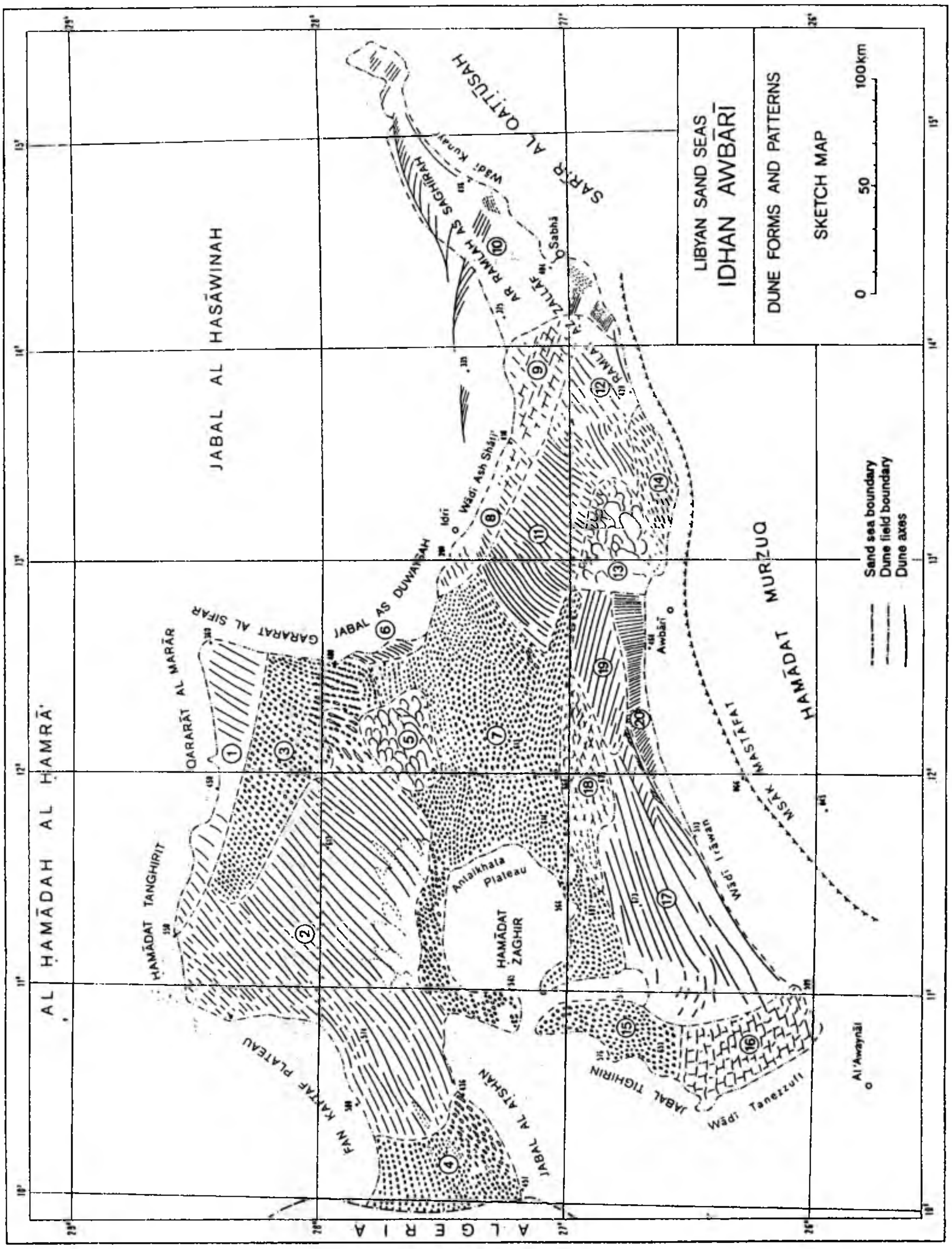


Fig.47 Dune fields in Idhan Awbarii
(In circles the numbers of the dune fields described in the text).

Wadi As Shaykah appears on the western flank of this depression and, towards the east, Wadi As Shati occupies a flat depression. The sabkhas surface is at an altitude of 350 m, which is the lowest level of Idhan Awbari. On an arbitrary division, the main part of Idhan Awbari terminates at 13°30'N, and then transits into Ramlat Zallaf which extends eastwards for a distance of 200 km to 15°30'E and covers an area of 6 890 sq kms. To the north, Ramlat Az Zallaf is bounded by the Quaternary outcrops appearing south of Wadi Al Shati and Wadi As Sayl. From 14°E, the Ramlat Az Zallaf changes its direction towards the NE, and, in the north, it is delineated by ancient wadis. The southern limit of Ramlat As Zallaf is defined by a nearly straight line connecting Wadi Kuntar on the northeast flank and Wadi Hayah to the southwest. This line corresponds approximately with the 420 m contour. Towards the southeast, Ramlat Az Zallaf adjoins the vast area of Sarir al Qattusah.

Towards the southwest, the southern limit of Idhan Awbari is formed by a broad, flat depression stretching along the northern side of the Mesak Mastafat escarpment. The eastern part of this depression is known as Wadi al Hayah (formerly as Wadi Ajal) and Wadi al Irwan. Along this depression, terrain rises gradually from 460 m in the vicinity of Awbari to 560 m on the western flank of Wadi Irawan. The mid - western part of the sand sea is interrupted by the Hamadat Zaghir, which includes the Antalkhata Plateau. The south - western limit is marked in the northern part, by Jabal al Atshan and, in the southern part, by Jabal Tighirin.

The western flank of Idhan Awbari extends to the Algerian border between latitudes 27°15'N and 28°N. This belt forms a gently undulating depression with little variation in the altitude. The lowest point at ca 420 m, is located in the central part of this depression, whereas the highest elevations of ca 500 m occur both on the northern and southern limits of this belt. The southern boundary of this depression is formed by the footslopes of Jabal al Atshan and the northern boundary by the rim of Fan Kaftaf Plateau.

The central part of Idhan Awbari constitutes a rocky plateau known as Hamada Zegher. Hamada Zegher represents a nearly flat plateau of Carboniferous rocks, bounded by a small escarpment. The terrain elevation rises from 500 m on the northeastern flank to 580 m on the southwestern flank. On the eastern part of Hamadah Zegher, called the Antalkhata Plateau, undivided Tertiary continental deposits outcrop. This plateau has an area of 1400 sq km, and mean altitude of ca 530 m.

The sand sea floor gently rises from a level of ca 400 on the eastern part up to 550 m on the western border. There is also a difference in a north - south direction: the southern border is higher than the northern border. The general slope of the sand sea is SW-NE.

4.1.2. Geomorphology of the sand sea

Idhan Awbari presents a variety of dune forms and patterns; hence, there is very differentiated geomorphology in sand desert areas (Linsenbarth, 1986a). The commonest forms are the longitudinal sandridges, domical and reticulate sandridges and the so called deflation rims. On the margins of the sand sea, composite linear dunes, linear dunes and crescentic chains can be discerned. Geographically, the sand sea can be divided into several parts: northwestern, central, eastern and southern.

4.1.2.1. The northwestern part

The northwestern part is occupied by longitudinal sandridges and by domical dunes. The largest field of sandridges is in area 2; this has a very distinct NE-SW dune trend. At the distance of 121 km, there are 31 sandridges. The length of ridges varies between 60 and 85 km; width, between 2 and 3 km and spacing ranges from 3 to 4 km. The crests of the highest ridges are 120 m higher than the the dune bases. The interdune areas are at a level of ca 500 m while the crests are over 550 - 600 m a.s.l.. In the middle part of this field, the linear dunes, arranged in braided forms, are superimposed on the crests of sandridges. The western, wide and gentle slopes are covered by short, slightly arcuate linear dunes, diverging obliquely from the dune crests. Towards the west, the dune width diminishes to 0.4÷0.6 km; the spacing varies between 1.0 and 2.5 km, and instead of long sandridges, segmented dunes appear. In the northern part of this area, the sandridges grade into composite linear dunes, with feathered tails developed on the northwestern flanks of dunes. In several parts of the northwestern corner of this field, star dunes are developed on sandridges. In the central - southern part of this field, the ridges are thinner and the interdune areas are connected and a rocky basement is also evident.

The domical sandridges appear in this part of the sand sea in areas 3 and 4. In area 3, the domical sandridges are aligned in chains, which have the same orientation as the longitudinal sandridges in area 2. The dune chains are spaced at 2.5 km intervals and the average distance between dunes in chains is ca 1.8 km. On the eastern flank of this field, the orientation of the dune chains changes their orientation to NNE-SSW. In the vicinity of the Qarart al Sifar escarpment, the dune relative heights are as much as 100 m.

The western part of a sand sea, near the Algerian border (area 4), is covered by domical sandridges arranged in chains. They are spaced 1.5÷2.5 km apart and the distance between the dunes in the chains averages 1.8 km. Towards the west, the star dunes occur on the domical ridges. Along Jabal Atshan, there is a cluster of dunes assembled into barchanoid sandridges.

The composite linear dunes constitute the northern part of Idhan Awbari (area 1). In the western part, the dunes are feathered, while, towards east, they have braided forms. Dune lengths range from 5 to 15 km and spacing varies between 2.5 and 3.0 km.

Between the northern and central part of the sand sea, area 5 is easily defined by its very characteristic pattern of deflation rims. The dune ridges form rims of sandridges surrounding large depressions. The longer semi - arms, oriented NE-SW, are 5 ÷ 15 km long, and the shorter - trending ESE-WNW are 3 ÷ 4 km long.

4.1.2.2. The central part

The central part of Idhan Awbari, surrounding Hamadat Zaghir, is covered by domical sandridges. In some parts, domical sandridges are arranged in chains, while, in others, they are randomly arranged. Their diameter varies between 1 and 2.5 km and spacing, between 1 and 3 km. Dune heights range between 50 and 100 m, but, in the south of Hamadat Zaghir, the dune peaks are elevated 160 m over interdune areas. In several cases, the star dunes are superimposed on the domical sandridges. In the eastern part of area 7, the domical sandridges are arranged in distinct chains, running subparallel to the longitudinal sandridges in area 11. To the north of Hamadat Zaghir, there is a 10 km - wide belt of domical sandridges, arranged in chains with star dunes superimposed. The chain - like pattern of domical dunes also occurs in the southeastern part of area 7, but the dune trend is here E-W. The dune chains form the extention of the longitudinal sandridges which appear in area 19. Within area 7, the density of sandridges and their diameters indicate gradual change; in the northeastern part, the dunes are more densely spaced (1.2 km) and are lower, while towards southwest the spacing is larger (up to 2 ÷ 3 km) and the dunes are much higher.

To the south, area 7 adjoins area 18 in which the reticulate dune pattern occurs. Reticulate dune patterns show of two dune trends: NE-SW and E-W. Interdune areas form circular or elliptical fields. Generally, in this part of the sand sea, the dune pattern is very complicated. Star dunes are developed at the intersections of two dune trends. The interdune areas lie at an elevation of ca 500 m, whereas the summits of star dunes are over 620 m.

4.1.2.3. The eastern part

The eastern part, extending in an easterly direction from area 7, is covered by different dune forms which have different trends. Area 11, presents very distinct and regular NE-SW trend of longitudinal sandridges. On the eastern part of this field, at a distance of 30 km, there are 11 ridges arranged in a parallel straight line pattern. The length of ridges varies between 25 and 35 km, their width ranges between 1.0 and 1.4 km, and their spacing is ca 2.8 km. The interdune corridors are defined by the 450 m contour line and the dune crests by by the 500 m contour. Towards the west, the sandridges swing slightly towards the south and their spacing is not so regular. The sandridges are segmented and finally grade into the chains of domical dunes of area 7. In the south - eastern part of area 11, there is a belt of short longitudinal sandridges which has the same orientation as the longer sandridges in the northern part of this field. Between these two parts, there is a very distinct zone without sandridges, and the dunes in the southern part of this field do not represent the projection of dunes in the northern part, but are shifted in EW direction.

In area 12, partly comprising Ramlat Zallaf, the longitudinal sandridges are segmented, and their trend oscillates between NE-SW and ENE-WSW. Areas 8 and 9 contain rather marginal dune fields of the sand sea. In area 8, linear dunes are developed, while in area 9, a second, NW-SE trend can also be detected. In the southern flank of this part of the sand sea (area 14), in the vicinity of Wadi Al Hayah, barchanoid sandridges are developed. The short barchan - like segments 2 ÷ 5 km long are spaced 1.8 ÷ 2.5 km apart. Linear dunes and crescentic chains appear on the windward slopes.

Area 13 represents a very characteristic dune field on which occur sandridges which are arranged in rim pattern occur. These sandridges surround the large deflation areas. In this pattern, the longer semiarms are oriented NE-SW and the shorter ESE-WNW. The lengths of sandridges range from 5 to 15 km, and from 3 to 4 km.

4.1.2.4. The southern part

The southern part extends towards the south and southwest from area 3. Longitudinal sandridges are the dominant dune forms. Area 19, known as Irq al Qattus, is composed of 11 sandridges trending ESE-WNW and spaced 2.2÷2.6 km apart. The dune length varies between 25 and 60 km; width ranges between 0.8 and 1.2 km, and the interdune corridors are 1.2÷1.6 km wide. The dune crests are composed of 2 or 3 linear dunes arranged in braided form and, on the gently sloping northern flanks, secondary oblique dunes may be present. Towards the west oblique linear dunes cross the interdune areas creating the *grid pattern* characteristic of area 18.

Between Hamadah Zagher and Wadi Irwan, a vast area of sand sea (area 17) is covered by very large longitudinal sandridges. There are 11 sandridges, oriented ENE-WSW, spaced at ca 3 km in the eastern part, an interval which gradually increases to 4 km in the western part. The dune length varies between 35 and 85 km, and the width of the dunes and interdune corridors averages 2 km. On the northern flanks, oblique linear dunes appear as secondary forms. The interdune corridors are at a level of between 500 and 515 m, whereas the crests rise to elevation between 640 and 675 m. The maximum dune heights reach 140 m. In the northern part of this field there are very large interdune areas, ca 22 km long and 3 km wide, in which the rocky basement is exposed.

The southern limit of area 17 is defined by a very characteristic huge sandridge which forms a sand barrier or embankment. This sandridge is some 80 km long and consists of a few segments, 20÷35 km long. The dune crests are composed of 2÷3 linear dunes arranged parallel with the dune axes. The width of the crestal part is 1.2÷1.4 km, whereas the width of the ridge at the dune base varies between 3 and 5 km. On the eastern part of this ridge, there are linear dunes, attached on the northern flank, and trending at a very acute angle to the main ridge.

In the western part of area 17, the sandridges appear on the rocky basement; they here have an E-W and ESE-WNW alignment and the sandridges are segmented. The south - western flank of the sand sea consists of areas 15 and 16. Area 15 represents an elongated dune field ca 55 km long and 20 km wide, aligned along Jabal Tighirin. The dune diameter varies between 1.0 and 1.2 km and the spacing ranges from 1.5 to 2.5 km. The dune heights are up to 120 m over the basal part of the dune. Towards the south, the domical sandridges become segmented sandridges, which finally transit into a grid pattern in area 16. This pattern combines two dune trends which cross at right angles and form closed interdune areas or hollows. The ridges in this pattern are longer in the E-W direction (6÷15 km) and shorter in the N-S direction (1÷3 km). The elevation of interdune areas varies between 580 and 600 m, while the dune peak elevations are up to 640 m.

North of Wadi Irawan, a very characteristic sand field occurs (area 20). This elongated field (85 km×4 km) is covered by short linear dunes, which are slightly arcuate and oriented N-S in the eastern part and NE-SW in the western part. Dune length varies between 4 and 8 km, and the spacing between 0.6 to 1.0 km. Dune crests are not visible on space photographs; hence, these dunes seems to be rather flat and low.

4.1.3. Hyphotetical morphogenesis and morphochronology

Idhan Awbari dispays a variety of dune forms and patterns. Analysis of the interrelationships between these forms and their location within the sand sea indicates that several factors were responsible for the development of particular forms. The principal factors governing dune form are: terrain topography, sand origin and deposition and wind regime. All these factors should be related to the long period of sand sea development and its several phases of development. Only such an approach can give a the general conception on the hypothetical morphogenesis and morphochronology of the sand sea.

The shape of the sand sea is determined by terrain topography, which also affects the sand deposition and local wind regime. Analysis of sand sources indicates that the *major part of the sand* which is ioncorporated in the sand sea *has autochthonous origin* and was derived from the surrounding plateaux and mountains. The sand derived from the mountains during pluvial phases was transported towards the central part of the sand sea where it was deposited. During the arid phase, the dune development started and reflects the influence of both erosional and depositional wind action. It seems that the extent of the deposited sand probably was not equal in different parts of sand sea and must depend on the origin of the sand, and the terrain topography. During the arid phase or phases, the particular dune forms were developed under the wind regime specific to the locality. Additionally, the aeolian sand has been formed as a result of wind erosion of the surrounding bedrocks and subsequent transport towards the Awbari area. The main sources of sand were the Paleozoic rocks and the Miocene continental deposits.

Different parts of Idhan Awbari had their own particular genetic background, which is manifested in the diffferent dune forms and patterns now present here. The present Author hypothesises that the primary forms of compound dunes were created in the earlier arid phases in the Libyan Sahara development and that *the different dune trends are in fact related to different arid phases*. The *sand source* is *mainly autochthonous* with a small amount of *allochthonous* sand. Owing to the basinal character of the Awbari sand sea, sand accumulated in this basin during the long pluvial phases, forming a vast alluvial plain or plains. It may be assumed that large rivers flowed through the Awbari basin and their courses can be deduced from the abrupt changes in the dune trends. Such abrupt changes occur, for example, along the the southern limit of area 11 and in areas 7, 19 and 13. The former rivers were able to prevent dune development and transfer of the sand. Other examples of the rivers acting as barriers to dune development exist in South Africa and in Niger. In the Aouker dune field in Niger, the dunes of the southern edge of the erg were dissected and partly destroyed by the Karakoro system of wadis (Grove and Warren, 1968). In the Namib Desert in the South West Africa, the Kuiseh river forms the northern boundary of the longitudinal sand dunes (Hanna, 1969). Elsewhere, it is obvious that the dunes are capable of blocking the course of rivers, e.g. the Senegal river was blocked by sand which was responsible for the development of ephemeral lakes about 150 000 years ago (Grove and Warren, 1968). Langford (1989) has pointed that aeolian topography can both change the existing drainage networks or create aeolian forms.

As referred to earlier, the sandridges are complex erosional - depositional forms, in which the interdune areas were eroded from the alluvial plain, and the eroded sand was trapped towards dune crests which form the upper part of large sandridges. Many of the longitudinal sandridges present the *matural forms* i.e. the forms which are fully developed. The degree of the sandridges maturity may be related the interdune areas. In the case when the interdune areas are formed by the underlaying rocks, the dune cannot develop any further, because the rocky basement is itself providing no more sand. Therefore if there is no external source of incoming sand, the aeolian process is restricted to particulate sandridges without inter - dune sand transfer. It seems that the *primary forms are resistant to wind erosional action*; on the other hand, there is evidence that the longitudinal sandridges, which appear in the southern part of area 2, have wider interdune areas, showing that further erosional action is influencing the basal parts of the sandridge.

The destruction and/or the reworking of the previous forms, in different climatic periods results in development of new dune forms created under different climatic conditions and/or wind regime. As an example, a completly different dune trend is obvious in area 19. In this field the sandridges are oriented nearly perpendicular to the sandridges present in the adjacent area 11. Area 3 provides an example with domical sandridges arranged in chains. This area was probably influenced by pluvial activity which partially destroyed the previous sandridges, and added the alluvial sand which, later, in the consecutive arid period, was used in the development of the present forms.

The rim pattern of sandridges occurs in the areas which are influenced by the very strong winds, which come from two directions and diverge in these areas. Large deflation hollows were created as a result of such action.

A majority of the sand dunes occuring in Idhan Awbarii represent compound forms, consisting of semi - fixed primary forms and secondary forms developed on their flanks. Analysis of satellite images indicated that crescentic chains and linear dunes are the most commonest secondary forms.

In many cases, star dunes are developed on the uppermost parts of the sandridges. These secondary forms represent the active part of the sand sea. The orientation of these forms is related only to the present wind direction; hence, these forms are oriented obliquel or perpendicular to the sandridges crests or brinks. Owing to the multidirectional wind regime which prevails here and which results in reversed flow directions, sand movement and transport seems to be restricted to particular forms only, or with only very limited transfer of sand between the primary forms or no transport at all.

The relative dune heights increase from east to west, which indicates a long term transfer of sand towards the west. The dune forms are related to the amount of sand. On the eastern flank, with only a limited sand cover, only linear dunes and crescentic sandridges occur; there are succeded by longitudinal sandridges and finally by domical sandridges. This sequence of dune form change results from wind direction, sand transport direction and increasing thickness of sand.

4.2. IDHAN MURZUQ

4.2.1. Location, topography and sand sources

Idhan Murzuq (Fig.48) is situated in the structural Murzuq basin, which has been created by geotectonic events. This means that the morphology of the sand sea is being controlled by structural characteristics and topographic location. Idhan Murzuq extends over an area of ca 71,000 sq km and has an elliptical shape elongated NE-SW with the major axis 320 km long and with the NE-SE oriented minor axis extending for 260 km. The sand sea is bounded to the north by the foreland of Hamadat Murzuq.

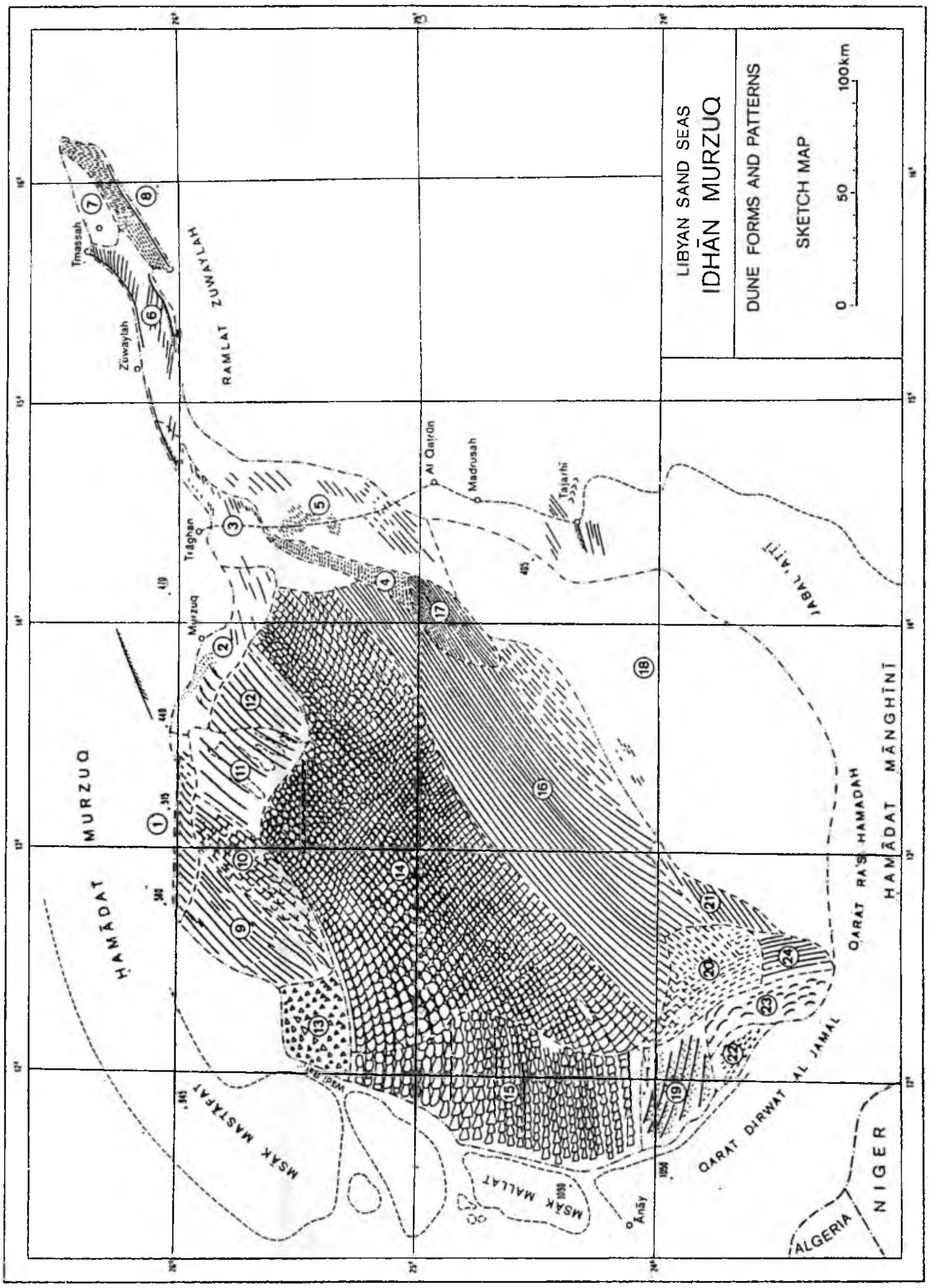


Fig.48 Dune fields in Idhan Murzuq
(In circles the numbers of the dune fields described in the text).

The continuous escarpment, which is actually a homogenous unit subdivided only in terms of having different geographical names in different places, forms the western edge of the Murzuq sand sea. The northern part of this escarpment is Masak Mastafat (1250 m); the central part is named Masak Mallat and the southern part Qararat Dirwat al Jamal. The foreland of these escarpments has an altitude of 700÷750 m. South of Col d'Anui the escarpment rises to its maximum height of 1250 m. The total length of the NS-trending escarpment is about 500 km. To the south, the sand sea is limited by Hamadat Manghini (1222 m) and, to the east, by the foreland of Jabal Bin Ghanimah. In the NE, Idhan Murzuq merges into Ramlat Zawaylah which extends up to 16°15'E.

The basin surface rises gently towards the southwest. On the northern margin of the sand sea, the elevation is about 460 m, whereas on the southwest flank, it rises to 800 m. All the valleys, which intersect the plateaux surrounding the Murzuq basin, are arranged radially to the centre of the basin.

4.2.2. Geomorphology of sand sea

The northern region of Idhan Murzuq, which is defined by the areas 1, 2, 9, 10, 11 and 12, is occupied by longitudinal sandridges (Linsenbarth, 1986b). Within this region, two subregions can be separated: the northern, with areas 1 and 2 and the southern, with areas 9, 10, 11 and 12. In areas 1 and 2, rather short sandriges trending NE-SW occur. Area 1 is covered by sandridges which are 10 ÷ 15 km long, 1.2 ÷ 1.5 km wide and spaced at a 2.5 km interval. The southeastern flanks are steeper, whereas on the gentle northwestern flanks, short linear dunes representing secondary forms, may be recognized. Area 2 represents a somewhat different dune field. It is rather a sandsheet, representing the marginal form, partially covered by crescentic chains and linear dunes. The northwestern flank of this field is covered by densely - spaced crescentic chains (5 crests per km), while, towards the south, there are composite linear dunes in fan - like form trending NE-SW, and, on the eastern flank, the composite linear dunes in braided forms are oriented ENE-WSW.

Between the northern and southern subregions there is a narrow but distinct zone free from the dunes. The longitudinal ridges in area 9 have a NE-SW orientation; they range in length between 30 and 50 km; their width is 1.2 km and spacing varies between 4 and 5 km. The gentle northwestern flanks are covered by linear dunes and crescentic chains. Some of the dune crests are composed of a few linear dunes arranged in braided form. Between the ridges, there are very wide (up to 4 km) interdune corridors. Towards the east (area 10), the sandridges are shorter, and their length ranges from 4 to 10 km. The interdune areas form rather short closed features which contrast to the elongate interdune corridors of area 9. Sandridges of somewhat different form occur in area 11. They are irregularly spaced at 3 ÷ 6 km apart and their length varies between 25 and 40 km. Their crests are composed of a few linear dunes aligned parallel to the dune axes. On the gentle western flanks, occur short linear dunes, arranged obliquely to the dune axes. The interdune corridors are very wide (up to 5 km), and are without dune forms. Area 12 consists of straight and equally spaced sandridges, varying in length between 30 and 40 km and spaced 3 km apart. Linear dunes, arranged obliquely to the dune axes, are present on the western flanks of these ridges.

Area 13 is located between the northern and central region of Idhan Murzuq; here pyramidal sandridges have been developed. The dunes are spaced between 2 and 3 km apart and their bases range from 1 to 2 km. in width. In the eastern part of this field, the dunes are arranged in chains. The secondary forms occur only on the northern and western dune slopes. The superimposed linear dunes are disposed radially from the dune summit, thus creating forms which are transitional between domical and star dune forms.

The central region consists of two large dune fields: 14 and 16. Area 14 represents a reticulate pattern of sandridges and area 16 is covered by extremely long sandridges. In both areas, the main dune trend is NE-SW. In area 14, a grid pattern of reticulate sandridges is present. The main trend is presented by very long and regularly spaced sandridges which trend NE-SW, whereas a second trend which dissects the main trend, follows a SE-NW direction. The reticulate pattern contains closed interdune areas, which form elliptical areas in the northwestern part, and rectangular in the eastern part. The elliptical forms of the interdune areas have dimensions of 1.5x3.0 km to 2.5x5.0 km. In the northwestern part, in the large interdune areas (up to 5x15 km) the rocky basement of Murzuq basin is easily discerned on satellite images.

Area 16 presents a large field covered by longitudinal sandridges. These sandridges have a NE-SW trend in the northern part, which swings slightly in the lower latitudes. The sandridges have a very obvious pattern of parallel dunes extending over a distance of 135 km. The 55 km - wide dune belt exhibits 20 subparallel longitudinal sandridges. The ridges are up to 1.5 km wide; the spacing varies between 1.2 and 1.8 km in the western part of this field, but, towards the east it increases to 2.5 km. The ridges are composed of several linear dunes, thus representing the composite form. In the western part, the linear dunes which form the sandridges are arranged nearly parallel to the dune axis, whereas, towards the northwest, this pattern changes slightly into subsidary linear dunes which trend obliquely to the ridge axis. The principal dunes appear on the northwestern side, while the shorter linear dunes lie on the northwestern side. In the eastern part of area 16, the ridges are divided into segments some 5 to 20 km long.

In the western region of Idhan Murzuq, there is a large variety of dune fields, dune forms and dune orientation. Area 15 presents dune forms similar to those appearing in area 14, but the dunes have an abrupt change of orientation and have a different shape. The dune pattern is the reticulate, with the main ridges oriented E-W, and shorter ridges, N-S. The grid pattern is larger than that of area 14. The ridges of the main trend are spaced at distances of 3 to 5 km, whereas the shorter ranges, between 3 and 10 km. The ridges are 1.0 to 1.5 km wide and are covered by linear dunes. On the main ridges, there are 4 to 5 linear dunes superimposed on the ridge. Towards the west, the lateral ridges are only partially developed; hence they do not close the interdune areas. This pattern of sandridges represents a backbone - like shape, i.e. the main ridge has short but wide arms which diverge on both sides.

The southern part of the western region (areas 19, 20, 21, 22, 23 and 24), displays a large variety of dune forms and orientations. Area 19 is covered by large sandridges, oriented WNW-ESE and spaced at 4 km intervals. The gently sloping eastern flanks of dunes are covered by crescentic chains, arranged parallel to the dune axes. In area 22, large barchans with axis oriented NE-SW, are aligned in chains. In area 23, barchan - like sandridges having axes oriented in the opposite direction have been developed. The barchan ridges are up to 10 km long, and spaced at a distance of 3 ÷ 5 km. Longitudinal sandridges have been developed in areas 21 and 24, but possess a different trend. In area 21, the dune trend is NE-SW, whereas, in area 24, the dune is in a NNW-SSE direction. In both fields, the ridges are spaced ca 2.5 km apart. In area 20, the two trends of dunes are superimposed. In the northeastern part, the dunes trend in a SW direction, wheras, in the southwestern part, the dunes are arranged NNW-SSW, and, in the northwestern corner of this field, these two trends cross.

In the southeastern part (area 18) a huge sand sheet is present, and, on the eastern flank, a sarir development is probable. In the vicinity of area 16, only traces of old sandridges, oriented NE-SW, can be detected.

Ramlat Zuwaylah constitutes the northeastern extention of Idhan Murzuq, and may be regarded as a separate erg. Ramlat Zuwaylah can be classified as an elongated erg formed in a terrain depression. Generally, the erg represents a sand sheet partially covered by two kinds of dune: crescentic chains and linear dunes. Satellite images suggest that the northeasternmost part (area 7), represents a sand sheet without dune forms. The northeastern part of area 8 is covered by segmented linear dunes, rather densely spaced and oriented NNE-SSW. The southwestern part consists of crescentic chains with axes oriented NNW-SSW. In area 6, there are two belts of linear dunes, a northern and southern belt, both having an E-W trend which is oblique to the depression axis. Areas 3 and 5 are sand sheets which have some clusters of linear dunes. The elongated area 4 is covered by very densely spaced crescentic chains (6 crests/km) with axes oriented NNE-SSW. Towards the south, linear dunes are developed (area17) which have a general trend NE-SW.

In the middle part, Ramlat Zuwaylah is bounded to the south by a large sandridge barrier some 50 km long, which has gentle northern slopes covered by short oblique linear dunes. Similar sand barriers occur on the northern flank of the erg.

4.2.3. Hypothetical morphogenesis and morphochronology

The sands which form Idhan Murzuq have been accumulated in a rather very past time. Ballar (cited by Desio, 1971) stated that the white erg of Tajarhi, representing the eastern flank of Idhan Murzuq, was formed during „the second interpluvial". Ballar did not give any date for this period, but, in comparison with other dates from North Africa, this period seems to correspond with the peak of the Riss glaciation about 150,000 B.P., which was associated with an Atlantic Sea regression and aridity over North Africa.

The last interglacial „Sangamon-Eem", which is dated ca 125,000 B.P. was responsible for the Atlantic transgression and thus for the pluvial phase. During that period, the dunes were lowered and quilled and the dune soils were reddened. The last (Wisconsin) glaciation commenced ca 75,000 B.P and terminated ca 10,300 B.P. The Wisconsin major glaciation occured in at least three phases or stadials separated by interstadials. The early Wisconsin stadial can be dated at around 70,000 to 60,000 B.P. The middle Wisconsin stadial was located ca 50,000 to 40,000 B.P. and the late Wisconsin stadial began ca 25,000 B.P. (with maximum aridity around 18,000 B.P.).

Assuming, that the glacial stadial are correlatable with aridity, and interglacial stadial correspond with pluvial conditions, these changes should have an expression in sand sea development. If the assumption that the interpluvial which was responsible for the formation of erg Tajarhi corresponds with the Riss glaciation, the Idhan Murzuq sand sea was developed within the period span between 150,000 B.P. and the present.

Preliminary analysis of dune fields, relating to the different dune forms (Fig.48), clearly indicates that different zones have different dune trends. Excluding Ramlat Zawaylah, the main part of Idhan Murzuq can be divided into three regions with different dune orientation: a northen region comprising areas 1, 2, 9, 10, 11, 12 and 13; a central region constituting the major part of the sand sea which, incorporates the areas 4, 5, 14, 16, 17 and 18, and a south - western region which includes areas 15, 19, 20, 22, 23 and 24. The abrupt changes in dune form and orientation cannot be accounted for solely by changes in wind regime, hence, the Author hypothesises that *each region or subregion has a different sand origin and history of development*. The main *sources of sands* incorporated in the Idhan Murzuq are *autochthonous* and were derived from the surrounding mountains during pluvial phases. During the first stage of erg formation, huge amounts of sands accumulated in the basin. The first long arid phase created the main forms of dunes in the Idhan Murzuq - the long sandridges trending NE-SW. The subsequent pluvial phases degraded the preexisting forms, but with different energy and territorial extend. Probably less affected were the dune fields in the central part of the sand sea, and those most changed were dune fields in the areas located in the vicinity of the mountains surrounding the sand sea to the north, west and south - west. The limits of these regions probably also reflect the limits of pluvial activity in the consecutive phases of sand sea development. It is very probabile that during one of the arid phases, strong winds were blowing from a NW direction, thereby developing the reticulate dune network in the central - western part of the sand sea.

The dune pattern in the northern region indicates at least two phases of dune development. The southern subregion, consisting of areas 9, 10, 11 and 12, was developed earlier than that on the northern represented by dune fields 1 and 2. The dune - building sand was transported from the northern subregion to the south, and, following the pluvial phase, the new sand was deposited in this subregion, and the present forms of dunes were developed.

The pyramidal dune forms in area 13 should be attributed mainly to wind activity both in terms of erosion and deposition. The chain pattern of sandridges, can be related to the subdivision of pre - existing longitudinal sandridges into segments. Such segmentation probably took place during the early stage of sandridge development, within one of the pluvial phases. After that, the winds created the present dune forms. It seems that this part of the sand sea was influenced by mutidirectional wind flow, generated by topographical barriers, and a high frequency of wind variability.

The differences in the dune fields in the southwestern part of the sand sea depend mainly on the sand environment, terrain topography and wind regime. In respect of the Author hypothesis, there was *multitemporal deposition of sand, connected with partial destruction of the previous aeolian forms in this region.* It is very difficult to present a scenario of dune development in this part of Idhan Murzuq. During the consecutive pluvial phases, there was probably further sediment deposition and destruction of previous forms. The direction of sediment transport and the amount of transported sand were also governed by the local topography and ancient stream systems. It is highly probable that the pre - existing rivers blocked the development of dunes in the successive arid phase or phases.

The hypothetical morphogenesis of Erg Zuwaylah is as follows. The terrain depression, in which erg Zuwaylah has been developed, has an orientation similar to the depression controlled by erosion levels in the Precambrian basement. The *sand origin* can be both *autochthonous and allochthonous*. The dune forms and their density indicate that there is a rather thin cover of sand which is unsuitable for dune formation. The crescentic chains and linear dunes are both developed under the influence of the present wind regime. The change of linear dune orientation in areas 5 and 6 also indicates the change of wind direction, which, in area 5, has stronger northerly component.

An excellent example of the interrelationship of dune forms, terrain topography and sand environment is afforded by fields 4 and 17. The shape of those fields may be attributed to the pre - existing river valley, in which different type of sediments were deposited. The crescentic chains in area 4 are probably to be correlated with a more coarser sand fraction, which permits the development of such dune forms. As the result of wind action, the finer sand was transported to the southern part, where the densely - spaced linear dunes were developed. These dunes seem to be much younger than the sandridges in area 16.

4.3. THE GREAT SAND SEA

4.3.1. Location, topography and sand sources

The Great Sand Sea (Bahr al Ramal al Azim) is the northeastern most sand sea in Libya and constitutes the western part of the Libyan Desert appearing both on Libyan and Egyptian territory (Fig.49). The boundary between these two parts is defined by the meridian 25 E. The Libyan part of the Great Sand Sea, here referred to convenience as the Great Sand Sea, covers an area of 80 150 sq km (Linsenbarth, 1986c).

The Great Sand Sea can be divided arbitrarily into two parts: a northern and a southern, which can be distinguished both by geographical location and by a different geomorphologic character. The northern part occupies an area between 27°45'N and 29°45'N, and eastwards of 22°E to the Libyan - Egyptian border, which is defined by meridian 25°E. The southern part extends into the northeastern wing of Ramlat Rabyanah. This part of the sand sea forms a belt of the sand sea between 22°15'E and 23°15'E which is ca 100 km wide.

The northern flank of the Great Sand Sea is limited by the low escarpment of the Qardabah plateau which trends E-W. On the western flank, the sand sea is limited by the axis of a low topographical depression trending NNW-SSE. The southern boundary of the northern part is defined by the sarir area and by Jabal al Hawaish. The southern part is bounded on the western side by the sarir area of As Sarir and from the east side by Jabal al Qardabah.

The terrain surface rises gently from ca 60 m at 29°45'N up to 150 m at 28°N and 250 m at 25°30'N. The maximum width of the Libyan Great Sand sea in an E-W direction is ca 310 km, while the maximum length in a N-S direction is ca 470 km.

The preliminary analysis of dune trends and forms appearing in the Great Sand Sea indicated that there are two different regions governed by different factors responsible for the present geomorphology. The boundary between these two regions nearly coincides with the previous division of the sand sea into a northern and southern part. This boundary is delineated by the abrupt change of dune patterns which occurs between areas 9, 10 and 14 from the northern part and areas 12 and 13 from the southern part.

Both parts of the sand sea overlie the alluvial plain deposited during the pluvial phase or pluvial phases. The sand deposited as alluvium was mainly transported from the south and only to a small extent from the north.

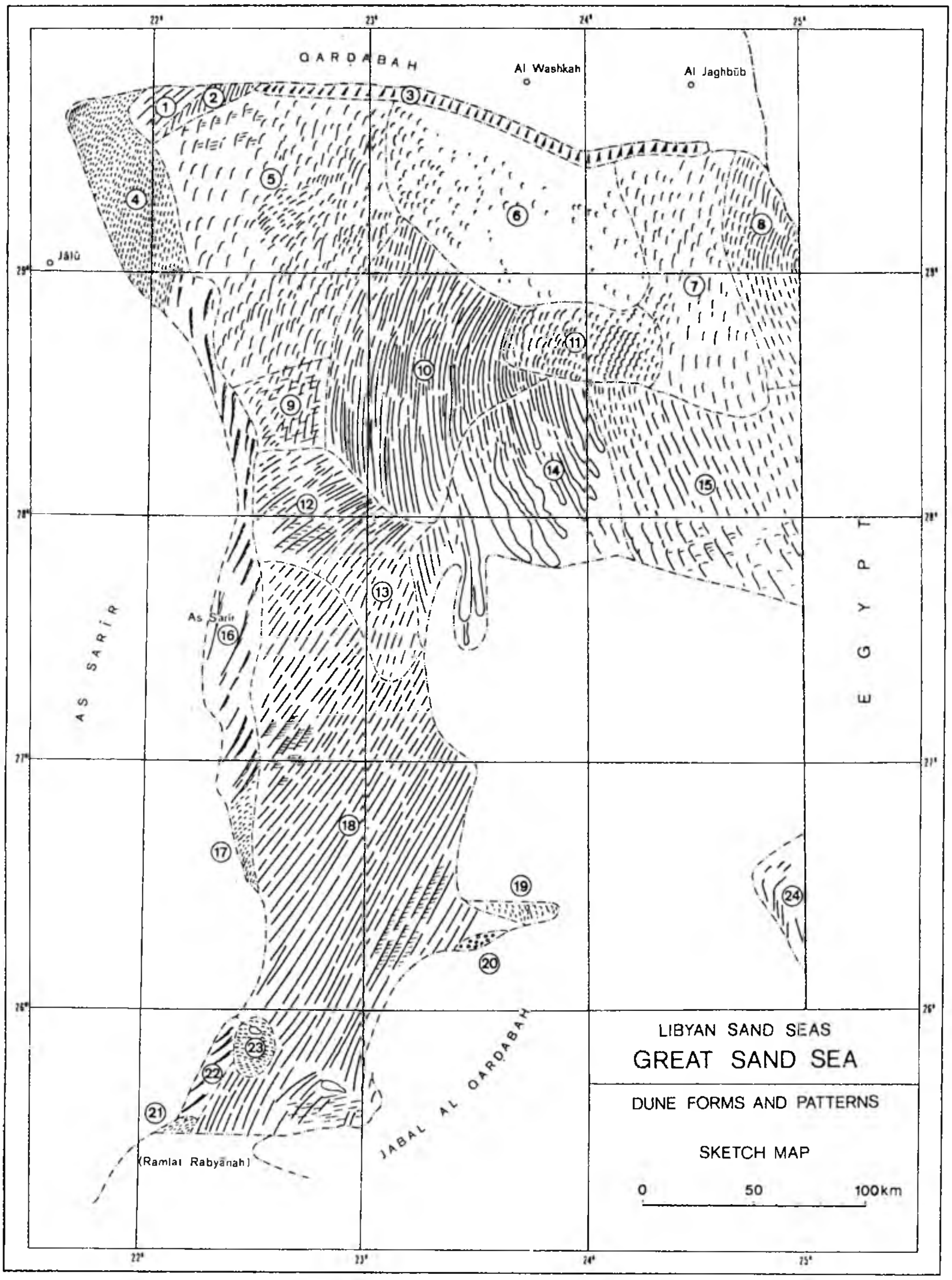


Fig.49 Dune fields in the Great Sand Sea
(In circles the numbers of the dune fields described in the text).

The present topography indicates that the main transport of sand was directed from the mountains to the south through the Rabyanah sand sea and through the present southern part of the Great Sand Sea further to the north, along the depression boundring the sand sea from the west.

The Great Sand Sea is the only sand sea in Libya which has well documented sand deposits underlying the dunes incorporated in the present erg (Wright and Edmunds, 1971). The documented thickness of sand deposits, varies between 150 and 500 m. The lowest deposits occur in the northern part of the sand sea, while the thickest deposits are located on the sarir area west to the sand sea.

4.3.2.Geomorphology of sand sea

4.3.2.1. The northern part

The northern and southern parts of the sand sea display quite different types of dune forms and trends. The commonest forms appearing in the northern part are hooked dunes, isolated megabarchans, crescentic chains and partially segmented seif dunes and sandridges, whereas the southern part is mostly occupied by longitudinal sandridges. There is a distinct change of dune trend between the northern and southern part: in the northern part the trend changes from N-E to NNW-SSE, whereas, in the southern part, there is only one predominant trend, NNE-SSW (Linsenbarth, 1986c).

The dune forms appearing in the sand sea may be divided into main forms and marginal forms boundring the erg. The main forms represent consecutive transitions of forms at the response to sand environment and wind regime. Area 5, is covered by hooked dunes resulting from the two main wind directions, NW and N. The longer arms (up to 8) are oriented N-S, while the shorter arms (up to 3 km) strike to E-W. The width of these dunes ranges between 0.4 and 1.2 km. In the northwestern corner of this field, the dunes are spaced 3 to 5 km apart, the spacing decreasing towards the southeast to 1.5 km. Within area 5, there is a small field with short crescentic chains which have a different orientation. This part strongly correlates with the topographical summit of this area, as marked by the 400 m contour line. It may be assumed that in this part of the area, stronger winds form more transverse forms. The hooked dunes appear again in the eastern part of the sand sea (areas: 7, 8 and 15). In area 7, the dunes are spaced at a distance between 2 and 4 km and their length varies between 3 and 10 km. In the southern part of area 7, the dunes change slightly, from an orientation of NS into NNW-SSE. In area 8, the hooked dunes are more densely spaced, their arms are shorter, and a more distinct change of dune trend into NNW-SSE is discernible. The longest hooked dunes (between 5 and 20 km) appear in area 15, but have rather short secondary arms here. They have a very distinct direction, NNW-SSE.

The central - northern part of the sand sea (area 6), is composed of barchan - like dune forms, which can be regarded as asymmetrical barchans with elongated south - western arms. These forms can be classified as transitional forms between pure barchans and hooked dunes. In the northwestern part of this area, dune length varies between 2 and 4 km, and spacing between 2 and 3 km. Towards the southwest, the length of dunes decreases and the dunes appear only very infrequently. These forms mostly reflect the sand environment, with a scarce availability of dune building sands. The dunes seem to be fixed onto a sandy basement consisting of a coarse grained sand supply, which forms the sand sheet. In contrast to the very mobile small barchans, the megabarchans are rather immobile, which is emphasised by a very low spectral response of the areas between the dune horns, which appear on satellite images in very dark tones.

South of area 6, the elliptical area 11 is covered by typical crescentic chains spaced at 1.5 km intervals. Comparison with the surrounding dune fields indicates that such forms can be created only as a response to a sand environment, because the wind regime cannot be changed on such a local scale. This field is probably covered by much coarser sand, i.e. that suitable for the formation of crescentic chains with large spacing between the chains.

The central part of the sand sea (area 10) is covered by low, slightly arcuate sandridges. They are formed by segments, 2÷4 km long in the northern part and 7 to 15 km long in the southern part of this field. In the southern part of this area, the dune segments form very long chains up to 60 km long. Dune width ranges from 0.8 to 1.0 km and the spacing from 1.2 km in the northern part to 2.6 km in the southern. The northern ends of these dunes have a slightly hooked form; thus, these dunes can be treated as transitional forms between the pure hooked dunes as appear in area 5, and longitudinal sandridges as appear in the southern part of the sand sea.

In the western part of the sand sea there is rather small area (9), depicting two trends of dunes. One the N-S trend, consists of hooked dunes, (which corresponds with the trend of hooked dunes in area 5), whereas the second trend (not so distinct and represented by linear dunes) corresponds with the dune trend in area 12. The *linear dunes are superimposed on hooked dunes* thereby indicating that the linear dunes are younger than the hooked dunes and are formed under the present day wind regime.

Area 14, being very conspicuous, is covered by large residual megadunes. The NW-SE striking megadunes are very long (up to 70 km); their width ranges between 1 and 4 km and spacing between 3 and 10 km. Their shapes are irregular, and on space images, they appear as rather low, old residual sandridges, degradated within a long period. Between the sandridges there are large interdune areas which expose the alluvial basement, which do not possess any smaller dune forms. Both the dune forms and the lack of smaller sand forms in the interdune areas prove the scarcity of aeolian sand in this area.

The sand sea is surrounded by several types of marginal dune form. The northern border of the sand sea constitutes a very long belt of linear dunes (area 3). These dunes have a fan - like shape and they were probably formed as alluvial fans, oriented and dipping to the south. The dunes present composite forms created from several linear dunes which range in length between 8 to 15 km. In the tail part, the width of dunes reaches 4 km. The western part of the northern boundary is composed of a small field of linear dunes (area 1) and a field of crescentic chains (area 2), with a wavelength between 0.6 and 1.0 km, and crest - to - crest spacing of 0.5 to 0.8 km.

The northwestern flank of the sand sea (area 4) comprises crescentic chains with an average wavelength of 1.5 to 2 km, spaced at 0.5 to 0.6 km apart. The dune axes strike towards the SE, and thus correspond with a resultant drift potential at Jalu (Fig.57). The rather closely spaced dune crests indicate that these dunes were formed in an aeolian sand environment with rather medium grained population, quite distinct from the coarser grain size population occurring in the similar fields in the eastern part (area 11). The very close crest - to - crest spacing does not agree with Lancaster (1985) observations that low wind velocities create such dune spacing.

The main part of the western boundary of the sand sea is marked by very distinct large linear and composite linear dunes which form conspicuous boundary forms (area 16). The composite linear dunes are very well depicted even on space photographs taken during Gemini missions. In the northern part of this field, the linear dunes are oriented N-S, and, towards south, the orientation changes to a more NNE-SSW direction. The dunes are composed of 2 or 3 linear dunes, the principal linear dunes appearing on the western flanks. The cross - section of the dunes is asymmetrical, with steeper flanks on the west and gentle slopes on the east. Crescentic chains occur on the gentle slopes of dunes. In the interdune areas, the giant zibars (zibars) are present. In the northern part, the length of linear dunes varies between 20 and 30 km and the dunes are spaced 5 to 10 km apart; they are arranged in a downwind staggered pattern. Towards the south, the dunes are longer (up to 40 km) and have widths of 0.6 to 1.0 km. The tail parts of dunes present in the north have a fan - like form. The dune heights vary between 30 and 40 m.

4.3.2.2. The southern part

The southern part of the Great Sand Sea contains completely different forms and trends. At the northern boundary of the southern part, (areas 12 and 13) there is an abrupt change of dune trends. These two fields together with the main field (18), have dunes which trend NNE-SSW. Area 12 is covered by longitudinal sandridges which trend NE-SW. The dune lengths vary between 7 and 15 km and spacing between them is 2 to 4 km. The dunes have a width of 0.6 to 2.0 km and their heights are up to 35 m over the interdune surface. Dune ridges are composed of one or two linear crests. On the east - southern slopes, short linear dunes appear with E-W orientation. Towards the southeast, these dunes grade into the field of arcuate sandridges (area 13) composed of segments 6 to 8 km long. These dunes represent braided forms created from several linear dunes which, in their northern ends, form wide tails. The dune spacing ranges between 2.0 and 2.5 km. The interdune areas are covered by giant zibars with axes oriented towards SSW.

The largest dune field (area 18) is composed of straight sandridges, divided into segments 5 to 15 km long and spaced at 1.5 to 3.0 km apart. The dunes have an asymmetrical cross - section, with comparatively gentle slopes on the eastern side. The dune crests are composed of several linear dunes forming a braided pattern. Below 27°N, the dunes are longer (up to 65 km) and are divided into longer segments, between 12 and 27 km. Also, the dune spacing is larger (3 to 4 km). In the northern part of this field, the width of dunes ranges between 0.6 and 1.2 km and in the southern part between 0.8 and 1.2 km. The interdune areas are covered by zibars.

4.3.3. Hypothetical morphogenesis and morphochronology

In the Author's opinion, the dune trends and forms which appear in the Great Sand Sea are strongly related to the sand environment and wind regime. Generally, the dune trends in the northern part correspond with the wind regime which can be reconstructed on the basis of sand roses of Jalu and Al Jaghbub. The dune trends in the southern part correspond with the wind rose in Al Kufrah. The dune forms also reflect very strongly the influence of sand environment in both areas. The dune forms in the northern part can be used as evidence of a rather limited availability of dune - building sand. It seems that, apart from the huge amount of sand incorporated in the underlaying alluvium, only the uppermost part consists of loose dune - building sand. The underlying alluvial sands are not mobile because the upper layer of consolidated sand is resistant to sand movement by wind. The large spacing of the dunes in areas 5 and 6 of the Great Sand Sea (Fig.49), where hooked dunes and megabarchans occur, confirms this assumption. The scarity of aeolian sand in the north - western part, may be interpreted as the result of a very long period of wind activity, which depleted this area of dune - building sand. As the results of the selective action of wind the sand particles moved by the saltation process were transported into more southerly and southeasterly areas . The sand transported from the northern part provided a sand supply to the Egyptian part of the Great Sand Sea and to the southern part of the Libyan Great Sand Sea, and, further to the south, to Ramlat Rabyanah.

The location of the boundary between an open desert and sand sea area is not very well defined. In the Author's opinion, the most important factor governing this boundary line is the sand environment and, to a lesser extent the wind regime. In the case of the Great Sand Sea, the areas on the west constitute a large sarir area, called As Sarir. As Sarir also overlies the large alluvial plain,

but the uppermost part constitutes a gravel surface. Such a surface was developed during a very long period as the result of the wind regime which reflects the location of a subtropical high - pressure cell over this part of Libya. On the sandflow map presented by Wilson (1971), the sand flow peak is located over the area of As Sarir (28°N and 18°E). Due to the clockwise outward flow around the high pressure cell, the winds have moved the aeolian sand from this area, causing a depletion of dune - building sand in this area and developing a flat and vast sarir surface. Thus, as the result of this mechanism, the sarir area does not represent the sand supply source. If there is any sand supply, it is caused by stronger winds which carry the coarser sand grains mainly in creep mode. Hence, the composite dunes can be built by coarse sand, transported from the sarir area during strong winds, and from the sand transported by more gentle winds blowing from the N and NE, i.e. from the interior of the sand sea.
The question is still unresolved as to, why the line delineating the boundary between the open desert and sand sea has defined its position. In the Aouthor's opinion there are two other factors which should be taken under consideration: *a change in the terrain slope gradient* and *the geological factors influencing the sand environment*. In this part of the desert, there is a low topographical depression trending NNW-SSE, the axis of which is parallel with the sand sea western boundary between 29°30'N and 28°N (then passing through areas 12 and 13). The axis of the depression very precisely corresponds with the large geological fault depicted on the tectonic map of Libya (National Atlas, 1978 p. 37). These two factors seem to be responsible for the sand environment changes. The sand environment changes can appear both during the deposition of sand in the geological past and during the present development of the desert area. It is possible that the deposition of pluvial sand, in this topographical low occured in a pluvial phase later than in the surrounding areas, hence the sand may have different characteristics. Futhermore, the later deposition of sand could destroy the preexisting aeolian forms, thereby creating a new sand surface, which was prepared for the development of new forms during the preceding arid phase. This hypothesis is supported by the different dune trends in areas 12 and 13, which form the boundary between the northern and southern parts of the Great Sand Sea. Of course, a change of wind regime in the consecutive phase should also be assumed. It is very surprising that the two transitional areas (12 and 13), strongly correspond with the erosion levels of the Precambrian basement (National Atlas, 1978 p. 47)

Another boundary between the sand sea and open desert appears to the south of areas 14 and 15, which are adjacent to the sarir area. Analysis of satellite images indicates that there is no change of terrain slope, neither were any geological faults recognised. This sarir area is strongly depicted by a different albedo and, it can be seen on colour space photographs, by a much darker brown colour. These changes can be caused by the sand environment. On the lithofacies map of post - Eocene formations, these areas are defined as

„unconsolidated sands with some shales and mudstones" (Wright and Edmunds, 1971, p.468). It seems that this sarir area has such a surface, and that the wind - blown sand is only in transition over this area. This can be related to the stronger saltating process over a gravel surface and also to the fact that the sand flow is undersaturated, both to the limited amount of sand from the wind direction and to the acceleration of the wind over the sarir area. Sarir areas are never areas of sand deposition; on the contrast, they areas of deflation and sand transfer.

As a final hypothesis relating to the dune forms in the northern part of the Great Sand Sea it is considered, that this part of the sand sea is in the *a stage of depletion* and the sand forms which appear result from both the very limited amount of dune - building sand available and from the two - directional wind regime (or a wide unidirectional wind regime). The local changes of sand environment are responsible for particular dune formations. Generally, it can be stated that the *northern part of the Libyan Great Sand Sea is in a regressive stage and that this erg can be classified as an erg with negative sand budget.* The active parts of the sand sea are marked by smaller dune forms which appear in the interdune areas.

In the southern part, the braided forms of dune crests and the interdune areas covered by zibars or crescentic chains confirm the active stage of sand sea development and an unlimited amount of dune building sand. In comparison with the northern part, it seems, that sand incorporated in the erg was deposited at a later stage or in the phase of pluvial deposition of the alluvial plain. The upper part, which consists of active sand taking part in the dune building process, is much thicker than is the case in the northern region of the erg. There is a rough correlation between the shape of the main dune fields and the erosion levels of the Precambrian basement. This area is separated on the lithofacies map of Post - Eocene formation, and defined as „the formation composed of an upper and lower series of unconsolidated sands with an intervening horizon of shales and mudstones" (Wright and Edmunds, 1971).

In the southernmost part, there are topographical barriers infuencing the dune pattern. Area 23 is covered by crescentic chains which are oriented towards the south. This area correlates with the highest level of the Precambrian basement, which is covered by a thin sand layer, on which crescentic chains were developed.

The dune forms were tailored by persistant winds blowing mostly from the NE sector. The dune trend corresponds precisely with the resultant wind direction calculated for Al Kufrah Fig.57). The resultant wind potential is also very high (RDP = 142). Thus, both the resultant wind direction and resultant wind potential can be held responsible for the development of the longitudinal sandridges in the southern part of the Great Sand Sea. The general dune trend is the same as that in the Egyptian part of the Great Sand Sea.

The marginal forms occur mainly on the western border of the sand sea from the As Sarir side. Longitudinal sandridges (areas 16 and 22) and crescentic chains (areas 17 and 21) are both present. The hypothetical morphogenesis of these forms is the same as for that in the northern part, adjacent to the As Sarir area. On the eastern side, there is a small field with crescentic chains (area 19) and small field of star dunes (area 20). The field of star dunes can mainly be related to the topographical barrier in south and east, and the changing and accelerating wind regime which thereby created conditions for the development of star dunes.

The morphochronology of the Great Sand Sea can be reconstructed on the basis of investigations performed by Benfield and Wright (1980), Di Cesare *et* al. (1963) and Desio (1971).

The Great Sand Sea constitutes a large alluvial prism which deepens towards the north. In accordance with geomorphologic development, the sand deposited in this alluvium plain has a pluvial origin and was transported in the past from the south and to a lesser extend from the north. Geological investigations (Benfield and Wright, 1980) indicate that the oldest formation base is covered by the sediments of Oligocene (up to 200 m); Lower and Middle Miocene, called Marada Formation (400 m), and Post - Middle Miocene, called the Calanshio Formation (up to 200 m), all comprising unfossiliferous sands, fine to coarse grained sands, pebbly in part, which were developed under fluvial conditions. The uppermost layer consists of Holocene/Pliocene sand, gravels and calcretes up to 30 m thick. The deepest post - Eocene strata (ca 1000 m), is present in a topographical depression trending NNW-SSE.

Benfield and Wright (1980) concluded that the Calanshio Formation was deposited in a continental alluvial plain environment in which fluviatile channel sand sedimentation was dominant. The Calanshio formation type of deposition may well have continued through the Pliocene and into Quaternary times. Quaternary sediments undoubtedly occur in the post - Middle Miocene time, but, owing to a lack of data no proper distinction can be made between the Quarternary and the older formations. The only well - documented sequence, in the eastern Sirt basin, in the Quarat Uedda, was provided by Di Cesare *et al.* (1963). This formation comprises four horizons of lacustrine sediments separated by three horizons of aeolian sand. The entire thickness of these deposits varies between 11 and 40 m. The sequence increases in thickness to the south and it shows a transitional passage southwards to more fluviatile sedimentation within every lacustrine member. The aeolian layers represent 4 or 5 pluvials and 3 to 5 interpluvials or arid phases with dating possibly ranging down into the Pliocene. Desio (1971) suggests that pluvials could correspond with the four classic glacial phases of the Pleistocene, but this point of view is not accepted by other scientists (Ragnon and Williams, 1977; Street and Grove, 1976; Nicolson and Flohn, 1980). The data obtained by DiCesare *et al.* (1963) suggest the following chronology of the Great Sand Sea formation:

a) during the first pluvial (Upper Miocene and Pliocene - the fluviatile sediments of Calanshio Formation) there was a deposition of gravel and sand (paleosarir) transported by rivers from the south;
b) during the first interpluvial or dry period of uncertain age, the erosion of the older sediments and a development of the first ancient erg may be assumed;
c) during a second pluvial phase, rivers flowing from the south eroded the former sediments and deposited new ones. Within this period a total levelling of the erg may be assumed;
d) in a second interpluvial, the reworking of deposited sands and development of a second erg may be assumed;
e) a third pluvial is characterized by caliche development with brown chert;
f) within a third interpluvial phase, the erosion of former sediments and development of a third ancient erg of quartz sand occured. This interpluvial phase probably correlates with an interstadial phase between 45,000 and 35,000 BP which is characterized by a warm and dry climate (Mc Burney, 1968);
g) a fourth pluvial was characterized by deposition of caliche with brown chert and corresponds with ground water recharge between 35,000 and 14,000 B.P. in the Eastern Sirt Basin (Benefield and Wright,1980). This pluvial period agrees with a number of age determination (30,000 and 25,000 B.P.) of the groundwater in the „Nubian" sandstone in the western desert in Egypt;
h) a fourth interpluvial, is characterized by a reworking of previous sands and corresponds with the period of groundwater recharge in the Eastern Sirt Basin, dated at 14,000 and 8,000 B.P. (Benfield and Wright, 1980);
i) a fifth pluvial, corresponding with groundwater recharge in the Eastern Sirt Basin, in the period between 8,000 and 5,000 B.P., was marked by erosion of the caliche and refilling of the streams.

The current arid period, which commenced around 5,000 ÷ 4,000 B.P., is associated with the development of the sand dunes of the Great Sand Sea and the deflationary sarir plains on the west. The dunes overlie a dissected surface which is contemporaneous with the fourth pluvial period (8,000 ÷ 5,000 B.P.). It is, however, also possible that the main dunes were formed during the very intensive arid phase documented in other parts of Libya in the period between 14,000 and 8,000 B.P.

The sands incorporated in the present sand dunes were derived from parent material which derives from the older sand formations underlying the modern erg.

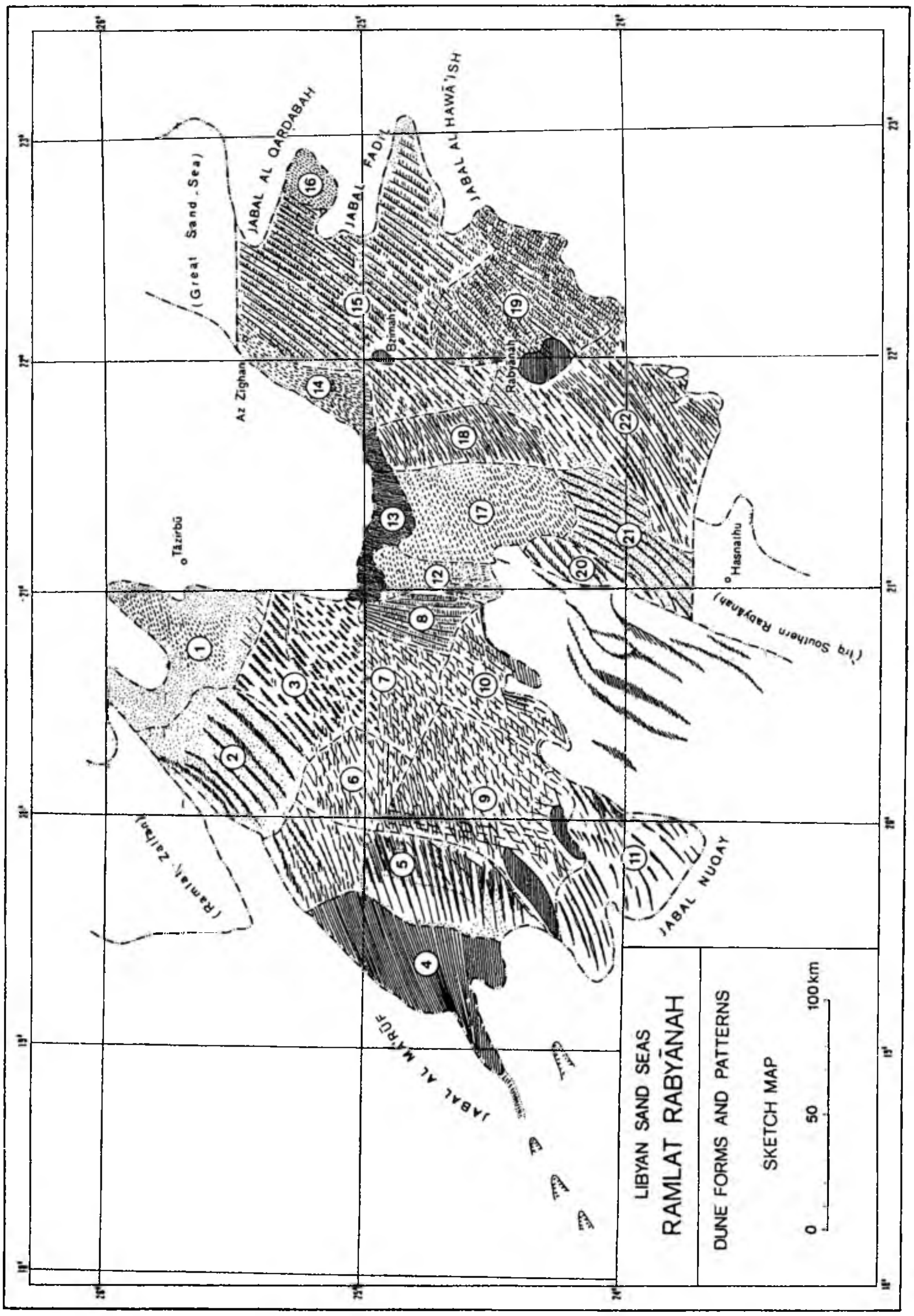


Fig. 50 Dune fields in Ramlat Rabyanah
(In circles the numbers of the dune fields described in the text).

4.4. RAMLAT RABYANAH

4.4.1. Location, topography and geomorphology

Ramlat Rabyanah represents the southern extention of Ramlat Zaltan from the west and the Great Sand Sea from the east (Fig.50). A topographic depresion running along meridian 21°30'E divides the sand sea into two parts: western and eastern. The western part extends over 26,100 sq km and the eastern, 24,620 sq km. Each part contains different dune forms and patterns, resulting from differing morphogenesis (Linsenbarth, 1986d).

4.4.1.1. The eastern part

The eastern part can be divided into three regions: a northeastern region, which consists of areas 14, 15, 16, 19 and 22; western region which is formed by areas 12, 13, 17 and 18 and a southwestern region, forms areas 20 and 21. Each of these has specific dune forms, patterns and trends.

The northeastern region constitutes a southern extention of the Great Sand Sea, and the main trend of sandridges is similar to that of dunes there. Thus, the dunes continue their NE-SW trend, slightly changing orientation in the southern part of this region towards ENE-WSW. Area 15 is covered by longitudinal sandridges, with well pronounced linear crests. The dunes are asymmetrical, with gentle slopes on their eastern sides. The segments of sandridges are 10÷25 km long, 0.6÷0.8 km wide and spaced 1.2÷3.0 km apart. The interdune areas are covered by short linear dunes trending in an EW direction. The dunes in area 19 present similar forms of sandridge, but their crest is not so readily detected on satellite images and the interdune areas can not be distinguished. Towards the southwest of Rabyanah (area 22), the sandridges again show sharp crests formed by single or multiple linear dunes. At their northeastern tailed ends, the dunes form fan - like shapes. The dunes are usually 10 to 35 km long, 0.6 to 0.8 km wide and are spaced at 1.2 to 1.5 km intervals. Area 18 consists of very dense sand - ridges composed of multiple parallel linear dunes. Dune spacing ranges from 1.0 to 1.5 km.

The marginal forms appear only in the central - northern part (areas 13 and 14) and in the northeastern flank of this region (area 16). The eastern border is marked by mountains without any characteristic marginal forms. Area 13 represents very densely - spaced linear dunes (4 dunes per kilometre), whereas areas 14 and 16 are covered by crescentic chains.

The sandridges in the northern region seem to have had a similar morphogenesis but with small differences, relating to the sand origin and supply. The sand environment is created by the alluvial sand underlying the active upper part of erg. There are no exposures of rocky basement in this region. The sand deposited in the alluvium represents *allachthonous fluvial sediment*, reworked during aeolian processes into *autochthonous aeolian sand.* Of course, besides the autochtonous aeolian sand, the sand transported by aeolian processes from northern parts of the Great Sand Sea also has to be accounted for. The differences in the sandridges in area 19, can be related to the *multiple and multitemporal sand sources*. It is possible that, in addition to the larger amount of the fluvial sand as transported from the south in the same pluvial period or in another, there was additional input of sand from Jabal Al Hawaish. The sandridges seem to be at another stage of development when compared with those in areas 15 and 22. The orientation of linear dunes, present in the interdune areas, indicates a distinct influence of winds from E, which do not correlate with the sand rose of Tazirbu (Fig.64).

The western region contains completely different dune forms. Area 13 represents a field of very densely spaced crescentic chains, with axes oriented W-E. The crescent modules in crescentic chains are very open; thus, these forms are similar to the linear dunes which form rather transitional forms between crescentic chains and linear dunes. In the northern part, area 17 is covered by barchans arranged in chains, whereas towards towards the south typical crescentic chains occur. In the northern part of this field, the dune wavelength varies between 0.8 and 1.5 km, and the dune density is 2 crests per kilometre. Towards the south, these parameters decrease, the wavelength ranging from 0.8 to 1.0 km and the density to 3 crests per kilomtre. In area 12, the crescentic chains have parameters similar to those of area 17, but, in the southern part, the dune pattern is more chaotic and, additionally, several subdued sandridges trending N-S may be recognised.

Area 21 is covered by rather low sandridges representing a different form of dune. The dunes, trending NNE-SSW, are different from the dunes in adjacent field 22, where they follow a NE-SW direction. The sandridges do not have crest lines and appeared to be the old subdued and rounded sandridges affected by present winds, which is revealed by linear dunes developed on their ridges and arranged obliquely to the dunes axes. The length of the dunes varies between 10 and 40 km, the width ranges from 1.0 to 1.5 m, and the dunes are spaced 2 to 3 km apart. The interdune areas are easily defined and covered by short linear dunes (2 to 4 km long), trending NNE-SSW direction.

The sandridges in area 20, have a different orientation and display forms quite different to those in area 21. They are typical composite linear dunes with the principal dune (sandridge) on the NW side and with minor dunes on the SE side. The widely - spaced dunes are oriented NNE-SSW and their length varies between 20 and 40 km. In the tail part, the dunes form a fan - like shape with widths up to 2 km.

The completely different dune forms present in the western and south western regions can be attributed mainly to differences in the sand environment. The sand environment may be assumed to be diffferent for eiother of two reasons. The first is that the underlying alluvial sand was probably transported in the pluvial periods such that, due to the close location to mountains, the sediment incorporated in alluvium has different granulometric characteristics (which depend on the length of the transport medium). The second, is that the aeolian sand incorporated in the areas 12, 13, 17 and 18, was transported mainly from the sarir area of As Sarir, and to a lesser extent from the Great Sand Sea. It may be assumed that this sand represents a much coarser population of grain sizes. In such an environment, the winds from the NNW direction are capable of forming the barchan chains and crescentic chains which are present in areas 12 and 17. The gradual changes of barchan chains into more - densely spaced crescentic chains confirms the hypothesis that, in the coarser sand, the barchan or transverse forms are larger and more widely spaced, whereas, and in the areas in which medium sand dominates, the forms are more densely - spaced and the dune crests are better developed (Lancaster, 1985). As the result of the segregation process during sand transport, the coarser grains are deposited behind the finer grains, which are transported to the downwind region.

The dune forms and trends evident in areas 20 and 21 appear to confirm the earlier - proposed assumption, that the sand in this region was deposited in another pluvial phase. The dune trend is completely different to that seen in area 22 which must be attributed not only to the changes in the wind regime. The abrupt change of dune patterns between areas 21 and 22, can be related to a different sediment environment, wheras the dune trend to a different wind regime in the period of sandridge development. Area 21 is covered by rather low sandridges comprsing different dune forms and it is possible that winds of rather small energy are reworking the upper parts of these sandridges. The different trends and forms of sandridges in area 20 can be attributed to local changes in the sand environment and to changes in the wind regime due to the vicinity of the mountains and escarpments adjoining this field to the west. The composite linear dunes seem to be very similar to the marginal forms of dunes which are present on the western border of the Great Sand Sea (areas 16 and 22). The stronger winds from the northwestern sector contributed to the development of the principial forms, while the present, rather light winds, are responsible for the minor dune forms. Analysis of dune patterns and trends in areas 20, 21 and 22, and their interrelationship, provides a basis for the hypothesis that the sands in area 21 were deposited later than the sands in areas 20 and 22. This hypothesis is supported by the similarity of the trend of sandridges in areas 20 and 22, which were developed in the same period of dune building, i.e. during the arid phase. There was probably a similar dune pattern in area 21, which was later destroyed during the next pluvial phase. The sand transported during this pluvial phase affected only the narrower zone corresponding with the present fields 21, 17 and 13. During the subsequent arid phase, the sandridges in area 21 were developed.

4.4.1.2. The western part

The western part of Ramlat Rabyanah displas a wide range of dune forms and trends. It is rather difficult to divide this part of the sand sea into homogeneous regions or subregions. Preliminary analysis of these forms and patterns leads the Author to propose a division into two regions: a central region consisting of areas 3, 6, 7, 8, 9 and 10; western region covering areas 4 and 5, and three subregions referring to the particular dune fields appearing in areas 1, 2 and 11.

The central region is characterised by sandridges which have quite different stages of development and which have the inprinted traces of different trends of dunes. The general trend of the principal sandridges is NE-SW. In area 3, the NE-SW - oriented sandridges are divided into 5 to 20 km - long segments. The width of ridges varies between 0.4 and 0.6 km and spacing ranges from 2 to 3 km. In the interdune areas, short linear dunes extending N-S are present. In areas 6 and 7, the sandridges are shorter with lengths ranging from 3 to 10 km, widths varying between 1.0 and 1.4 km and spacing between 1.5 and 2.5 km. In area 6 and in the southern part of area 7, linear dunes striking equatorially occur in addition to the principal ridges. In area 8, the sandridges are longer, up to 35 km, and their trend has changed to a NS direction. The segmented sandridges, 8 to 12 km long, 1 to 2 km wide and spaced 1.5 to 2.0 km apart, appear in area 10. The interdune areas are covered by short linear dunes trending E-W.

In area 9, this grid pattern is more distinct. Besides the main NE-SW trend of segmented sandridges, there is a second trend (E-W) of linear dunes crossing not only the interdune areas, but also the sandridges. On the western flank of this field, a few subdued sandridges trending NNE-SSW can be discerned on the Landsat Photomap of Libya

In the dune morphology, the general inprinted NE-SW trend of sandridges sugests that these forms were developed during this same arid phase. The short linear dunes, which appear in the interdune areas in field 3, may be related to the present winds from the north. Towards the south, this trend gradually changes to the E-W direction, thus reflecting the influence of easterly winds. Such change in wind direction can result from a general change of wind pattern, attributable to the influence of the mountains located to the south. A third trend inprinted on the western margin of area 9 seems to represent the remnants of the oldest, subdued sandridges, developed in an earlier arid phase. By contrarst, the dune trend in area 8, which differs from the general NE-SW dune trend in this region, may be related to a mutitemporal deposition of sand in the later pluvial phase, during which the earlier dunes were destroyed and new forms developed in the following arid phase.

In areas 4 and 5, constituting the western region, there is only a single significant ENE-WSW dune trend. In area 5, the linear dunes are arranged in composite forms. These types of dune composition can be named *bunch - type dunes*, which have no principal dune, but are developd from a bundle of linear dunes which merge at one point. The length of dunes varies between 15 and 35 km and their spacing between 2 and 3 km. The tails of the adjoining dunes are merged at their eastern ends. Area 4 represents very closely - spaced linear dunes, also extending also NEE-SWW. The dunes are 10 to 40 km long, 0.2 km wide and are spaced 0.4 to 0.5 km apart.

The composite linear dunes in area 5 and linear dunes in area 4, represent the contemporary results of dune formation under the present wind regime. Both fields are located on the upwind flank of the sand sea and hence, they are developed in pure aeolian sand deposits, without underlying alluvial sediments. The sand incorporated in these dunes represents a medium - size population which is the most convenient for sand transport in the saltation mode.

Outside Ramlat Rabyanah, between 18°E and 19°E, and 24°and 25°N, in a southwest direction from area 4, there are very characteristic compound megabarchans, on which smaller barchans are superimposed. They are indicative of wind direction and sand migration. Similar forms were recognised by Mainquet (1984a) in Chad on the southern margin of the Tibisti massif.

The northern subregions represented by areas 1 and 2, can be regarded as marginal dune fields. Area 1 is covered by crescentic chains. The crescentic chains have a wavelength of between 0.5 and 1.0 km and a density of 4 crests per kilometre. This field of crescentic chains adjoins the sarir area, from which the remaining sand which can be transported by wind, is shifted to this field. Hence, the crescentic chains are developed in a sand environment characterized by a rather coarse population, which permits the formation of zibar - like forms under the influence of strong winds from a N-W direction.

Area 2 presents one of the most conspicuous fields of dunes in Libya. This field can be classified as a very thin sand sheet on which composite megadunes have developed. The megadunes form *spear - headed figures* stretching NE-SW. Over a 35 km - wide area, there are 7 megadunes varying in length from 50 to 90 km. They average 2 to 4 km in width and are spaced 3.5 ÷ 8.0 km apart. The megadunes are composed of oblique linear dunes spaced 0.3 to 0.4 km apart and varying in length from 5 to 7 km. Some of these megadunes have very distinct crests which are excellently reflected from SE side on satellite images. The interdune areas are without any small dune forms. The morphogenesis of such forms is unknown. Each megadune can be regarded as a separate dune field, bacause, between the megadunes, there are large sand - free areas showing the rocky basement with a thin sand cover. The shape of the megadunes can be attributed to terrain depressions in which aeolian sand was deposited.

The southern part, here called a subregion and restricted to area 11, represents a dune field unique to the Libyan sand seas. This field is covered by chains of gigantic arc - shaped barchans which trend ENE-WSW in the northern part and NE-SW in the southern part. The length of the dune chains varies between 5 and 40 km and their spacing between 3 and 6 km. In the northern part, the dunes are 0.4 to 1.0 km wide, whereas in the southern part their width is up to 2 km. The ridges represent braided forms composed of several linear dunes. These forms were probably developed as a result of a more complex wind regime, generated in the vicinity of mountains. The different forms in the northern and southern parts of this field can be related to the channelling of winds between the topographical barriers.

4.4.2. Hypothetical morphogenesis

Analyis basis on the interpretation of satellite images indicates that each part of Ramlat Rabyanah has had a different genesis and chronology of sand sea development. In the eastern part, very distinct forms were developed, mainly the effect of a sand environment. The eastern part has a thicker cover of both underlying alluvial sand and of active aeolian sand. The larger amount of aeolian sand can be related to sand migrating from the Great Sand Sea. This sand is presently involved in the reconstruction of previous sandridges; hence, these sandridges may be be classified as *active sandridges under the process of current development*. By contrast, the sandridges occuring in the western part seem to be subdued, and the main dune fields can be treated as *dune forms in the process of redeposition or redevelopment*. The linear dunes, indicating present day wind forces, cross the old dune trends. In comparison with the Great Sand Sea there is probably only a limited aeolian sand supply. It seems that sand which is involved in the transportation process from the Great Sand Sea, is trapped by the sandridges in the eastern part; hence, only a very limited amount of sand transported from this direction is deposited in the western part of Ramlat Rabyanah.

The western part of the sand sea may be taken as an *example of sequence of dune deveopment*. On the most windward side, on the boundary with the sarir area, there are crescentic chains. Later, there are sandridges which are developed in the thicker part of the sand cover. Then follow the linear sandridges, which pass into very densely - spaced linear dunes. Further to the southwest, which is on the furthest downwind direction, there are barchan chains and compound barchans, which are generally associated with a limited amount of sand supply, and which indicate the travelling forms which are present on the sarir or sand sheet surfaces.

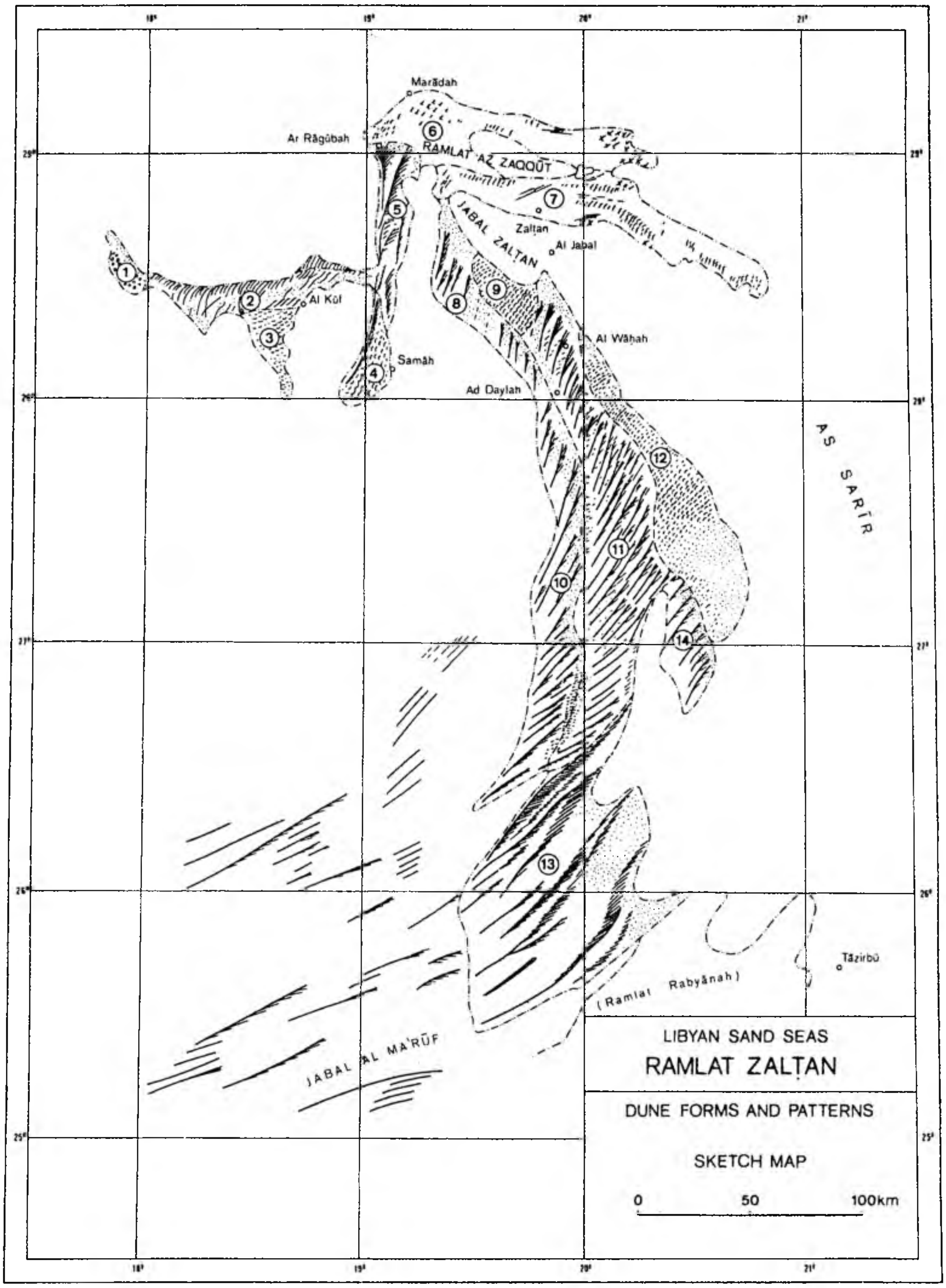


Fig.51 Dune fields in Ramlat Zaltan
(In circles the numbers of the dune fields described in the text)

4.5. RAMLAT ZALTAN

4.5.1. Location and topography

The sand sea, named in this investigation as Ramlat Zaltan (Fig.51), can be conveniently divided into three separate geographical parts: a the northern part, constituting Ramlat Az Zaqqut, a western - extending in a westerly direction from Al Kuf, and the main part - extending form Ramlat Az Zaqqut towards the south as far as the Ramlat Rabyanah (Linsenbarth, 1986d). The total area of these three parts amounts to 28 ÷ 410 sq km.

The dune morphology in the northern part, called Ramlat Zaqqut, was investigated by Linsenbarth (1987a) who used Metric Camera photographs. In 1994, an additional investigation was performed on the basis of MOMS-02 data, which are characterised by very high resolution and stereoscopic capability (Linsenbarth, 1995). These space photographs were very usefull for the detection of the smaller aeolian forms, such as linear dunes and crescentic chains, which are present in this area. There are also very prominent star dunes. In the north, this area is bounded by the escarpment of Jabal al Maqayl and on the southern side by Jabal Zaltan. In this area there are many residual hills of various height, the highest of them reaching the same elevation as Jabal Zaltan. These flat - topped rocky hills represent the local topography and create topographical barriers for dunes develoment. The main part, elongated in a NWW-SEE direction, falls gradually from 160 m on the western flank up to 120 m on the eastern. This elongated part represents a topographical depression. A number of wadis terminate in this depression.

4.5.2. Geomorphology and morphogenesis

4.5.2.1. Northern part - Ramlat Zaqqut

Area 6, which constitutes the northern part of the erg, is covered mainly by star dunes, crescentic chains and linear dunes. It is rather difficult to divide this area into separate fields of particular forms because, due to the terrain topography, these forms grade from one to another or coalesce. In area 7, the main dune forms are represented by the linear dunes, crescentic chains and the star dunes. The morphometric parameters of these forms, together with as the hypothetical genesis of their development, were given in chapter 6. Thus, in this chapter analysis will be restricted to the determination of the interrelation between these forms in the erg.

The sand incorporated in the dunes and sand sheets appearing in this erg are autochthonous and derived from the surrounding mountains. The local sand is transported and reworked by aeolian processes under the influence of a wind regime adjusted the local topography. The wind regime in this area can be reconstructed from the sand roses of Hun and Jalu (Fig.59 and 60). The drift potential for both stations is rather high (362 VU and 414 VU respectively), but the resultant drift potential is rather low, which indicates a complex wind regime. Such a wind regime creates excellent conditions for the development of reverse and star dunes. As well as the wind regime, terrain topography is the second main factor governing the location of star dunes. The latter are located on the higher parts of plateaux or in the vicinity of escarpments. The star dunes in this area belong to the highest sand dunes which reaching to 180 m. The linear dunes are located in the lowest parts of the depressions, while the crescentic chains occur on the downwind flanks of the erg. The general trend of linear dunes corresponds with the axis of depression and with the resultant drift direction.

Analysis of dune forms appearing in Ramlat Zaqqut leads to the conclusion that these forms are developed under the present dune regime and that this *erg can be treated as a modern active erg.*

4.5.2.2. The western part

The western part trending which trends in a W-E over a distance of 80 km occupies the area below 28°30'N. The sand desert area is bounded on both northern and southern sides by a rocky escarpment, so it is obvious that the sand field area is limited by the terrain topography. The easternmost flank is formed by star dunes in the pattern of a few short chains trending NE-SW (area 1). The average diameter of the domical dunes is 1 km, and the spacing ranges from 1.2 to 2.0 km. Area 2 is covered mainly by linear dunes which are oriented NE-SW and developed as braided forms. The length of the dunes varies between 3 and 15 km, the width averages 0.4 km; spacing varies between 1.2 and 1.5 km. Areas 3 and 4 represent crescentic chains. The axes of the crescentic chains in area 3 are oriented NWW-SSE and the axes of chains in area 4 have the orientation NW-SE. Area 5 represents a connection between the western part and Ramlat Az Zaqqut. In this area, there are several composite dunes which trend N-S. The length of the dunes varies between 25 and 58 km. The principal dunes are on the western side, whereas shorter, complementary dunes are developed on the eastern.

The western part of the sand sea represents present - day dune activity under the influence of contemporary winds. The wind regime seems to be controlled by local topography. The sand source is autochthonous, which is very clearly depicted on satellite images as a very dark red colour, corresponding with a similar colour of rocks on the northern side. The sand deposits may be related partly to colluvial deposits and partly to the alluvium reworked by aeolian activity.

4.5.2.3. The southern part

The main part of the sand sea, extending towards the south, is the largest part of Ramlat Zaltan. This extends over 360 km from 28°45'N to 25°15'N. The slightly arcuate elongated belt of sand desert is 80 km wide in the middle part and covers an area of 20 330 sq km. The northern flank of this area is called Ramlat Sazynut, which, on the northeastern side, is bounded by Jabal Zaltan. From 28°15'N towards south, the sand fields adjoin the As Sarir area extending to the Great Sand Sea. The western border is marked by the Qarat Bilhidan and Qarat An Nagah, constituting the eastern extention of Al Haruj Al Aswad.

The basic dune forms appearing in this part of the sand sea are composite linear dunes (areas 8, 10 and 11) and very conspicuous longitudinal megadunes (area 13). Area 8 constitutes a connection between the northern and southern parts of the sand sea. This area represents a sand sheet partially covered by composite linear dunes which trend N-S. The dunes are arranged in arrow - shaped form with well - marked western principal dunes. The eastern sides of the dunes are composed of shorter linear dunes oriented obliquely at 30° to the main dune axis. The length of dunes varies between 5 to 15 km, their width is up to 1.6 km in the tail part. The dunes are spaced from 2.5 to 4.0 km apart. The interdune areas are marked by zibars. Towards the south, the dune pattern is less distinct.

In area 10, the composite linear dunes are more distinct and longer. Their trend is NNE-SSW in the northern part and but, towards the south, this changes to NE-SW. The dune length ranges between 15 and 30 km and spacing between 1.5 and 3 km. The dunes pattern is staggered downwind. The dune have very acute and narrow southwestern ends, whereas the northeastern parts represent the fan - shaped forms up to 3 km wide, and the tails of adjacent dunes merge. Interdune areas are covered by zibars.

In area 11 the dune forms and patterns are very similar to those described in area 10. The composite dunes, present in this field, are arrow - shaped sand bodies. In the northern part of this field there is a significant N-S trend, while, towards south, this trend changes its direction into NNE-SSW. As in the previous cases, the principal dunes occur on the western side and the subsidiary dunes, which occur on the eastern side, are arranged obliquely to the main axis of the principal dune. The length of the dunes varies between 15 and 50 km, the width between 1.0 and 1.5 km, and spacing between 2 and 3 km. Low crescentic chains or zibars appear in the interdune areas. Similar, but shorter, composite linear dunes occur in area 14.

Owing to completly different dune forms which are present here, area 13 can be treated as a subregion in this part of the erg. In contrast to other areas discussed earlier, this field represents a sand sheet covered with megadunes. The megadunes can be also classified as separate dune fields, because the interdune areas between them are sand free, or sand is present only as a very thin sheet. The megadunes are shaped like long swords or like long shuttles. The length of these megadunes varies between 60 and 104 km; the width is up to 10 km and they are spaced at an average of 15 km apart. These megadunes are formed by linear dunes arranged obliquely or parallel to the main axis. Some of these dunes are similar to the very large composite dunes with the largest principal dune on the west side.

Areas 9 and 12 are covered by crescentic chains. In area 9, the dunes are very densely spaced - 4 crests per 1 km, and have a wavelength of 0.8 km. To the west the crescentic chains grade into the tail parts of the composite linear dunes of field 8. Area 12 is covered by the crescentic chains of parameters similar to those of 9. In the southeastern part of area 12, the crescentic chains are less significant and grade into a sand sheet.

The shape of this elongated erg is controlled in the northern part by the terrain topography, but, below 28°N, on both western and eastern sides there are sarir areas. Hence, the location of the erg can be related to the low terrain depression in which, during past geological time, the river transported sediments in a northerly direction. In such a case, the sand incorporated in the erg probably originated both from alluvium sediments and from aeolian sand transported from sarir areas to both east and west.

There is also a distinct correlation of the erg shape with the tectonic features manifested in this area. A large tectonical fault appears along the meridian 20°N and constitutes the boundary between areas 10 and 11 (National Atlas, 1978). Along this fault, there is an abrupt change of erosion levels in the Paleozoic rocks, and the contour lines of erosion levels indicate a depression. There is also some correlation with erosion levels of Precambrian basement, in the upper part of the erg, mainly between 27°40'N and 28°30'N, where the outline of the erg exactly follows the contour line of the erosion level. Area 13 corresponds closely with the closed contour line of the erosion level of the Precambrian basement.

4.6. SOUTH - EASTERN LIBYAN SAND SEAS

4.6.1. Irq Southern Rabyanah

Irq Southern Rabyanah constitutes the southern extention of Ramlat Rabyanah, and, in an arbitrary division, extends from 23°45'N to the south and terminates at 21°N (Fig.52). Irq Southern Rabyanah covers an area of 16 490 sq km. Its length is more than 320 km and the average width is some 60 km. The general direction of this erg is NNE-SSW, while its constituents - various types of linear dunes - show a NE-SW trend (Linsenbarth, 1986f). The ground elevation rises steadily from 500 m at 23°30'N to 700 m at 21°N.

The western limit of the erg is defined by the nearly straight line of the eastern escarpment of the plateau surrounding Jabal Nuqay and the eastern boundary is marked by the Hamadat al Faoudi, Hamadat al Aqdamin and Hamadat Ibn Batutah which are prominent structural escarpments (Williams and Hall, 1965; Pesce, 1971). Clusters of unnamed hills crop out in the centre of the erg (Fig.52).

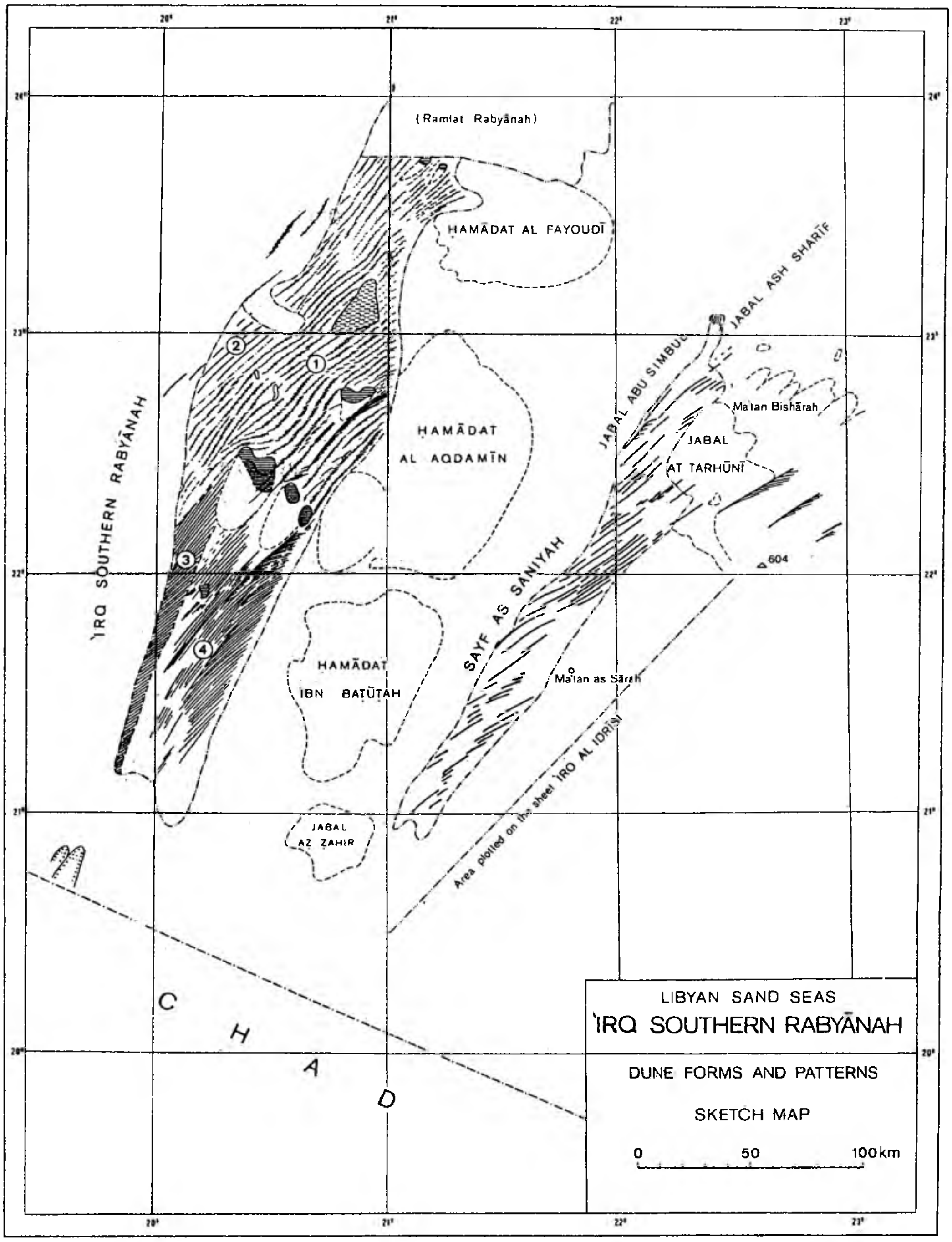


Fig.52 Dune fields in Irq Southern Rabyanah and Irq Sayf as Saniyah (In circles the numbers of the dune fields described in the text).

From a geomorphologic point of view, Irq Southern Rabyanah forms a long trough which seems to be an erosional feature, thus forming a depression created in the Nubian sandstone. The erg is covered by two types of dunes : linear sandridges (areas 1 and 2) and linear dunes (areas 3 and 4). The sandridges which cover the northern part of the erg terminate on the rocky hills present between 22°15'N and 22°30'N.

In the northern part (area 1) large sandridges occur. The ridges are composed of short linear dunes which are arranged obliquely (35°) to the axes of ridges. The length of these sandridges varies between 15 and 40 km; their width from 1.0 to 1.2 km and spacing ranges from 2.2 to 2.5 km. The ridges do not have crest lines and seem to be rather low. Towards the south, the dune dimensions decrease. The sand - free interdune areas present a low albedo. In several cases, the dune trend is slightly changed as the result of the influence of topographical barriers such as rocky hills in this field.

Area 2 is covered by sandridges of a trend similar to that in area 1, but these dunes have a comple different morphology. They are linear, spear - like dunes, with razor crests which are very striking on satellite images. The dune length varies between 15 and 25 km; the width averages 0.2 km and the spacing ranges from 1.8 to 2.5 km. The interdune areas are partially covered by zibars and the rocky basement is partially exposed.

Area 3 represents a very extensive belt of linear dunes, some 145 km long and 15 km wide. The dune trend within this field is NNE-SSW. The dune length ranges between 15 and 35 km; their width averages 0.2 km and spacing ranges between 0.4 and 0.6 km. Area 4 is covered by similar linear dunes, but the dune pattern is not so regular. The linear dunes are up to 55 km long. The dune crests are very well displayed on satellite images. South of 21°20'N, the crests are not visible and the dunes seem to be little more than sand streaks. On the eastern flanks, the dune trend is NE-SW whereas, towards west, it changes to become more NNE-SSW. On the western flank of area 4, there is a dune - free zone in the vicinity of area 3. Outside the erg, in the vicinity of the Chad border, there are two small fields covered by compound barchans, which are similar to those which are present in the southwest of Ramlat Rabyanah.

Dune forms appearing in Irq Southern Rabyanah, are controlled by sand environment, terrain topography and wind regime. The erg represents a topographical depression which can be attributed to structural and/or erosional origin. As the result of stream erosion of the quartzitic Nubian sandstone, the bulk of sand was deposited in the pre - existing depression and and some of this was transported towards the north. As the result of aeolian activity during arid phases, the sand was retransported towards the south; hence, the sand incorporated in the erg has both an autochthonous and an allochthonous origin. The northern part, namely area 1, represents an old remnant erg, while the southern part, represented by fields 3 and 4, may be regarded as a modern active erg. The sand incorporated into the sandridges of area 3 was derived from a surrounding sandstone plateau during pluvial phases and deposited in an ancient erg, following which, during a later arid phase or phases, the sandridges were developed. The present sandridges, without crest lines, have subdued ridges which have been lowered during previous pluvial phases. Owing to the limited sand transport from the north, the NE winds reworked the sandridges during the present or even a previous arid phases, thereby forming a pattern of densely - spaced linear dunes which were developed on these sandridges. The absence of smaller dune forms in the interdune areas confirms that there must have been a limited supply of sand from external sources. On the contrary, the linear dunes in area 2 can be regarded as contemporary active dunes, developed in a somewhat different sand environment. It may be assumed that there was a local source for the sand which was deposited in this area, which owing to the topographical obstruction boundring this field on the northern side, was not transported during pluvial phases to the north. The pronounced crest lines of linear dunes and zibars in the interdune areas confirm the present activity in this field.

Areas 3 and 4 follow the same trend of dune orientation, but in this part there are no remnants of old sandridges. The large hills occurring on the boundary between area 1 and areas 3 and 4 create a topographical barrier against which the development of sandridges was terminated. The sand incorporated in the linear dunes of areas 3 and 4, seems to be aeolian sand transported by wind from the northern part of the erg. Comparison of these two areas indicates that area 3 has a sand cover much thicker than that of area 4. This is obvious from the thin sandsheet and lack of dune forms on the western and southern margins of this field. It can be also accounted to the wider open area between the mountains escarpment from west and the mountain ridge in the middle of the field. Between the other rocky hills, there are rather narrow open passages through which sand is transported. On the western flanks, these hills form sand and wind shadows on the lee - side. The sand shadow effect is revealed by a very dark tone on satellite images. A very distinct line which divides areas 3 and 4, passes through the western side of the topographical barrier, and represents the wind direction. If this is true, it is indirect evidence that linear dunes are arranged obliquely to the dominant wind direction.

Irq Southern Rabyanah presents an excellent example of sequences in the dune pattern development within the erg. Following the wind direction, on the windwardmost flank there are sandridges; thereafter, then linear dunes gradually pass into sand streaks and, finally, on the downwind, flank barchan forms develop. The sequence of dune forms can be attributed both to a dimnishing sand cover and to the diminishing size of the sand grains.

4.6.2. Erg Sayf As Saniyah

Erg Sayf As Saniyah is located between Irq Southern Rabyanah and Irq Al Idrisi in a prominent terrain depression (Fig.52). From the west, the erg is bounded by the escarpments of Jabal Abu Simbul, Hamadat Al Aqdamin and

Hamadat Ibn Batutah, whereas, from the eastern side, the erg is limited by Jabal al Tarhuni and Jabal Al Hadid. The northern boundary is represented by Jabal As Sharif. The erg represents an elongated field of sand, some 270 km long and 20 ÷ 30 km wide, and covers an area of 5 600 sq km. The erg is oriented NNE-SSW, and is covered by linear dunes oriented NE-SW. The dune length ranges between 5 and 55 km, with the longest dunes appearing in the central part of the erg and the shorter ones on the southwestern flank.

The sand incorporated in the erg mainly has an autochthonous origin and was derived from the surrounding Nubian sandstones. The sand was and still is partly transported from the Kufra region. This is confirmed by the sandrose of Kufra (Fig.65). The resultant wind direction strongly corresponds with the axis of the erg. The linear dunes are oriented obliquely both to the erg axis and to the resultant wind direction.

4.6.3. Irq al Idrisi

The name of the erg was proposed by Pesce (1971) in memory of Al Idrisi, the leading Islamic geographer and cartographer of the Middle Ages (ca 1099 ÷ 1154).

Irq al Idrisi is the largest southeastern erg in Libya (Fig.53). It commences at the Libyan - Egyptian border at 23°N and extends SW for a distance of 520 km. The southwestern part of Irq Al Idrisi crosses the Libyan - Chad border for a distance of 50 km and terminates on the gently rising surface which forms the plateau of Erdi Korko. The Libyan part of Erq al Idrisi covers an area of 26,040 km, while the part of the erg in Chad covers an area of 2,360 sq km. The erg may be divided into a northern part, which consists of several wings, and a southern part, which forms a rectangular field 340 km long and some 45 km wide (Linsenbarth, 1986g).

The northern part of the erg is composed of several *sand corridors* which pass between high mountains: Jabal al Bahrt, Jabal Arknu (1435 m) and Jabal Al Awaynat (1852 m). The sandy corridors between the mountains have the widths of between 3 and 18 km. The dune fields are located in terrain depressions. The northeastern part of the erg encompasses Jabal Al Bahrt; the northern wing is wider and the southern, between Jabal Al Bahrt and Jabal Arknu, is very narrow. Area 1 is covered by crescentic chains with a wavelength of between 1.0 and 1.5 km and spacing from 0.6 to 1.0 km. The linear dunes are between 8 and 12 km in length. Area 2, located on the western side of Jabal Al Bahrt, is covered by linear dunes, which are slightly arcuate in form and trend NE-SW. The very densely - spaced linear dunes, at distances of 200 m apart, are between 5 and 25 km long. Towards the south, the dunes are channelled into two sand corridors surrounding an unnamed rocky hill of 3 km diameter and at a distance of 15 km; on the lee - side of this hill, two sand streams coincide. In field 4, which commences between Jabal Al Bahrt and Jabal Arknu, only linear dunes are present.

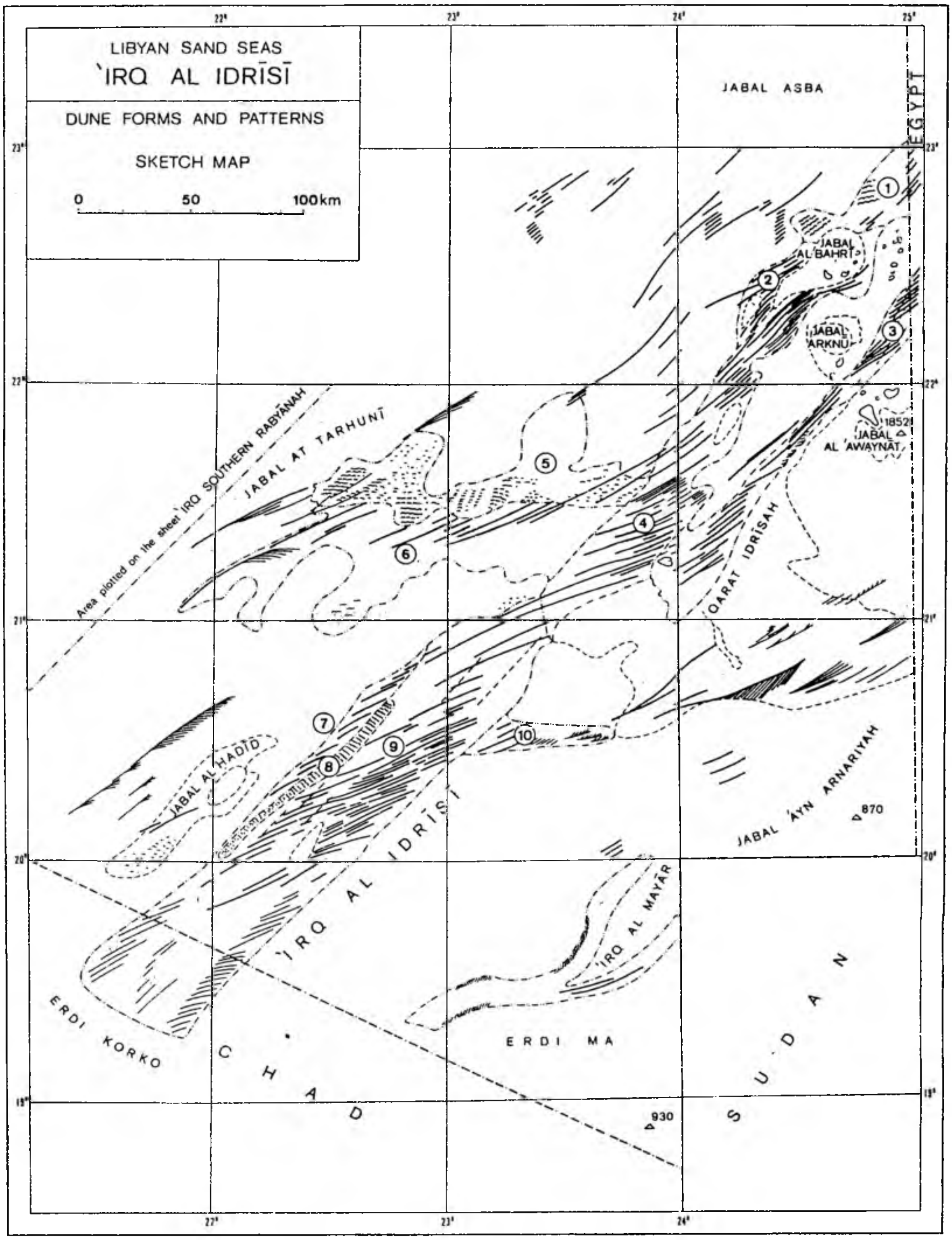


Fig.53 Dune fields in Irq al Idrisi
(In circles the numbers of the dune fields describet in the text).

In the northern part of this field, their trend is NE-SW, whereas towards the south, this direction changes into NNE-SSW. The dunes are up to 40 km long, 1 to 4 km wide and are spaced at distances of 2÷3 km apart. As well as the composite forms of dunes, there are also single linear dunes ca 100 m wide. Area 3, located between Jabal Arknu and Jabal Al Awaynat, is also covered by linear dunes. The length of linear dunes varies between 15 and 20 km; they are ca 0.2 km wide and are spaced at the distances of between 0.8 and 1.2 km. The linear dunes overlie the old subdued sandridges which trend parallel with the corriodor flanks.

At 21°50'N, the corridor neck is only 4 km, south of which large composite linear dunes occur which have a NE-SW orientation. The dunes are spear - shaped with an acute forehead at the SW end and with a feathered tail. In the central part these dunes are ca 0.4 km wide. The feathered tail parts of these dunes tend to merge together. In the southeastern part of area 3, the composite linear dunes have braided forms. Some of these dunes terminate on the rocky outcrops.

The western wing of the erg, extending from the main erg body between 21°N and 22°N, towards west up to the Jabal At Tarhuni, consists of two dune fields. The northern field (area 5) represents a sand sheet covered by crescentic chains, whereas on the southern field (area 6), the linear dunes appear. On area 5, on the northwestern flank, the transverse dunes which have a wavelength 0.8 to 1.2 km are spaced at distances from 0.5 to 0.8 km, whereas, towards the south, the spacing is larger (up to 1.2 km). The linear dunes in area 6 range in length from 15 to 60 km; their width is generally less than 1 km, but in tail parts the dune width rises to 3 km. The dunes are spaced from 2 to 7 km apart but they do not cover the entire sand sheet.

The main part of the erg (areas 7, 8, 9 and 10) covers an area of 9,880 sq km and consists of linear dunes and crescentic chains. In the northern flank of area 9, the linear dunes are longer, up to 80 km, and less frequent. Below 20°40'N, two belts of linear dunes may be distinguished. The linear dunes in the eastern belt are relatively long and in the western belt appreciably shorter. Area 7, located on the western side of the erg, is also covered by linear dunes. In both areas the dune trend is ESE-WNW while the erg axis is NE-SW. In area 10, which forms the eastern wing of the erg, the linear dunes and composite linear dunes are arranged in a staggered pattern. Between areas 7 and 9, there is a long narrow field of crescentic chains (area 8). This is some 102 km long and 5 km wide. The axes of crescentic chains are oriented parallel to the main erg axis, i.e. NE-SW. The crescentic chains or zibars have wavelengths of 0.3 to 0.5 km and are densely spaced at distances of 0.3 to 0.4 km.

Linear dunes, mostly in composite form, are present outside outside the erg. On the eastern side of the erg, between 20°30'N and 21°N and 24°E and 25°E, several large composite dunes were detected. On the northern side of the erg, there are also several solitary linear dunes, which are sometimes arranged in groups. All these dunes are located in depression corridors. To the west of Jabal Al Hadid there is very long solitary dune composed of several segments. The total length of this dune is 65 km.

The boundaries of Irq Al Idrisi are controlled by terrain topography. The main part of the erg, between 19°20'N and 21°N, with perfectly straight edges, looks like a structural trough. Pesce (1971) assumed that the erg is an erosional feature. Pesce believed that the residual hills which appear in the middle of erg have the same structural relief on their flatted - tops as the facing scarps of the erg. He postulated that the erg is a result of fluvial erosion of the Nubian sandstone, which produced the bulk of sand, which fills the pre - existing depression. Hence, the sand incorporated in the erg seems to be both of fluvial and aeolian origin. During pluvial phases, the sediments were transported towards the northeast, and, during arid phases, were re - transported towards the southwest as a result of aeolian activity. Most of the aeolian sand which accumulated in the active dune fields was (and is) transported from the Egyptian Sahara through the sand corridors between the mountains on the northeastern side of the erg. These huge mountains represent both topographical barriers for wind and sand transport. The wind regime is effected by mountains and is channelled through gaps between them. The channelled wind accelerate and hence, the sand transport is increased there. The effect of these topographical barriers of sand deposition on the lee - ward side is very characteristic. The so called sand free zones, which are spindle - shaped, are developed thereby creating a *shadow effect* (El Baz, 1984). The largest is the spindle - shaped sand - shadow of Jabal al Awaynat, which is 150 km long and 80 km wide, the second is the shadow of Jabal Arknu - 95 km long and 35 km wide, and the third, slightly S-shaped shadow of Jabal Al Bahrt - 48 km long and 22 km wide. There are also many smaller topographical heights which cause similar shadow - like effects, all plainly visible on satellite images.

The linear dunes which appear in sand corridors are arranged obliquely to the corridor flanks. This angle varies between 30° and 35°. There are no direct data relating to the wind regime in this region, but the extrapolation of the wind rose from Kufra (Fig.57 and 65), together with the corrosion marks on the terrain surface outside the erg, indicate that direction of dominant wind is NE-SW. Thus, it may be assumed that all linear dunes which appear in Irq al Idrisi are oriented obliquely, ca 30°÷35° to the resultant wind direction. The ancient subdued sandridges are oriented along the erq, i.e. to the depression axis, and they may be related to the different wind directions in the period of dune formation and/or to erosion of the underlying sediments in the earlier stages of an erg formation. The densely spaced and rather thin linear dunes can be related to the present wind action on the areas which have large amounts of sand, while the larger forms such as composite linear dunes or sandridges, which are much wider, have a longer history of development. Owing to their vertical cross - section, the sandridges in some areas in Irq Al Idrisi, were named *seif dunes* (Bagnold, 1933; Pesce, 1971)

Two other phenomena may be adduced from the analysis of the location of dunes within such an elongated erg, bordered by escarpments from both sides. Along the western flank of the erg, which is more prominent than the eastern flank, there is a corridor free from sand dunes, which can be interpreted as the effect of deflection of the wind direction from the escarpment. The linear dunes, at their western end, are slightly bent southwards, thus reflecting the influence of this change in wind direction. Another phenomenon occurs on the eastern flank. Along the erg border there is long sandridge, which forms a sand embankment similar to such forms which occur in Idhan Awbari and Ramlat Rabyanah.

The appearance of crescentic chains in the northern part of the western wing (area 5) may be correlated with the largely monodirectional wind and to the coarser sand in this part of the desert. The crescentic chains or large zibars which occur in area 8, probably have a different morphogenesis. Due to the fact that this elongated field occurs between two fields which are covered by linear dunes, the development of such forms cannot be attributed to a change of dune regime, but only to the sand environment. The Author hypothesises, that this area is located along an old river channel; hence, the sand incorporated in this field has a different origin and is therefore different in terms of its granulometric character and parameters. The wetness of the uppermost part of sand deposits suggests that this area may have a different underground water level. The wetter sand in this area may be more resistant to sand transport; hence, only the strongest winds blowing selectively from one direction are likely to be responsible for the building of transverse forms, like zibars or crescentic chains.

This erg can be classified as a modern active erg, with dune forms which are both *sand passers* and *sand trappers*. The whole area of the erg seems to be a transit area for the sand transported from Egypt to Chad through the Libyan part of Irq Al Idrisi. On the other hand, the linear dunes which occur in this area, are probably sand trappers, and during the sand transport along the erg, some amount of sand is trapped by dunes which are growing in height and are elongated; thus, lesser amounts of sand are exported from the erg than imported, and in this sense the erg has *a positive sand budget*. This is also indicated by the very few dune forms present in the southernmost part of the erg, thereby reflecting the limited amount of aeolian sand suitable for dune development.

5. GEOMORPHOLOGIC COMPARATIVE ANALYSIS OF LIBYAN SAND SEAS

The Libyan sand seas, which are subject of the present investigation, represent different types of sand seas. Individually they are characterised by different environmental factors and parameters. The aim of this part of the investigation is to define the differences between the sand seas and the factors which control their parameters. In this context the following parametres and factors are considered: *sand sea shape and aeral extent; sand sources and change in the sand cover; forms and patterns appearing in particular sand seas in relation to the environmental controls*. The differences in morphogenesis and morphochronology were also considered.

Libyan sand seas may be classified into three groups on the basis of their location:

* sand seas located within structural and topographical basins (Idhan Awbarii and Idhan Murzuq),
* sand seas developed over alluvial plains (The Great Sand Sea),
* sand seas developed within topographical depressions (Irq al Idrisi, Ramlat Southern Rabyanah and partially the southern part of Ramlat Zaltan).

Ramlat Rabyanah can be classified partially as an alluvial plain and partially as a topographical depression erg.

The shape of the sand seas is mainly controlled by the surrounding topographical barriers (Idhan Awbari, Idhan Murzuq, Ramlat Zaltan), or by escarpments of erosional depressions (Irg al Idrisi). In the case of the alluvial plains there are no distinct boundarries between sand seas and surrounding sarir areas. These boundaries are mainly marked by the composite linear dunes (the western boundary of the Great Sand Sea and Ramlat Zaltan).

The sands incorporated within the sand seas may be classified into *immobile* and *mobile sands.* Such differentiation is very important in analysis relating to the geomorphology of sand seas and their development. The immobile sands constitute the cores of sand seas and they play an important role in sand sea formation. The mobile sands incorporated in the dune forms, in the sand sheets and interdune areas, take part in the processes of dune initiation and their development.

The immobile sands are mainly unconsolidated sands which underlie the mobile sands. These immobile sands were accumulated in the pluvial phases of desert development, when sediment was transported by rivers to the places of accumulation, such as the central parts of topographical basins, alluvial plains or topographic depressions.

In the case of the sand seas of basin type, the sediment was transported by the rivers from the sourrounding mountains into the basin centre. These types of sand seas can have a *multisources origin*, because the sand was transported from the various parts of the surrounding mountains which may have the different geological make - up. Such types of sands incorporated in alluvial sediment can be classified as *autochthonous*. The examples of sand seas with autochtonous sands are Idhan Murzuq and Idhan Awbari.

In the case of the alluvial plains, the sand was transported by rivers from remotely placed sand sources, and such sediments can be classified as *allochthonous*. The Great Sand Sea is an example of an alluvial plain with allochthonous type of sediments presented by unconsolidated sand, which was transported from the mountains to the south of the Libyan Sahara and deposited mainly in the southern part of this sand sea. The northern part was probably filled by sands delivered by the drainage system issuing from Cyrenaican mountains. The Ramlat Rabyanah may also be regarded as an erg with an alluvial core, created by allochthonous sediment.

In the depression - type of erg, the sand can have both autochthonous and allochthonous character. In the case of Irq al Idrisi, which represents an erosional depression, the sand was derived from Nubian sandstones and was probably delivered at least in part, through river transport, during pluvial phases.

The mobile sands incorporated in the sand seas have been derived from internal sources, i.e. from the alluvial core, or have been supplied from the external sources by wind transport. In many cases, the mobile sands incorporated in the sand seas have a mixed origin.

In a case of the sand seas without external sources, the mobile sand takes part in the process of dune formation and development, and is partially transported along the resultant wind direction. When the sand sea is closed by the topographical barriers located on the downwind side of erg, the active sand then accumulates in that part of the sand sea. As the result of such action, the upwind parts of sand seas are depleted of mobile sands. In contrast, the amount of mobile sand on the downwind part of the erg increases, and the active sand thickness tends to increase across the sand sea. Idhan Murzuq is a good example of such a sand sea. Here, the NE parts are depleted of sand, while a huge amount of sand has accumulated in the SW parts.

In the case of the sand seas with external sources of mobile sand, the situation is different. Usually, on the upwind side of the erg, the sand is integrated (trapped) into dune bodies and only a part is transported across the erg. When the erg is open on the downwind side, some sand is transported outside the erg. The classical examples of such ergs are Irq al Idrisi and Irq Southern Rabyanah, both characterised by a diminishing sand cover in the downwind direction.

The amount of active sand incorporated into the sand sea can be also characterised by the *sand budget*, which can be *positive* or *negative*. When the amount of exported sand exceeds the amount of imported sand, the sand sea is said to have a negative sand budget. In the case when there is no transport of sand outside the erg, the sand sea has positive sand budget (or, the sand content of the erg remains the same) there is a tendency for deposits to shift in the downwind direction). The northern part of the Great Sand Sea is an excellent example of an erg with a negative sand budget; also Idhan Awbari which has a very limited supply of mobile sand from NE. It seems also that Ramlat Rabyanah represents an erg with a positive budget because the amount of sand transported (mainly from the southern part of the Great Sand Sea) is probably larger than that transported to the Irq Southern Rabyanah.

The dune forms and patterns which are present in the different sand seas have a strong correlation with the sand sources and wind regime in particular sand seas and neighbourhoods. As indicated by the detailed studies of wind regime presented in App. 4, the western sand seas, Idhan Awbari and Idhan Murzuq, are controlled by multidirectional winds of high potential. By contrast, the sand - moving winds in the vicinity of the eastern sand seas: the Great Sand Sea, Ramlat Rabyanah, Irq al Idrisi, represent typical bimodal wind regimes with medium drift potential. The northern part of Ramlat Zaltan is influenced by multidirectional winds of rather high drift potential. These differences are reflected in presence of different dune forms and trends in those areas.

The western sand seas are characterised by the large, mainly compound, dune forms which have resulted from a huge amount of sand being incorporated in these ergs, and the wind regime. The main forms, present in these regions are the different types of compound dunes, such as longitudinal sandridges and domical sandridges. Idhan Awbari is the only sand sea in the Libyan Sahara where the domical sandridges occur. Idhan Murzuq is occupied by a grid pattern of sandridges whereas in Idhan Awbari, the so - called deflation rims occur. The large compound forms occuring in both sands seas, namely the grid pattern of sandridges, confirms the hypothesis that such forms are controlled by the secondary winds generated by these dune forms, which acted as local terrain relief. These forms do not occur in the other Libyan sand seas.

In the eastern sand seas and in the southeastern ergs, the dominant dune forms are presented by different types of linear dunes, both simple and composite, and by the crescentic chains which reflect the control by bi - modal wind regimes in this part of the Libyan Sahara.

The northern part of Ramlat Zaltan may be classified as an area of contemporary dune development, where the most conspicuous dune forms are represented by star dunes, crescentic chains and linear dunes.

In several sand seas the abrupt change of dune trend and form cannot be explained by the relationships discussed above. Author hypothesises that these changes can be relatted to the multistage or multicyclic development of sands seas and environmental conditions prevailng during these cycles. The more detailed hypothetical scenario of sand seas formation and development is discussed in Chapter 6.

Table 4

CHARACTERISTICS OF LIBYAN SAND SEAS

NAME OF SAND SEA	MORPHOST-RUCTURAL TYPE OF SAND SEA	SAND SOURCES	CYCLES OF EVOLUTION, (mono-multi)	SAND BUDGET	PRESENT WIND REGIME	DOMINANT DUNE FORMS
IDHAN AWBARI	structural and topographical basin	autochthonous (multisources)	multi	negative	multi-directional	- longitudinal sandridges - domical forms - star dunes
IDHAN MURZUQ	structural and topographicai basin	autochthnonous (multisources)	multi	negative	multi-directional	- longitudinal sandridges - reticulate pattern of sandridges
GREAT SAND SEA	alluvial plain	allochthonous	multi	negative	bi-directional	- longitudinal sandridges - linear dunes, composite linear dunes - hooked dunes - barchans
RAMLAT RABYANAH	alluvial plain	allochthonous	multi	positive	bi-directional	- longitudinal sandridges - linear dunes - composite linear dunes - crescentic chains
RAMLAT ZALTAN	topographic depression (alluvial plain)	autochthonous	mono	negative	multi-directional	- star dunes - linear dunes, composite linear dunes - crescentic chains
SOUTHEASTERN SAND SEAS	topographic depression	autochthonous + allochthonous	mono	positive	bi-directional	- linear dunes - composite linear dunes - barchans and crescentic chains

Within the sand seas, the effects of topographical barriers are perfectly reflected e.g.Ramlat Rabyanah, Irq al Idrisi, and Irg Southern Rabyanah and Ramlat Zaltan. On the up - wind side of the topographical barriers, the change of the dune trends, tailored to the shape of barriers is obvious, whereas on the downwind side the so called *sand shadows* appear. The main characteristics of the Libyan sand seas are given in Table 4.

6. HYPOTHETICAL SCENARIO OF SAND SEA DEVELOPMENT

This hypothetical scenario of sand sea development is based on the results of studies performed on the study area of Libyan Sahara, where a remote sensing approach was applied for geomorphologic investigations (Fig.54). A detailed analysis of dune forms and patterns undoubtedly indicated that the development of sand seas is referrable to several cycles, corresponding with the phases of climatic fluctuation in that part of the Libyan Sahara. Each cycle consists of two phases: an arid phase and a pluvial (wet) phase.

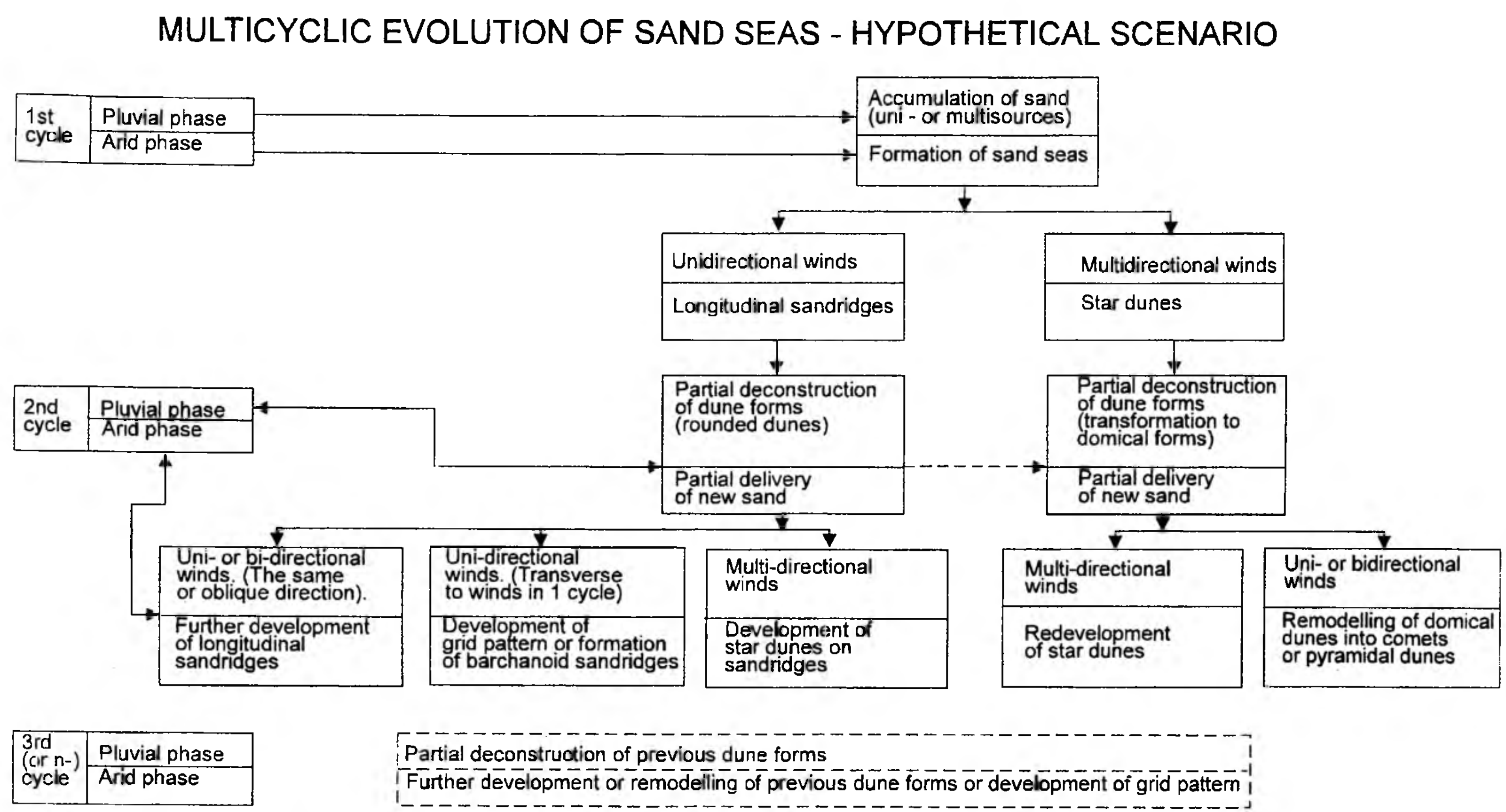
It is assumed that the sand seas have been formed in areas where the sediment accumulated during rather long pluvial phases. This sediment constitutes the source material for sand seas formation during the consecutive arid periods.

The formation of sand seas commenced during the first arid phase. This was the result of wind action on the accumulated fluvial sediments which started to be reworked as aeolian sand. Assuming that, at that time, the wind energy was much stronger that of the present, and that winds were much more unidirectional, the longitudinal ridges were initiated. The Author agrees with the hypothesis that in the first phase of dune formation, the dune forms resulted from both erosional and depositional action of wind, which thereby created the longitudinal sandridges, which are regularly spaced at a distance of between 1.5 km to 3.0 km. Considering the large size of contemporary sandridges, it may be assumed that the first phase of arid period was of considerable duration.

In the case when, during the first arid phase, the wind was multi - directional, rather than uni - directional, the other forms of dunes, such as the star dunes, must have been initiated.

The demobilisation of aeolian sand and sand forms occurred presumably within the first pluvial phase. The dune crests were lowered and dune forms rounded and subdued. It is assumed that secondary forms which were superimposed on the primary forms, were eliminated at that time.

The rivers were then reactivated and as a result of this, some parts of dune fields, especially those located on the boundary of the sands seas, were completely destroyed. The newly - available sediment was transported to the accumulation areas. Among elsewhere, the interdune areas must have been a very convenient places for sediment accumulation.



Note. All processes (both within the pluvial and the arid phase) depend on the duration and energy of particular climatic factors. Consecutive phases of sand sea development depend on the forms developed in the preceding phase and their interaction with new climatic factors.

Fig.54 Hypothetical scenario of sand seas development.

On the onset of the second arid phase, the sand sea represented a completely different situation. The drying climate reactivated the aeolian processes of dune redevelopment. These processes were a function of contemporary wind regime, which might have been the same, as that of the first arid phase or was much changed. In case of the same wind regime (mainly the same wind directions), the earlier dune forms were reactivated in a similar way and started to grow. But it is necessary to emphasise that the earlier dune bodies, might have acted as a local terrain relief, self - generating secondary winds, all of which influence the processes of dune formation.

In the case of a different dune regime, expressed both in the changes of wind potential and wind direction, the dune forms created during the first arid phase became remodelled. This process depended on the dimension of the dunes created during preceeding phase and on the wind energy and direction.

Let us consider the case of the huge longitudinal sandridge and the new wind regime with flow transverse to these ridges. The sand will be transported over the windward flanks, to be partially deposited on these flanks and partially transported through the sandridge to the interdune area beyond, where the new dune forms may be developed in a direction transverse to the sandridges. If that process was prolonged, the linear dunes joined the successive sandridges and a grid pattern will be formed (e.g. Idhan Murzuq). As a consequence, due to the new orographic conditions, cellular secondary wind flows can be generated, thus remodelling the grid pattern into more elliptical interdune hollows.

When the wind regime in the consecutive arid phase is changed into multidirectional type, the process of remodelling of the sandridges will be concentrated on the uppermost parts of sandridges, where the star dunes can be developed and other reversed dunes along the crests (e.g. Idhan Awbari).

In the case of the linear dunes or low sandridges, subdued during the wet periods, and with a new transverse flow in the consecutive arid phase, new linear dunes will cross the previous dune forms. If this process is prolonged, the new linear dunes can be more prominent than the old forms (e.g. Ramlat Rabyanah and eastern part of Idhan Awbari).

In the case of the huge domical sandridges, the scenario of their development, seems to be different. Within the first arid phase, characterised by multidirectional and high energy wind regime, star dunes were developed which can reached large dimensions. During the wet phase, the arms of star dunes were partially destroyed and subdued. Finally, the star dunes changed into a more domed form. In the consecutive arid phase, marked by a mutidirectional wind direction but a lower wind energy, the process of dune rebuilding commenced. On the flanks of domical sandridges, secondary forms were developed, whereas on the most upper parts, star dunes were developed, a reflection of the stronger winds in the upper parts of dune. The domical sandridges in the western part of Idhan Awbari are classical examples. Some of these domical sandridges have the form of „a comet", with a marked head part and tail part, thereby indicating the dominant wind direction in a multidirectional wind regime.

In the proposed scenario, the most important factors influencing remodelling of the old dune forms or development of new forms are the climatic controls within the arid and wet phases in each cycle of sand sea development, and the duration of those phases. Firstly, in the arid phase, the large dune forms are developed; in the consecutive wet phase the demolition of such forms is not so drastic (as in the case of smaller dune forms), and the main bodies of dunes will persist. Hence, in the next cycle, during the arid phase, the remaining (existing or remnant) dune forms act as the terrain relief, modifying the wind circulation and influencing the remodelling processes of the dune forms.

In each cycle of the sand seas development, the balance of sand supply should be considered. With reference to the wind direction and sand sea location, the new sand sources are very important. A similar scenario will be maintained in successive cycles of sand seas development.

The formation of the dunes in sand seas commenced during the first interglacial dry period of uncertain age. In chapter 4 the hypothetical age of particular sand seas in the Libyan Sahara was given. In accordance with this hypothesis the age of Libyan sand seas is in the order of 20,000 B.P. (last glacial maximum) to 150,000 B.P. (Riss glaciation), but it possible that the formation of the huge Saharan sand seas started ca 2 million years B.P. as suggested by Wilson (1971). The later pluvial phases correspond with the interglacial periods, while the dry, dune building phases - with glacial periods. There is no doubt that the most intensive processes of the dune formation in the Libyan Sahara, as well as in other Saharan sand seas, occured during the long arid interval between 20,000 B.P. and 12,000 B.P. The differences in the sand dune forms in the Libyan sand seas indicate that the formation in western sand seas namely Idhan Awbari and Idhan Murzuq started earlier than in the eastern sand seas.

Assuming the changes in climatic control persist in the future, the proposed hypothetical scenario of sand sea development can be used for the prediction of the future development of the sand seas and sand transportation along deserts. In turn this knowledge may enable us to predict where desertification may be progressive and what it might be counteracted.

7. FINAL CONCLUSIONS AND RECOMMENDATIONS

1. The application of remote sensing data and techniques to the geomorphological studies of sand desert areas shows that investigations can be carried out both on a regional and a local scale. A regional scale study reveals the desert landforms and processes in a wider geomorphic context; it demonstrates the connection between certain desert landforms by comparative analysis. The small - scale satellite images provide an opportunity to investigate the desert areas on a regional scale and to perceive features that are not readily perceptible either on site or at a larger scale. High - resolution remotely sensed data have been used for local scale studies of selected dune fields and dune forms.
2. The remote sensing data offer a unique possibility for multiple - scale of geomorphical investigation: i.e. at the macro-, meso- and micro scale. In digital form, the same space remote sensing data, may be applied to multi scale investigations. The space images displayed at different scales on a monitor screen represent an objective portrait of desert areas, which are relatively free from unintentional bias which is commonly introduced in the traditional generalization procedure of mapping.
3. Small scale remote sensing data were used for geomorphological mapping of the Libyan sand seas (total area of 303 000 sq kms.) and for measurements of the morphometric parameters of dune fields and dune forms. The high - resolution data were applied for the local geomorphic studies of selected dune fields in Idhan Awbari, Idhan Murzuq, the Great Sand Sea and Ramlat Zaltan. Multitemporal data were used for analysis of sand desert dynamics.

The evaluation of different remote sensing data, as acquired by different sensors and from different space platforms over the Libyan Sahara, and processed by different techniques, has led to various conclusions and recommendations relating to the application of remote sensing for particular aspects of desert morphology (see Chapter 2.2.9).
4. Due to the fact that Libyan sand seas represent such a variety of geologic and geomorphic features (tectonic and topographical basins, alluvial plains and topographical depressions) as influenced by the different climatic conditions both in the past and present time, the studies confirmed that the Libyan Sahara exemplifies a very representative test study area.
5. The elaborated maps of the Libyan sand seas and the processed remote sensing data, have together created the basic material for detailed morphological studies of particular dune forms and sand seas. The analytical results of the applied remote sensing data indicate that individual geomorphic processes leave the distinctive imprint upon desert landforms, what is now reflected in specific dune forms and patterns. This variety of dune forms and patterns indicates the complexity of geomorphic processes responsible for their formation and development. Satellite images have revealed the distinctive characteristics of each

stage of dune development, e.g. the secondary dunes superimposed onto primary forms or recently developed dunes superimposed on the subdued older forms.

The integrated investigations based on remote sensing data and other supplementary data regarding geology, sand sources, terrain topography and climatic conditions, both present and past, has facilitated the of study the dune forms, their morphology and development. The studies conducted on the Libyan Sahara test area confirmed many of the existing theories and hypotheses regarding the geomorphology of sand deserts and has also led to the definition of some new ideas relating to the formation of particular dune forms and sand seas and their development. In many previous investigations, which related only to small areas or to certain environmental conditions, the resulting theories or hypotheses did not possess a sufficiently wide overview and it is hardly surprising, with hindsight, that so many of them have proved to be so controversial. Generally, most of the previous studies may be considered to have been too biassed towards geomorphological process and to the short term responses of dune forms to diffferent climatic and geomorphological controls.

6. The approach to the analysis of similar dune forms appearing in the different localities and different climatic conditions taken in this investigation has led to the differentiation of the factors which control the formation and development of particular dune forms. The results of investigations carried out in the Libyan Sahara, have confirmed that it is necessary to differentiate and classify the dunes into: simple, composite and compound forms. The studies indicate that **all factors** controlling sand desert formation and development (geology, sand sources, sand characteristics, terrain topography, wind regime and climatic factors considered both spatially and on a temporal scale) should be included in any comprehensive morphogenetic and morphochronologic investigation.

The factors controlling the dune development may be divided into stable and unstable. The geology, terrain topography and sand sources, can be accounted to the stable factors, while climatic characteristics including wind regime - to the unstable factors. The performed investigations have revealed that the terrain topography plays a very important role both in the location of sand seas and particular dune fields and in the generating of the secondary flows of surface winds which control the development of certain types of dune forms. The formation of huge compound dune forms creates the new orographic conditions, selfgenerating secondary winds which in turn influence further development of dunes.

The wind regime plays a very important role in dune initiation, formation and development. Particular dune forms depend on the drift potential and drift direction as well as on the resultant wind potential and direction. Also the frequency of the wind directional variability (hourly, daily, monthly, seasonally or yearly), taken together with the drift potential from each direction in each phase, plays a significant role in the formation and development of particular dune forms. For example the straightness or sinusoity of linear dunes results from these factors.

The investigations have shown that the complex interrelation of the many different factors plays very important role in the creation and development of particular dune forms. The same controlling factors, acting with a different energy or in diffferent proportion, or in different time scale, can produce quite different dune forms; on the contrary different factors operating in an another relationship can produce quite similar forms. The detailed findings and conclusions regarding particular dune forms and their hypothetical morphogenesis were presented in Chapter 3.

7. The results of geomorphological investigations carried out on the test area of the Libyan Sahara sand seas suggest that it might be possible to determine the dune trends and patterns within particular sand seas and to perform a comparative analysis of sand sea geomorphology as a function of the factors which control their formation and develoment. The remote sensing approach permitted the definition of areas within particular sand seas which have a different morphogenesis and morphochronology, as expressed by the different dune forms and trends, sand characteristics and in the abrupt changes of dune trends and patterns. The results of investigations show that particular Libyan sand seas are in different stages of development and are controlled by different factors (see Chapters 4 and 5).

8. The remote sensing approach has revealed the multicyclic development of sand seas, which is reflected in the particular dune forms, mainly the huge sandridges appearing in Idhan Awbari and Idhan Murzuq. The sequences of superimposed dune forms, created in the different cycles of sand sea development, were detected in all sand seas studied. The investigations confirm that a proper interpretation of present - day desert landforms is impossible without a full appreciation of the manifold influences of the geologic, geomorphic, and climatic changes during the Pleistocene. The remote sensing approach to geomorphologic investigations of sand desert areas allows to propose a hypothetical scenario of sand seas (see Chapter 6).

9. The multitemporal remote sensing data afforded the possibility of carring out preliminary studies of sand sea dynamics and of classification of the sand seas into areas of stabilized sand dune processes (Idhan Awbarii and Idhan Murzuq) and contemporary sand dunes development (Ramlat Zaqqut).

10. The results of these studies indicate that the Libyan sands seas generally possess a negative sand budget and that, there is a general tendency for sand to be transferred to the in S and SW.

11. On the basis of these studies, the following recommendations regarding future application of remote sensing to the geomorphic investigations of sand desert areas may be formulated:

- investigations of particular sand seas should be based on high resolution remote sensing data acquired in a stereo mode;
- digital photogrammetric techniques, based on digital orthophotography and a digital terrain model, will improve geomorphologic investigations of sand dune forms and sand seas;
- the dynamics of sand desert areas should be studied on the basis of multitemporal high resolution images referenced to ground control points;
- for very local studies of sand dune dynamics, aerial photographs or terrestrial photographs should be applied (Linsenbarth,1974; Ostaficzuk 1962, 1978; Sitek, 1971).

All these investigations based on the remote sensing or photogrammetric methods should be supplemented by local measurements of wind and other climatic parameters and field studies of sand characteristics.

The permanent remote sensing monitoring of sand desert areas may permitt the determination of the morphodynamics of this presently active landscape and so to predict the future tendency of sand sea development and its influence on both regional and global environmental change.

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APPENDICES

1. GEOMORPHOLOGIC EVOLUTION OF LIBYAN SAHARA
2. ENVIRONMENTAL AND CLIMATIC CHANGES IN AFRICA IN LATE PLEISTOCENE AND HOLOCENE
3. WIND CIRCULATION IN NORTHERN AFRICA
4. SAND MOVING WINDS IN THE LIBYAN SAHARA
5. BASIC PRINCIPLES OF AEOLIAN SAND MOVEMENT
6. DESERT SANDS AND THEIR CHARACTERISTICS

Appendix 1

GEOMORPHOLOGIC EVOLUTION OF THE LIBYAN SAHARA

1. EASTERN AND CENTRAL PART OF THE LIBYAN SAHARA

Morphologic history of Libyan Desert commenced with its emergence from the sea. The first to emerge was the Libyan Sahara, which had already emerged at the beginning of Mesozoic; then came Tripolitania at the beginning of the Eocene, and, finally, Cyrenaica and Sirte which emerged towards the end of the Miocene (Desio, 1971).

The geomorphologic evolution of Libya started during the Lower Tertiary (Paleocene and Eocene) when the Tethys Sea, the sea ancestral to the Mediterranean, occupied the area of what is now the Sahara. At the beginning of Eocene, the Tethys Sea was connected by the Trans - Saharan Seaway with the South Atlantic Ocean (Gulf of Guinea). This Trans - Saharan Seaway passed through Libya, Algeria, Mali, Niger and Nigeria. At that time, Libyan territory was divided into two parts by the large Paleosirtic Gulf (Fig.55), which profoundly effected the morphologic evolution of the surrounding continental territories, as well as the drainage system. At the beginning of the Tertiary (Paleogene) the Paleosirtic Gulf was 300 ÷ 500 kilometres wide. A marine transgression to the Sirt basin (Paleosirtic Gulf) reached its maximum during late Eocene times, when it reached the foot of the Tibisti massif.

The presence of the sea had a great influence on the climatic conditions of the Libyan Sahara being surrounded in south by the mountain barriers of Uweinat (Jabal al Awaynat), Tibisti, Hoggar and Tassili, which acted as areas of atmospheric humidity. Desio (1971) believes that at the beginning of Paleocene, there was a tropical climate with rather high rainfall in the vicinity of the Paleosirtic Gulf, perhaps one comparable to the present climatic conditions of Nigeria.

To reconstruct the drainage system of the Libyan Sahara at that time, we must assume that there were water courses flowing from the surrounding mountains to the sea. The orographic configuration of Libya, evidence for which is supported by the interpretation of satellite images, indicates that two principal drainage areas existed at that time in the southern part of the Libyan Sahara. The western area originated in north Tibisti and extended to the northwest and north, and the eastern, was situated between the Jabal Eghi and Jabal al Awaynat mountains and oriented towards north and northeast. The two large catchment areas merged near Jabal Eghi and Jabal Nugay, where they formed the northern extremity of the northeast span of the Tibisti i.e. where the Rabyanah Sand Sea is presently located. The system of large wadis, more or less complex in structure, slopes gradually towards the north. It may be assumed that, during the Paleogene, the active water courses were developed and transport of pebbles took place over distances of several hundred kilometres. This is confirmed by the comparison of the gravel of Sarir Tibisti and the sand from Erg Al Jaghbub, which constitutes most of the northern part of the Great Sand Sea. The mineralogical and petrographic evidence indicates an origin of the sediments from the granite, dioritic and syenite from the massifs of the south.

A marked change of the depositional environment occurred during the Oligocene. In the early Oligocene (close to the Eocene boundary) marine conditions prevailed in the Sirte Gulf, and marine sedimentation occurred there. Only on the southwestern margin did fluvial conditions occur at that time. Uplift of the southern part of the Sirte basin, resulted in a regression of the sea. The shore line, recognized by geologists, had retreated at least 150 km to the northeast. Hence, the maximum extension of the non-marine fluvial sedimentation took place within Oligocene times. Towards the end of the Oligocene, marine transgression recommenced and the shore line advanced southwestward as latitude 27°30'N. In Lower and Middle Miocene times, trangression and regression alternated. These alternations are reflected in the differing lithologies, showing an overall gradation from alluvial sands and sandstones in the south and southwest via an intermediate shoreline complex to a completely marine sequence in the north (Benfield and Wright, 1980). In post - Miocene times, a major regression of the sea occurred, due to a significant fall in the level of the Mediterranean. The post Middle Miocene sea retreated beyond the present coast line. The lowering of the Mediterranean Sea level would have resulted in terrestrial erosion over a very vast region, and it seems possible that the erosion surface at the base of the Calancio Formation was created at that time (Benfield and Wright, 1980).

The disappearance of a large area of the sea (hence the increase of land area) has affected the climate conditions resulting in the reduction of the annual rainfall. The drying - up of the climate caused the reduction of the water courses which originated in the mountain regions and terminated at the southern shores of the ancient Mediterranean. Major inputs of sand were directed northwards in the southwestern part of the basin and to the north and northeast in the vicinity of As Sarir, showing the existence of two large water courses at that time hereabouts.

During the Miocene, Cyrenaica (north to the Libyan Desert) emerged, taking the form of a large asymmetric dome, with steep dips on the north side and very gentle ones on the south. The Jabal al Akhdar largely rose from the sea, leaving only a narrow arm of the sea in the Sabkhat al Qunnayyin and al Jaghbub area, which later disappeared in the Pleistocene. The uplift of Cyrenaica and general emergence from the sea continued into the Pliocene.

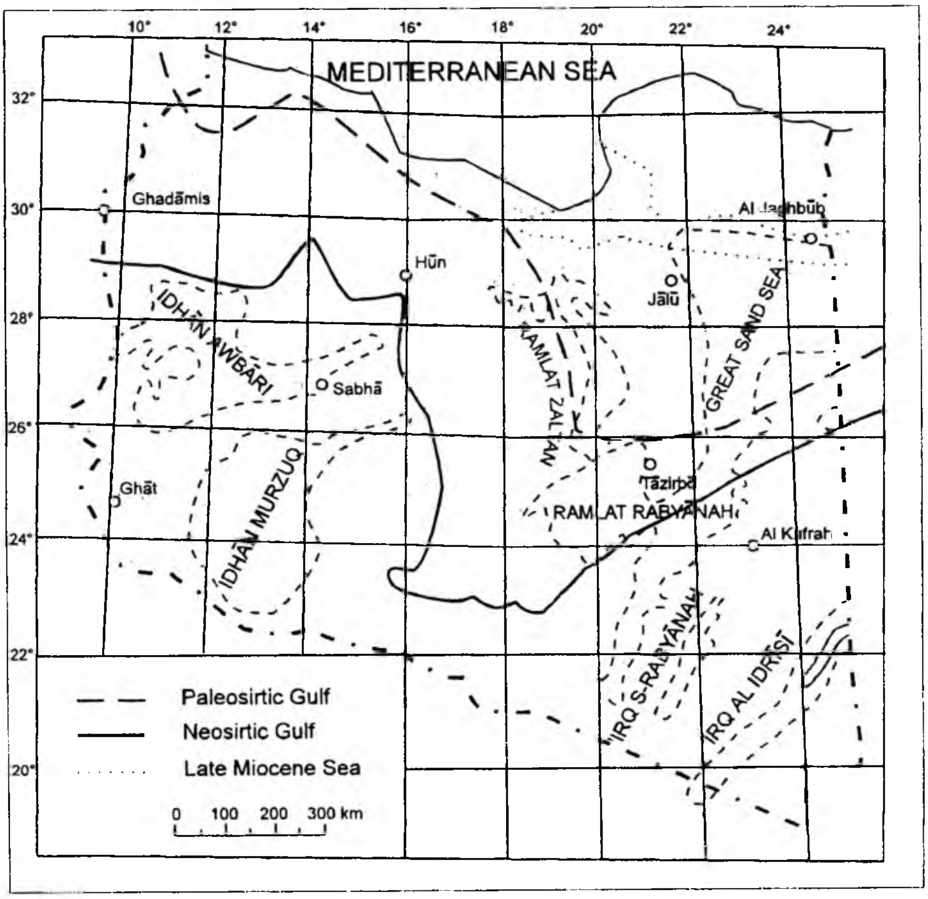


Fig.55 Extension of the Paleosirtic Gulf and of Neosirtic Gulf during the Middle Miocene and the shore of the Late Miocene Sea (modified after Desio, 1971).

At the end of the Miocene, the water courses, which were more extensive on the gentle southern slopes, flowed into the long arm of the sea. The sea remained in the al Jaghbub depression (20 metres below sea level) until the end of the Miocene. In northwestern Sirtica, a long NW-SE trending ridge emerged during the late Miocene. The ridge blocked some parts of the lower water courses, which originally reached the sea after having drained the surface of the former Paleosirtic Gulf, which had recently emerged from the sea. In the Sirtica region probably only a small bay of the sea remained in the Pleistocene, which did not extend further than the 30° parallel to the south. In the Sahabi area (Sabkhat al Quanayyin), the gravelly-sandy Sahara deposits of sarir type are located further north than in any other part of Libya. The aeolian-type sandy gravels are estuarine deposits which fill the depression and cover ancient erosion surface of the Late Miocene.

In the Early and Middle Miocene this area possess a tropical or subtropical climate. At Sahabi and in the area situated to the south, a tropical climate persisted into the Pliocene. The large part of the present Libyan Desert had a warm, humid climate comparable to a monsoon climate. Such conditions were suitable for pediplain development. The land surface represents an aggradational pediplain within the basin and, towards the west, this is succeeded by an erosional pediplain of Late Tertiary to Quaternary age. The aggradation did not become significant until Early Pleistocene times, when sea level returned to its former position. These surfaces have been modified and regraded during Holocene erosion cycles.

The recent history of the Libyan part of the Great Sand Sea (Calanshio Sand Sea) can be reconstructed on the basis of the documented Quaternary sequence of the Garet Uedda Formation (Di Cesare *et al.,* 1963) in the area bordering the Calanshio Sand Sea, between Al Jaghbub and long. 22°30'E. This formation overlies an eroded surface which intersects Middle Miocene rocks and also lacustrine sediments which contain fossils of Pleistocene to Quaternary age

The documented sequence comprises of four horizons of lacustrine sediments separated by three horizons of eolian sands. The thickness of the outcrop zone varies between 10 and 40 m, and increases to the south. Also, the sequence shows a transitional passage southwards to more fluviative sedimentation, notable in respect of each lacustrine member. Di Cesare *et al.*, (1963) assumed that the aeolian layers represent 4 or 5 pluvials and 3 or 4 interpluvials of arid phases with dating in the possible range of Pliocene to Quaternary. During the 1st pluvial deposition of gravel and sand by streams coming from south (Paleo - sarir) occurred and the lacustrine deposits were found at the northern limit of the present Great Sand Sea. Then, during the 1st interpluvial, the older sediments were eroded and the deposition of the first ancient erg occurred. During the 2nd pluvial, a degradation of the former erg, erosion of the former sediments and deposition of new sediments may be assumed. The 2nd interpluvial is responsible for the creation of the 2nd ancient erg (quartz sand) which was again eroded during the 3rd pluvial period. The 3rd interpluvial was responsible for deposition of caliche with brown chert. The signs of the 4th interpluvial are not recognized. During the 5th pluvial, erosion of the caliche occurred and the infilling of streams took place (sarir).

Towards the end of the Paleogene and at the beginning of the Neogene, the continental climate, with a reduction of precipitation, was a characteristic of the Libyan Sahara. Geomorphologic events during the Neogene are extremely difficult to reconstruct because, between Paleogene and the beginning of Quaternary, the evolution was continuous. The variation of the climate caused changes in the topography.

The movement of the coast-line of the Paleosirtic Gulf towards the north was gradual. The river courses lengthened over the gently slopping plain which emerged from the sea. The former sea-bottom, covered by loose marine deposits, was reworked by rivers. In the south, the rivers reworked the alluvial deposits of the preceding period. In the southern part, two ancient drainage systems were redeveloped; they became wider and generated new branches. The demobilisation of the tableland was progressive. The fluvial system was formed by the junction of the former wadis from the Tibisti and Nugay which crossed the Sarir Calanshio in a NE direction. One main river course probably goes NE, as confirmed by the fossil river courses under Erg al Jaghbub (Di Cesare *et al.*, 1963) and which can partially be detected on satellite images. The second main river course swings towards the Gulf of Sirte (Wright and Edmunds, 1971). This early Holocene system called „Ye-Ba-Eg" consists of Yebique, Bardaque and the rivers from the Eghei (Nugay) mountains such as Oyouroum and Tideti, which flow into the Sarir Tibisti (Pachur, 1980). According to Pachur, the formation of such a fluvial system can only be explained by autochthonous precipitation in the flat areas beyond the Tibisti mountains. The morphology and petrography of the fluvial deposits indicate that they were transported over a distance of about 800 km from the northern edge of the Tibisti, at a gradient of about 0.6%. The fluvial deposits of the proto - Ye-Ba-Eg river are over 10 km wide in Sarir Tibisti and narrow to about 100 m in the Calanshio. It is assumed that this river flowed only periodically. The fluvial deposits of Wadi Bahar Beloma contain acidic volcanic pebbles which are only known from the interior of Tibisti.

2. WESTERN PART OF THE LIBYAN SAHARA

The western part of the Libyan desert is known as Fezzan. The western sand seas, Idhan Murzuq and Idhan Awbari (Fig.5), are located within the Murzuq basin. The Murzuq basin forms essentially one geological unit, a large cratonic structure, which is underlain by Precambrian rocks (Zaluski and Sadek, 1980). The basin is filled with deposits which range in age from Cambrian to Quaternary, with a total thickness of more than 3000 m in the central part. The Murzuq basin is bounded on the southwest and south by the high crystalline massifs of Hoggar and the Tibisti which form a part of the African shield, and on the north by the Al Qarqaf arch which was uplifted during the Caledonian orogenesis. Towards the west, this basin is separated from the Jllizi (Pelignac) basin in Algeria by the Tiomeboka Anticline which is a late - Caledonian and Middle Devonian uplift. During the Jurassic and the lowest part of the Cretaceous, a braided river system crossed the basin. This had a palaeotransport toward the northeast. The whole sedimentary cycle was probably finished by the Early Cretaceous when a new cycle of erosion started. During the Upper Cretaceous, Paleogene and most of the Neogene, there was a long period when continental conditions prevealed during which there was very little sedimentation.

The time between the Early - Middle Creaceous and the late Tertiary was a period of weathering and erosion, generally under arid and semiarid conditions. The morphologic process was one of pediplanation. It certainly created the main morphologic features of this area: the Msak escarpment (Msak Mallat), the pediplane on its northern side, and the hamada surface which slopes gently southwards. The flat area in the eastern part named Sarir al Qattusah, may be one of the peneplains produced at that time (Desio, 1971).

Before the onset of the Tertiary, slight uplift of the central part of the Murzuq basin occurred, resulting in erosion of the Msak plateau which trends eastwards and northeastward towards the present Awbari depression. Then a gradual sinking of the central parts of the basin took place, which resulted in the tilting of the Msak plateau towards the east.

During the Paleogene, large drainage system was developed around the mountains of Manghini and Tunimo to the south and Tassili n'Ajjer to the west. The principal water courses which originated in these mountains, probably also flowed to the northeast and east into the Paleosirtic Gulf. The upheaval and emergence of the Haruj (Al Haruj al Aswad), which took place in the Oligocene, effected radical changes in the drainage system of Fezzan. The upheaval of the Haruj created a natural obstruction of the eastern part of the drainage system. The topographic basins of Murzuq and Awbari, formerly open to the sea, became closed basins. The water from the rivers formed vast lakes in the lower eastern parts of these basins, in which gravel and sand from the surrounding mountains were deposited. At a later stage of relief development, during deep-water erosion, some escarpments and broad wadis were formed (e.g. Msak Mastafat and Ra's Shirdan escarpments, Wadi adh Dhamran, Jabal Traghan etc.). The newly-originated depressions were filled with Tertiary to Quaternary continental deposits.

By the end of the Neogene and during the transition to the Quaternary, there was increased fluvial and lacustrine deposition. Two large lacustrine basins developed: one on the northern side of the Msak escarpment the second in the interior of the Murzuq basin. The deposits of both basins show a two - part sequence: sand in the lower part (the first phase of deposition) and algae limestone with algae which remains in the upper part. The relatively thick lacustrine deposits probably belong to a period close to the Tertiary - Quaternary boundary. These continental lacustrine deposits are discontinuously developed along the margins of the Murzuq and Awbari basins. At the end of the Tertiary period a large lacustrine basin formed on the southern side of the Qargaf uplift. The preserved lacustrine deposits are only a few metres thick. They close a very long cycle (megacycle) of erosion which produced a highly mature landscape, which is generaly recognised as a pediplane.

Sedimentation continued during the Pleistocene. This process increased significantly during the pluvial Quaternary periods, where large rivers existed in those times (Wadi Drwan, Wadi Tanozzuff, Wadi Tarat, Wadi Antahurnqian etc.). The escarpment of the Msak Mastafat was formed by action of rivers which flowed towards the Taita and Awbari depressions.

The succeeding cycle of erosion is characterized by strong eolian activity both in terms of erosion and sedimentation; the sand dunes of the Idhan Awbari and Idhan Murzuq were formed at that time. The most impressive product of this wind and water erosion is Wadi ash Shati, which is a distinct endeheric depression expending in an E-W direction.

A new cycle of erosion started after the deposition of the Al Mahurzak formation. In the more humid periods, the relief was modelled by the deepening of wadis which issued from the Qargaf uplift. In the drier periods mainly escarpments and flat-topped hills were formed. Deflation was responsible for the removal of the lower Quaternary deposits and thus for a gradual deepening, which totalled about 120 m. A gradual deepening of the depression was interrupted by periods of lacustrine sedimentation.

After the deposition of the Tertiary - Quaternary sequence, two or three cycles of erosion and deposition took place, during which the Quaternary and recent deposits were laid down: slope deposits, lacustrine deposits, old and recent wadi deposits, fluvio-eolian deposits and sabha deposits (Pachur, 1980). Whichever type of deposit was formed depended largely on the prevailing climate. In the last stage of desertification, the large depressions were covered by eolian sands belonging to the Idhan Awbari and Idhan Murzuq sand seas.

Appendix 2

ENVIRONMENTAL AND CLIMATIC CHANGES IN AFRICA IN THE LATE PLEISTOCENE AND HOLOCENE

1. NORTH AFRICA

The late - Pleistocene and Holocene environmental and climatic fluctuations in Africa may be correlated with the high and middle - latitude glaciations. The Pleistocene Ice Age started ca 3,000,000 B.P. The formation of major ice sheets over North America and northern Europe began at a later, as yet uncertain date. In North America there were at least four major glaciations; in Northern Europe at least five; in the Alps six or seven. The glacial periods were interrupted by temperate or warm interglacials. The last glacial (Wisconsin - Wurm) began ca 75,000 B.P. and terminated ca 10,300 B.P., depending upon latitude. This period was interrupted by a long interstadial with partial deglaciation and repeated climatic oscillation, which were intermediate between those of the glacial and interglacial norms. The last glacial was preceded by last interglacial dated ca 125,000 ÷ 75,000 B.P.

Within the past 20,000 years, major climatic changes occurred. The glacial maximum was at about 18,000 years B.P. and the climatic optimum on about 6,000 years B.P. (Fig.56). Since the beginning of this century, it became generally accepted that a tropical pluvial, a wet period, correlates with a high-latitude glaciation or Ice Age. However, recent research (Ragnon and Williams, 1977; Street and Grove, 1976; Williams, 1985) has modified this hypothesis, and confirm that, at the peak of the last glacial, at ca 18,000 B.P. tropical Africa was experiencing. However, Street and Grove (1976) suggest that the maximum of African aridity was between 15,000 and 12,500 B.P. rather than at 18,000 B.P. They argue that the glacial phases correspond with colder ocean waters, which simply means less precipitation i.e. aridity is a function of lack of rainfall and is not temperature controlled. The pluvial phase corresponds with a time of high solar radiation which occurred about 10,000 B.P.

According to Nicholson and Flohn (1980), the three major episodes occurred in the climate of Africa during the past 20,000 years:

* a period of activity and dune building ca 20,000 ÷ 12,000 B.P. in which Sahara advanced considerably southward,
* a first moist, lacustrine period ca 10,000 ÷ 8,000 B.P.,
* a second moist, lacustrine period toward ca 6,500 ÷ 4,500 B.P.

On the basis of their C-14 dating, Geyh and Jakiel (1974) concluded that the late Pleistocene and early Holocene were determined by a humid periods, and were followed by a relatively arid interval between 11,700 and 10,500 B.P.

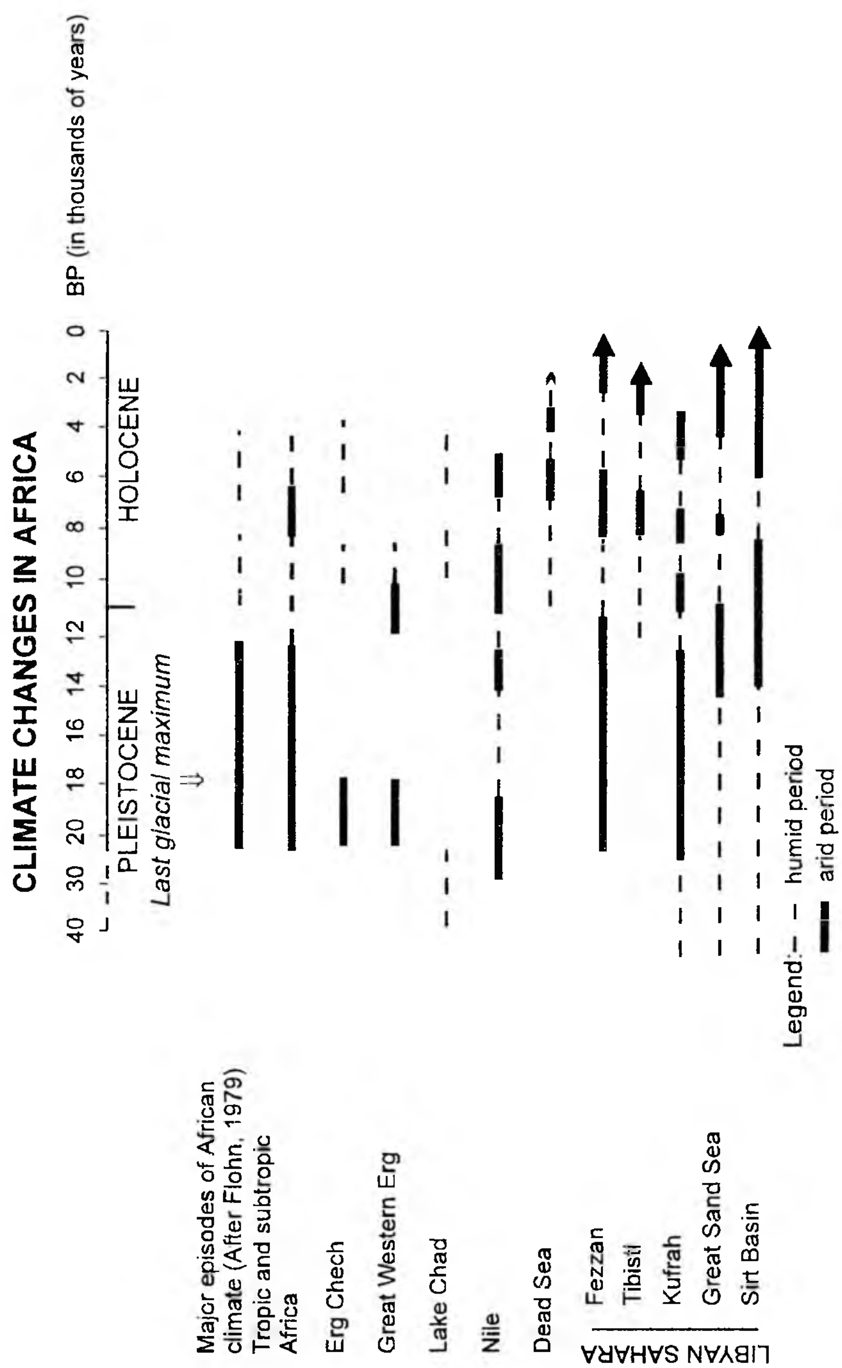


Fig.56 Climate changes in Africa in the late Pleistocene and Holocene (based on data cited in App.2).

Thereafter, a remarkable humid phase with good conditions for vegetational growth prevailed around 8,700 B.P. Between 7,100 B.P. and 6,000 B.P. an arid phase prevailed. Around 6,000 B.P. another change to humid phase commenced which lasted until ca 4,700 B.P.

Tropical and subtropical aridity marked the period between 20,000 and 12,000 B.P., which roughly corresponds with the last glacial maximum (Wisconsin or Weichsel). The maximum aridity was probably achieved between 18,000 and 14,000 B.P. The predominant feature of this episode was the formation and expansion of sand dunes along the margin of the present Sahara. The dunes extended to about 14°N in the East and 10°N in the West; thus, the Sahara was expanded by 200÷500 km along its entire east - west extent. Sarnthein (1978) estimated that, at that time, the sand dunes covered 50% of the land between 30°N and 30°S, whereas, today, only about 10% of this area is covered by active sand deserts. The existence of dune systems is very well documented in Senegal, Mauritania and Chad, where the dunal horizons are bracketed by lacustrine deposits dated at 22,000 and 12,000 B.P.

During the intertropical cold dry phase from 20,000 to 12,500 B.P., the aggrading Nile was a braided, seasonal river (Adamson *et al.*, 1980). The two major lakes in the upper White Nile basin, Lake Victoria and Lake Mobutu (Lake Albert) were without outlet in the period from 25,000 to 18,000 B.P and from 14,000 to 12,500 B.P. The glacial maximum coincides with the minimum level of Lake Mobutu (25,000 ÷ 18,000 B.P.), but the overflow of lake Mobutu between 18,000 and 14,000 B.P. suggests that glacial aridity in this region was less prolonged than in Ethiopia. The flow of Nile, like that of the other great rivers of Africa, became very reduced during the low-precipitation, arid phases of the Pleistocene. In Central Sudan, the source - bounding dunes were formed downwind of the Nile distributory channels, and aeolian reworking of sandy the alluvium of Kordoofan and Darfur has been recognized (Warren, 1970). In the Chad basin, lake levels were high until about 20,000 B.P., but very low thereafter. In Nubia and Egypt, the wadis were intermittently active during the period 17,500 and 12,000 B.P. The floods of both the White and Blue Nile are well documented for the period between 12,500 and 11,000 B.P. (Adamson *et al.*, 1980). The marked increase of aeolian dust which was blown into the Atlantic from the Sahara, and the Chad basin between 25,000 and 18,000 B.P. also reflects the aridity of this period.

During most of the late Pleistocene, ca 20,000 ÷ 14,000 B.P. relatively wet conditions prevailed over most of North Africa, with the exception of northern Egypt and Libya. In the Nile Valley, close to the Wadi Halfa, the maximum dune-forming activity lasted from more than 19,000 to 17,800 B.P. and/or 16,500 B.P. (Sarnthein, 1978). At the peak of the high-latitude glacial, towards 18,000 B.P., a brief arid interval occurred in some of the northwestern and northeastern regions. Dunes were formed then in the Great Western Erg, in the

Erg Chech in Algeria, in southern Egypt and in Cyrenaica. The arid interval responsible for dune forming, may have been limited to two or three millennia in North Africa and for six to eight millennia in the southern Sahara. The major phase of the Great Western Erg, was identified as between 19,800÷17,500 B.P. (Sarnthein, 1978). The Erg Chech activity ended at 17,700 B.P. Increasing aridity prevailed over most of North Africa between 12,000 and 10,000 B.P.

In the early Holocene, a lacustrine episode began; this reached and has a maximum between 10,000 and 8,000 years ago. In the mid-latitudes and tropics, the end of the last glacial period was marked by a great increase of rainfall which, some believe, corresponds to a peak in solar radiation which occurred around 10,000 B.P. The increased radiation would have led to increased evaporation from the sea surface, and thus increased rainfall. In the equatorial regions, evidence of high solar radiation and high rainfall during the early Holocene is apparent in the record of the Nile sediments. There are many traces of this lacustrine period throughout the southern Sahara, and the Sahel. During the peak of the lacustrine episode, between 10,000 and 8,000 years ago, many lakes were deeper and more expansive than at the present time. For example: Lake Chad stood 38 m above its present level (Mabbutt, 1979). Numerous radiometric data have established the peak of the Niger/Chad lake phase as 9,000 to 8,000 B.P. The peak of the pluvial period corresponds with the largest extention of Mega-Chad, as marked by the 320 m level (Grove and Warren, 1968). At Nabta, three Holocene lake phases are dated of 9,000, 8,600 and 7,500 B.P. (Wendorf and Schield, 1980). In Sudan, the White Nile floods occurred around 8,400÷8,100 B.P. and 7,000 B.P. and the Blue Nile flood around 7,500 and 6,900 B.P. As a result of these floods, the dunes of central Sudan were partly submerged beneath Blue and White Nile alluvia, and the dunes of Kordofan and Darfur were fixed by vegetation (Warren, 1970). A very high runoff from Senegal existed about 11,000 and 8,000 B.P, and a very high Niger discharge is dated from 13,000 to 11,800 B.P. In south Ghana (Lake Bosumtwi) the water level was very high between 10,000 and 8,300 B.P. and again from 7,000 to 5,000 B.P., indirectly confirming a dry episode around 7,000 B.P. The expansion of monsoonal rains during the early Holocene in the tropical latitudes permitted an extensive spread of moist savanna - type vegetation over the Sahara in North Africa. Throughout the early Holocene the Dead Sea shows a record of sedimentation from a humid headwater area.

The Saharan highlands of Tibisti, Taisili and Hoggar experienced lacustrine phases in the period of high level and expantion of lakes along the southern fringes of the Sahara. The rivers on these massifs, which had previously carried coarse sand and gravel, began depositing finer grained and well sorted alluvia which are indicative of the presence of a dense vegetation cover (Nicholson and Flohn, 1980).

Rongon (1980), reasoned that whereas increased aridity prevailed in Northwest Africa during the lacustrine episode, eastern Algeria, Tunisia and

Libya were relatively humid. There is a considerable evidence for a dry phase in Northwest Africa. At Laghout, aeolian sands were deposited from about 12,500 to 8,000 B.P. The Ouraghla dunes were active between 9,500 and 7,900 B.P. In the Algerian Maghreb, there was a dry episode between 14,000 and 9,000 B.P. The Chotts and Atlas regions of Tunisia were relatively arid between 16,000 and 8,000 B.P. (Nicholson and Flohn, 1980). However, within this arid period, brief humid phases occured, which correspond with the peak of the tropical lake episode: 9,280 B.P. in the Tunisian pediment and 8,500÷7,500 B.P. in the Algerian Maghreb and in Mauretania 7,320 B.P. The Great Western Erg and Erg Chech were apparently inactive after 10,000 B.P.

Between the first humid phase in the early Holocene and the second humid phase of the Neolithic, a brief arid episode towards 7,000 B.P. occurred in most of tropical and subtropical Africa (Geyh and Jakiel,1974). After 7,000 B.P., the climate of the Chad basin became drier, but lakes persisted in basins for some time (Grove and Warren, 1968). The termination of a long arid phase in numerous parts of North Africa is dated ca 6,000 B.P (Geyh and Jakiel, 1974).

The second lacustrine phase occurred in the period from 6,000 to about 4,700 B.P. During this period, increased rainfall prevailed in the semiarid - subtropical - south of the Sahara as well in the tropics further south. At that time, wetter-than-present conditions affected both the temperate and tropical margins of the Sahara, resulting in considerably shrinkage of the desert belt (Nicholson and Flohn, 1980). The climate was much wetter than that of today, with many lakes existing, even in the interdunal depressions. Northwestern Africa and the northern fringes of the Sahara were characterized by rather arid conditions. In Central Sahara and most of North Africa, evidence for this humid lacustrine phase ca 6,500 ÷ 4,500 B.P. is strong. After the arid millennium, around 7,000 B.P., the runoff from the northern part of the Tibisti massif continued to feed playa lakes in the Libyan desert (Pachur, 1980). Between 28°N and 30°N, the lakes in this region, were numerous until about 8,000 B.P. and from 6,000 to 4,500 B.P. In the present hyper-arid areas of Kufra and Tibesti, rainfall at that time amounted to 200 ÷ 400 mm annually (Jakel, 1978). The lakes were surrounded by a mixture of Mediterranean and Sahelian vegetation. In Tassili, near Ghat (Idhan Awbari); semi-arid vegetation prevailed between 5,500 ÷ 4,500 B.P. This period was succeeded by an arid period with deposition of aeolian sands, which marks the present state of the sand desert areas. At the Neolithic, lakes were present in the Erg Chech (6,000 ÷ 3,000 B.P.) and dunes of southern Morocco became inactive around 6,000 B.P.

The differences in the climatic conditions in tropical Africa and North Africa are probably related to the influence of the late Pleistocene ice sheets. The Pleistocene glacials were only relatively moist in mid - latitudes with inverse correlation between temperature and precipitation. In lower latitudes the glacial maxima were dry, the interstadials moist. The aridity of the glacial maxima may be explained by a reduced evaporation over cooler ocean surface waters. Tropical Africa was imposed primarily by the presence of ice masses, while northern Africa was predominantly effected by the changes between the „ice-build-up" circulation to an „ice-melt" circulation which occurred around 18,000 B.P.

The correlation of glacial periods with dune-building phases has been demonstrated both in Australia and in Asia. Wasson (1986) showed that a lacustrine phase began about 50,000 B.P., with oscillations beginning about 30,000 B.P. and continuing, with decreasing amplitude, to about 20,000 B.P. The dune buiding maximum occured between 25,000 and 13,000 B.P., with a peak between 20,000 and 14,000 B.P. In Australia the period of dune building coincides with the main phase of Würm/Wisconsin Glaciation with a peak at around 18,000.

The Thar desert dunes began to format least as early as least 20,000 B.P. and were stable by the middle to late Holocene (Wasson *et al*, 1983). The dunes were certainly growing during the Last Glacial Maximum.

Probably, the earlier glacial phases were characterised by widespread dryness (Goudie, 1983). The evidence from deep sea cores suggests that the last glaciation was comparable with the multiple glaciations which preceded it.

The sediments in the deep sea cores provided important information on aeolian sedimentation phases. The analysis of the sedimentology of the Atlantic cores along West Africa at 12°÷27°N. indicated that around 18,000 B.P., riverborne sediment were completely absent, whereas, the imput of aeolian silt was greatly expanded (Sarnthein and Koopman, 1980).

The Wisconsin major glaciation occurred in at least three phases or stadia, separated by interstadia, during which the ice sheet shrank. These phases caused the climatic variations manifested in Africa. The late Wisconsin stadial in central North America began some 25,000 years ago and reached its maximum some 15,000 to 20,000 B.P. The middle Wisconsin Stadial can be placed close to 50,000 to 40,000 BP, and early Wisconsin Stadial to around 70,000 to 60,000 B.P.

At the beginning of the Holocene, around 10,300 B.P., the continental glaciers had not yet disappeared and most Scandinavia was still under ice. At that time, all of Canada remained glaciated. The world sea level was at least 40 m lower than today. Deglaciation, expressed by a general thinning and retreat of the ice margins, proceeded rather slowly. Deglaciation did not proceed without interruption. For example: the regrowth of the North American glacier is recorded at ca 8,500 B.P. The Scandinavian ice cap has completly melted away by 8,000 B.P and the remnants of the northern Canada ice cap persisted until 5,000 B.P. At this time, the sea closely approached its present level. Glacial fluctuations were clearly reflected in the climate oscillation in Africa. Certainly, the termination of glaciers in Canada coincides closely with the beginning of an arid phase over most of the Sahara.

Following the thermal optimum, when glaciers appear to have been at their minimal, glacial readvances were noted between 4,700 and 4,200 B.P. The second readvance was initiated between 2,900 and 2,400 B.P., with cooler conditions persisting until 300 AD. The third readvance was initiated shortly after AD 1,550 with a maximum glaciation in mountain areas shortly after 1,850 AD. Since 4,700 B.P. these minor glacial stades are commonly grouped and referred to as the „Little Ice Age" or Neoglacial period.

Thus, a correlation of glacial and interglacial phases with arid and pluvial climate conditions is evident. The aridity of tropical and subtropical Africa correlates with the glacial maxima, while the high (pluvial) lake levels in Tropical Africa all coincide with glacier recession. The same applies to the significantly higher Nile discharge as recorded by floodplain aggradation.

Reviews of the climatic changes in Africa have indicated a general correlation between the glacial maximum and aridity, but also local differences, which depend mainly on the latitude and distance from the sea.

2. HUMID AND DRY PERIODS IN THE LIBYAN SAHARA

During the Quaternary, the whole of the Libyan Sahara was subjected to repetetive climatic change, humid and dry periods alternating with other. During late Pleistocene and Holocene, the environmental and climatic changes in the Libyan Sahara, can be reconstructed on the basis of pluvial and arid phases documented in Tibisti region, and in Kufra, Sirt and Murzuq basins. The climatic variations are expressed in the various levels of terraces within the valleys, and in the great variation of water levels of lakes. According to the most-recent investigations, the climatic changes during the last 30,000 years contain three main phases of pluvial conditions:

I	40,000	÷	20,000 B.P.
II	12,000	÷	8,000 B.P.
III	6,000	÷	3,000 B.P.

2.1. The first pluvial phase (40,000 ÷ 20,000 B.P.)

The best evidence for a pluvial regime in the period between 40,000 and 20,000 B.P. is provided by the isotopic dates for ground water in the Fezzan: Traghan and Al'Awanyat (Petit-Maire *et al.*, 1980). In the Sirte basin, the radiocarbon age of groundwater ranges between 35,000 and 14,000 B.P., but it should be emphasised, that a few samples sugest an age of between 19,000 and 14,000 B.P. (Benfield and Wright, 1980).

The first pluvial phase is also confirmed by dates from the Egyptian Sahara, where, in Bahariyah, the high-carbonate lake sediments are dated at about 28,000 B.P. The lakes of Southern Tunisia and Chad are also dated to this period. The second transgression of Lake Chad (level 350 ÷ 400 m) is dated at between 30,000 and 21,500 B.P. (Mabbutt, 1979), whereas the lakes in Southern Tunisia are dated around 25,000÷20,000 B.P. (Rognon, 1980).

2.2. The interpluvial phase (20,000 ÷ 12,000 B.P.)

The lack of evidence for a pluvial regime in the period between 19,000 B.P. and commencement of the second pluvial phase around 12,000 B.P. sugests that this period, was a the long, arid interval, suitable for sand desert development. The various lines of evidence support the notion of a general rise in temperature and/or increase in aridity which started around 14,000 B.P and reached a peak around 11,000 B.P. (Benfield and Wright, 1980).

2.3. The second pluvial phase (12,000 ÷ 8,000 B.P.)

The second, early Holocene, pluvial phase can be very well reconstructed on the datings from Fezzan and Tibisti. In Fezzan (Awbari, Sabha, Ghat and Murzuq) the isotopic dates for ground water span the period between 12,000 and 7,000 B.P. (Petit-Maire et al., 1980). The fresh water sediments in the centre of Sarir Tibisti accumulated between 12,000 and 7,500 B.P. (Pachur, 1980). Pachur assumed that the increased runoff occurred about 8,800 B.P. with a minimum precipitation rate of 200 mm per year. The middle terraces in the Tibisti valleys have the deposits dated at 12,000 and 7,800 B.P. (Pachur, 1980). Also, the bones of large mammals (giraffes, elephants) found in Wadi Oyouroun on the west side of Jabal Nugay, near the Tibisti mountains, are dated at 7,500 B.P. The lake deposits on the Pleistocene Lake in Kufra are dated to 9,000 ÷ 8,000 B.P. (El Ramly, 1980), and the 4th pluvial (level 400 m) can be correlated with the same level of Lake Chad in Niger (Pallas, 1980).

The pluvial phase in the Libyan Sahara corresponds very strongly with the documented dating from desert areas in Egypt. In the Gilf al Kabir, the stillwater sediments are dated at ca 8,500 B.P. and east of the Gilf al Kabir the sediments are dated to 10,100 and 7,800 B.P. (Pachur, 1980).

2.4. The interpluvial phase (8,000 ÷ 6,000 B.P.)

After the Early Holocene second pluvial phase, there was a short, but intense dry interval in the Libyan Sahara around 8,000 ÷ 7,500 B.P. According to Jäkel (1980), this arid phase occurred between 7,300 and 6,000 B.P. in the Tibisti region. He described this period as arid by comparison with the preceding pluvial phase, but wetter than the present.

2.5. The third pluvial phase (6,000 ÷ 3,500 B.P.)

This pluvial phase is very well documented in many regions of the Libyan sand seas. The radiocarbon ages of ground waters in the Kufrah and Sirte basins span the period between 8,000 and 5,000 B.P. (Benfield and Wright, 1980). In Fezzan (Ghudwah, Umh Al Aranib) the isotopic date of ground water is around 3,000 B.P. (Petit-Maire, 1980). According to Benfield and Wright (1980), the maximum rainfall in the Sirte basin was ca 5,500 B.P. There is a considerable evidence for this pluvial phase from the Tibisti mountains and

Sarir Tibisti. The deposition of thelower terraces in the Tibisti mountains took place between 6,000 and 3,000 B.P. (Hagedorn and Pachur, 1980). The second stage of the freshwater lake in Sarir Tibisti is dated to 5,000 B.P. (Pachur, 1980). The sabkhas were formed about 5,000 B.P. The bones of large mammals found in Wadi Oyourum on the west side of Jabal Nugay were dated to 5,100 B.P. (Pachur, 1980). In the youngest pluvial deposits in Wadi Behar Belama, where found there and dated to 3,400 B.P. The dunes in the Yebique valley in Tibisti, which have a vegetationed cover with calcicrusted roots dating to 3,400 B.P. provide further evidence for this phase. Radiocarbon dating of the prehistoric relicts of a Neolithic shepherd indicate that the central Sahara plains had an extensive vegetation cover between 5,700 and 4,100 B.P. (Rognon, 1980).

2.6. The present arid phase

2,500 ÷ 2,000 B.P. is generally assumed to be the end of the pluvial phase. In the vicinity of Tibisti, the lakes finally dried as around 1,900 B.P. as shown by cessation of growth of tamarisks on the lake floors.

It is generally assumed that there have been no significant pluvial episodes since 3,500 B.P. (Benfield and Wright, 1980). Geyh and Jakel (1974) pointed out that, the humid climate was succeeded by arid conditions during the period of 4,700 B.P. and 3,700 B.P. Geyh and Jäkel (1974) and El Ramly (1980) belive that, until 1,000 B.P., there were regular alternations of humid and arid phases each of duration 700 ÷ 800 years.

Appendix 3

WIND CIRCULATION IN NORTHERN AFRICA

1. NORTH AFRICAN CIRCULATION SYSTEM

1.1. Seasonal changes of wind circulation

Wind circulation in Northern Africa has the most significant influence on the development of sand dune forms and patterns which appear in the Sahara desert. The wind circulation in Libya constitutes a part of the North African circulation system, which is characterized by seasonal changes.

Surface wind circulation in North Africa, including Libya, has to be evaluated in for seasons (Griffiths and Soliman, 1972):

- winter season: December ÷ February,
- spring season: March ÷ May,
- summer season: June ÷ September,
- autumn season: October ÷ November.

The pattern of winds in the Sahara depends upon the yearly oscillation of the zone of low pressure called the *Intertropical Convergency Zone* (ITCZ), which follows the apparent movement of the sun with a time lag of six weeks to two months (Dubief, 1979). This consists of a string of low pressure centres. The position, extension, and depth of this zone changes over the yearly cycle. In winter, the ITCZ zone is situated along the latitude of the Gulf of Guinea in the west whereas, to the east it is at a somewhat lower latitude. In summer, the ITCZ comes in contact with the whole of the Sahara desert.

1.2. Winter season

In the winter season, the following factors play a significant role in wind circulation:

* the belt of subtropical high pressure cells extending from the Azores to southern Asia over the northern Sahara and Arabia,
* the low pressure area over the central and eastern Mediterranean,
* the equatorial trough over cenrtal Africa,
* the ITCZ (The Intertropical Convergence Zone), and to a lesser extent,
* the Arabian high (as a part of sub-tropical high pressure belt),
* the Balkan high, in conjunction with the great anticyclone over central Asia.

The *high pressure cells* are the main factors governing wind circulation in North Africa. During winter, the Azoran high (1024 mb) is at its nearest to the African coast; simultaneously, the Saharan high is strongest and the ITCZ is situated in the vicinity of the Equator. Wind circulation in the Western Sahara, between 14°N and 26°N and 6°W and 19°W, is controlled by the Azora and

Saharan pressure cells, and by ITCZ. Generally, these are the moist northerly and northeasterly *trade winds*.

During the winter season the greater part of Northern Africa is influenced by large - scale subsidence caused by the *Sahara high* (1020 mb), a condition responsible for dynamically-heated, desiccated air. The Sahara high is generally centered over Libya, but sometimes it occupies a more westerly position (27°N, 5°W). The furthest eastermost position is on the border between Libya and Egypt. The surface winds blow counterclockwise, out from the centre of high pressure cells, thereby creating the pattern of wind circulation. Southward of the line connecting the high pressure cells (Azores, Sahara and Arabia), the outward flow results in the windflow from the NE and E, whereas northward of this line, the winds generally blow from the SW and W. The resultant winds are rather weak and increase with decreasing latitude. In some countries (Nigeria, Sudan) this wind is called *the harmattan.* Due to the clockwise out flow around the high pressure over Libya, surface winds blow in various directions in winter, but wind velocities are rather low there. Calm periods prevail for as much as a quarter of the time.

In the winter season, the *the moving Mediterranean depressions* represent another major fact influencing wind circulation. During the winter, the Mediterranean Sea is warmer than the northern coast of Africa so it supplies a huge amount of vapour to the air and creates strong wind storms. Some of these depressions are single-centred whereas others are complex. In front of these depressions, S and SW winds blow across northern Libya. The depressions persist for an average of two or three days over the middle and eastern Mediterranean. When these depressions are deep, the SW winds occuring in the front of these depressions may reach a gale force, thus causing strong sandstorms. The NW winds at the rear of these depressions may also have a strong force, especially in the northern part of the Sahara, but, generally, they are weaker than the winds at the front of the depression.

In the winter the eastern part of Northern Africa is influenced in the winter time by the *Siberian anticyclone,* which extends westwards over the Balkans.

1.3. Spring season

The main meteorological characteristics of the spring season are :

- southward shift of the depression centres,
- passage of the cold fronts from west to east.

In the spring, a southward shift of the tracks of the Mediterranean depressions may take place. The centres of the depressions move west to east along the the coast line of North Africa or even further south, where they are known as desert depressions. The storm winds preceeding these depressions are called *Khamsin.*

Counterclockwise circulation around the lows, results in southwest winds (between 20°N and 30°N) or in southeast winds in the northern part of the Sahara (between 30°N and 34°N). The circulation around the lows results in SE winds in western Libya and in NE winds over south - eastern Libya.

The second meteorological event is the passage of cold fronts from west to east across the northern desert, thereby creating strong winds. Winds in advance of the front are from the SW and those behind the front from the NW. The winds in the advance appear as little as only a few hours ahead. Due to the fact that a large part of the Libyan Sahara is situated in a zone between the high and low pressure cells in spring, this season is characterised by the highest drift potential.

1.4. Summer season

The summer season, which starts at the beginning of June and terminates at the end of September, generally presents the inverse of the winter situation. The pressure regime and wind circulation are mainly influenced by:

- high pressure ridges over central Africa,
- thermal lows over southern Sahara,
- thermal lows over the Arabian Peninsula.

The Azores high, which is stronger in the summer than it is in winter, is located farthest from the African coast (approx.30 N and 32 W). This high expands over part of the European continent and gives rise to northerly winds on its southeastern flanks. These winds, which were called the *Etesian winds* during Classical times, penetrate the Sahara and the Middle East after passing over the eastern part of the Mediterranean. They have the character of maritime trade winds. Being rapidly heated over the continents, they dry out there (Dubief, 1979). The northern Etesian winds turn towards the northeast, later to the east and, finally, show the same character as the harmattan. These winds cause drought in northeastern Africa, mainly in Libya and Egypt. In eastern Libya and in Egypt, these steady Etesian winds (northwestern, northern and northeastern) blow persistently as part of a circulation around the huge Arabian low. In June, the Sahara high is replaced by a ridge of high pressure which extends northwestward into the Mediterranean Sea. Sometimes, the high pressure cell develops over the central part of the Mediterranean. Due to intensive solar heating, a thermal low develops in the central Sahara (approx. 20°N and 0°W). In summer time, the Intertropical Convergence Zone (ITCZ) is shifted northward and oscillates between latitude 17°N and 13°N.

In western Sahara, the northward-shifted position of the Azores high results in a more northerly wind flow. The outward flow into the Central Sahara thermal low, both from Azores high and the high pressure ridge over the Mediterranean, results in the development of *trade winds* over the deserts of Northern Africa. Over Mauretania and the eastern Atlantic, the true trade winds blow from a northerly direction. Over the deserts of Algeria and Libya they have a northeasterly and easterly direction. Over eastern Libya, the thermal trough is developed. During July, August and September, the southern margins of the Sahara are affected mostly by the southwest monsoonal wind, resulting from the cyclonic circulation associated with the Intertropical Convergence Zone (ITCZ).

1.5. Autumn season

The autumn season is the other transitional season. In late October, the desert depressions (Khamsin-like) begin to cross the desert from west to east, causing a breakdown of the settled summer regime. The early depressions in September are infrequent and usually occur only in the west and centre of the Sahara. In comparison with spring, the depressions are much less vigorous and slower in their eastward motion. The most vigorous depressions appear in November; these cause a greater frequency of thunderstorms.

From August through to November the thermal low over Asia weakens, and the centre of high pressure near the eastern part of northern Africa shifts from the Mediterranean Sea, southward into Libya. Generally, in this season, the winds are weak and directionally quite variable.

2. ATMOSPHERIC CIRCULATION DURING THE LATE PLEISTOCENE AND HOLOCENE IN NORTH AFRICA

The atmospheric circulation in the Late Pleistocene and Holocene was responsible for the formation of the sand seas and dune patterns; hence, such information is very important in the reconstruction of sand desert morphogenesis and morphochronology.

The presence of the ice sheets in the late Pleistocene and Early Holocene, ca 20,000 and 12,000 B.P., strongly influenced the global thermal variations and hence the atmospheric circulation patterns. According to Nicholson and Flohn (1980), the main changes would have taken the form of displacement as well as the weakening or intensification of the present circulation features, and changes between zonal (east - west) flow and meridian flow. The presence of the northern continental ice sheets caused an increase of the temperature gradient between tropical and temperate latitudes and a greater zonal contrast between land and water. This resulted in an equoterward displacement of circulation features and in the intensification and shrinking of the Hadley cell and associated Subtropical High, and additionally should also have caused strong westerlies. The massive ice sheets must have acted as barriers to the development of mid - latitude cyclone tracks, displacing them southward of the continental margins.

Nicholson and Flohn (1980) introduced the concept of differentiation between zonal and meridional circulation modes. He suggested that *meridian circulation types dominated during glacial periods* with weaker westerlies and strong north-south components, in boths troughs and cellular elements: cyclones and anticyclones. Basically, during the interglacials, a strong east - west flow was developed with weaker troughs.

The ice sheet, which existed at the late Pleistocene glacial maximum, induced a strongly baroclinic zone along and to the south of their margins, which diplaced the westerly flow south of their present position into North Africa (Nicholson and Flohn, 1980). The increased hemispheric temperature gradient probably caused the displacment of the subtropical high southward, over the Atlantic, and caused an increased subsidence in the subtropical high presure cells. Williams (1985), suggested that, if the marine and continental aerosol inputs during the last glacial (ca 20,000 B.P.) were respectively 5 and 20 times higher than present, the winds would also be stronger. He suggested that wind speeds were 1.3 to 1.6 higher than present winds. Sarnthein (1978), in his analysis of glacial and interglacial wind regimes over the Sahara and the adjacent Atlantic, concluded that the trade winds at 18,000 B.P. were much stronger than these of today, with wind speeds of 20 m/s in contrast to a much weakened Harmatan.

In the period between 10,000 and 8,000 B.P., the European ice sheet had considerably diminished, but much of the Northern American ice sheet still remained. This surely warmed the North Atlantic and decreased the northern hemisphere temperature gradient. These events provoked a weakening and northward displacement of the North Subtropical High. These circumstances developed a quasi-stationary trough over North-America and a secondary European trough. The increased inter-hemispheric thermal contrast caused the northward displacement of the ITCZ. The frequency and intensity of the mobile upper-level troughs of the westerlies should have only slightly diminished with respect to the glacial maximum (ca.18,000 B.P.), and reached somewhat more southerly latitudes than those of the present (Nichlson and Flohn, 1980). The northward advance of the ITCZ and re-establishment of the SW monsoons, caused more frequent development of the Saharan depressions and storms, and their persistence through most of the year over North Africa.

The northern hemisphere reached a thermal maximum around 6,000 B.P. at a time when the southern hemisphere was cooling. At that time, the European ice sheet disappeared and the North American (Laurentide) ice sheet diminished. These events caused a gradual warming of the northern hemisphere, a decrease of the temperature gradient, a weakening of the subtropical high over the Atlantic and its poleward diplacement. The increased thermal contrast between the hemispheres, (thermal maximum in the northern hemisphere and cooling on the southern) resulted in the northward displacement of the ITCZ and, hence, the summer monsoon would have reached northward into the southern Sahara.

Appendix 4

SAND MOVING WINDS IN THE LIBYAN SAHARA

1. METEOROLOGICAL DATA

The general rules governing the wind regime and its seasonal changes in Northern Africa were discussed in the App. 3. Additional a row meteorological data relating to the Libyan territory, which were collected by the Author from the Meteorological Department in the Ministry of Communication in Tripoli, constituted the basic material for a more detailed analysis of wind regime in the region of the Libyan sand seas. The data assembled refer to the following meteorological stations, situated in the vicinity of sand seas in Libya: Ghadamis, Hun, Jalu, Al-Jaghub, Ghat, Sabha, Tazirbu and Al Kufrah. The location of meteorological stations in the vicinity of sand seas is shown on Fig.5.

The data include meteorological records of the percentage frequency of wind direction and speed. These data refer to the years 1967 ÷ 1984. The wind velocities were registered in accordance with wind velocity classification of WMO data in 11 classes: 1 ÷ 3 knots, 4 ÷ 6 knots, 7 ÷ 10 knots, 11 ÷ 16 knots, 17 ÷ 21 knots, 22 ÷ 27 knots, 28 ÷ 33 knots, 34 ÷ 40 knots, 48 ÷ 50 knots and wind with speed higher than 56 knots. The wind directions were listed in 16 directional categories.

The data were used for computation of sand moving winds which may be characterised by *drift potential*, *resultant drift potential* and *resultant drift direction*. The obtained results were graphically presented in the form of so called *sand roses*.

2. DRIFT POTENTIAL, RESULTANT DRIFT DIRECTION AND RESULTANT DRIFT POTENTIAL

The relative quantities of potential sand movement can be determined by the parameters introduced by Fryberger (1979), namely *drift potential, resultant drift direction and resultant drift potential.* The *drift potential (DP)* is a measure of the relative amount of potential sand drift at a station for a stated period of time, and can be determined for various compass directions. The *resultant drift direction* (RDD) expresses the direction of the net trend of sand drift, while the *resultant drift potenital* (RDP) expresses the magnitude of the net sand drift potential along the resultant drift direction. For convenience, the units of sand drift potential were named by Fryberger (1979), as *vector units* (VU), because the wind velocites can be regarded as vectors.

Fryberger (1979) modified Lettau's equation for the determination of relative amount of sand migration called *sand drift*. Lettau's generalised equation for the rate of sand transport (q) (Fryberger, 1979), can be presented as follows:

$$q = v_*^2 (v_* - v_{*t}) ;$$

where:
v_* - shear velocity;
v_{*t} - threshold shear velocity.

Assuming that the shear velocity is proportional to the wind velocity for a given height (Belly, 1964), the equation can be expressed as:

$$q = v^{*2} (v^* - v_t^*) ;$$

where:
v^* - wind velocity at 10 m height (the standard WMO anemometer height);
v_t^* - threshold wind velocity at 10 m, which is needed to keep sand in saltation.

Transformation of Latteau's equation is made with the assumption that the shear velocity is proportional to the wind velocity for a given height (Belly, 1964). Hence, the wind velocities measured at a 10 m height, the standard WMO anemometer height, can be substituted for shear velocities. Using Bagnold's equation, and applying the figures determined by Belly (1964), a value of 11.6 knots was obtained for v_t^* This value is related for sand to 0.30 mm average grain diameter.

Following Fryberger (1979), an assumption was made that wind speed and direction component occur in nature for a certain amount of time which is proportional to its percentage in the meteorological summaries, which average - out the very many observations taken over 17 year period. Lettau's equation, modified by Fryberger (1979), expresses the relative amount of sand which is potentially moved by the wind during a certain period of time. Hence, the proportionate amount of sand drift (Q), equivalent to sand drift potential (DP), may be expressed as :

$$Q = DP = v^{*2} (v^* - v_t^*)\, t;$$

where :
t - time the wind blew (expressed as a percentage in N summaries);
$v^{*2}(v^* - v_t^*)$ - weighting factor.

For evaluation of the relative amounts of sand drift, the weighting factors should be computed for each velocity category. For computation of v^*, the mean velocity of the winds in each category was taken, and for v_t^* a value of 12 knots, rather than 11.6 knots, was chosen. The relative amount of the sand drift was computed by multiplying the weighting factor by the percentage of wind in a particular velocity category for all 16 directions.

The meteorological data obtained from the Meteorological Centre in Tripoli were listed separately for particular months; hence the sand drift potential was calculated both for particular months and for an annual average. The results obtained were used for the evaluation of the sand drift potential at each station, during the period of the recorded meteorological data and for the analysis of seasonal (monthly) changes of sand drift potential. Such data are of the highest importance in the studies of sand dune development.

3. SAND ROSES OF THE LIBYAN SAHARA

The sand roses are the most convenient method for the expression of sand drift potential, and are calculated for 16 compass directions. The sand roses present graphically the amount of sand drift potential, as well as the directional variability. The arms on the sand roses, oriented towards the circles centre, are proportional in length to the sand drift for a given direction.

The resultant drift direction (RDD) was also computed and presented on the sand rose in vector form. The resultant drift direction, expresses the direction in which sand movement will trend as a resultant of the winds from all directions. The resultant wind potential expresses the net sand potential in a given period.

The *directional variability* of the effective winds, can be expressed by the *index of the directional variability* (IDV), presenting the ratio of the resultant drift potential (RDP) to the drift potential (DP). The ratio of the *resultant sandflow* (resultant drift potential) *total potential sandflow* (drift potential) was named by Wilson (1971) as *unidirectionaliy index*. For small directional variability, IDV is higher and for greater directional variability, lower.

On Fig.57, the annual sand roses for all the meteorological stations used in these investigations are presented. On this figure, the values of drift potential, resultant drift potential and for the index of directional variability are also given. The highest annual drift potentials are recored at Sabha (DP = 655 VU) and at Ghadamis (DP = 588 VU); these stations are located in very high wind energy zones (greater than 500 VU). The sand roses for those stations indicate the complex wind regime or a multidirectional wind prevalence. Drift potentials calculated for meteorological stations in Al Kufrah (DP = 468 VU), Jalu (DP = 414 VU), can be classified as zoness of high wind energy (between 400 VU and 500 VU). Stations in Hun (DP = 362 VU) and Al Jaghbub (DP = 270) belong to the medium energy wind environment zone (200 VU ÷ 400 VU), while the lowest wind energy zone is represented by the stations in Tazirbo (DP = 86 VU) and in Ghat (DP = 116 VU). In term of the directional classification of winds, the winds in Al Kufrah and in Jalu are located in complex wind regimes, while the stations in Hun and Al Jaghbub in an obtuse biomodal category.

Of course, the annual sand roses can give only very limited information on the sand moving winds. For a more detailed analysis, i.e. one which deals with the relationship between sand moving winds and dune forms or patterns, the monthly information has the highest importance. The monthly sand roses were computed and presented in the same manner as were the annual sand roses.

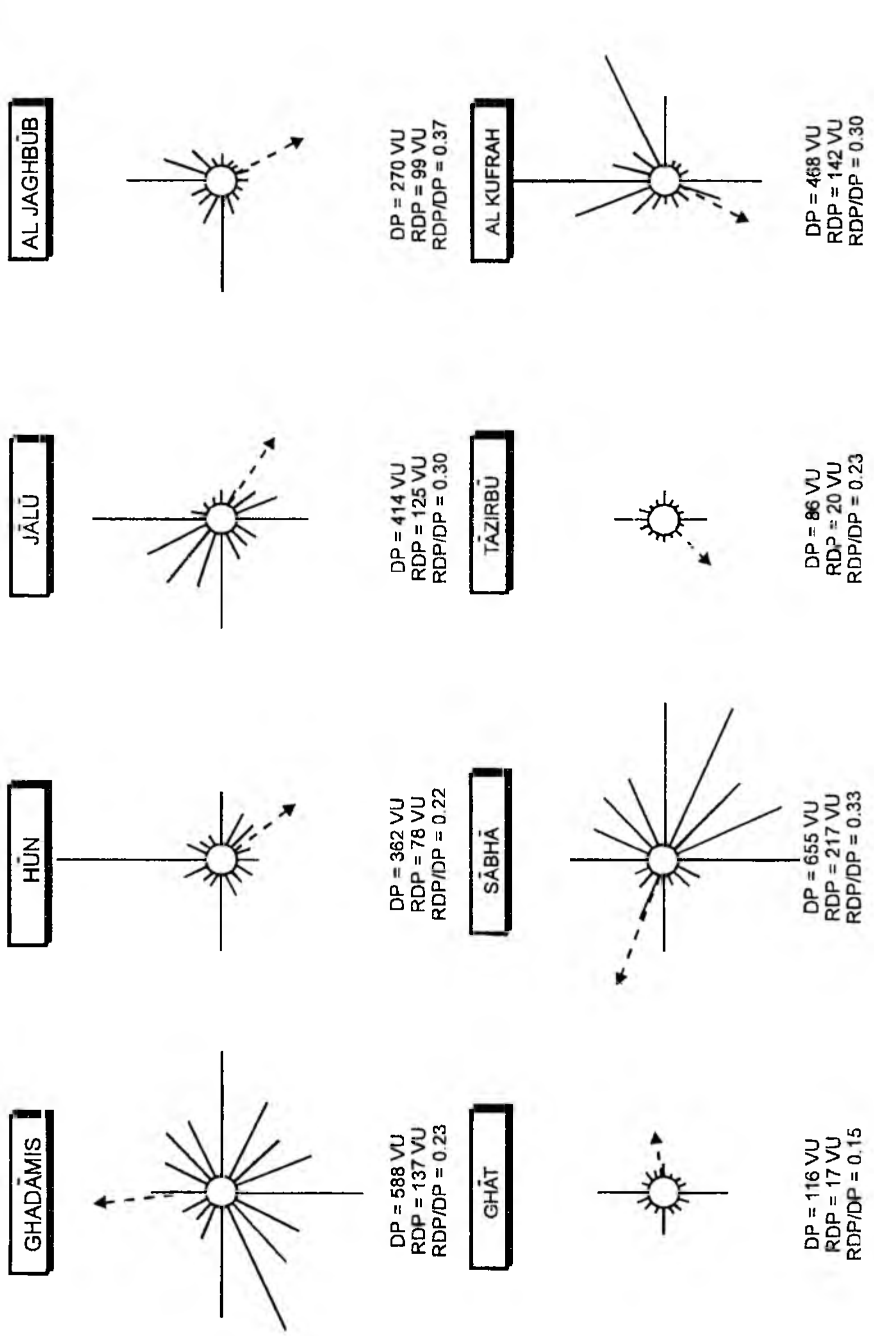


Fig.57 Annual sand roses for meteorological stations in the vicinity of the Libyan Sahara sand seas.

The monthly sand roses for particular stations are depicted on the following figures: Ghadamis - Fig.58a, Sabha - Fig.63a, Ghat - Fig.62a, Hun - Fig.59a. Jalu - Fig.60a, Al Jagbub - Fig.61a, Al Kufrah - Fig.65a and Tazirbu - Fig.64a. It is obvious frome these that there are significant differences beetwen the seasons of the years and between even particular months. Apart from the differences within particular stations, there are are large differences between the stations.

4. MONTHLY VARIATIONS OF SAND DRIFT POTENTIAL

Monthly variations of sand drift potential for meteorological stations in Libya, are presented graphically on Fig.66a and 67a. The highest drift potential occurs in the spring, between March and May, with the peak of sand drifts in April. On figures 66b and 67b the percentage of sand-moving winds recorded by meteorological stations on Libya is shown. The percentage of sand moving winds varies between 10 and 50%. There are significant differences between seasons of the year as well as differences between particular stations. These data correspond very well with the appearance of sandstorm days in the period 1956 ÷ 1985 (Fig.68a and Fig.69a). On these figures, the frequency of days with sandstorms in particular months is depicted. These figures have been combined with the tables which present the average number of sandstorm days per year and the maximum number of sandstorm days in particular months. For the western stations, located in the vicinity of sand seas, the maximum number of sandstorm days was recorded in Ghadamis (11 days in April 1968) and in Hun (10 days in March 1969). The highest average number of sandstorm days also occurs in Ghadamis (16 days) and in Hun (14 days).

For the eastern stations, the maximum number of sandstorm days also occurs in March and April (10 days in April at Jalo and 10 days in April at Al Jaghbub). The average number of sandstorm days within the year is lower in this region and varies between 2.57 in Tazirbu and 12.43 in Al Jaghbub.

A comparison of diagrams presenting the mean number of sand storm days (Fig.68a and 69a) indicates that the western stations have more sandstorm days than the eastern stations and that the climax of sandstorm phenomena in western region is in April-May, whereas, in eastern stations it is in March ÷ April.

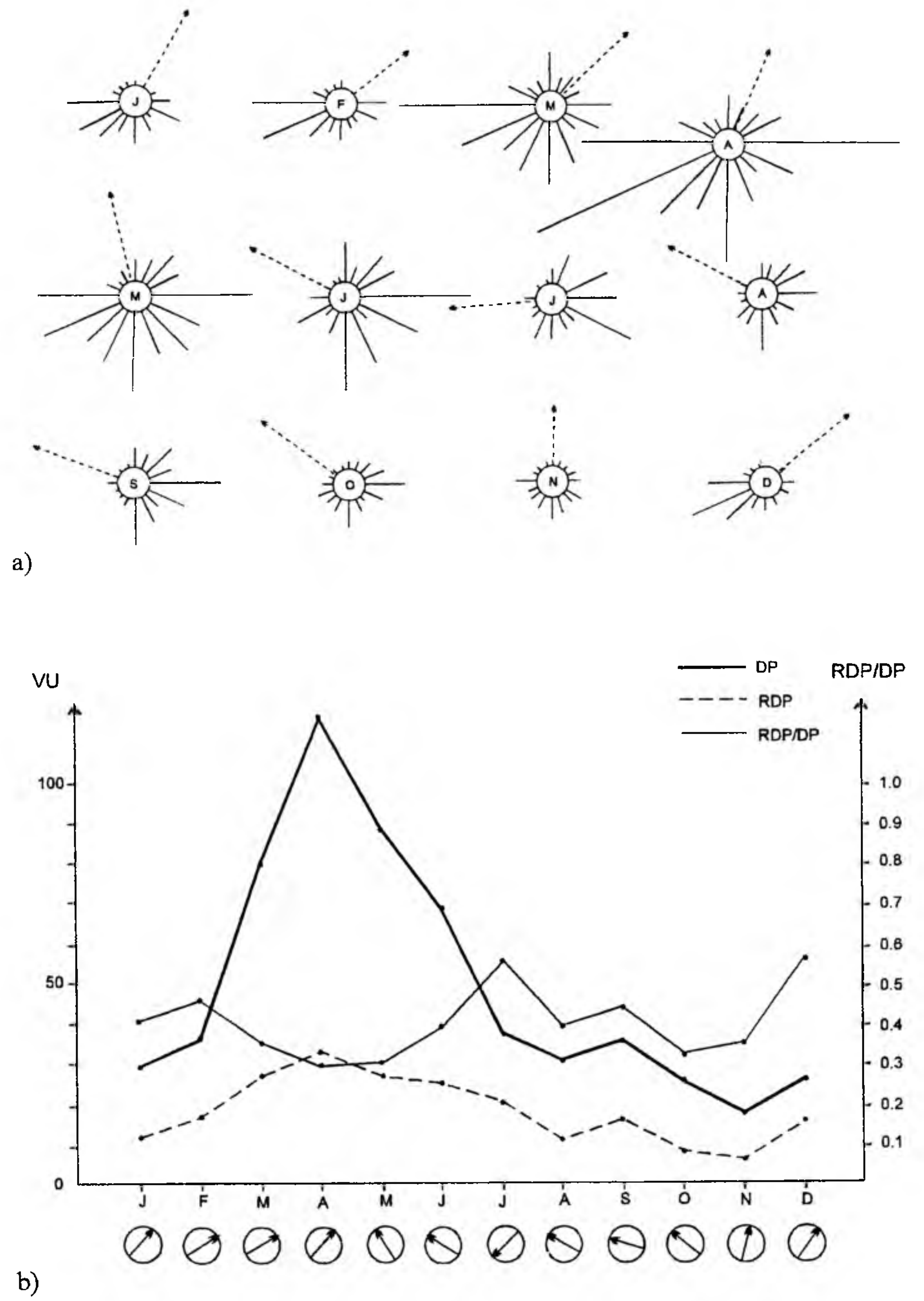
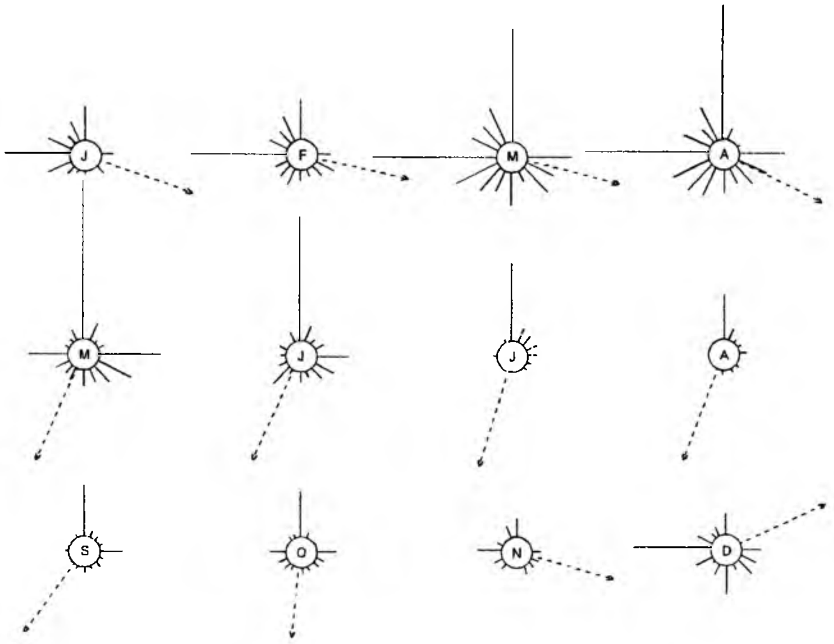
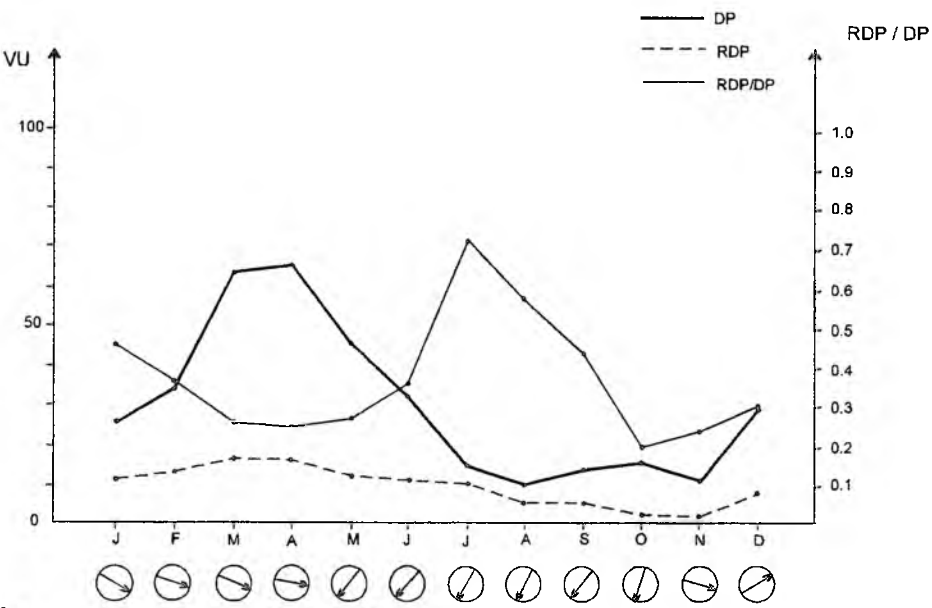


Fig.58 Characteristics of the sand moving winds in Ghadamis:
a) monthly sand roses, b) monthly variations of drift potential (DP), resultant drift potential (RDP) and index of directional variability (RDP/DP).



a)



b)

Fig.59 Characteristics of the sand moving winds in Hun:
a) monthly sand roses, b) monthly variations of drift potential (DP), resultant drift potential (RDP) and index of directional variability (RDP/DP).

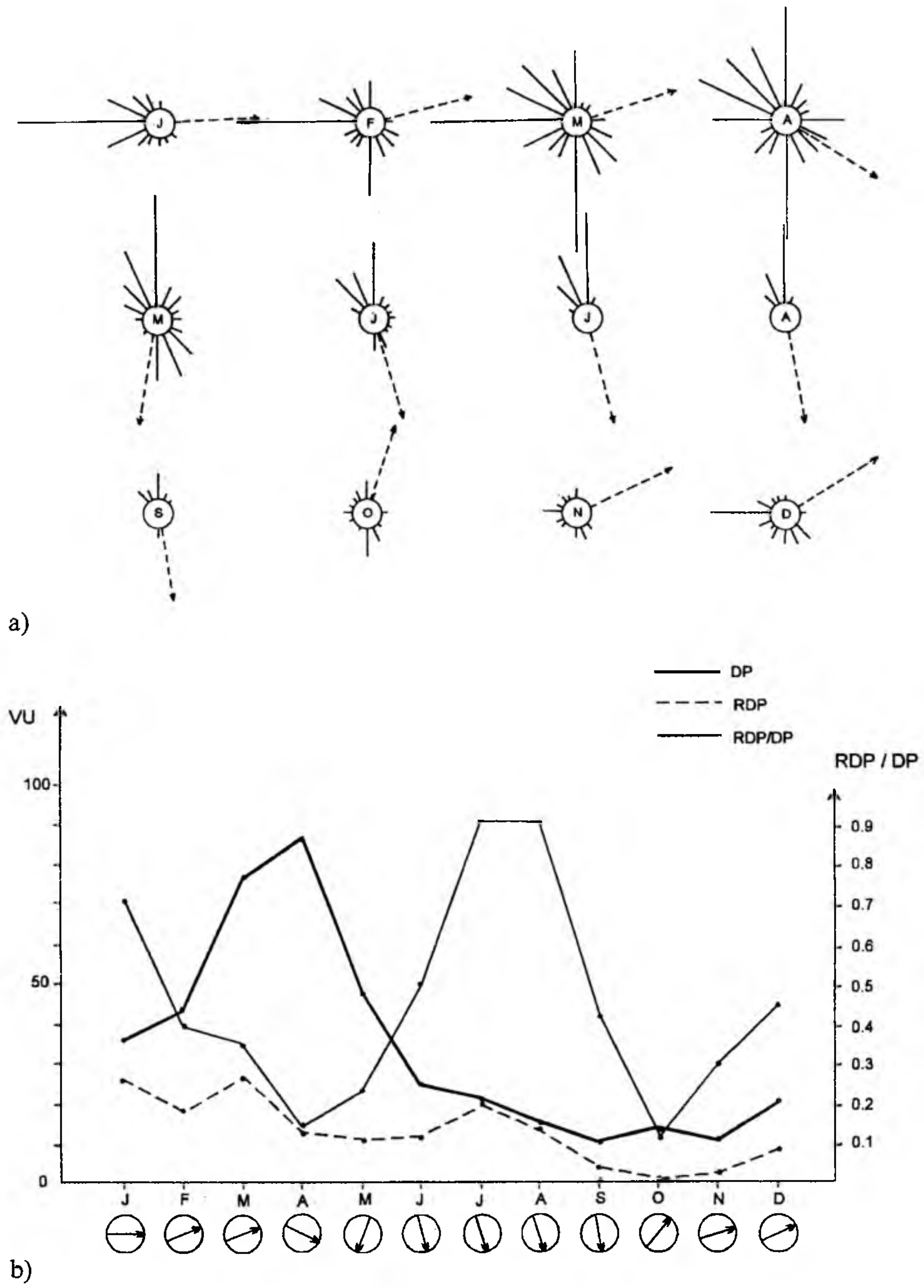


Fig.60 Characteristics of the sand moving winds in Jalu:
a) monthly sand roses, b) monthly variations of drift potential (DP), resultant drift potential (RDP) and index of directional variability (RDP/DP).

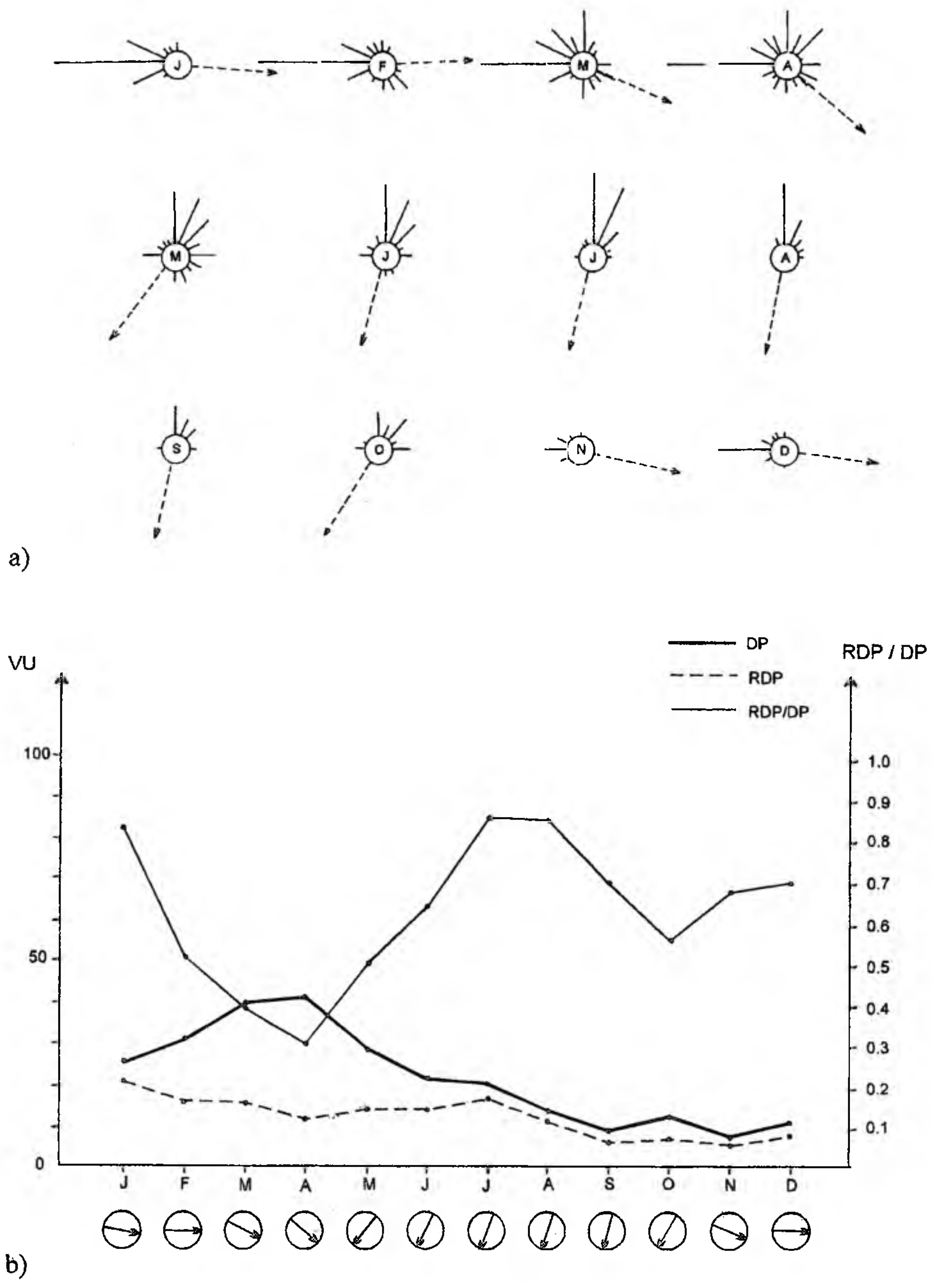


Fig.61 Characteristics of the sand moving winds in Al Jaghbub:
a) monthly sand roses, b) monthly variations of drift potential (DP), resultant drift potential (RDP) and index of directional variability (RDP/DP).

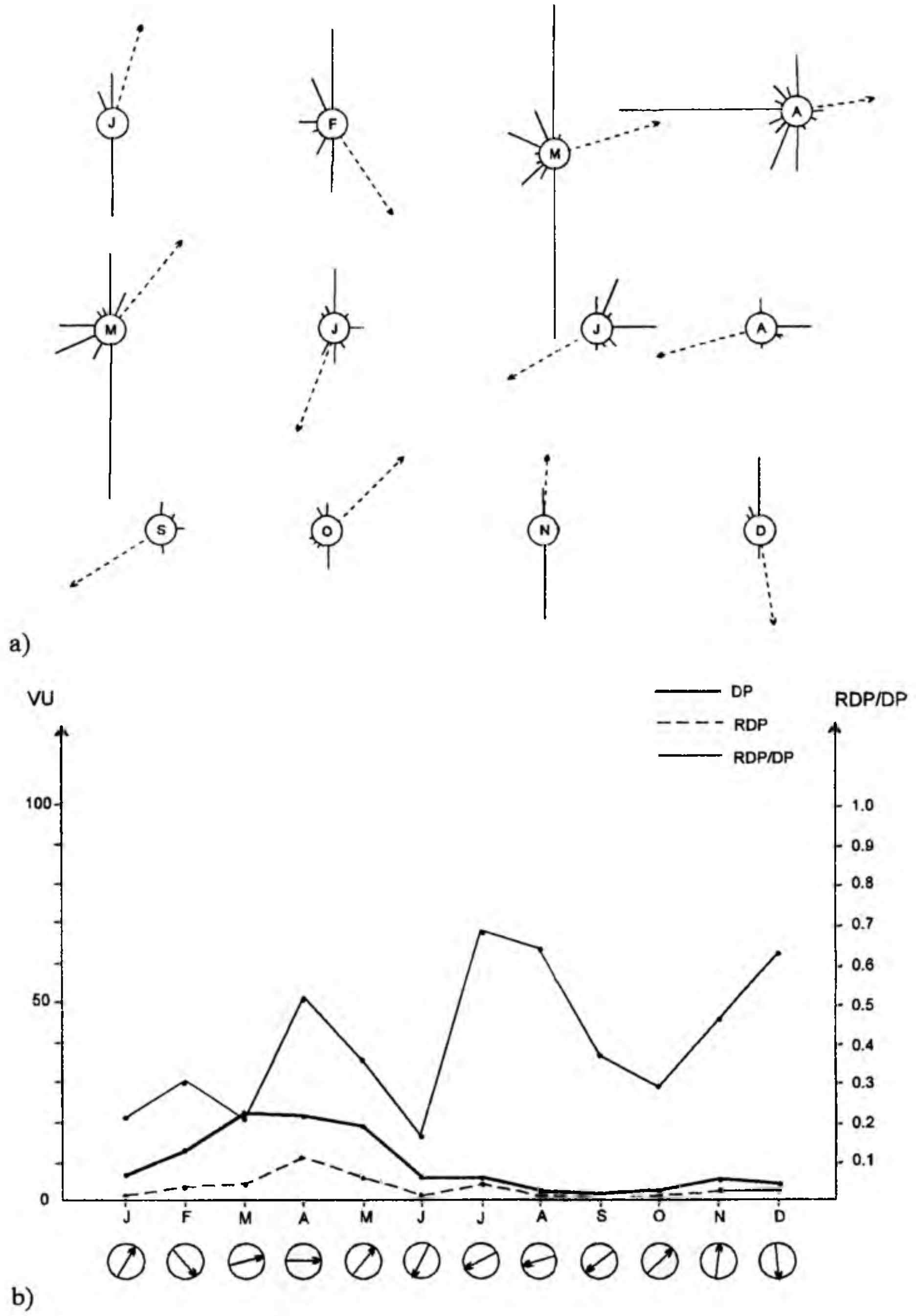


Fig.62 Characterisctics of the sand moving winds in Ghat:
a) monthly sand roses, b) monthly variations of drift potential (DP), resultant drift potential (RDP) and index of directional variability (RDP/DP).

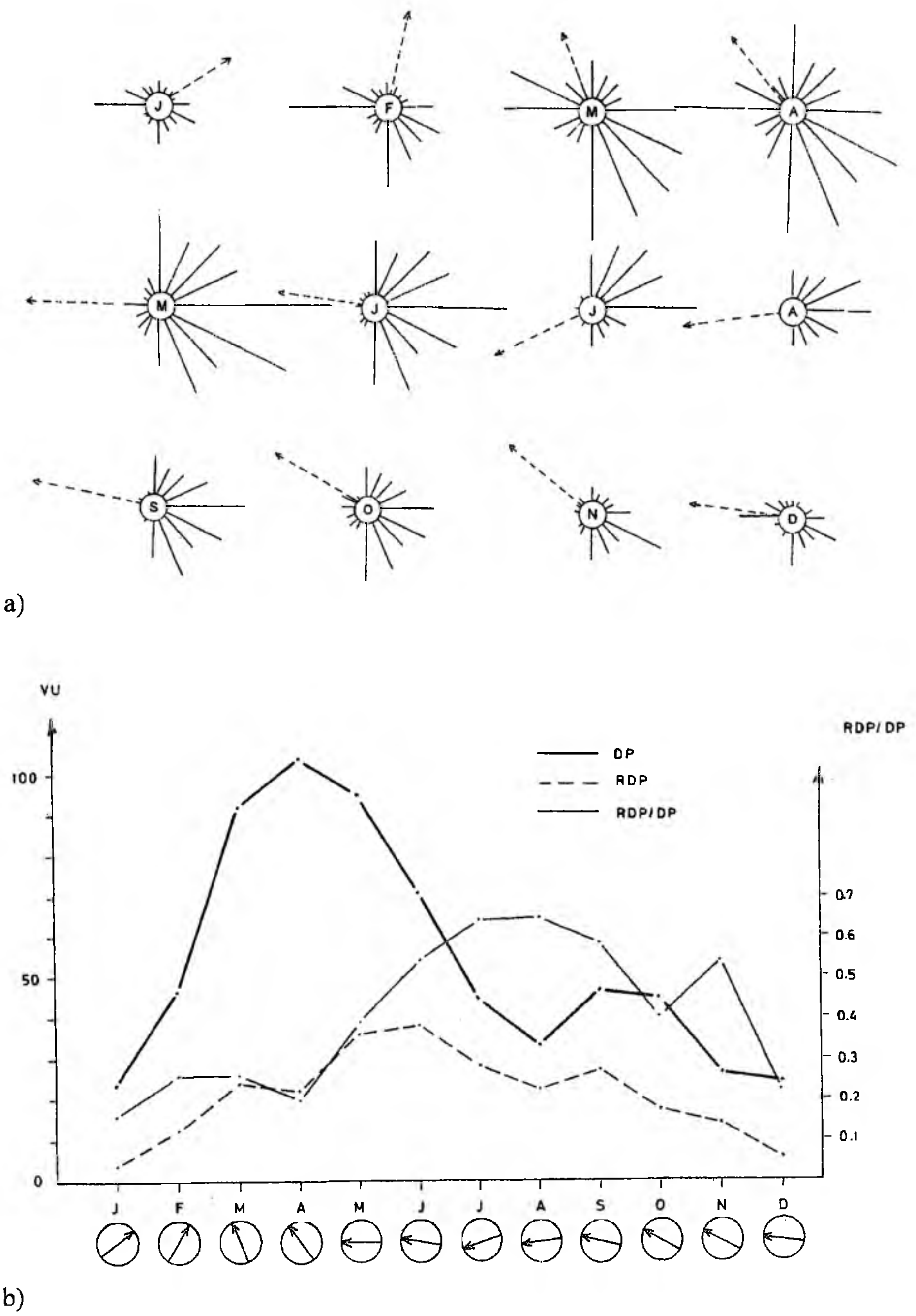


Fig.63 Characteristics of the sand moving winds in Sabha:
a) monthly sand roses, b) monthly variations of drift potential (DP), resultant drift potential (RDP) and index of directional variability (RDP/DP).

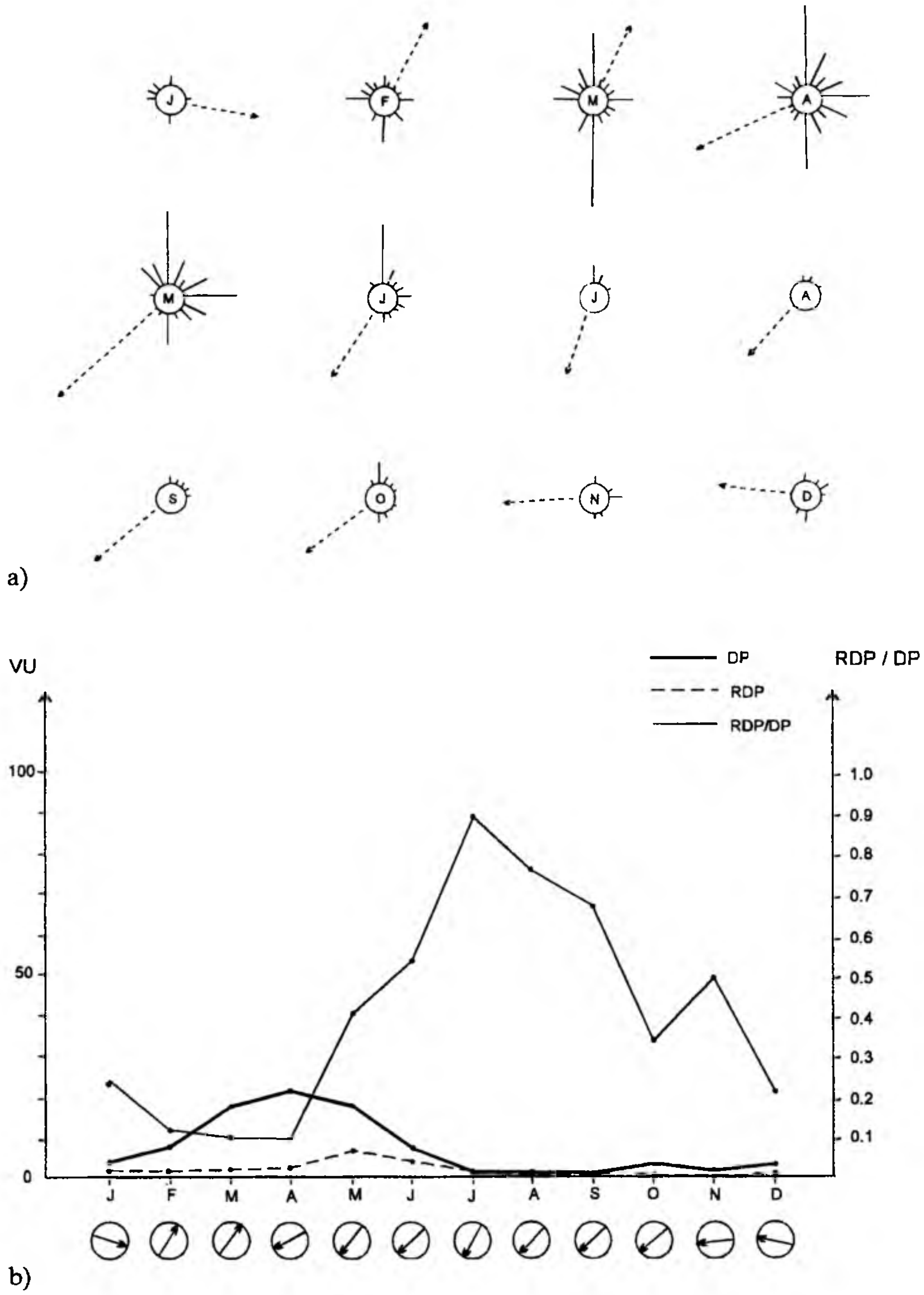
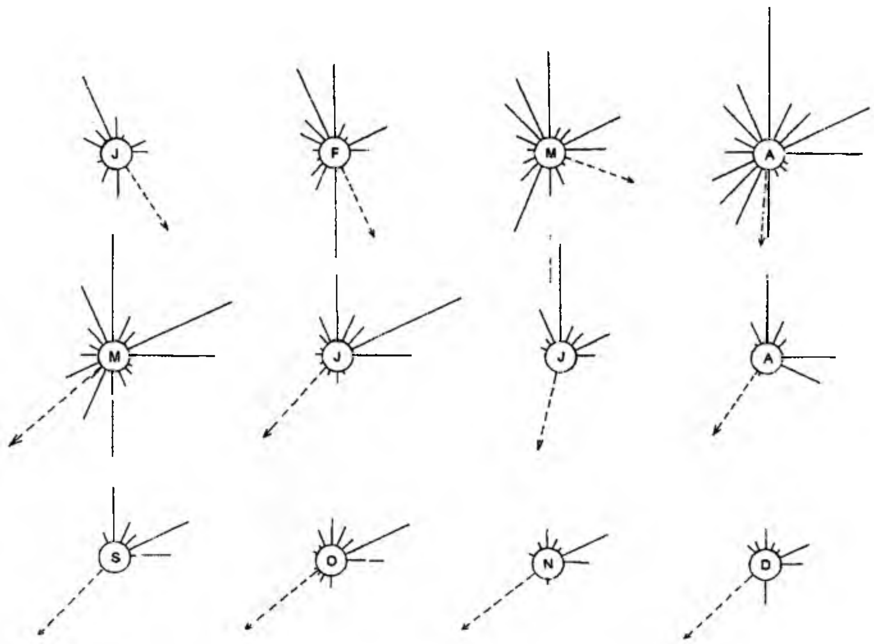
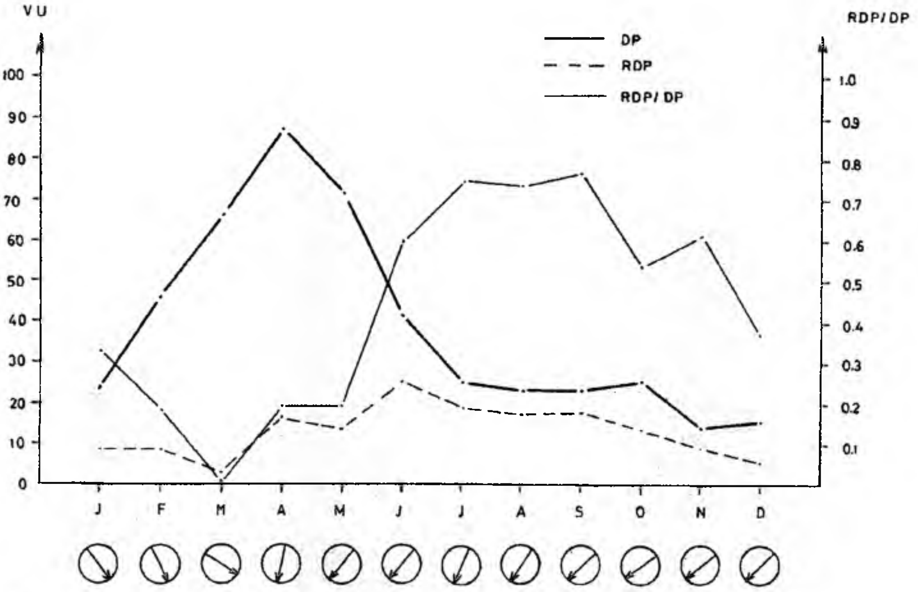


Fig.64 Characteristics of the sand moving winds in Tazirbu:
a) monthly sand roses, b) monthly variations of drift potential (DP), resultant drift potential (RDP) and index of directional variability (RDP/DP).



a)



b)

Fig.65 Characteristics of the sand moving winds in Al Kufrah:
a) monthly sand roses, b) monthly variations of drift potential (DP), resultant drift potential (RDP) and index of directional variability (RDP/DP).

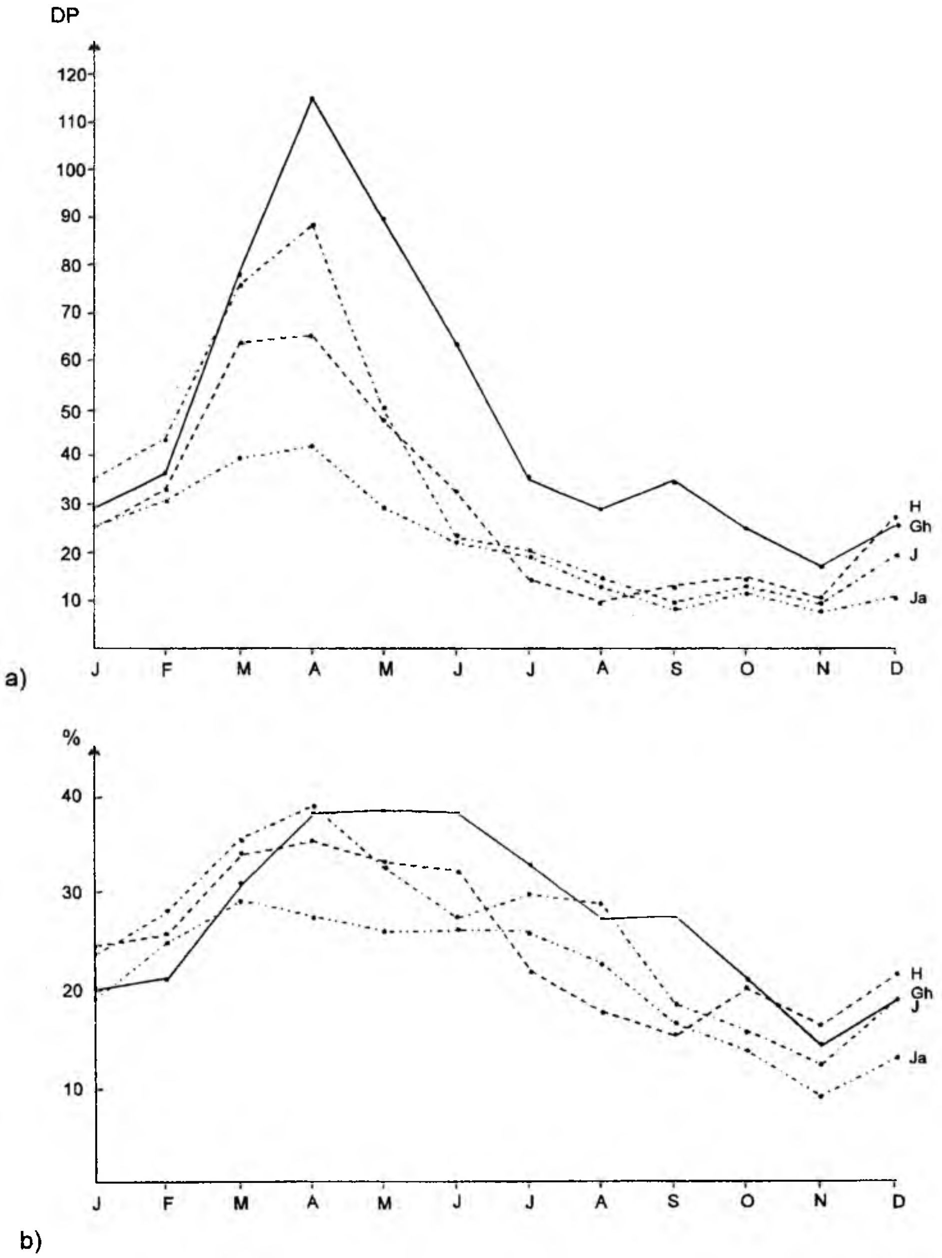


Fig.66 a) Comparison of monthly variations of drift potential, b) comparison of the percentage of sand moving winds for the meteorological stations in: Ghadamis(Gh), Hun (H), Jalu (J) and Al Jaghbub (Ja).

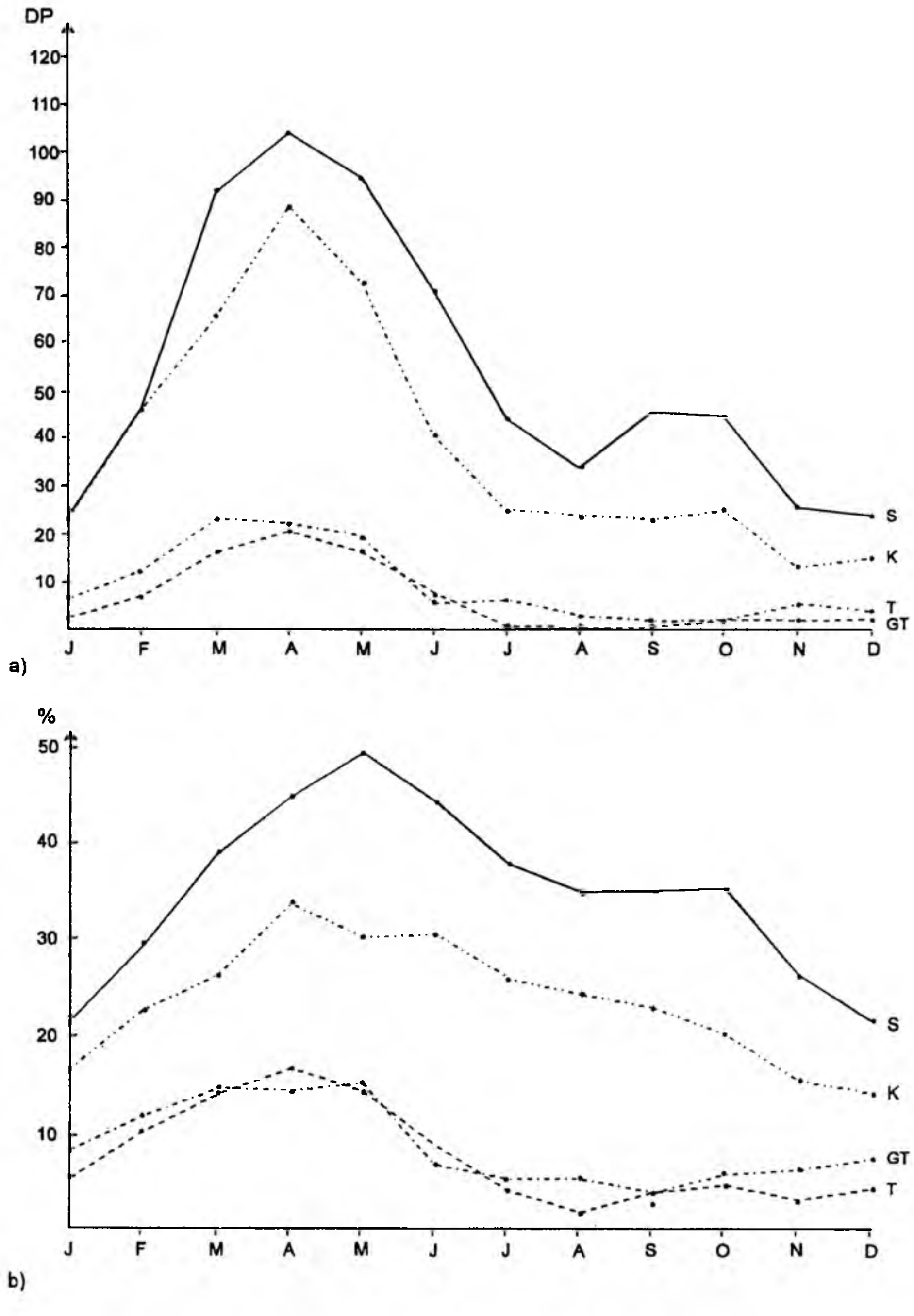
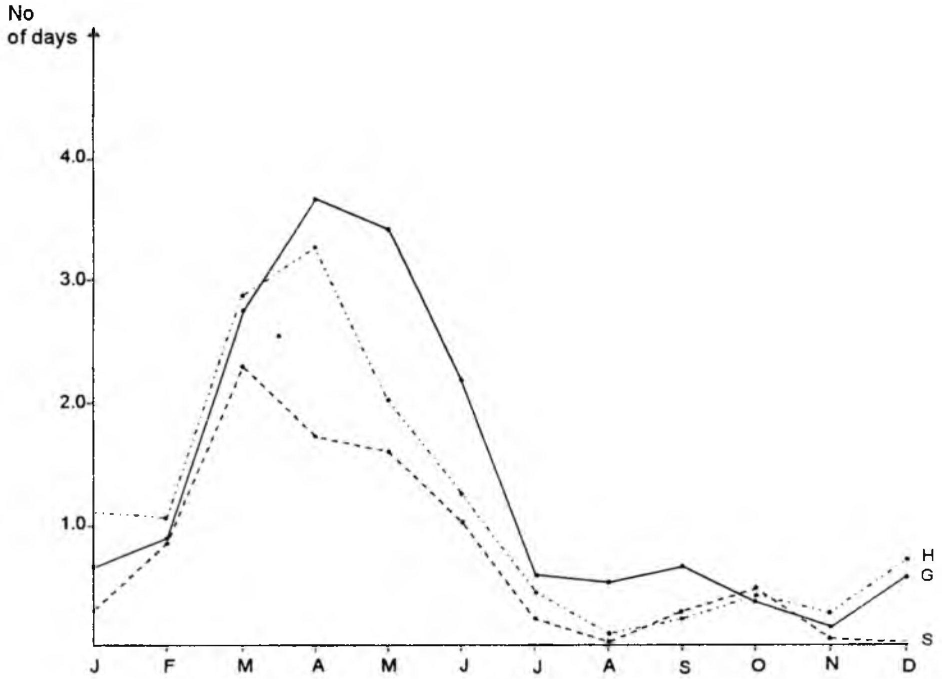


Fig.67 a) Comparison of monthly variations of drift potential, b) comparison of the percentage of sand moving winds for the meteorological stations in: Sabha (S), Al Kufrah (K), Ghat (GT) and Tazirbu (T).



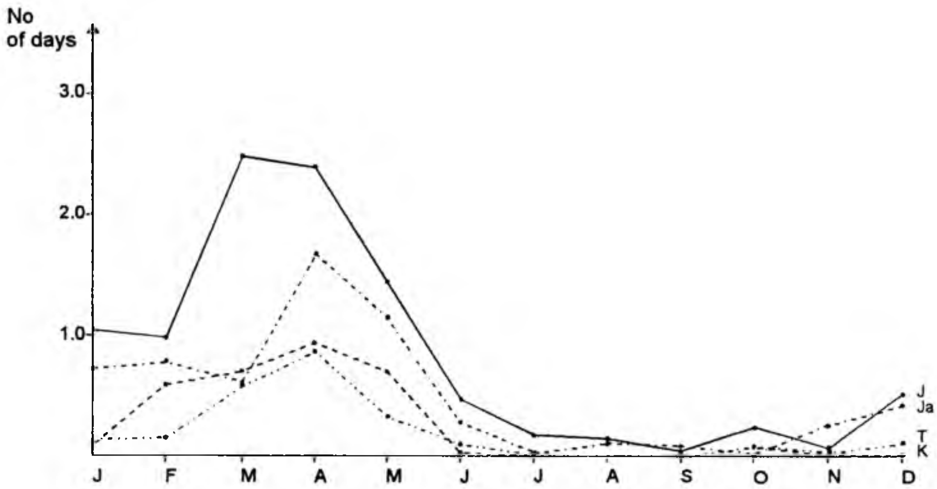
a)

Meteorological station	Maximum number of sandstorm days recorded in one month												*)
	J	F	M	A	M	J	J	A	S	O	N	D	
GHADAMIS	2	4	9	11	9	6	5	2	3	2	1	3	16.2
SABHA	2	3	4	5	5	6	2	1	2	2	1	1	9.4
HUN	3	3	10	9	6	6	2	2	1	4	2	3	14.3

*) Annual average number of sandstorm days

b)

Fig.68 Sandstorm days in the western part of Libyan Sahara (in the period 1956-1985). a) Mean number of the sandstorm days in Ghadamis (G), Sabha (S) and Hun (H), b) Maximum number of sandstorm days recorded in one month.



a)

Meteorological station	Maximum number of sandstorm days recorded in one month												*)
	J	F	M	A	M	J	J	A	S	O	N	D	
JALU	5	3	6	10	5	2	2	1	1	1	1	3	10.3
AL JAGHBUB	3	2	6	5	4	2	1	1	1	1	2	2	7.4
TAZIRBU	1	1	3	4	2	2	0	0	0	1	1	1	2.6
AL KUFRAH	1	3	4	4	4	1	0	0	0	1	1	1	3.2

*) Annual average number of sandstorm days

b)

Fig.69 Sandstorm days in the eastern part of Libyan Sahara
(in the period 1956-1985).
a) Mean number of the sandstorm days in Jalu (J), Al Jaghbub (Ja), Tazirbu (T), Al Kufrah (K), b) Maximum number of sandstorm days recorded in one month.

5. THE RELATIONSHIP BETWEEN DRIFT POTENTIAL, RESULTANT DRIFT DIRECTION AND POTENTIAL, AND THE INDEX OF DIRECTIONAL VARIABILITY

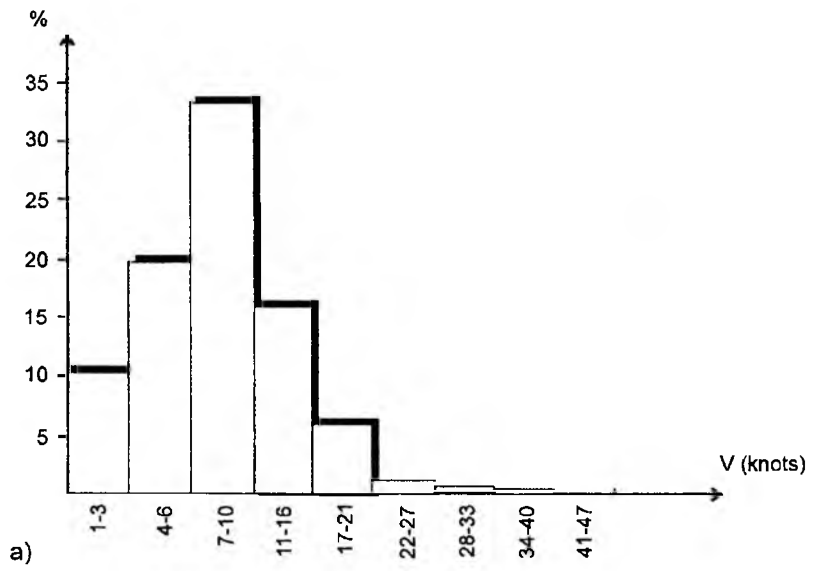
The relationship between drift potential, resultant drift direction, resultant drift potential and the index of directional variability presents the best set of information available for the successful analysis of sand dunes and patterns. The Author has proposed a new form of diagram, presenting these relationships for each meteorological station. On such diagrams, (Figs 58b to 65b) the monthly variations of the drift potential, resultant drift potential and the index of directional variability (RDP/DP) are presented, together with the monthly resultant wind directions (arrows in circles).

For example: in Sabha (Fig.63b) the highest drift potential occurs in April (ca 105 VU), but the resultant wind potential is rather low (ca 22 VU). This is expressed by the low value of RDP/DP thereby indicating the high directional variability. The converse of this situation is in August, when drift potential is low (ca 33 VU) but the resultant drift potential is nearly the same as it is in April (ca 23 VU), hence the ratio RDP/DP is much higher (ca 0.65), thereby indicating the unimodal or prevailing wind direction.

The wind regime in Kufra, as presented on Fig.65b, reflects a slightly different situation when compared with Sabha. The drift potential is also highest in spring time with a peak in April (ca 88 VU), but with lower resultant drift potential. From May to November there is a little higher resultant drift potential oscillating between 10 VU and 20 VU and a much higher ratio between RDP and DP, thereby indicating a persistence of unidirectional winds.

6. THE INFLUENCE OF THRESHOLD VELOCITY CHANGES

As is obvious from Fig.66b and 67b, only about 20% of the recorded winds take part is sand movement. On Fig.70a, the percentage of winds recorded in the meteorological station at Kufrah in particular wind categories is given. The winds recorded in the first three categories ($1 \div 3$, $4 \div 6$ and $7 \div 10$ knots) did not result in sand movement. These winds represent 63% of the total. The drift potential (DP) calculated for wind categories between 11 and 47 knots in presented in Fig.70b. The highest wind potential appears with respect to the wind category between 17 and 21 knots, which represents only 7% of all recorded winds. The influence of winds recorded in the category between 11 and 16 knots is rather small, which is attributable to the previously-assumed threshold wind velocity of 12 knots. By contrast, this category has the largest percentage of all recorded sand-moving winds which amount to 15%. The assumed threshold velocity of 12 knots relates to sand of an average grain diameter of 0.30 mm and for a flat surface.



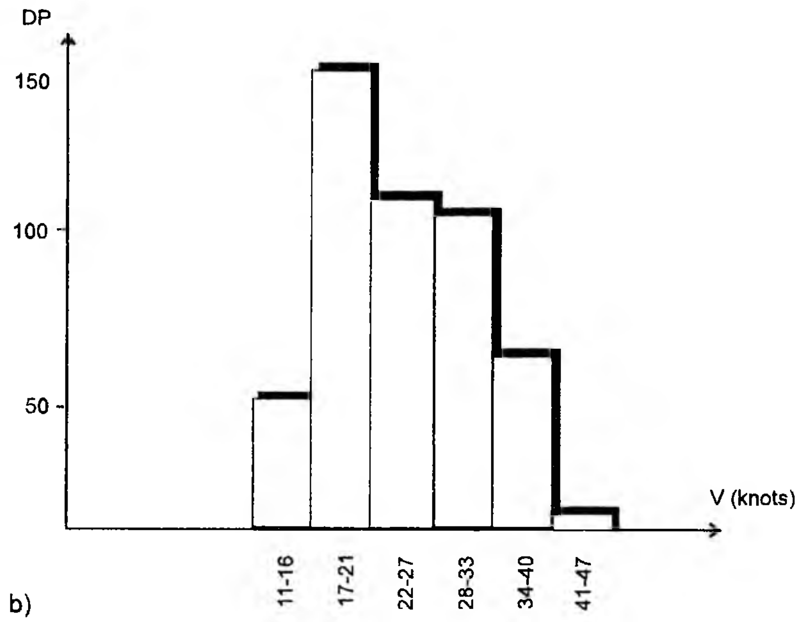


Fig.70
a) Percentage of winds recorded in particular wind categories in Al Kufrah,
b) values of drift potential (DP) calculated for the given wind categories.

If the grain diameters are smaller, so also the value of threshold velocity is smaller; hence, bigger the influence of the winds recorded in the category between 11 and 16 knots. The same situation is evident in the case of threshold velocity over the windward side of a dune, where, in the upper part of dune, the threshold velocity is smaller. Also one must remember that the computation of weighting factors was based on the assumption that the frequency of wind velocities are equally distributed within a given velocity category, which is probably unrealistic.

In consideration of the problem set out here, it is obvious that sand drift can locally be changed within this same wind regime. These changes occur mainly in the fields of large dune forms, when, in this same wind regime large differences can occur between the dune base and dune crest, thereby increasing the drift potential in the upper part of the dune. It is necessary to remember that winds which are not able to move sand at the dune base, may be able to move the sand in the crest region.

7. ANALYSIS OF THE WIND REGIME IN THE REGIONS OF SAND SEAS

7.1. The region of the western sand seas

The wind regime in the region of the western sand seas, namely Idhan Awbarii and Idhan Murzuq, can be evaluated on the basis of the meteorological data collected in the stations at Sabha, Ghadames and Ghat. The annual sand roses for Sabha and Ghadames show a very high drift potential (Sabha DP = 655 VU, Ghadames DP = 588 VU). The drift potential in Sabha has the highest value from all Libyan meteorological stations (RDP = 217), whereas at Ghadames it is lower (RDP = 137 VU). The sand roses for all stations in this region indicate a mutidirectional wind regime (this is especially the case at Ghadamis and Ghat), as confirmed by a rather low index of RDP/DP - 0.23 and 0.15.

The analysis of monthly variations indicates large differences in drift potential, resultant drift potential and resultant wind direction. The spring months are characterized by the highest drift potential (between 80 and 116 VU) and the resultant wind direction, from the SE in Sabha and the SW in Ghadames. Within those months, the winds blow from opposite directions. The strong SE winds depicted on the sand roses for Sabha confirm the influence of acounter-clockwise circulation around the lows in this season. The strong winds from the SW (as recorded at Ghadamis) are related to the passage of cold fronts from east to west in this part of the Libyan Sahara.

The summer months are characterized by rather low drift potentials (between 26 VU and 47 VU) except in June (up to 71 VU) which may be regarded as the transitional period between the spring and summer seasons. In this period, winds blow predominantly from NE and SE.

The autumn season displays only very low activity of sand-moving winds. The drift potential varies between 18 VU and 45 VU. In this period, the winds blow mainly from the SE direction. Thunderstorms related to desert depressions are infrequent at this time.

The number of meteorological stations in this region of the Libyan Sahara is insufficient for detailed analysis of sand-moving winds, but the general trends can certainly be useful as indicators for dune development.

7.2. The region of the eastern sand seas

The wind regime in this part of Libya can be evaluated on the basis of the general wind circulation in the Northern Sahara and on the basis of meteorological data from the stations at Hun, Jalu, Al Jaghbub, Tazirrbu and Al Kufrah. The annual sand roses calculated for these stations (Fig.57) indicate both the drift potential and drift direction as well as the resultant drift potential, thereby expressing the general trend of sand-moving winds.

The resultant drift potential has an orientation similar which is for all the northern stations: Hun, Jalu and Al Jaghbub and indicates that NW winds are in the majority (winds blowing from the sector between 310°÷330°). The highest drift potential (414 VU) and the highest resultant drift potential is at Jalu, which gives a RDP/DP ratio of 0.30.

For analysis of sand dune development, the seasonal changes noted at these stations are very important. (Fig.59a, 60a, 61a along with monthly sand roses). At all stations, the prevailing direction of sand moving winds in winter is from the W (260°÷280°). The highest drift potential is caused by the W winds (Hun: 10÷15 VU, Jalu: 7÷15 VU and Al Jaghbub 6÷13 VU).

In spring the W direction (260°÷280°) of sand drift is maintained, but another direction of similar drift potential becomes evident, from the N (350°÷10°). The sand drift potential from the W direction is a little lower in March and in April when compared with winter time (Hun: 63 ÷ 65 VU, Jalu: 79 ÷ 87 VU, and al Jaghbub: 39 ÷ 41 VU). In May, the sand drift potential is very low and oscillates between 25 and 48 VU. The decreasing tendency of the drift potential can be observed in the succeeding months.

The summer season is characterized by a lower total drift potential in particular months (Hun: 10 ÷ 48 VU, Jalu: 10 ÷ 24 VU and Al Jaghbub: 10 ÷ 23 VU), with a decreasing tendency through this period. The sand - moving winds blow generally from the N but over a wider range of compass directions (290°÷10°), indicating the influence of the NW and W winds. In Al Jaghbub, the prevailing wind direction ranges between 350°÷50°, indicating N and NE windflows.

The short autumn season (October-November) has a generally lower drift potential (Hun: 14 ÷ 16 DP, Jalu: 11 ÷ 15 DP and Al Jaghbub: 8 ÷ 13 DP). The principal wind direction is from the N (350°÷10°) at Hun. At Jalu, winds from the S and SE also occur. At Al Jaghbub, in October, the winds blow from the NE sector, while, in November, from the NW.

In the southern region of the Libyan sand seas, meteorological data are available from only two stations in Tazirbu and Al Kufrah are available. Data from these stations can be used only for a general estimation of the wind regime in this region. In comparison with the northern region, the resultant drift direction changes to a SW trend. In Al Kufrah both DP and RDP have a high value (DP = 468 VU, and RDP = 142 VU); the RDP/DP is 0.30.

At Al Kufrah, the winter season is characterized by a large directional variability, with a wide NW-N sector between 290° and 10°, and a narrow sector in the S between 170° and 190° in February. At Tazirbu, a very low multidirectional DP is recorded (3 ÷ 8 VU).

The spring season has the highest DP (Al Kufrah: 40 ÷ 87 VU), with the strongest winds blowing from the N, NW and S in March, the NE in April and the NE and E in May. In summer, the drift potential is much lower, but with very well defined wind directions from the N and NE and from the E (between 340° and 100°). The highest monthly DP is in June (42 VU) and the lowest in September (25 VU). In autumn, the wind direction is similar to that of summer, with a majority of sand-moving winds in the NE quarter.

The comparison of the sandmoving winds in the western and eastern parts of the the Libyan Sahara indicates the multidirectional nature of the wind regime in the western part. It is mainly multidirectional in the norhern part, but a more bidirectional wind regime affects the eastern part. The wind regime in the Libyan Sahara represents the transition between the wind regime in the central and western Sahara, and the eastern part of the Sahara in Egypt, where unidirectional or bidirectional winds prevail. The changes of wind regime are reflected in the dune forms and patterns of the sand seas of the Libyan Sahara.

Appendix 5

PRINCIPLES OF AEOLIAN SAND MOVEMENT

1. MODES OF AEOLIAN SAND TRANSPORT

The processes involved in the movement of sand form the basis for the understanding of dune forms and sand desert geomorphology. These processes were studied by various authorities (Bagnold 1936a, 1936b, 1938, 1941, 1973; Belly, 1964; Horikawa et al., 1984; Kawamura, 1951; Tsoar, 1985), and a brief review of the basis principles of sand movement is given in this appendix. Sand is moved forward by the wind in two closely related processes: *saltation* (Latin verb: *saltare* which means to leap or dance) and *surface creep*. Saltating grains move by being propelled into the near surface moving air layer by the combined action of aerodynamic lift and impact of other saltating grains returning to the surface. The results of various investigations have indicated that coarser grains move by creeping and their movement depends in large measure on the bombardment by saltating grains.

Most of the sand moves by saltation, but a small amount is moved along by surface creep. Bagnold (1941) found that creep was usually about one quarter of the total load. This figure was confirmed by Chepil (1945) who found that, in the case of sand grain in the diameter of between 0.25 and 0.83 mm, the creep load to the total load ratio was 0.249 whereas, for the sand between 0.15 and 0.25 mm, it was 0.157. These figures indicate that creep becmes more important as the proportion of coarser grains increases. Chorley and Schum (1984) estimated that 84% of fine sand (0.15 ÷ 0.25 mm) is moved by saltation whereas for the medium coarse sand the corresponding value is 75%. Wind tunnel tracer experiments reported by Willetts and Rice (1985), indicated that creep amounts for 25% of the transport rate and that many of the transported grains have progressed by a *combination of saltation and creep*.

Besides saltation and surface creep, the fine particles of sand travel by *suspension* in the wind. These fine particles can be lifted by turbulence, and are therefore diffused to greater heights. The small particles are carried away more quickly and are separated from the coarser grains. It should be emphasised that there is no very precise size limit for particles which travel by suspension. This limit is certainly related to wind speed, and varies within 0.05 and 0.1 mm.

Aeolian saltation is a process whereby the grains alternate between *grain/air* and *grain/ground* interactions (Rumpel, 1985). The grain/ground interaction is responsible for the sucessive saltation. The saltation mode of sand movement results from two forces acting on the sand particle (White, 1985). The forces exerted on the sand particle are resolved into two components, one parallel to the direction of the mean wind flow, called *drag* (tangential) and the other normal to the flow, called *lift*. The drag consists of *surface drag* (viscous skin friction) and *pressure*, due to the pressure difference in the upwind and downwind side of sand grain. The skin friction drag is the viscous stress acting tangentially on the grain surface. The normal force is caused by unequal pressure distribution across the individual grain surface and is predominant near the surface. As well as *the aerodynamic forces*, there are also *inertial forces*, such as the grain weight, which acts opposite to the lift force. The *adhesive forces* operate between the grains and other surfaces. Between neighbouring grains, there are *cohesive forces*, which have to be taken under consideration in case of fine sand grains.

Hunt and Nalpanis (1985) pointed out that, in saltation, the only significant factors acting on the sand are drag and gravity; lift due to spin is observed to be unimportant and turbulence has no effect on particle trajectories.

The key factor in the movement of sand is the *shear force* exerted on the surface by the wind. The *shear stress* (τ) or *drag force*, that air exerts on the sand surface, can be expressed as:

$$\tau = \rho v_*^2;$$

where:

ρ - air density (» 1.22 kg/m^3)

v_* - shear velocity, drag velocity or friction velocity.

2. INITIATION OF SAND MOVEMENT

The wind acting on sand particles has a speed which is generally called *drag speed* or *drag velocity* (v_*).The critical speed at which sand particles begin to saltate is called *the threshold speed* or *threshold velocity* (v_{*t}). The critical velocity at which particle movement begins was referred to as the *fluid threshold* by Bagnold (1941). This fluid threshold velocity called also *the static threshold drag velocity* (El-Baz and Wolfe, 1982) can be expressed by the equation:

$$V_{*t} = A\left[\left(\frac{\sigma - \rho}{\rho}\right) gd\right]^{0.5};$$

where:

v_{*t} - the static threshold drag velocity

σ - the bulk density of sand (for quartz σ = 2650 kg/m^3 or 2.65 g/cm^3)

ρ - the density of air (1.22 kg/m^3 or 0.0012 g/cm^3)

g - the acceleration of gravity (980 cm/s)

d - the grain diameter

A - the coefficient, which for particles above 0.1 mm diameter was found to be 0.1.

It should be noted that the density of dry air is a function of temperature: for 0°C it is 1.292, whereas at 40°C, it drops to 1.128.

The *static drag velocity* is the critical value of the wind velocity which has to be reached, in order to initiate the movement in the hypothetical case of an extremely flat sand surface. In a such case, the movement of the sand grains is caused only by the drag of the wind. When particle movement starts, the surface is bombarded with saltating grains, and a further movement is initiated; hence, the sand movement can be maintained at lower velocities than were required for the initial movement. The impact of grains already in motion on an immobile sand surface, helps the wind drag to put the latter into motion, and this process repeats itself. This velocity, which was called by Bagnold (1941) *impact threshold,* is also known under the name *dynamic threshold drag velocity* (El-Baz, 1982). For the definition of impact threshold, the same formula as that for the fluid threshold can be used but the coefficient A should be 0.08 for grains larger than 0.25 mm. Chepil (1945) found A to be about 0.085 for grains larger than 0.1 mm.

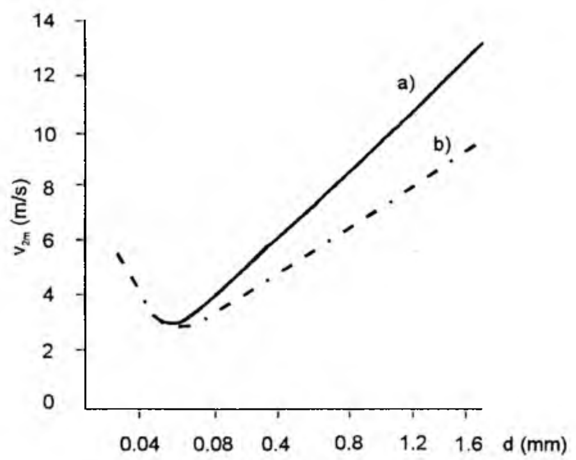


Fig.71 Relationship between threshold velocities
(a) fluid threshold velocity, b)impact threshold velocity)and sand grain diameters
(adapted from Bagnold, 1941).

As a function of sand grain diameters the relationship between fluid threshold and impact threshold is presented in Fig.71. As can be seen, both fluid threshold and impact threshold vary within grain diameter and, generally, for movement of the bigger grain diameter, a higher wind velocity is required. Only for sand particles below 0.06 mm do the fluid threshold velocities rise again. This phenomenon was discoverd by Bagnold as a result of wind tunnel experiments. It is related to an increased interparticle cohesion, with weak chemical bonds, and to lower values of surface roughness and to a greater water retention. Sand grains smaller than 0.06 mm are transported mainly in suspension but, in some cases, the ballistic impacts of saltating grains can cause the movement of such small grains.

The threshold velocity for quartz grains larger than 0.25 mm, can be computed from the simplified formula given by Pye and Tsoar (1990):

$$v_{*t} = 146\, d^{0.5};$$

3. AIRFLOW OVER A SAND SURFACE

Aeolian sand movement is related to the drag velocities on the surface, but the wind velocity varies with the height over the ground surface.

The part of the atmosphere where the wind velocity changes from v_0 to v_∞ is called *the atmospheric boundary layer.* The height of the atmosphere boundary layer depends on the roughness of the ground surface and varies between hundreds of metres and one kilometre. Very close to a smooth ground surface is a thin layer characterized by *laminar flow* which is called *the laminar sub-layer.*

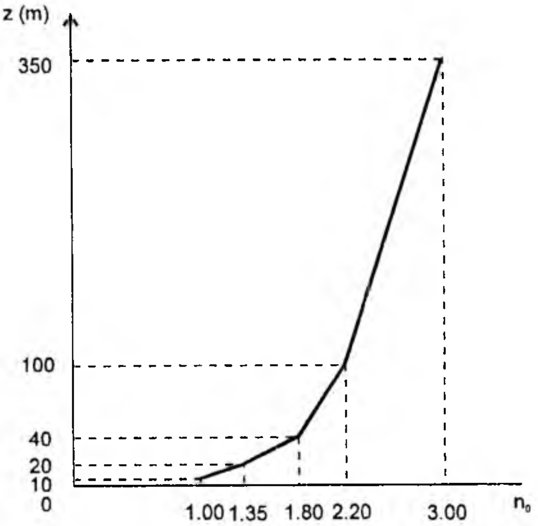


Fig.72 Relation between the coefficient of wind velocity (n_o)
and height over terrain surface (z). Sources: Polish standard PN-70/b-02011.

The thickness of this sub - layer is a function of shear stress and wind velocity. Near to the ground, the wind velocity is affected by the viscous force, but at a certain height, the viscous force diminishes and the wind flow is effected mainly by an inertial force and pressure gradient forces. At a higher level, the inertial forces become predominant and the laminar flow is changed into *turbulent flow*, characterized by chaotic irregular wind motion. Generally, the aeolian flows involved in aeolian processes have a turbulent character.

As the effect of the interaction of these forces, the wind speed near the ground increases sharply with height and, eventually, the wind speed increases asymptotically with height (Fig.72).

Bagnold's (1941) experiments indicated that there are different velocity profiles measured without saltation over a fixed bed and during saltation over a mobile bed (Fig.73). The data for the velocity profiles without saltation, define straight lines described by logarithmic lows and cross in z_0.

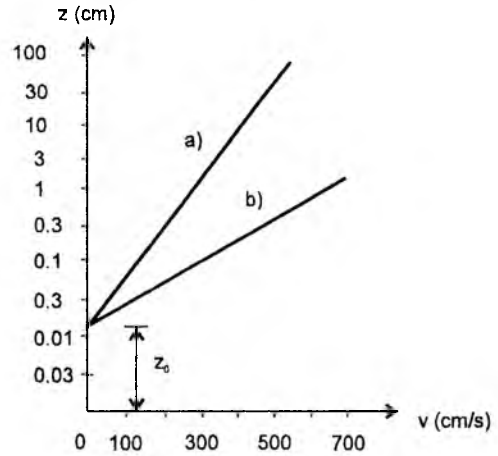


Fig.73 Wind velocity profiles near ground without saltation, for two friction velocities: a) 0.26 m/s and b) 0.61 m/s. (Adapted from Bagnold, 1941).

At ground surface, the wind velocity is zero. When the wind passes over a surface it is retarded at its base by friction and there is a very thin layer above the base within which the velocity is zero. The thickness of this layer depends on the roughness of the base. Bagnold (1953) estimated that for small grain sizes the thickness of this layer is approximately 1/30 of the diameter of sand grains and White (1985) estimated this thickness as 1/9 of the diameter of the large grains. The height below which the velocity is zero, is called the reference height (z_0) and may be determined by the empirical formula given by Zingg (1953):

$$z_0 = \frac{0.081 \log d}{0.18};$$

The roughness of the ground surface has a significant effect on wind velocities and hence on sand movement. *The larger particles prevent the movement of the smaller ones by protecting them from the impact of the wind.* The surfaces covered by coarser sand form the sand sheets which are very effective in reducing aeolian erosion. In the case when the surface is covered with an armour of scattered boulders, aeolian erosion is very effectively reduced. Such areas are called *regs* or *sarirs* and they appear in several places in Libya (Sarir Tibisti, Sarir Calanshio, Sarir al Qattusah).

As long as the sand is in a static position, the wind velocity distribution in relation to the height can be expressed by the equation:

$$v = C \log \frac{z}{z_0};$$

where: z - height above the sand surface for which the wind velocity (v) is determined and C is a coefficient which, according to Karman's development equals:

$$C = \frac{2.3}{K} v_*;$$

where K is the Karman constant. Taking the value of 0.40 for K, the equation for v can be formulated as:

$$v = 5.75\, v_{*t} \log \frac{z}{z_0};$$

Zingg (1950) proposed a similar formula, but with a different constant value:

$$v = 6.13\, v_{*t} \log \frac{z}{z_0};$$

During saltation the velocity profiles near the ground surface are changed. The velocity gradients converge at a higher point than in the case of the wind passing over a stabile sand surface. Bagnold (1941) indicated that as a result of saltation load, the wind velocity profiles associated with a range of shear velocities, converge in so called focal point. The coordinates of the focal point are z' and v' where z' is the average height of the focal point and v' is the mean wind velocity at height z'.

The coordinates of the focus point may be estimated by the following equation given by Zingg (1953):

$$z' = 10\, d \text{ (mm)};$$

$$v' = 10.3\, d \text{ (m/s)}$$

The Zingg's equation implies that the focus point is related to the grain size of the saltating sand grains.

The results of experiments performed by Belly (1964) for sand of 0.44 mm and 0.30 mm diameters, indicated that focal point coordinates agree quite well with Zingg's formula, but for sand with 0.145 diameter (Kadib, 1964) these values did not agree. Gerety (1985) concluded that it is difficult to interpret the apparent correlation between focal height and the mean size of sand in transport.

The height z' of the focal point during saltation implies larger roughness elements during saltation. Bagnold concluded that the roughness felt by wind is not due to the form or grain roughness of the bed, but rather due to drag on the wind by the saltating grains (Gerety, 1985).

The relationship between wind and drag velocities for the sand in saltation can be expressed by rewriting Bagnod's logarithmic low with the focal point as origin:

$$v^s = 5.75\, v_{*t} \log\frac{z}{z'} + v' ;$$

where: z' - height of the focal point and v' - velocity measured at height z'.

On the basis of the direct measurement of the total shear stress on the bed during saltation, Zingg (1952) adjusted the value of the Karman constant in Bagnold's equation, which then has the form:

$$v^s = 6.13\, v_{*t} \log\frac{z}{z'} + v' ;$$

As the result of experimental measurements with sand of 0.30 mm grain diameter Belly (1964) obtained: z' = 0.3048 cm and v' = 274.3 cm/s. For that data and for threshold shear velocity v_{*t} = 16 cm/s the equivalent wind velocity measured at a meteorological station at a height of 10 m will be 21.5 km/h (11.6 miles/h).These figures will vary with the grain diameter, which influences the threshold shear velocity v_{*t}.

4. FLYING DISTANCE AND FLYING HEIGHT OF SAND PARTICLES IN SALTATION

Saltating grains move by being propelled into the near surface moving air layer by the combined action of aerodynamic lift and impact of other saltating grains returning to the surface. The path of a single saltating grain is shown on Fig.74. The trajectory starts with an almost vertical movement. Then, when the grain enters the faster wind above surface, its path is modified and pushed forward. When its initial upward movement is decreased, it is pulled down by gravity. The trajectory bends over and reaches the surface.

The trajectory started at the *launch angle* (or *take off angle* or *ejection angle*) which, according to Bagnold, is about a right angle. Sorensen (1985) estimated that the mean launch angle for the entire sand population varies between 43° and 46°, but the grains dislodged from the rest by impinging grains start at angles from 52° to 54°. Hunt and Nalpanis (1985) reported that the launch angles vary between 34° and 41°. Jensen and Sorensen (1986) in concluded that the the height of jumping grains does not depend much on the wind speed once saltation is established and that the vertical component of the mean launch velocity decreases with the grain size. It is approximately inversely proportional to the grain diameter.

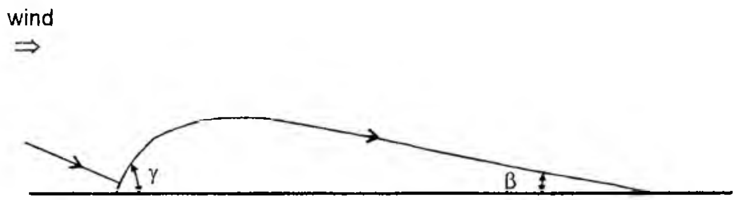


Fig.74 Trajectory of a single saltating grain: γ - launch angle, β - approach angle.

The *impact angle,* also called *the approach angle,* depends on the grain size and is in the range of 15° to 25° (Sorensen, 1985). On the basis of experiments in which saltating grains were photographed using a high speed cinema camera, Willetts and Rice (1985a) indicated, that approach angles ranged predominantly from 10° to 20° with a few extraneous values up to 40°. The mean impact angle increases with the grain size: 9.6° for fine the fraction, 11.7° for the medium fraction and 12.7° for the coarse fraction. The impact angle decreases with wind speed.

The results of wind tunnel experiments and field observation show that the saltation mode is influenced by the grain size and shape, the wind velocity and character of the surface over which saltation take place. The saltation layer may be defined as the maximum height and maximum distance of saltating grains. The saltation layer is not precisely defined in the upper limit, but the maximum saltation length is normally between ten to twelve times saltation height (Owen, 1980). The saltating grains modify the wind velocity profile near the surface and reduce the friction velocity below the fluid threshold.

Mathematic approaches to the flying distance of sand particles have been made by Bagnold (1941) and were confirmed by photographic observations of the sand paths. Owen (1980) developed an experimental formula for the length of the mean saltation path:

$$l = 10.3 \, \frac{v_*}{g} \, ;$$

The flying height depends on the character of the surface. Bagnold (1941) found that sand bounced to a maximum height of two metres above bouldery areas, and nine centimetres above sand. Generally, saltating curtains of sand consist of a dense layer with few odd particles travelling above it. In accordance with the data given by Chorley and Schum (1984) sand particles moving in moderately strong winds of 15 m/s can reach a maximum height of 2 m on boulders and 9 cm on a sand surface and a mean height of 63 cm on boulder surfaces. Owen (1980) stated that the flying height, h, may be expressed as a function of wind velocity v and a gravity constant g:

$$h = 0.82 \frac{v_*}{g};$$

A similar formula was given by Hunt and Naplanis (1985) but they used a different coefficient:

$$h = 1.1 \frac{v_*}{g};$$

The flying height also depends on the grain shape. It was found that angular grains have a lower mean height jump than spherical grains (Jensen and Sorensen, 1986). Apparently, more - angular grains are less effective in their ability to rebound from a surface than spherical grains.

5. THE DISLODGEMENT RATE OF THE SALTATION MODE

Neither the rate at which surface particles are dislodged and replaced nor the speed of dislodged particles in a given wind condition, were understood until recently. Belly (1964) defined the sand distribution in relation to different grain sizes.

Inter - saltation collisions have a large influence on sand movement in the saltation mode. Based on saltation mechanism, photographed with high speed film camera, the experiments of Willetts and Rice (1985b) permitted the determination of the influence of descending grains on the dislodgement of sand particles. They found that 50% of fine grain descents and 82% of coarse grain descents, give rise to dislodgement of one or more grains. The mean rates of dislodgement are ca 3 grains per coarse grain collision, 1.5 for medium grains and 0.9 per fine grains. Hence, the role of the *coarser grains appears to be the generation of the new saltation sequences.*

Successive saltation plays a major role in the saltation mode of sand transport. Successive saltation is the only process with a high probability of transferring energy from a horizontal into a vertical grain movement. Successive saltation contains a mechanism which confines the range of impact angles between 10° to 16°.

The population of grains moving at a given time, contains grains in one part of a continuous sequence of saltation, others in isolated saltation and the rest in foreward creep. The observations indicated growth in height in successive saltations and a close relation to different fractions - high for the fine fraction and low for the coarse fraction.

Sorensen (1985) noticed that the size distribution of the saltating sand depends much more on the size distribution of the bed material than on the wind speed. Experiments indicated that *for medium drag velocities (46 cm/s) the saltating grain is slightly finer than the bed material,* and *for larger drag velocities the saltating sand is coarser than the bed material.* Hence saltating sand becomes coarser as the wind speed increases. He noted that for the large drag velocities ($v_* > 100$ cm/s), saltation seems to have reached a stage where the size distribution of saltating grains is almost independent of wind speed.

Successive saltation, which is responsible for the largest amount of sand transport, *is theoretically impossible if the ground over which saltation occurs is inlined at about 15° against the wind direction* (Rumpel, 1985). These theoretical findings are confirmed in nature. The windward slopes of dunes usually range between 10° and 15°. This leads to the concept that *windward-sides of the dunes are tilted up by the deposition until successive salatation is entirely subdued.*

6. SURFACE CREEP

Surface creep is a movement of sand particles, where they have continuous contact with the surface. Generally, surface creep is restricted to coarser grains, which cannot be moved as a saltation load. Sand grains ranging in size between 0.5 to 2.0 mm move mainly by creep. Grains smaller than 0.5 mm can be moved by surface creep before beginning to saltare (Willetts and Rice, 1985b). Generally, the larger grains are pushed forward by the impact of saltating grains. Sand particles larger than 2 mm are moved only under extreme wind velocities such as sandstorms, which occur very often in the Libyan Sahara. There is a very large difference between the movement of grains in saltation and surface creep; according to Tsoar and Pye (1987), this relation is between 200 and 400 fold relationship.

7. THE EFFECT OF MOISTURE ON THE THRESHOLD VELOCITY

Both laboratory investigations and field experiments (Belly, 1964; Horikawa *et al.*, 1984) demonstrated that moisture clearly increases the value of the threshold velocity. Sand moisture can have its origin in the atmosphere humidity or from other sources such as precipitation and/or the capillary rising of underground water. The moisture absorbed by the sands holds the grains together. After wetting, the cohesive forces become significant. In desert areas, such as the Libyan Sahara, the variation in atmosphere humidity is rather negligible, but in the case where the water content of sand attains a value of 2% or 3%, the changes in threshold values are considerable. Accordance to Belly (1964), the threshold velocity for wet sand (v_{*tm}) is as follows:

$$v_{*tm} = A(1.8 + 0.6 \log W)\, v_{*t} \,;$$

where W is the water content expressed in percent, A is 0.1 for the fluid threshold and 0.08 for impact threshold.

When the atmosphere is responsible for the sand moisture, Belly (1964) prefers to use the formula:

$$v_{*tm} = A\left[1 + \frac{1}{2}\left(\frac{h}{100}\right)\right] v_{*t} \,;$$

Sarre (1988) found that with moisture content of 16% the sand transport rate was reduced to 67% of the dry value, and the transport was reduced to almost zero when the moisture content increased to 22%.

8. THE EFECT OF SLOPE SURFACE ON THE THRESHOLD VELOCITY

When the surface over which the sand is transported is not flat, but is inclined as in the case of sand dunes, the threshold velocity is changed. Theoretical analyses have been presented by several authors (Howard,1977; Dyer,1986). It was found that there is a slow increase of threshold velocity on the upwind slope (between 1 to 1.2), and a larger decrease on the windward slope (between 1 and 0.2). Experiments performed on Saharan dunes (Hardisty and Whitehouse, 1988) confirmed the theoretical values of changes in threshold velocity.

9. AIRFLOW AND SAND TRANSPORT OVER DUNE SLOPES

9.1. Factors controlling airflow and sand transport

The relationship between dune profile, wind speed and wind direction has essential influence on the rate of sand transport over the dune surface. When the atmospheric boundary layer flows over a region with changes in surface elevation, such as sand dune, both the wind speed and surface shear stress can change significantly, even when the dune slope is gentle. For steeper windward slope (up to 40°) the increase of wind speed can be 100% or more (Hunt and Nalpanis, 1985). The sand particles are more rapidly entrained and transported over the dune top and deposited both on the lower part of the upwind slope and on the downwind slope.

It was found that different processes control the air flow over sand dunes in different regions of the flow (Hunt and Nalpanis, 1985). Near the surface, there is an inner layer (also called the boundary layer), within which the perturbation in velocity is affected by the changes in the shear stress and by the pressure distribution set up over the dune by the flow in the outer region (Fig.75).

Recent investigations performed by Lancaster (1985), Sweet and Kocourek (1990), and Tsoar (1983) introduced new elements into the understanding of the aerodynamic interaction between the dune and the winds acting on it. The main factors controlling those interrelations are: dune shape and height, and the incidence angle between the primary wind direction and the dune crest or brink line. Different processes govern those relations on the windward flank and on the lee-flank.

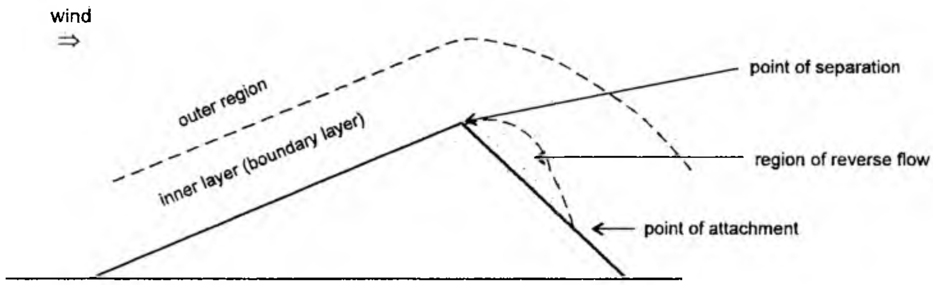


Fig.75 Flow zones above a dune surface and airflow on the lee-side.

9.2. Airflow and sand transport on the windward flank

Both laboratory tests and field investigations confirm the increase in wind velocities on the windward flanks, both on transverse and linear dunes. The increase of wind velocity is characterized by *the speed up factor* or by *the amplification factor.*

The speed-up factor ,S, as introduced by Mason and Sykes (1979), is defined as:

$$S = \frac{v_c}{v_b};$$

where:
v_c - wind velocity at dune crest;
v_b - wind velocity at the base or plinth of the dune.

The change of the wind velocity over the dune can also be expressed by the amplification factor (A_v):

$$A_v = \frac{v_d}{v_s};$$

where :
v_d - mean velocity at the height z above the dune base;
v_s - mean velocity of the undisturbed wind flow on the same height z above the flat surface.

The results of laboratory tests (Tsoar *et al.*,1985) and field investigations (Lancaster, 1985; Tsoar, 1986a) confirm that the speed-up factor or amplification factor depends on dune shape and height. The dune shape is usually defined as: 2 h/l, where h - dune height, l - the horizontal distance from the dune upwind base to the dune crest.

Wind tunnel investigations (Tsoar *et al.*, 1985) indicate that, for the convex bunt profile, (which corresponds with the crossection of barchan or transverse dune profile), the amplification factor for the lower model is 1.4 and for the higher model the amplification factor rises to 1.7. Very similar results were obtained by Lancaster (1985) during his field investigations. Lancaster's investigations have shown that, for transverse dunes ranging in height from 1 to 18.5 m, the speed up factor varies between 1.35 and 2.2, with the lowest value corresponding with the dune shape 2 h/l = 0.25 and highest for the dune 2 h/l = 0.53. These figures indicate that **for dunes with steeper windward flanks the speed-up factor is higher, and, for gently sloping windward flanks, lower**.

For linear dunes, the speed - up factor is strongly related to the incidence angle. On the basis of their tunnel investigations Tsoar *et al.*(1985) reported that, for a triangular profile, which corresponds with the cross-section of longitudinal dune, the amplification factor decreases at the dune base to 0.7, then it rises and, at the dune crest, reaches its highest value of 2.0. Lancaster (1985) reported that the increase of velocity is strongly related to wind direction. For winds blowing perpendicularly to the crest, the speed - up factor has the highest value of 1.8, whereas for the winds blowing at an acute angle, the speed up factor drops to 1.33. The dune height has no special influence on the speed - up factor, but when the winds blow more parallel withj the dune axis, the dune shape becomes more dominant. For linear dunes, the speed - up factor should be determined in the direction of the primary winds; thus, for the same linear dune, for different wind directions there will be a different dune shapes (2 h/l). Hence, a different acceleration of wind speed between dune base and dune crest. Even a small change in wind direction results in a different speed - up factor; hence, in different sand movement. As the wind blows nearly parallel with the dune axis, the acceleration on the windward side becomes progressively less.

The change of airflow on the windward flank of the dune has a very significant consequence in sand transport; hence, on the dune development or movement. It should be emphasised that the rate of sand transport is proportional to the cube of the excess of wind velocity over the threshold velocity of sand movement. In the case when the speed - up factor equals two, the potential sand movement will increase to third power, it means the potential sand movement on the crest will be 8 times greater than on the dune base. When, on the dune base, the wind speed is only a little higher than the threshold velocity, hence along a slope, the acceleration of the wind will result in even more effective action of the wind in the upper parts, both in erosion and transportation. Even if, at the dune base, the winds are below the threshold velocity, due to the speed - up factor, the winds in the upper parts can exceed the threshold velocity and are thereby able to move sand. When analysing those changes in sand transportation, grain size characteristics should also be taken under consideration.

In accordance with the figures for the speed-up factor given by Lancaster (1985), the effect of a velocity speed - up will be greatest for those directions of the incident winds which are characterized by low or moderate wind velocities and also in the case where such wind directions are oriented perpendicularly to the dune crest. The final effect of this mechanism can change the mode of the effective wind regime, especially in the case of the multidirectional wind regime.

When the wind velocity in the basal areas of dunes is much larger than the threshold velocity the differences in sand transport are smaller. In contrast, for the lower incident wind velocities, the ratio between the sand transport, at the base and crest of the dune, is greater. When, at the base of the dune the wind velocity is very low, i.e. lower than the threshold velocity, there can be an infinite increase of sand transport over the crest.

The potential sand transport rates also vary with dune height, because the speed - up factor is a function of dune height. Lancaster (1985) reported that, for wind with a velocity of 5.6 m/s, sand movement at the crest of a transverse dune will be 19.6 times greater than thatat the upwind base, if the dune is 5 m high, and 121 times greater, if the dune is 20 m high.

9.3. Airflow and sand transport on the lee - flank

The changes of wind speed velocity and wind direction on the lee-side play a very important role in the process of dune formation (Tsoar, 1985; Hunt and Nalpanis, 1985). The movement of sand particles on the lee - side of the dune depends mainly on the dune shape, the sand particle size and the wind speed. Let us assume that the wind blows perpendicular to the dune crest. In the case of a dune with a triangular profile (e.g. a longitudinal dune) the streamlines are separated from the surface over the point of separation, which corresponds to the dune crest (Fig.75). As the result of a separation phenomenon, the air layer that was on the dune surface (the so called boundary layer) rises to a greater altitude and cuts off contact with the dune surface. The separated streamlines are then reattached to the surface at a certain distance downward on the lee - slope, the point of attachment being, where the wind velocity is zero. From that point, the streamlines move into two contrary directions: streamlines which continue movement in the general wind direction, and streamlines which are moving to the crest.

Between the point of separation and the point of attachment, reversed flow occurs as a consequence of the vortex that is formed there (Fig.76). The position of the point of attachment depends on the slope of the dune profile and on the wind direction in relation to the crest line.

For sand dunes with a slope less than about 1/4, or where the ratio of dune height H to half-length L is about 1/3 to 1/2, there is no significant region of reversed flow on the lee-side and the thickness of the inner layer is of the order of 1/20 of the half-length of the dune (Hunt and Nalpanis, 1985).

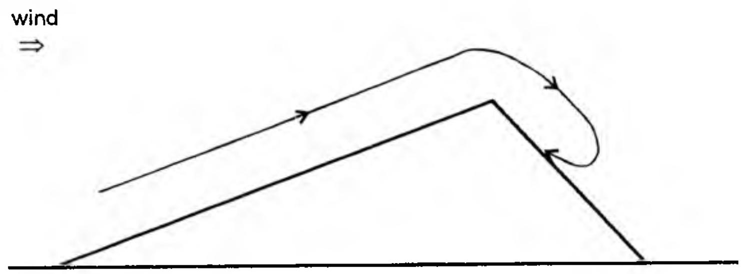


Fig.76 Inverse flow on the lee-side of a dune.

The change of wind velocities and wind direction on the lee - side influence sand movement and also sand deposition. The drastic fall of the wind velocity on the lee - side leads to the deposition of sand in that part of the dune. As it was emphasised earlier, the movement of the sand particles on the lee - side of the dune depends mainly on the dune shape, particle size and wind speed. In the case of the convex dune profile, particles in saltation remain close to the dune surface, and have greater saltation speed at the crest. In the case of a dune with a sharp crest and separated flow over a lee - side slope, particles in saltation travel as flat parabolic trajectories which leave the surface of the dune and descend into the wake. Because the sand particles leave the surface, drag no longer acts on them before they land, and, therefore, even when different size particles are ejected at similar velocities, they land in quite different positions.

In the case when the wind blows obliquely to the dune, the process of separation is more complicated. In that case, there occurs both the change of wind direction in the horizontal plane and a change of wind velocity and its direction in vertical plane. When the wind blows obliquely to the dune, the wind has to be divided into two components: one perpendicular to the crest and another parallel to the crest line. The separation of wind flow is created only by the component perpendicular to the crest line. This component is a function of the incidence angle. The less acute the incidence angle, the larger is the perpendicular component. Hence, the larger is the separation vortex. The position of the attachment point is controlled by the magnitude of the wind and the incidence angle.

The horizontal distribution of the wind flow on the lee - side is as follows: between the line of attachment and line of separation lies the inverse flow region and, here, are two components of wind direction; one parallel to the crest line and the another perpendicular. At the line of attachment, the perpendicular component has zero velocity, and only the component which is parallel to the crest remains.

The magnitude of the wind component parallel to the crest line at the line of attachment Cp may be expressed by the equation:

$$C_P = v_z \cos\alpha;$$

where: v_z - the magnitude of the wind at the crest line, and α - the incidence angle of the wind. When the wind passes over the dune crest, the separation phenomenon occurs. Hence the air layer that was on the dune surface cuts the contact with the dune surface and rises to a greater altitude. The separated flow then gains momentum from the faster layers of the air above. When the wind returns to the dune level at the line of attachment, it has a higher velocity typical of a higher altitude. The results of experiments performed by Tsoar (1986a) indicated that an incidence angle $\alpha = 30° \pm 10°$ is the optimal to cause the maximum increase of the deflected wind on the lee - side.

The aeolian processes on the lee-faces are not very well documented from the field investigations, and no universal model exist among intersted parties. However, recent studies (Tsoar, 1986a; Sweet and Kocurek, 1990) have provided more data for the behaviour of the airflow and sand transport on the lee-flanks of dunes. Sweet and Kocurek (1990) concluded that the airflow on the lee - face is a function of dune shape, incidence angle of primary winds and atmospheric thermal stability. On the basis of both tunnel experiments and field investigations, the following generalized model of airflow on the dune lee - side may be drawn. Dunes which have high aspect ratios and low aspect ratios have different airflow patterns. For transverse flow ($i = 90°$) the airflow pattern is mostly influenced by the shape of dune flanks. On the dunes with a high aspect ratio (0.2), and a lee - face with a near - angle of repose, flow separation occurs and a weak back eddy is developed. The separated flow is characterized by a zone of low wind velocity. For high velocites on the crest line, the velocities on the lee - side drop to 5 ÷ 10%, and, for the lower velocities, the reduction of velocities behind the reattachment line varies between 20% and 50%. Sweet and Kocurek (1990) recordeded that dunes with a low aspect ratio have a relatively higher flow (the ratio between wind velocity on lee - flank and wind velocity on crest) than is the case for high aspect ratio dunes. The airflow separation results in the dominance of grainflow and grainfall in sand transport down the lee - slope.

In the case of dunes with gentle lee - slopes the transverse flow remained attached and undeflected. The airflow velocity on the lee - side generally dropps and reaches the lowest value on the dune base (ca 60%). Sweet and Kocurek, (1990) noted that, for dunes with a similar aspect ratio, the rounded brinks favour the attached flow. Such forms can be classified as *zibars,* which are low - relief bedforms with rounded brinks.

In the case of oblique primary airflow, the flow on the lee - side remains attached and is deflected along the slope. This tendency is well documented both by tunnel experiments and by field investigations. At an incidence angle of ca 30° there is a component of return flow at the foot of the lee - face, but a larger component of airfow is subparallel to the lee - face. Some distance along the lee - face (ca 1.3 ÷ 1.5 of the dune height), the flow direction turns approximately to the primary wind direction.

The relative flow speed on the lee - face is a function of the incidence angle and dune shape. The flow speed on the lee - face is expressed by the following formula:

$$v_l = v_c \cos\alpha;$$

where:

v_l - wind velocity on the lee - side

v_c - wind velocity on the crest

α - incidence angle

9.4. Avalanching on the dune lee - side slip faces

On the dunes which develop slip faces, sand is very often transported in the mode of avalanching. Slip faces of dunes maintain the *angle of repose*, which represents the condition of balance between the intergrain kinetic friction and the force of gravity. When gravity exceeds the intergrain kinetic friction, the sand grains are pulled downslope. The avalanching of sand occurs when the tangential force exceeds the static friction force. The tangential force depends on the slope angle and the weight of the grain. The tangential force which is required to initiate the avalanching process is greater than the force which is required to maintain sliding; this is similar to the fluid threshold velocity and impact threshold velocity.

Laboratory experiments and field investigations have shown that the angle of repose for medium-fine sand varies between 32° and 34°. The sand passing over the crest is deposited on the upper part of the lee - side, causing oversteepening of this uppermost side. This mode of sand transport is called *grainfall.* The deposition from this mode of transportation is in the zone between the dune brink and the line of the maximum length of the saltation path (few tenth of centimetres). When the gravity force accesses the intergrain kinetic friction, the process of avalanching starts, causing failure of the upper most part of the slip face. This process of sand transport is referred to as *sandflow* or *grainflow.*

10. THE RATE OF SAND TRANSPORT

The rate of sand transport may be determined on the basis of mathematical models which, predict the mass transport of sand, or on the basis of field measurements which, to date, are very limited. The rate of sand transport is represented by q, which has units of mass, per unit width, per unit time (kg/m/s or g/cm/s).

The rate of sand transport per unit width and unit time, q, is proportional to the cube of the shear threshold velocity v_* and can be determined by Bagnold's (1937) formula:

$$q = C\left(\frac{d}{D}\right)^{0.5} \frac{\rho}{g} v_*^3 ;$$

where:

- D is the standard 0.25 mm sand diameter;
- d is the grain diameter of the sand in question;
- is the density of air, g is the gravity constant ($\rho/g = 1.25\ 10^{-6}$ cgs);
- C is a constant related to the degree of sorting of sand.

The latter ranges from 1.5 for well sorted sand, through 1.8 for the naturally graded sand, to 2.8 for poorly sorted sand (one with a very wide range of grain sizes), and 3.5 for a pebbly surface. The coefficient C indicates that the sand transport rate is higher over the poorly sorted sand or gravel surface, thereby reflecting the phenomenon of the saltating mode. The Bagnold formula did not initially include the threshold term. Later Bagnold (1956) improved his formula:

$$q = \alpha C\left(\frac{d}{D}\right)^{0.5} \frac{\rho}{g}\left(\frac{v_z}{v_{*z}}\right)^3 ;$$

where:

- α is a constant value [0.174/ln(z/z')],
- v_z is the mean velocity at height z,
- v_{*z} is a threshold velocity at height z.

The other workers: Kawamura (1951), Zingg (1953) and Horikawa *et al.* (1984) developed several formules, based on results obtained in wind tunnels. According to Kawamura, the rate of sand movement is given by:

$$q = K\frac{\rho}{g}\left(v_* - v_{*t}\right)\left(v_* + v_{*t}\right)^2 ;$$

where:

- v_* is the shear velocity,
- v_{*t} is the threshold shear velocity,
- K is a constant value.

For sand of average grain size 0.25 mm, Kawamura obtained 2.78 in wind tunnel experiments.

On the basis of experimental results obtained in wind tunels, Zingg (1953) modified the Bagnold formula, thus :

$$q = C\left(\frac{d}{D}\right)^{0.75}\frac{\rho}{g}v_*^3;$$

with C = 0.83.

Lettau and Lettau (1978) developed the following formula which was applied by Fryberger (1979).

$$q = C\left(\frac{d}{D}\right)^{0.5}\frac{\rho}{g}v_*^2\left(v_* - v_{*t}^s\right);$$

where - C is universal constant for sand (about 6.7), v_* is shear velocity, v_{*t}^s is impact threshold shear velocity or the minimum shear velocity required to keep the sand in motion.

There are also several formulas for estimation of sand transport based on experimental measurements. In 1941, Bagnold had given the following formula:

$$q = 1.4 \times 10^{-5}\, d^{0.5}\, v_*^3;$$

where d - sand grain diameter (cm), v_* - wind shear velocity (cm/s)

Zakinov, cited by Cooke and Warren (1973), has given the following relationship:

$$q = 0.160\left(v_{lm} - 4.1\right)^3 \text{ (g/m/s)};$$

where v_{lm} is the wind velocity at one metre above the ground surface (in cm/s) and 4.1 is the threshold velocity (in cm/s).

On the basis of 500 measurements, Borowka (1990) developed the following experimental formula:

$$q = 0.000301\, v_{lm}^{4.68};$$

where v_{lm} is the wind velocity measured 1 metre above the ground.

The flux density of saltating grains is limited by the ratio between the friction velocity measured over the saltation layer and the threshold friction velocity, known as the transport stage (Bagnold, 1973). The greater concentration of grains in the saltation layer lowers the residual shear stress transmitted to particles. It was found that **when the friction velocity is approximately three times higher than the threshold value, the transport stage approaches zero** and the concentration of grains in the saltation layer reaches saturation, defined by a ratio of 0.9 (Pye and Tsoar, 1990). For sand with a mean size of 0.18 mm, the saturation can be achieved at v_{*t} = 60 cm/s.

The sand transport equations provide only a tool for estimating the maximum rate of sand movement, because other factors which are not incorporated in those equations can significantly change the transport rate.

Appendix 6

DESERT SANDS AND THEIR CHARACTERISTICS

1. SAND SOURCES

In sedimentary environment such as the desert sands, the sediment is derived from some source area, then is transported to the site of deposition by some agent, and is finally deposited.

The main sources of sand are different types of rocks and alluvial deposits, and the principal agents of transport are water and wind. If the type of transportation is considered, the desert sands can have fluvial or aeolian or mixed origin. The physical and chemical weathering of rocks prepare loose detrital sediments, such as clay, silt, sand and gravel for the processes of gravity, running water and wind action. The sources of desert sand are the sand particles released by the breakdown of igneous, metamorphic, clastic sedimentary and carbonate rocks. Globally, the quartz and silicate grains released by weathering and erosion of crustal rocks constitute the majority of the sand in the sand desert areas. Sandstone and sandy conglomerates are also very important source of sand grains. They contain more than 60% of quartz, and more than 10% of feldspar. Whitney *et al.* (1983) estimated that much of the sand in the An Nafud in Saudi Arabia was derived from poorly-cemented Paleozoic sandstones which outcrop in the surrounding areas.

The particles which result from weathering, are transported to depositional sites: *alluvial fans* or *alluvial plains*. When the sediment eroded from the hillslopes is not delivered directly to a channel, it may be accumulated at the base of the slope to form a *colluvial deposit,* otherwise called *colluvium*. The colluvial deposits may be stored temporarily at the slope base, before moving into a stream or river. After that, it may be stored as *alluvium* in the floodplain and bed of the stream, before moving out of the drainage system. Agents of transportation may sort the discrete particles such as gravel, sand, silt or clay. Sediments of small size may be transported through the river systems over great distances. During the transportation process, the individual grains of sediments may be deposited and eroded many times before they leave the system. During great floods, large quantities of sediments may be flushed from the system, a reflection of the channel wideness and deepness (e.g. the Nile floods).

Some aeolian sands originate in place from weathering of bedrocks. Examples of such source material are the residual sand sheets on the Nubian Sandstone in the Eastern Sahara, and the dune sands extending south from the Kharga Depression and other oasis depressions in Egypt, which have originated by salt weathering of sandstones in the depressions (Mabbutt, 1979).

But most of the desert sands are derived by wind acting on the poory sorted fluvial sheets; hence, the desert sands represent an improved result of the sorting process which have already begun on the parent material. There is little dispute to this general thesis. Most of the windblown sand is composed of quartz, which exists in large quantities in many igneous and metamorphic rocks as crystals of sand size. This tends to accumulate when the rocks are weathered away. Quartz resists chemical breakdown better than most minerals, which are normally taken away in a solution or converte to clay minerals. The quartzose sands which are common in desert areas, were eroded by rain and stream run-off, and then were transported by rivers and deposited in the plains (e.g. Great Sand Sea) or in the interior basins (e.g. Awbari and Murzuq basins).

2. SAND GRAIN PARAMETERS

The main parameters of the sand grains are: *grain size* and grain size distribution, commonly called *grain sorting* and *grain shape.*

In accordance with the terminology introduced by Wentworth, the sediments may be divided into the following main classes: *boulder, cobble, pebble, granule, sand, silt and clay.* With regard to the sandy sediments, a more detailed classification was introduced: *very coarse sand* (2 ÷ 1 mm), *coarse sand* (1.0 ÷ 0.5 mm), *medium sand* (0.5 ÷ 0.25 mm), *fine sand* (0.25 ÷ 0.125 mm) and *very fine sand* (0.125 ÷ 0.063 mm).

The size frequency distribution can be presented in tabular form, in the form of histograms, or as size-frequency curves or cumulative curves. The differences between the particular distribution of analysed sand grains and a normal distribution may be used for the interpretation of sands and also for differentiation between various sand characteristics.

Folk and Ward (1957) proposed a few statistical parameters which are commonly used for quantitative comparisons between the normal and a given distribution. These parameters are: *inclusive graphic standard deviation, inclusive graphic skewness* and *inclusive graphic kurtosis.*

The inclusive graphic standard deviation expresses the sorting value. Lower values indicate the better sorted sediments and higer value the poorly sorted.

The asymmetry of grain size distribution (Fig.77) is determined by skewness. In the case of positive skewness, there is a long tail of fine sand in comparison with a normal distribution, and for a negative value of skewness a long tail of coarser grains appears on the graph. Kurtosis represents the peakedness of grain size distribution.

The cumulative curves which approximate to a Gaussian profile are referred to as *mesocurtic*; curves which are strongly peaked are named *leptocurtic* and the flatter curves are called *platycurtic.*

These parameters help to classify the sand and the degree of sorting and to determine the mixed sand as: *unimodal, bimodal* or even *polymodal.*

The grain shape or form describes the external morphology of a grain including the *gross form* (*sphericity*), the *roudness* (sharpness of edges and corners), and the *surface texture* (roughness or smoothness).

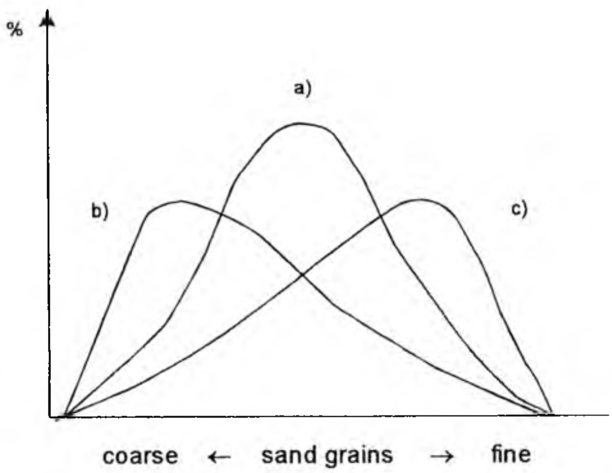


Fig.77 Asymmetry of the grain size distribution (*skewness*),
a) normal distribution, b) positive *skewness*, c) negative *skewness*.

The sand grain roundness is divided into the following classes: *very angular*, *angular*, *subangular*, *subrounded*, *rounded* and *well rounded*. The „well rounded" grains approximate roundness and sphericity.

In the remote sensing approach to sand desert investigations, *the grain surface texture*, which determines grain *microrelief*, is very important. This is the degree of the surface smoothness or roughness. The grain surface presents many microfeatures which are strongly related to the sediment source, weathering and the agent of transport. Some grains are very well polished and give very good light reflection. Generally, the grains which have been very gently abraded or have a coating of secondary silica, have a high degree of polish. Those grains, the surface of which contains irregularities (frosting) due to the abrasion and/or chemical etching, cause the scattering or diffusion of light.

The shape of sand grains is a function of the sand composition, sand source and medium and time of transport. Most of the sands incorporated in sand desert areas has experienced several cycles of erosion and deposition in different climatic conditions. As a result the grains are mostly sub-rounded to well-rounded.

The size grades as well as the roundness of individual grains provide a measure of erosional and depositional rates. Sediments with grains which are poorly sorted and angular in shape, with little or no rounding, are typical of areas which have undergone relatively rapid erosion and sediment transport and deposition. The opposite of this well sorted and well rounded sediments reflects a long history of the reworking processes within the repetetive cycles of erosion, transportation and deposition. There is little oportunity for size reduction through chemical or mechanical processes (solution or abrasion of grains), nor is there sufficient reworking by currents to sort the sediments.

The rounding of sand grains by aeolian abrasion depends on the hardness of the different sources of the material. The rounding depends also on the number of recycling processes, on the number of cyclic depositions and reworkings during erosion and transportation.

Considering the sand parameters, the sands may be divided into: *immature, mature and supermature. Immature sands* are formed when sediments are rapidly deposited in the subenvironments such as alluvial fans or river floodplains. In immature sands, the grains are usually angular and poorly sorted and represent a wide range of sand sizes. Mature sands are clay-free and the grains are subangular, well sorted and have nearly uniform particle size. The supermature sands are clay free, well-sorted and the grains are rounded.

3. CHARACTERISTICS OF SAND INCORPORATED IN AEOLIAN FORMS

The different aeolian forms such as sand sheets, interdune areas and dunes are characterized by different grain size and sorting. The pattern of grain size distribution indicates, to some extent, the mode of aeolian forms development and genetic relationships. The grain size and sorting can change across individual sand sheets and dune fields. There are also differences between the various types of dunes and also within particular dunes. Generally, the regs, sand sheets and interdune areas consist of poorly sorted and rather coarser sediments. Sorting of grain sizes depends mostly on the mode of transport. The certain modes of sand transport move grains of particular size predominantly. In case of poorly sorted fluvial sediments, with a wide range of particle sizes, the wind selects the intermediate-sized grains, leaving the coarser and finer material behind because they are less mobile. The selected grains can than be moved in saltation mode (between 2.4 and 2.7 Φ) and piled up to form the distinct dune forms.

The sand drifts and sand sheets, which cover vast areas of deserts, have a quite different grain size characteristics and genesis.

3.1. Sand drifts, sand sheets and giant ripples (zibars)

The sand drifts are generally developed by aeolian sand on the lee - side of desert shrubs or other small topographical barriers. They are associated with both deflational and depositional areas. In the Libyan Sahara they can be recognized on satellite images over vast areas of the south - eastern part of Libya, mainly in the vicinity of Al Kufrah, where they are associated with gravel deposits. The sand drifts are generallty characterized by unimodal grain size.

The sand sheets constitute an integral part of sand desert areas and other aeolian depositional systems. The American Geological Institute defined sand sheet as :*a thin accumulation of coarse sand and fine gravel formed of grains too large to be transported by saltation, and characterized by an extremely flat or plain-like surface broken only by small ripples.* They may exist as independent aeolian sand areas (e.g. the Selima Sand Sheet in Egypt or the sand sheets in Kuwait) or within the sand seas or sand fields, were they occur on the trailing edges, progradational edges and lateral fringes. Sand sheets represent widespread blankets of aeolian sediments. Generally, the sand sheets are marked by an extremely flat surface and there is an absence of any topographical relief. Sand sheets constitute areas of aeolian sand where dunes with slipfaces are generally absent. The surfaces of sand sheets can be rippled or unrippled. In some localities the sand sheets are covered by gentle undulations called *giant ripples*. Sand sheets represent significant *low - angle aeolian deposits*. Sand sheet deposits are easy to recognize in ancient sequences and are characterized by horizontal or low angle aeolian stratification (Kocurek &Nielson, 1986). The sand sheets have a range of a grain sizes which is different of that of dune fields.The grain size distribution is greatly influenced by the source of sediment, the nature of environment and the wind regime.

The outstanding example of sand sheet is the famous Selima Sand Sheet, which is practically flat, and covers at least 100,000 sq.km of southwestern Egypt and northern Sudan (Maxwell and Haynes, 1989). The sand sheet is typically only few centimetres to a few metres thick and overlies either wind-truncated sediments of alluvial, colluvial and mixed origin, or shallow bedrock (McCauley *et al.*, 1986). The sand sheet deposits are built of paired units of horizontally laminated sediment, each pair containing a lower zone of coarse silt and fine sand, and an upper zone, one grain thick, of coarse sand and granules. The sand sheet deposits constitute a single - paired unit or multiplicity of pairs, with each set a centimetre or less in thickness. Accoording to McCauley *et al.* (1986), the grain size represents an equilibrium of the surface with the strongest winds at each locality.

In many places, the modern, presently-active aeolian unit, overlies an older sand sheet which is only a few tens of centimetres thick (Breed *et al.*, 1987). The sand, incorporated in both the upper and lower sand sheet, has a bimodal distribution and is poorly - sorted. The modes of sand belong to the fine sand class (0.125 mm) and to the very coarse class (ca 1.5 mm). The histogram (Fig.78) illustrates the grain-size distribution typical for sand dunes, sand sheets and alluvium, representing three types of Quaternary sediments which mantle the eastern Sahara.

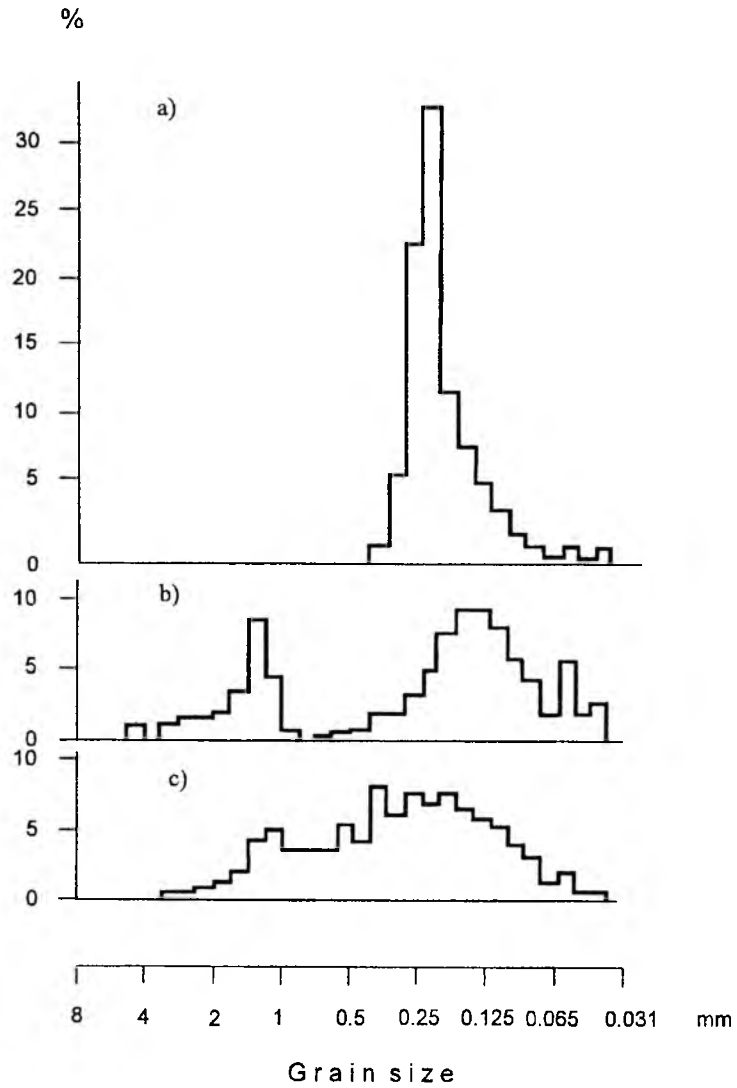


Fig.78 Grain size distribution for:
a) sand dunes,
b) sand sheets and
c) alluvium in the Eastern Sahara (after Breed *et al*, 1987).

The investigations of Maxwell (1982) in the southwestern desert of Egypt, confirmed a bimodal mixture of sand - size grains and coarser lag fragments which vary in diameter from 2.0 to 5.7 mm and are well sorted. The mean grain size of the sand - size fraction averages 0.38 mm and is larger than the average size dune sands. The fine fraction displays poor sorting. Coarse sand forms an armoured surface layer which effectively seals the sand sheet and leave it almost undisturbed.

On the basis of field investigations in the Kuwait desert, Khalf (1989) divided sand sheets into *smooth sand sheets* and *active sand sheets*. The smooth sand sheets occur mainly in flat - lying areas and form relatively flat surfaces covered with a thin blanket of sand. The sand sheets in Kuwait have a wide range of grain sizes. Smooth sand sheets are relatively rich in granules, while the active sand sheets are characterized by a relative abudance of coarse sand. Active sand sheets have a large fraction of medium and fine grain size, which accounts for more than 70% (Khalf, 1989). Generally, smooth sand sheets are coarser and more poorly sorted than active sand sheets. The smooth sand sheets, which are a mixture of two or three different size populations, are generally developed by the accumulation of the desert washout material in lowlying areas which is then reworked by wind action (Khalf, 1989). Active dune sheets are generally unimodal, but bimodal deposits also occur. Deflation of poorly-sorted smooth sand sheets removes the sand population which is then transported via the saltation process. The deflated material from the smooth sand sheets comprises two size populations: medium and fine sand. As the result of deflation, the sand sheet become coarser and this leads to an increase in the skewness of the sediment. Bi - modal smooth sand sheets consist of very fine sand class (about 0.125 mm) and very coarse sand (about 1.0 mm). Coarsening and positive skewness usually increases downwind (Khalf, 1989). The mobile sands of active sand sheets are generally finer and better sorted, due to the continuous wind deflation and transportation process. The sand grains range from 0.17 mm to 0.73 mm.

On the sand sheets, dunes with slipfaces are generally absent, but ripples can occur. Sand migrates in low ripples, the height and spacing of which depends on grain size and wind velocity. In some localities in southwestern and northwestern Egypt, trains of barchans appear on the sand sheets (Breed *at al.*, 1987; Maxwell, 1982) but these dune forms can be regarded as *migrating dune forms* which pass over the Selima Sand Sheet. Breed *et al.* (1987) assumed that the dunes are not necessarily built of sand which has been winnowed directly from the sand sheet, but may have migrated into the region and are now hundreds of kilometres downwind of their former position.

On the basis of studies of six modern sand sheets in North America, Kocurek and Nielson (1986) defined the following factors favourable for sand-sheet development: a high water table, surface cementation or binding, periodic flooding, a significant coarse - grained sediment population and vegetation.

Giant ripples occur in many sand sheets, also, sometimes, in the interdune areas so called *giant ripples* occur. They have no slipfaces and do not avalanche. In the plan view, giant ripples present striped or wavy patterns. Generally they are low - amplitude and long-wavelength bedforms, which are commonly called *zibars* (from arabic term *zibara*). Holm (1960) defined zibars as „rolling transverse ridges without slipfaces". Zibars were recognized by Holm in interdune areas in the central Rub'al Khali in Saudi Arabia. The term *zibar* was used by Warren (1970) for similar forms in the Qoz area in Central Sudan, where the zibars are spaced as much as 400 m apart. Also, undulations on the sandy plains in the Tenere Desert of Niger have been termed *zibars* (Warren,1972). In Algadones dune field in southern California, smaller zibars, which appear on the sand sheets in the interdune corridors, were recognized by Kocurek and Nielson (1986).

Generally zibars occur in areas where a significant portion of sediment is too coarse to form dunes with slip - faces but allows the formation of zibars. Sand incorporated in zibars is coarser than dune sand, the range in grain size being between 0.3 and 0.65 mm; the sorting is usually poor. The stoss and lee - slopes have extremely low angles, i.e. below 5°. The zibar spacing ranges from 60 m in Algadones dune field, through 400 m in Sudan to 500 m in Selima sand sheet. The zibar amplitude varies between 5m (Algadones) to 10 m (Selima sand sheet). The ratio between spacing and height of zibars is of the order of 50:1 (Breed *at al.*,1987).

Giant ripples are oriented generally transverse to the prevailing strong winds. In some cases, the slopes of giant ripples can be covered by minor ripples. As the result of wind action, the sand grains from the up - wind slope can be moved and deposited on the lee - slopes by saltation or creep movement. Such a process is responsible for the migration of entire zibars or giant ripples. Another types of giant ripples are the large scale, low - amplitude bedforms (*chevrons),* detected by Maxwell and Haynes (1989) on the Selima Sand Sheet in Egypt. The chevron patterns, as depicted on Landsat images, consist of irregularly spaced, light - toned, 0.5 to 2.0 km wide streaks generally oriented transverse to the dominant northerly wind direction, but with acute angles which create an *en-echelon* or *chevron* pattern. They have extremely low-wavelengths (130 ÷ 1200 m), low amlitudes (10 ÷ 30 cm) and their spacing varies between 0.5 to 6.0 km, with a mean of 3.0 km. The shape of these bedforms is asymmetric, the steeper slope being on the windward side (stoss side) the converse of the case of the transverse dunes. The difference between the dark and light chevrons is that the topmost layer of the dark chevrons contains a small percentage of 3.5 to 4.5 mm granules. Both dark and light chevrons show the bimodality. These forms have a 10 ÷ 15 cm thick layer of active sand on their lee - side. Along the crest the larger chevrons have 3 to 5 cm - thick layer of active sand which thickenes to 15 cm on the lee - side.

3.2. Sand dunes

There is some evidence that dunes of different morphologic forms are associated with sands of different grain characteristics. Studies relating to the various forms of sand dunes indicate different sand characteristics of linear dunes, sandridges, barchans etc. With regard to a particular dune type, there are usually differences in sand parameters between the crest, flanks and dune base. Usually, the windward flanks have sand charactersitics which differ from the lee - side.

3.2.1. Linear dunes and sandridges

The results of systematic sampling of linear dunes indicate not only some general trends in grain size characteristics, but also some differences which can be related to different locations and different types of particular dunes. These differences can also be related to the differences in sampling methods and to the unprecise location of samples on the dune surfaces. Also, the sand dune characteristics depend on the characteristics of the sand parent material.

Linear dunes, are much influenced by wind blowing from both flanks. Some of these dunes have no slip faces, whereas others have slip faces on one side; there are also some dunes which have seasonally - changed slip faces.

The linear dunes with low slip faces, which are only about 2 m high, have no differentiation of the sand grain population (Sneh and Weissbroad, 1983). The tops and bottoms of such dunes have nearly the same mean grain size and nearly the same sorting value. The dunes with slip faces a little higher (up to 8m) show only insignificant changes of statistical parameters between the bottom and the crest.

Bagnold (1941) identified three different sand populations in the Libyan seif dunes: *a basal zone, an accretion zone (the plinth)* and *a crest zone*. According to Bagnold, the basal zone consists of poorly sorted sand, the accretion zone comprises the median and is relatively constant, and sorting improves with height. As the result of avalanching on the slip face, the coarser grains are carried downwards towards the plinth; hence, the fining and better sorting of sand occurs towards the crest.

Much more detailed information regarding grain size characteristics and sorting over longitudinal sandridges was generated from the results obtained by Lancaster (1981), Livingstone (1987) and Sneh and Weissbroad (1983). In all these cases, the dune sands were shown to be characterized by a limited range of values for the mean grain size from 0.14 to 0.30 mm and distinct patterns of variations in grain size and sorting over the dune cross-profiles. A small range of grain size apparently indicates that the sand was moved predominantly only by saltation. The generalized results are presented graphically in Fig.79.

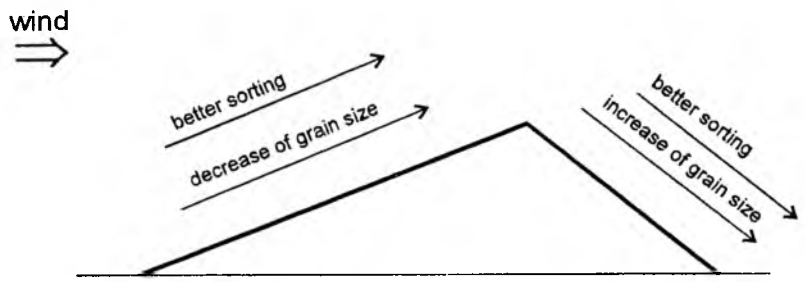


Fig.79 Grain size characteristics and sorting over a longitudinal sandridge.

The results of these mentioned investigations indicated that, on the windward flank, there is a regular decrease in the mean grain size from the base (ca 0.25 mm) through the plinth and flanks up to the crest (ca 0.15 mm). The progressive fining of sand is matched by decreasing values of standard deviation, which determine the sorting value. Generally, the finer crestal sands are also better sorted (fine, well - sorted) than the coarser base or plinth sands (moderately - sorted). Except for the dune base, the sand along the flank comes from one population rather than from a mixture of two or more populations. The mixing of two populations at the dune base results from the combination of saltation and creep load.

The slipface sands show coarsening of the mean grain size from the top to the bottom, and in contrast to the windward flanks, the slipfaces are always better sorted downslope. According to Sneh and Weissbroad (1983) the bottom sands of slip faces are better sorted than the bottom sands of windward (or rippled) flanks.

In their upper parts, the compound dunes have slip faces on both sides due to the winds blowing in two seasons from nearly opposite directions (Lancaster, 1981). Linvingstone (1987) reported that the grain - size values on the dune respond to seasonal changes of the wind.

In most deserts, the dune crests are finer - grained than the flanks and sediment surface on which they lie, but in some deserts, such as the Simpson Desert (Folk, 1971a), Kalahari (Lancaster, 1986) and Kuwait Desert (Khalf, 1989), these relations are reversed. Here, the dunes have coarser crests and finer flanks. According to Folk (1971a), all the dune samples average about 0.155 mm and the interdune samples average about 0.140 mm. The average grain size of a dune crest is 0.175 mm and dune flanks 0.15 mm. Generally, the crests are coarsest and best sorted, whereas the lee flanks are the finest and least well sorted. The windward flanks are mostly coarse, with lesser fines, whereas the leeward sands are a sub - equal mixture of a fine and coarse population. The windward flanks are better sorted than the leeward flanks.

Explanations for reversal of relationships in grain - size distribution have been given by Folk (1971a), who demonstrated the influence of the source material (Fig.80).

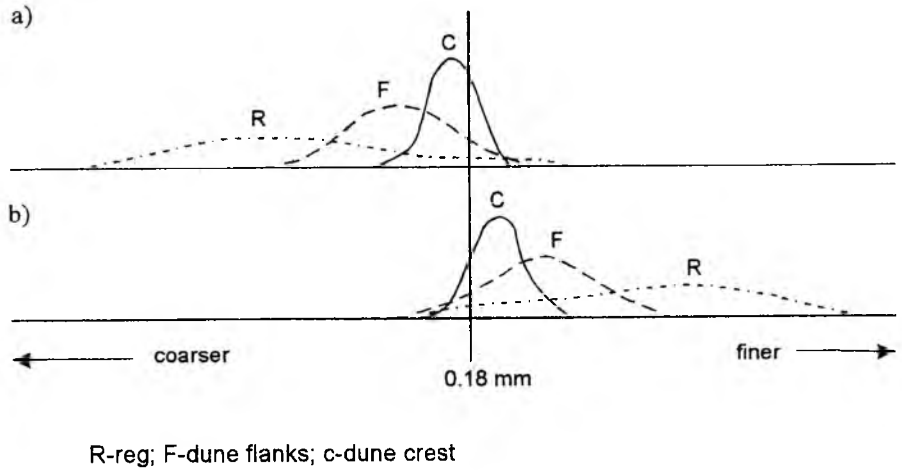


Fig.80 Influence of sand source material on the grain-size distribution (after Folk, 1971).

In the case when the fluvial plain contains much coarser material, the saltating sand will be finer than the average grain size of the source material, and the dunes will be finer than the reg. By contrast, if a fluvial plain contains much very fine sand and silt, the saltating sand will be coarser than the source material and the dune sands will be coarser than the reg. Udden (cited by Folk, 1971a), discovered that the *dune crest sands tend to evolve towards the „fine sand" size* range (between 0.125 and 0.25) mm and that this *grain size is selected out whatever the source material is*.

The grain size characteristics over linear dunes reflect sand movement over the dune surfaces. The sand-moving winds, blowing at an acute angle to the crest of longitudinal sandridges, result in an upslope and alongslope movement of sand. The sand movement results both from saltation and creep. As the result of such movement on the windward flank, the coarser fraction is left at the bottom, while the sands become progressively finer and better-sorted upslope. The windward flanks of the dunes are also influenced by erosional processes, which are common in the case of barchans and transverse dunes. These processes lead to further selection and removal of finer grains. In some cases, ripples can also be formed also on the lee-side of the dune, as the result of alongslope movement of sand in a direction parallel to the dune crest. On the slipfaces, the deposition of sand is formed by sandflow or by grainfall. Sandflow occurs on the lee-sides with high angle flanks (30°÷34°) and the grainfall is related to flow separation, which depends on the dune profile and wind speed. In many cases, the two processes occur simultaneously and the grainfall lamination might then be completely destroyed by sandflow expressed as discontinuous avalanching on the dune lee - sides.

3.2.2. Barchans and transverse dunes

In case of barchan dune types, the windward slopes and horns are coarser than the crest and slip - faces. The *encroachment deposits* are universally finer, and generally more poorly-sorted than the *accretion deposits* and their size distributions show less fine - tailed asymmetry (McArthur,1987). The windward slopes of dunes are commonly mantled with a thin veneer of accretion deposits. Accreation deposits result from a slackening of wind speed during sand transport, and they accumulate both during the deposition of the creep and saltation mode of sand transported on the windward flank. On the lee - side, the sediments are built prominently of high - angle encroachment beds produced by gravity adjustment of wind - deposited sand.

Most of the inland desert dunes are derived from wind acting on poorly sorted fluvial material. The abundance of fine grains in dry regions of fluvial sediments explains the great amount of fine material available in the desert dunes. During very strong winds, silt and very fine sand is carried by wind in suspension and resulting in the so called dust storms. When the wind ceases, the small particles fall out and form *laminae on the dune faces*. Later on, these can be incorporated to the dune, by avalanching of other grains. These fine particles from suspension are then subsequently buried, without winnowing and incorporated in dune deposits.

As has been mentioned earlier, the dunes are better sorted than their parent material, but the sorting depends on the type of dunes and the *recycling of sand grains within the process of dune development*. In the case of barchans and transverse dunes, the dunes migrate. The grains which fall over an alavanche face are buried, and, later, exhumed during the cycle of dune migration. The sand is sorted each time when the grains pass over the crest and down the flanks. During this work, the wind winnows the silt.

3.2.3. Reversed dunes and star dunes

In the case of reversed dunes or the star dunes which are formed by seasonally varying winds, the shape of the dunes may be reversed, and hence the recycling of the sand grains may be repeated many times. As the result of this oft-repeated recycling processes, such dunes are made of very well sorted sand.

The longitudinal sandridges have a "skin" upon a poorly sorted fluvial foundation, and the recycling of sand is related only to the thin upper layer of sand.

4. SAND COLOUR

That desert sands have a reddish hue is a commonly observed feature of sand desert areas. On the other hand is common observation that different sand seas represent different sand colours, and that, within particular sand seas, there are significant differences in sand colour (Norris, 1969; Walker, 1979; Folk, 1971a). For example, the sands of the Great Western Erg are noticeably darker than those of the Great Easten Erg (Capot-Rey and Gremion, 1965) On the basis of dune studies in the Simpson desert Folk (1971a), noted that airborne dust plays an important role in the reddening of dunes, and that the dunes attained redness from the dust which was initially red and was delivered from red lateritic soils developed under former humid conditions. Norris (1969) noted that much of the iron in the pigment is delivered internally from the dunes, and that such processes occur in modern deserts. Detailed studies related to the red colour in the dune sand have been performed by Walker (1979). The results of studies by Norris (1969) and by Walker (1979) indicated that the *dune sands are reddening with age and with distance of transport.*

In the process of aeolian tranport of sand, the fine grains are abraded more slowly than the coarser sand grains; hence, they have more indentations in which they can preserve the pigment. The fine grains show little evidence of abrasion, despite long distances of aeolian transport. As a result of the aeolian transport of grains, the finer grains are usually more heavily stained than the coarser grains. *Fine sand grains* incorporated in dunes are coated with more clay, contain more pigment and *are redder than the coarse grains*. In contrast, many authors have suggested that the additional iron oxide and clay coatings on grain surfaces are developed only when the grains are at rest. Hence, two opposing processes appear to act upon the sand: abrasion during sand transport and accumulation of pigment when the grains are at rest. Fine and very fine particles tend to accumulate more pigment when they are at rest than they lose by abrasion during sand transport.

But the reddening also depends on other factors, such as the amount of moisture and the availability of iron as well as the type and amount of clay minerals retained on sand grains. The amount of moisture plays a very important role increasing the reddening process. The moisture may be rain, dew or ground water. In arid areas, even seemingly - negligible amounts of rainfall can provide enough moisture to cause mineral alterations which result in reddening. The diurnal temperature changes, which are commonly extreme in the Sahara are very important, and, as result, they lead to frequent occurences of heavy dew.

The most important sources of iron for pigment in dune sands are the source rocks which consist of unstable iron - bearing minerals, particularly ferromagnesian silicates such as augite and hornblende. The dune sands derived from such rocks will redden faster than sand derived from rocks deficient in these minerals. In the case when the pigment in the dunes is carried in part by a clay - coating of sand grains, an increase in the amount of clay will cause an increase of reddening.

Wasson (1983) reported that the colour differences in the Strzelecki-Simpson dunefield can be accounted for by differences in the iron content and that the thickness of the grain coatings is also of some importance. In this dunefield, two primary colours have been distinguished: pale and brown, with only a narrow transitional zone. Wasson concluded that the *reddening of dune sand* in the Strzelecki dunefield largely *occurred prior to dune construction.* The red-brown dunes were built by pre - reddened sediments, whereas the pale dunes are built by pale sediments, and that both bodies of sediment were derived from different sources. Hence, the *dune colour ultimately reflects source sediment.* The red sediments from which the red dunes have been derived, are probably older than the pale alluvium from which the pale dunes were formed.

All of these factors, which influence the dune reddening are strongly interrelated, and, in particular differing environmental circumstances result in variations of sand desert colour.

5. CHARACTERISTICS OF THE LIBYAN SAHARA SANDS

The data regarding the granulometric characteristics of Libyan Sahara sands were collected from numerous sources. The general information related to the grain size distribution in Sahara desert as presented on Fig. 78 was taken from Breed *et al* (1987). Evidently, there are significant differences between sand present in the sarir areas, the interdune corridors and the dunes.

In his (1935, 1937 and 1941) papers Bagnold stressed repeatedly that the sands which form the crests of seif dunes were finer than that in their flanks. The crest sands were unimodal at about 0.2 mm and were better sorted than these of the flanks.

Additional information regarding sand particles characteristics were provided by Ahlbrandt (1979). A seif dune located in the Sabha region, of 15 m height, was characterized by a mean grain size of 0.24 mm on the base of slip-face. Another seif dune of 25 m height was characterized by the sand of 0.65 mm on the dune base, 0.27 mm on the windward side and 0.32 mm on the slip-face.

Histograms of grain size distribution in sand of the slip faces of seif dunes, interdune and sarir (McKee and Tibbitts, 1964) indicate large differences in the sand incorporated into these forms. Generally sarir is characterized by bi-modal sand ranging from 0.065 ÷ 4 mm with the peaks between 1 and 2 mm and 0.065 and 0.125 mm. The distribution curves are flatter than the normal Gaussian profile. In the interdune areas the grain size distribution ranges between 0.065 and 2 mm but indicates a unimodal character with the highest number of grains between 0.25 and 0.50 mm. The interdune areas are characterized by the flatest curves (platycurtic). There are large differences between windward sides (low angle strata) and lee-sides of seif dunes. On the low angle strata the grain size ranges between 0.065 and 1 mm, with the highest percentage between 0.25 and 0.50 mm. The distribution curves are very strongly peaked and may be regarded as leptocurtic with a peak between 0.125 and 0.25 mm (up to 70%).

The grain size of sands incorporated in the Libyan sand seas was also analysed within the help of the Geologic Map of Libya at ascale 1:250,000. These analyses indicate that in the Idhan Awbari and Idhan Murzuq the grain size varies between 0.2 and 0.5 m (65%÷75%) and that the grains are well-rounded and well - sorted.

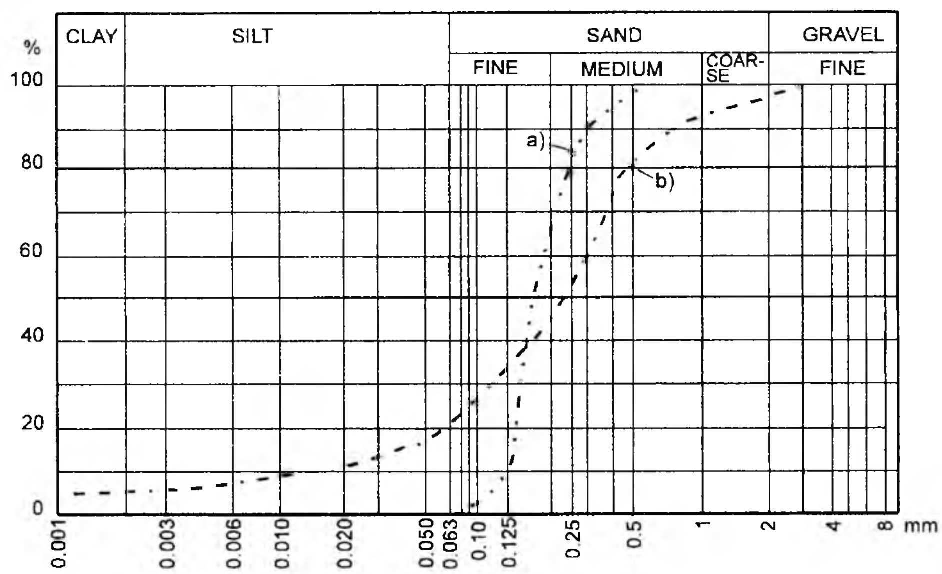


Fig.81 Grain size distribution curves
a) sand dune in Zaltan al Jadid (northern part of Ramlat Zaltan),
b) residual gravelly silty sand of sarir
(adapted from Domaci, 1985).

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TELEDETEKCJA W BADANIACH GEOMORFOLOGICZNYCH PUSTYŃ PIASZCZYSTYCH

(Rejon badań - Sahara Libijska)

Pustynie piaszczyste, stanowiące 20% obszarów pustynnych naszego globu, należą do obszarów o stosunkowo mało znanej i zbadanej geomorfologii.

Sahara, zajmująca powierzchnię około 6 milionów km^2, stanowi jeden z największych obszarów pustynnych na świecie. Jej wschodnia część, zwana Saharą Libijską, obejmuje kilka pustyń piaszczystych zwanych *morzami piasku.* Są to pustynie Idhan Awbari, Idhan Murzuq, Great Sand Sea, Ramlat Rabyanah, Ramlat Zaltan, a wśród pustyń południowo-wschodnich południowe skrzydło pustyń Ramlat Rabyanah oraz Irq al Idrisi. Wymienione *morza piasku* zajmują obszar o powierzchni około 303 000 km^2, i były dotychczas wyłączone z opracowań kartograficznych z wyjątkiem map geologicznych, na których zaznaczano je, jako obszary pokryte piaskami czwartorzędowymi.

Pustynie piaszczyste Sahary Libijskiej, w porównaniu z innymi pustyniami świata, charakteryzują się bardzo szerokim wachlarzem różnych typów wydm. Występują tam np. *wydmy gwiaździste* osiągające wysokość około 160m oraz *równoległe wydmy podłużne* osiągające długość 130 km i wysokość do 120 m.

Autor monografii podjął próbę wykorzystania zdjęć i obrazów satelitarnych do geomorfologicznych badań terenów pustyń piaszczystych Sahary Libijskiej. Celem badań Autora było przeanalizowanie możliwości wykorzystania różnego rodzaju teledetekcyjnych danych satelitarnych oraz technik ich przetwarzania, zarówno do sporządzania map tych rejonów, jak również w badaniach geomorfologicznych. Zakres badań geomorfologicznych objął badanie morfologii form wydmowych pokrywających pustynie Sahary Libijskiej, a także morfogenezy tych obszarów oraz określenia dynamiki zmian zachodzących w czasie.

Monografia składa się z przedmowy, 7 rozdziałów części zasadniczej oraz z 6 załączników.

W przedmowie omówiono problemy związane z pustyniami piaszczystymi i ich wpływem na pustynie obszarów uprzednio zagospodarowanych oraz na globalne i regionalne zmiany klimatu i środowiska.

W rozdziale 1. omówiono pustynie świata, ze szczególnym uwzględnieniem Sahary i Sahary Libijskiej (podrozdz.1.1); dokonano przeglądu dotychczasowych badań geomorfologicznych pustyń piaszczystych (podrozdz.1.2) oraz przedstawiono cel i zakres badań (podrozdz.1.3).

Rozdział 2. poświęcono roli oraz potencjalnym możliwościom zastosowania teledetekcji w badaniach geomorfologicznych pustyń piaszczystych.
W oparciu o analizę zasad określających ruch piasków (zał.5) oraz na podstawie charakterystyki piasków pustynnych (zał.6) określono zakres infor-

macji, możliwych do uzyskania dzięki odpowiednio przetworzonym danym satelitarnym (podrozdz.2.1). Zależy on od charakterystyki spektralnej obiektów, od rozdzielczości wykorzystywanych obrazów satelitarnych jak i od metody oraz techniki ich przetwarzania. Informacje spektralne umożliwiają wykrycie zróżnicowania piasków zarówno na obszarze dużych pustyń piaszczystych, jak i wybranych pól wydmowych a nawet poszczególnych wydm. Możliwość określania geometrycznych form wydm oraz ich cech morfometrycznych jest związana nie tylko ze zdolnością rozdzielczą danych satelitarnych, ale w dużym stopniu zależy od tego, czy są do dyspozycji zdjęcia stereoskopowe. Z kolei dane satelitarne pozyskane w różnych terminach umożliwiają badanie dynamiki zmian zachodzących w czasie zarówno w skali regionalnej jak i lokalnej.

Zbadanie przydatności teledetekcji do geomorfologicznych badań pustyń piaszczystych przeprowadzono na obszarze 5 takich pustyń, stanowiących rejon Sahary Libijskiej. Ta część badań Autora została opisana w podrozdziale 2.2 i obejmuje analizę różnych danych źródłowych oraz różnych technik przetwarzania danych teledetekcyjnych. Do badań w skali regionalnej wykorzystano fotomapy Libii w skali 1:250 000 sporządzone na podstawie obrazów satelitarnych Landsat MSS (podrozdz.2.2.1.1). Fotomapy te pozwoliły na opracowanie map pustyń piaszczystych w skali 1:1 000 000. Na mapach poszczególnych pustyń wydzielono pola wydmowe, charakteryzujące się jednakowymi formami i układami wydm piaszczystych. Fotomapy satelitarne zostały ponadto wykorzystane do określenia parametrów morfologicznych poszczególnych pól wydmowych. Barwne fotomapy pozwoliły na określenie zróżnicowania koloru piasków zalegających poszczególne rejony pustyń występujących na obszarze Sahary Libijskiej (podrozdz.2.2.1.2). Na wybranej scenie Landsat MSS, obejmującej północną część pustyni Murzuq oraz południowy fragment pustyni Awbari, przebadano różne techniki przetwarzania cyfrowych danych satelitarnych podnoszące wiarygodność geomorfologicznych badań obszarów pustynnych (podrozdz. 2.2.1.3).

Poza materiałami z satelity Landsat MSS, Autor w swoich badaniach wykorzystał dane zarejestrowane w trakcie kilku misji eksperymentalnych promu kosmicznego Space Shuttle: zdjęcia fotogrametryczne wykonane kamerą metryczną w 1983 roku, obrazy zarejestrowane w systemie MOMS-1 w 1983 roku, dane uzyskane w systemie MOMS-02 zarejestrowane w 1993 roku oraz dane radarowe SIR-A zarejestrowane w 1981 roku.

Zdjęcia z kamery metrycznej (podrozdz.2.2.2) pozwoliły na analizę form pustynnych występujących w rejonie Idhan Awbari oraz w rejonie Ramlat Zaqqut. Stosunkowo duża zdolność rozdzielcza tych zdjęć (30 m) oraz, dzięki modelowi stereoskopowemu, możliwość obserwacji terenu w przestrzeni trójwymiarowej, pozwoliły na szczegółową analizę zarówno dużych form wydmowych (primary forms) jak i mniejszych tzw. drugorzędnych (secondary forms), będących wynikiem aktualnych procesów wydmotwórczych. Dane te umożliwiły także rozpoznanie i przeanalizowanie niskich form tzw. *zibars* utworzonych z piasków gruboziarnistych pod wpływem wiatrów o dużej energii.

Zdjęcia MOMS-02 (podrozdz.2.2.4), charakteryzujące się największą zdolnością rozdzielczą (4.5 m panchromatyczne, pionowe oraz 13,5 m wielospektralne, ukośne) umożliwiły rozpoznanie i analizę bardzo spektakularnych form wydmowych w rejonie Ramlat Zaqqut (wydmy gwiaździste, wydmy liniowe i barchanowe) oraz w rejonie pustyni Great Sand Sea.

Zdjęcia z satelity SPOT (podrozdz.2.2.5) użyto do badań w rejonie Idhan Awbari. Wykorzystano je do przeprowadzenia analizy zakresu i wiarygodności informacji zróżnicowanych w zależności od przyjętej metody cyfrowego przetwarzania danych teledetekcyjnych. W badaniach korzystano z systemu Intergraph MicroStation.

W podrozdziale 2.2.6 przeprowadzono analizę teledetekcyjnych systemów radarowych pod kątem użyteczności do badania obszarów pustyń piaszczystych. Szczegółową analizę przeprowadzono w oparciu o zdjęcia radarowe SIR-A zarejestrowane nad wschodnią cześcią Idhan Awbari (podrozdz.2.2.6.4, rys.25a i rys.25b).

Cyfrowe połączenie obrazów SPOT ze zdjęciami satelitarnymi pozyskanymi z kamery metrycznej (podrozdz.2.2.7) wykazało, że powstała w ten sposób kompozycja ma znacznie większą czytelność, a więc i większe walory interpretacyjne niż każde z tych zdjęć osobno.

Badanie dynamiki zmian wydm piaszczystych w rejonie Idhan Awbari dokonano w oparciu o zdjęcia z kamery metrycznej z 1983 roku i fotomapy z 1957 roku (podrozdz.2.2.8.1). Przeprowadzone badania wykazały tylko nieliczne zmiany w formach drugorzędowych pokrywających duże formy wydmowe osiągające wysokość 120 m. Na tym samym obszarze porównano obrazy SPOT, zarejestrowane w 1987 roku, ze zdjęciami z kamery metrycznej z 1983 roku (podrozdz.2.2.8.2). Analiza wykazała niewielkie zmiany w formach drugorzędnych oraz wystąpienie kilku małych, słonych jezior. Z kolei porównanie obrazów MOMS-02 wykonanych nad Ramlat Zaqqut w 1993 roku ze zdjęciami satelitarnymi z 1983 roku, wykazało bardzo duże zmiany w istniejących formach wydmowych oraz powstanie nowych form w ciągu dziesięcioletniego okresu (podrozdz.2.2.8.3).

We wnioskach dotyczących możliwości wykorzystania danych teledetekcyjnych do badania pustyń piaszczystych (podrozdz.2.2.9), stwierdzono, że dane z satelitów Landsat i SPOT nadają się do badań prowadzonych w skali regionalnej (makro), natomiast dane wykonane w systemie MOMS-02, z uwagi na wysoką zdolność rozdzielczą oraz możliwość tworzenia modelu stereoskopowego, pozwalają na prowadzenie badań nawet w skali lokalnej (meso i mikro).

Analizę geomorfologiczną przeprowadzono w dwóch etapach. W pierwszej kolejności przeprowadzono analizę geomorfologiczną poszczególnych typów wydm występujących na obszarze Sahary Libijskiej w oparciu o klasyfikację wydm zaproponowaną przez Autora pracy (podrozdz.3.1.2 i rys.31), a następnie dokonano opisu i analizy poszczególnych pustyń piaszczystych (rozdz.4). Zarówno w pierwszym, jak i w drugim przypadku przeprowadzono próbę hipotetycznej morfogenezy poszczególnych form wydmowych jak i całych pustyń. Celem przeprowadzonych badań było opisanie relacji, jakie zachodzą pomiędzy poszczególnymi formami wydm a czynnikami wydmotwórczymi.

Rozdział 5 poświęcono analizie porównawczej poszczególnych pustyń, która wykazała, że pustynie te, oraz występujące tam formy wydmowe, powstały w kilku okresach wydmotwórczych (zał.1 i 2). Aktualne formy są wynikiem wielu czynników wydmotwórczych, do których w szczególności należy zaliczyć: ukształtowanie terenu, źródła piasku oraz warunki klimatyczne zarówno aktualne jak i panujące w okresie plejstocenu i holocenu. Formy powstałe w jednym okresie wydmotwórczym mogły zostać całkowicie lub częściowo zniszczone w następnych okresach lub zostać przemodelowane. Każda z analizowanych pustyń Sahary Libijskiej przestawia inne formy wydm, stanowiące końcowy wynik kompleksowego odziaływania wszystkich czynników wydmotwórczych rozpatrywanych w czasie i w przestrzeni.

Najbardziej intensywne procesy wydmotwórcze na obszarze Sahary Libijskiej nastąpiły w okresie zawartym w przedziale ca 20 000 - 12 000 lat temu, ale istnieje duże prawdopodobieństwo, że proces formowania pustyń libijskich rozpoczął się znacznie wcześniej; ok. 150 000 lat, a może nawet dwóch milionów lat temu.

W rozdziale 6 przedstawiono hipotetyczny model rozwoju pustyń piaszczystych (rys.54), a w rozdziale 7 sformułowano wnioski wynikające z zastosowania metod teledetekcyjnych w badaniach geomorfologicznych pustyń piaszczystych.

W załącznikach podano informacje uzupełniające, które wykorzystano w badaniach opisanych w rozdziałach od 1 do 7, a zwłaszcza w formułowaniu wniosków dotyczących morfogenezy form pustynnych. Informacje te dotyczą: ewolucji geomorfologicznej Sahary Libijskiej (zał.1), zmian środowiskowych i klimatycznych w Afryce Północnej w okresie plejstocenu i holocenu (zał.2), cyrkulacji wiatrów w Afryce Północnej (zał.3), wiatrów wydmotwórczych panujących w Saharze Libijskiej (zał.4), podstawowych zasad ruchu piasku pod wpływem wiatrów (zał.5) oraz piasków pustynnych i ich charakterystyki (zał.6).

АДАМ ЛИНСЕНБАРТ

ДИСТАНЦИОННОЕ ЗОНДИРОВАНИЕ В ГЕОМОРФОЛОГИЧЕСКИХ ИССЛЕДОВАНИЯХ ПЕСЧАНЫХ ПУСТЫНЬ
(Район исследований – Ливийская Сахара)

Резюме

Песчаные пустыни, составляющие 20 % пустынных территорий нашего земного шара, принадлежат к пространствам со сравнительно наименее знаной и исследованной геоморфологией. Сахара, охватывающая пространство 6 млн. км^2, является одним из наибольших пустынных пространств в мире. Её восточная часть, называемая Ливийской Сахарой, охватывает несколько песчаных пустынь, называемых морями песка (Idhan Awbari, Idhan Murzuq, Great Sand Sea, Ramlat Rabyanah, Ramlat Zaltan и юго-восточные пустыни: южное крыло Ramlat Rabyanah и Irq al Idrisi). Пространства этих пустынь, которые занимают около 303 000 км^2, были до сих пор исключены из картографических, а также геологических работ. На геологических картах обозначались они как пространства, покрытые песками четвертичного периода. Песчаные пустыни Ливийской Сахары, по сравнению с другими пустынями мира, характеризуются довольно широким диапазоном разных типов дюн (например, звёздные дюны, достигающие 160 м высоты, обширные продольные дюны, достигающие длины 130 км и высоты до 120 м).

Автор монографии предпринял попытку использования спутниковых снимков и изображений для геоморфологических исследований территорий песчаных пустынь Ливийской Сахары. Целью этих исследований было проанализирование возможности использования каждого вида данных космического зондирования и техник их обработки, как с точки зрения составления карт этих районов, так и использования этих материалов в геоморфологических исследованиях. Объём геоморфологических исследований касался определения морфологии дюнных форм, покрывающих пустыни Ливийской Сахары, а также определения морфогенезиса этих пространств и исследования динамики изменений.

Монография состоит из: введения, 7 разделов основной части и 6 приложений. В введении рассмотрены проблемы, связанные с песчаными пустынями и их влиянием на пустынные пространства предварительно

освоенные, а также на глобальные и региональные изменения климата и окружающей среды.

В первом разделе, состоящем из 3 подразделов оговорены мировые пустыни, с особым учётом Сахары и Ливийской Сахары (раздел 1.1), произведён обзор проводимых до сих пор геоморфологических исследовании песчаных пустынь (раздел 1.2), а также представлены цель и объём исследований (раздел 1.3). Ливийская Сахара характеризуется исключительно сухим климатом, который определяют очень низкие осадки, составляющие в среднем ниже 20 мм в год, очень высокие температуры, достигающие 58^{0}C и большая инсоляция, доходящая до 3 800 часов в году.

Раздел 2 посвящён роли и потенциальным возможностям применения дистанционного зондирования в геоморфологических исследованиях песчаных пустынь. На основе анализа принципов, определяющих движение песков (прилож. 5), а также на основе характеристики пустынных песков (прилож. 6) определён объём информации, которую можно получить из спутниковых данных (раздел 2.1). Объём этой информации является функцией спектральной характеристики и геометрического разрешения применённых систем дистанционного зондирования, как и применения соответствующих техник обработки исходных данных. Спектральные информации дают возможность дифференцирования песков как в масштабе больших песчаных пустынь, отдельных дюнных полей, так и дюн. Возможность определения геометрических форм дюн и их морфологических черт связана как с разрешающей способностью этих данных, так и с возможностью их получения в стереоскопической оптации. В свою очередь спутниковые данные, полученные в соответствующих интервалах времени, дают возможность исследования динамики изменений, происходящих как в региональном, так и в местном масштабе.

Исследование пригодности техник и методов дистанционного зондирования для нужд геоморфологических изучений песчаных пустынь проведено было на территории пяти пустынь, составляющих регион Ливийской Сахары. Эта часть исследований, описанная в разделе 2.2, охватывает анализ различных источниковых данных, а также различных техник обработки этих данных. Для исследований в региональном масштабе использованы фотокарты Ливии в масштабе 1 : 250 000, составленные на основе спутниковых изображений Landsat MSS (раздел 2.2.1.1). Эти фотокарты дали возможность составить карты песчаных пустынь в масштабе 1 : 1 000 000. На картах отдельных пустынь выделены дюнные поля, характеризующиеся одинаковыми формами и структурой песчаных дюн. Спутниковые фотокарты были также использованы для определения морфологических параметров отдельных дюнных полей, а цветные фотокарты дали возможность дюнных полей, а цветные фотокарты дали возможность дифференцирования цвета песков, залегающих в отдельных регионах пустынь, выступающих на территории Ливийской Сахары (раздел 2.2.1.2). На избранной сцене Landsat MSS, охватывающей северную часть пустыни Murzuq и южный фрагмент пустыни Awbari, были исследованы возможности применения различных техник цифровой обработки с целью получения оптимальных результатов с точки зрения морфологической интерпретации (раздел 2.2.1.3).

Кроме материалов со спутника Landsat MSS в исследованиях использованы были данные, зарегистрированные во время нескольких экспериментальных миссий космического парома Space Shuttle: фотограмметрические снимки, выполненные метрической камерой в 1983 году, изображения зарегистрированные в системе MOMS-1 в 1983 году, данные из системы MOMS-02, зарегистрированные в 1993 году, а также радиолокационные данные из системы SIR-A, зарегистрированные в 1981 году. Данные метрической камеры (раздел 2.2.2) дали возможность произвести стереоскопический анализ пустынных форм, выступающих в регионе Idhan Awbari, а также в районе Ramlat Zaqqut. Большая разрешающая способность этих материалов (30 м), а также возможность наблюдения и измерения стереоскопической модели, разрешили произвести более подробный анализ как больших дюнных форм (primary forms), так и меньших, так называемых второстепенных (secondary forms), являющихся результатом актуальных дюнообразующих процессов. Эти данные дали возможность также опознавания и анализирования низких форм, так называемых zibars, созданных из крупнозернистых песков под влиянием ветров, обладающих большой энергией. Данные системы MOMS-02 (раздел 2.2.4), характеризующиеся самой высокой разрешающей способностью (4.5 м в панхроматической версии и вертикально регистрированных изображениях и 13,5 м в многозональной версии или косых снимках), дали возможность опознавания и анализа очень эффектных дюнных форм в районе Ramlat Zaqqut (звёздные дюны, линейные и барханные дюны) и в районе пустыни Great Sand Sea. Данные из системы SPOT (раздел 2.2.5) были использованы для исследований в регионе Idhan Awbari . На основе этих данных проведён анализ различных методов цифровой обработки (на системе Intergraph MicroStation) под углом получения оптимальных информаций. В разделе 2.2.6 проведён анализ дистанционных радарных систем под углом изучения территорий песчаных пустынь. Подробный анализ проведён на основе радарных снимков SIR-A, зарегистрированных над восточной частью Idhan Awbari (раздел 2.2.6.4, рис. 25а и 25в).

Цифровое соединение изображений SPOT со спутниковыми снимками, полученными так называемой метрической камерой (раздел 2.2.7), показало значительное увеличение читаемости этих изображений, динамики изменений песчаных дюн в районе Idhan Awbari проведено на основе снимков, сделанных метрической камерой в 1983 году, и фотокарты с 1957 года (раздел 2.2.8.1). Проведённые исследования показали только незначительные изменения во второстепенных формах, покрывающих большие дюнные формы, достигающие высоты 120 м. На этой же территории сравнены были изображения SPOT, зарегистрированные в 1987 году, со снимками, сделанными метрической камерой в 1983 году (раздел 2.2.8.2). Анализ показал небольшие изменения во второстепенных формах, а также появление нескольких небольших солёных озёр. В свою очередь сравнение изображений, зарегистрированных системой MOMS-02 над Ramlat Zaqqut в 1993 году, со спутниковыми снимками 1983 г. показало очень большие изменения в имеющихся дюнных формах и появление новых форм в течение десятилетнего периода (раздел 2.2.8.3).

В выводах, касающихся возможности использования данных дистанционного зондирования в исследованиях песчаных пустынь (раздел 2.2.9), установлено, что данные со спутников Landsat и SPOT пригодны в исследованиях, проводимых в региональном (макро) масштабе, зато данные системы MOMS-02 ввиду их высокого разрешения, а также стереоскопической оптации, дают возможность проведения исследований в местном (мезо и микро) масштабе.

Геоморфологический анализ был проведён в двух этапах. На первом этапе был проведён геоморфологический анализ отдельных типов дюн, выступающих на территории Ливийскои Сахары, на основе классификации дюн, предложенной Автором работы (раздел 3.1.2 и рис.31). На втором этапе сделаны описание и анализ отдельных песчаных пустынь (раздел 4). Как на первом, так и на втором этапе были сделаны попытки гипотетического морфогенезиса отдельных дюнных форм, так и пустынь в целом. Целью проведённых исследований было поиск реляции, какие заходят между отдельными формами дюн и дюнообразующими процессами. Раздел 5 посвящён сравнительному анализу отдельных пустынь, который показал, что эти пустыни, а также выступающие на них дюнные формы возникли в нескольких дюнообразующих периодах (прилож. 1 и 2). Актуальные формы являются результатом многих взаимонакладывающихся дюнообразующих факторов, к которым следует зачислить рельеф, источники песка, климатические условия, как актуальные, так и господствующие в период плейстоцена и голоцена. Формы, возникшие в одном дюнообразующем периоде, могли быть целиком или частично уничтожены в следующих периодах или могли быть преобразованы. Каждая из анализируемых пустынь Ливийской Сахары представляет иные формы дюн, являющиеся конечным результатом комплексного воздействия всех дюнообразующих факторов, рассматриваемых во времени и пространстве. Наиболее интенсивные дюнообразующие процессы на пространстве Ливийской Сахары наступили в период, находящийся в интервале около 20 000 – 12 000 лет тому назад, но существует большое правдоподобие, что процесс формирования Ливийских пустынь начался около 150 000 лет тому назад, а может быть даже 2 миллиона лет тому назад. В разделе 6 представлена гипотетическая модель развития песчаных пустынь (рис.54), а в разделе 7 сформулированы выводы, вытекающие из применения методов дистанционного зондирования в геоморфологических исследованиях песчаных пустынь.

В приложениях даны дополнительные информации, которые были использованы в исследованиях, описанных в разделах от 1 до 7, а в особенности при формулировании выводов относительно морфогенезиса пустынных форм. Эти информации относятся к : геоморфологической эволюции Ливийской Сахары (прилож.1), изменениям окружающей среды и климата в Северной Африке в период плейстоцена и голоцена (прилож.2), циркуляции ветров в Северной Африке (прилож.3), дюнообразующих ветров, господствующих в Ливийской Сахаре (прилож.4), основным принципам движения песка под влиянием ветров (прилож.5), а также пустынным пескам и их характеристике (прилож.6).

Перевод: Róża Tołstikowa

ERRATA

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