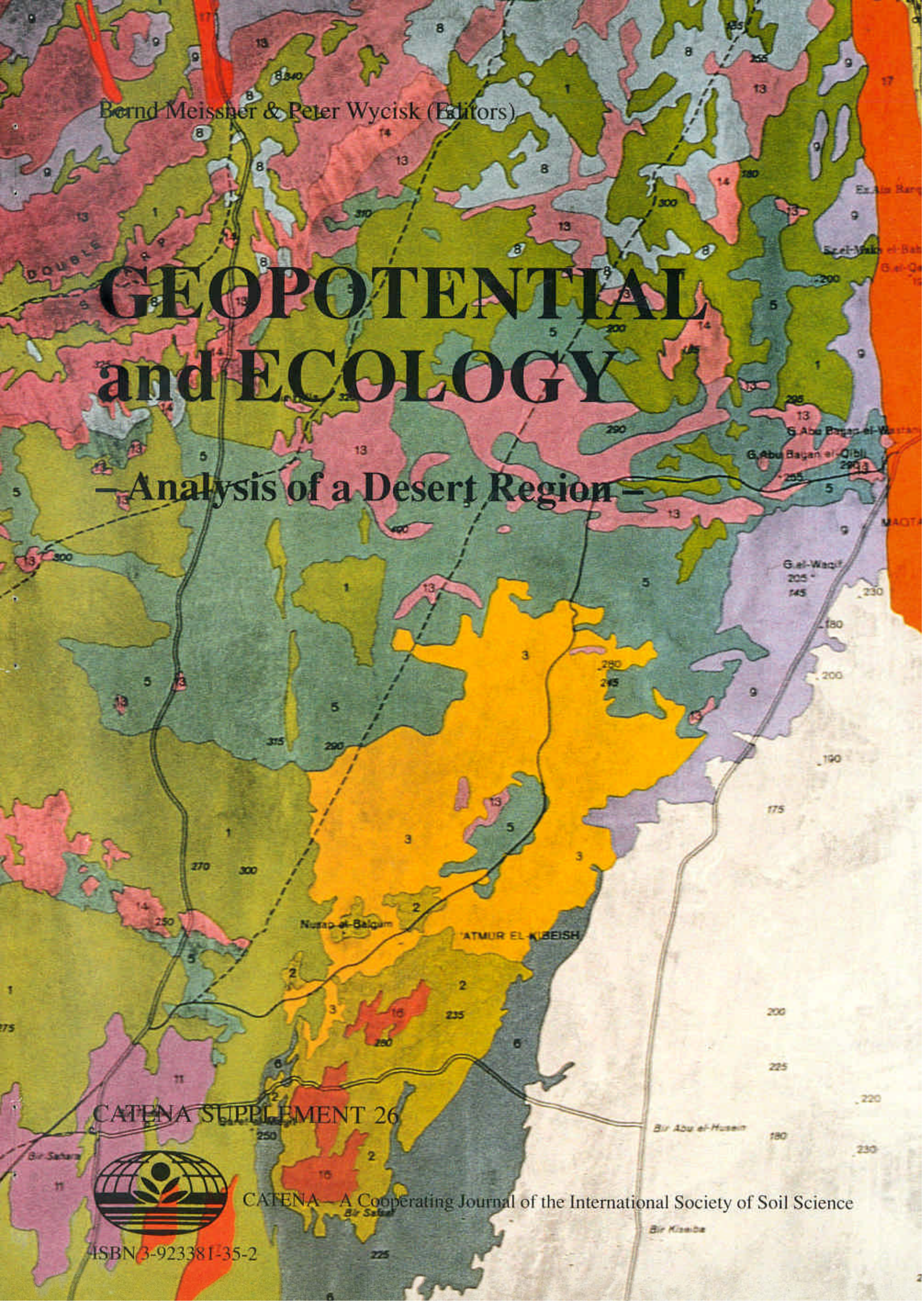

Bernd Meissner & Peter Wycisk (Editors)

GEOPOTENTIAL and ECOLOGY

– Analysis of a Desert Region –

CATENA SUPPLEMENT 26

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Geopotential and Ecological Maps of the Western Desert of Egypt
South-West Egypt 1 : 1 Mio.

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Geomorphology, Landscape Evolution and Paleoclimates of Southwest Egypt

S. **Kröpelin**

Summary

The southern part of the Western Desert of Egypt occupies the centre of the largest hyperarid area of the earth. It has the features of lithology-controlled scarp lands with ridges separated from each other by up to a hundred kilometres. Apart from the Gilf Kebir and the Egyptian Plateau, the region is devoid of strong relief, barren rocky platforms and sandy plains with isolated zeugenbergs being the main geomorphologic feature. Most of the surface is hamada desert and roughly half of the sedimentary cover consists of Quaternary dune, sand sheet, wadi and serir deposits.

The present morphology of the Western Desert is a palimpsest of forms inherited from ancient, contrasted morphogenetic systems and responses to climatic changes. The physiographic framework is largely the legacy of pre-Quaternary fluvial erosion. With most relief features removed by erosion, the large rocky plains may be regarded as remnants of ancient planation surfaces. During the Quaternary, running water has played a comparatively minor role in the large-scale shaping of landforms and has been restricted to the numerous, but short pluvials with episodic wadi erosion and deposition in the foreland of plateaus and escarpments, and playa mud formation in deflation hollows. For most of the Pleistocene, as for the present period, deflation is of prime importance with the products of aeolian erosion either hurled into the atmosphere as dust, accumulated in the form of sand sheets and dunes, or left behind as coarse lag deposits.

Aridity has been the dominating factor in the Quaternary climate in this central part of the eastern Sahara around the Tropic of Cancer. While little is known about the early and mid-Pleistocene environments and climatic conditions, recent data suggest wet phases characterized by a near-surface groundwater level around >300 ka, 210 ka, 175 ka, 160 ka and 140 ka. During the last interglacial (isotopic stage 5; 130–75 ka bp) several episodes of lake development occurred, the most pronounced ones showing an unparalleled faunal spectrum, between 130 ka and

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120 ka ago (possibly corresponding to the Eemian in the strict sense), and others between 104 ka and 70 ka. An uninterrupted aridity in the Sahara during early and mid-glacial times from 70–22 ka bp is still a matter of controversy, although there are several indications of more humid episodes around 45 ka and between 35 ka and 25 ka bp. During the peak of the pleniglacial (last glacial maximum at 18 000 ^{14}C-years bp, about 19 500 BC) as well as during the terminal Pleistocene, the Western Desert witnessed extremely arid conditions until the abrupt onset of humid conditions at 9500 bp (about 8800 BC). The early- and mid-Holocene wet phase lasted about 5500 calendric years with an estimated maximum rainfall of 100 mm per annum. This increase in precipitation created episodic stands of water and an extensive cover of ephemeral grasses and herbs which furnished a basis for faunal migrations, nomadic cattle-keeping and gathering activities, and inter-regional contacts of prehistoric man. Aridification and southward shifting of the Sahelian boundary started around 4500 bp (3300 BC), turning the Western Desert into uninhabitable land, and may not have been without influence in the forming of the pharaonic Nile culture at that time.

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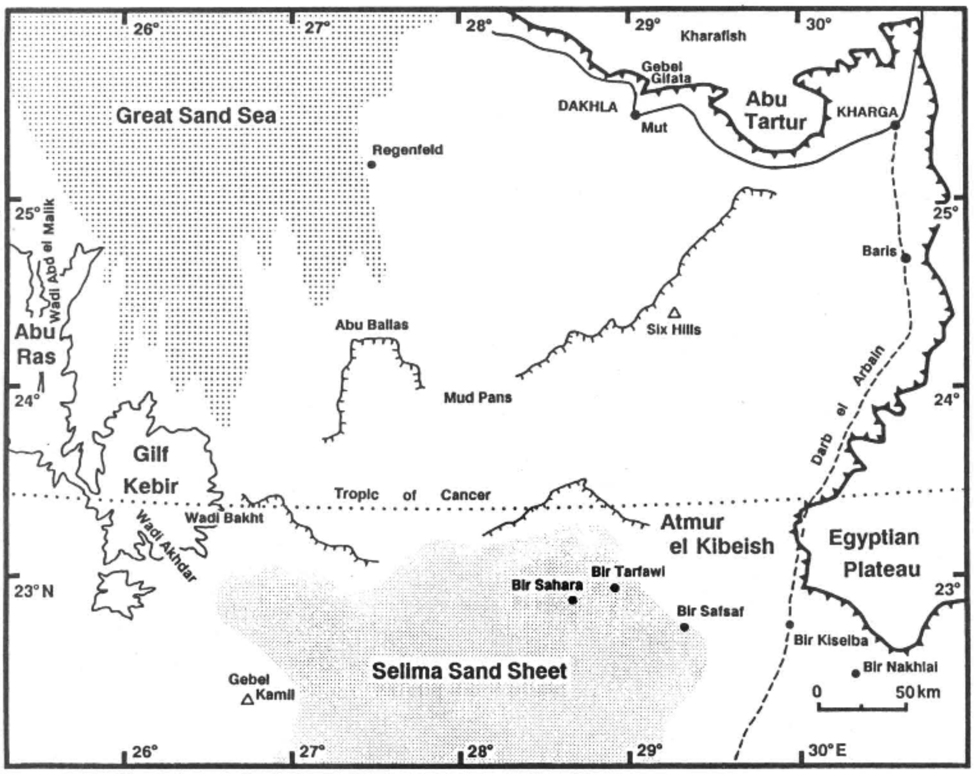

Fig. 1: *Locality map of the southern Western Desert of Egypt.*

1 Introduction

West of the Nile, the Western Desert of Egypt extends over an area of almost 700 000 km^2, more than two-thirds of the total area of Egypt. Its southern part, which is treated here, lies on both sides of the Tropic of Cancer and comprises the extremely dry core of the Eastern Sahara (fig. 1). This is the largest hyperarid area of the earth receiving less than 2 mm of average annual rainfall where the incident solar energy is capable of evaporating more than 200 times the amount of precipitation received (HENNING & FLOHN 1977, KEHL & BORNKAMM this vol.). Owing to the almost complete lack of surface water and vegetation, the area under discussion here is virtually uninhabited outside the oases of Dakhla and Kharga in the New Valley and the cultivated areas around Baris (cf. Topography sheet). The few places with near-surface groundwater all lie within a stretch between Bir Sahara and Bir Nakhlai, roughly parallel to the Sudanese borderline (HEINL & THORWEIHE this vol.). Their occurrence is caused by the crystalline basement of the Uweinat-Safsaf-uplift system which dams up the south-north migrating groundwater (KLITZSCH 1983a).

2 Research background

The ROHLFS expedition of 1873/74 initiated the exploration of the Western Desert and laid the foundation of Egyptian geology and geomorphology outside the Nile valley and the oasis depressions (JORDAN 1876, ROHLFS 1875, ZITTEL 1883). The discovery of the Great Sand Sea is still remembered by the locality name "Regenfeld" (rain field), where the party witnessed more than 16 mm of exceptional rainfall during 2–4 February 1874. Primary reconnaissance and basic surveying of the area mostly took place during the 1920s and 1930s (BAGNOLD 1933, BAGNOLD et al. 1939, BALL 1927, CLAYTON 1936, KEMAL EL DINE HUSSEIN 1928, KING 1912). Modern geoscientific exploration of the vast plains and plateaus which make up more than 99% of the surface of the accompanying maps began, however, only in the seventies (ISSAWI 1971, KLITZSCH 1978, SAID 1975). Fieldwork into the Quaternary geology of the Western Desert aiming at elucidating the paleoclimatic evolution during the late Pleistocene and the Holocene has mostly been carried out by the Joint Research Project "Geoscientific Problems in Arid Areas" at the Berlin universities in cooperation with the General Petroleum Company of Egypt (subproject E1; PACHUR & BRAUN 1980, 1982, PACHUR & RÖPER 1984, PACHUR et al. 1987), and a team from the University of Arizona in cooperation with the Egyptian Geological Survey (HAYNES 1981, 1984, 1985). A short expedition to the Gilf Kebir and Gebel Uweinat supported by NASA and the Egyptian Geological Survey resulted in a comprehensive volume treating several aspects of desert geology and terrestrial analogies for the surface of planet Mars (EL-BAZ & MAXWELL eds. 1982 and ISSAWI ed. 1981; EL-BAZ et al. 1980). The subsurface paleodrainage patterns and sand sheet deposits in the Egyptian part of the Selima Sand Sheet were the target of field trips supported by the U.S. Geological Survey (McCAULEY et al. 1982, 1986; BREED et al. 1987). Basinal sediments in the Dakhla oasis region have been studied by a Canadian working group (BROOKES 1989).

Archaeological studies into the history of Neolithic settlement of the Western Desert outside the oasis depressions of the New Valley have been the subject of several field seasons by the B.O.S. project of the University of Cologne (KUPER 1981, 1988) which also included multidisciplinary studies aiming to reconstruct early- and mid-Holocene environments in SW Egypt (GABRIEL 1986, KRÖPELIN 1989, NEUMANN 1989, VAN NEER & UERPMANN 1989). The Combined Prehistoric Expedition, an international group sponsored by the University of Dallas, the Polish Academy of Science and the Geological Survey of Egypt, has been conducting intensive excavations mostly into the middle paleolithic occupation of the Bir Sahara - Bir Tarfawi area (SCHILD & WENDORF 1981, WENDORF & SCHILD 1980, WENDORF et al. 1990).

Photo 1: *Barren rock surfaces with remnant hills are the main landform of the Western Desert of Egypt. Vehicles for scale.*

3 Geomorphology

The Western Desert is a huge platform with a mean elevation of 500 m a.s.l. consisting of thick layers of sedimentary rock largely unaffected by tectonic disturbances (SAID 1962). Sandstone with a slight northward regional slope and dip makes up the largest part of the exposed and subsurface strata. Carbonate strata are confined to the resistant limestone cap of the Egyptian Plateau (formerly called Libyan Plateau). Within the cut of the sheets, outcrops of crystalline rocks only occur east of Bir Tarfawi (WYCISK this vol.). Exorheic drainage lines are absent and internal drainage is restricted to the depressions. Devoid of strong relief, barren rocky surfaces with remnant hills are the main geomorphologic trait and most of the area can be considered a hamada desert (photo 1).

The highest elevations of the Western Desert are the crystalline Gebel Uweinat (1893 m a.s.l.; not figuring on the accompanying maps) and the Gilf Kebir Plateau (1000 m a.s.l.) in the southwestern corner of Egypt, and the Eocene limestone plateau (400 m a.s.l.) which occupies its south-central part. Other exposures are residual hills (or zeugenbergs, called nusab in Arabic), inselbergs, and minor quartzite ridges or dykes. The dominating negative topographic features are the oasis depressions of Dakhla and Kharga and the deep-cut valleys of the Gilf Kebir. Among the minor depressions on the bare rocky platforms are deflational hollows occurring in most cases in the leeward side of obstacles, and relict shallow drainage lines (HAYNES 1981).

The Western Desert has the characteristics of scarp lands, but differs from others in having ridges which are largely dissected and separated from each other by as much as a hundred kilometres (ABU AL-IZZ 1971). Most of the escarpments and minor cuestas on the sheets are roughly W-E- to NW-SE-trending with the exception of the slopes of the Gilf Kebir Plateau in the west and the limestone plateau in the east with its long north-south trending western edge. The formation of cuestas is closely related to lithology at which silicified or ferruginous sandstone or Tertiary limestone formations are the cliff-makers, whereas silt- and mudstone layers often build up the slopes and the lower-lying parts (cf. Lithology map; WYCISK this vol.).

There is also a close relationship between the geological structure and the drainage pattern of wadis which in most cases are controlled by tectonic lines; depressions tend to occur at junctions. The plains mainly consist of stripped surfaces while truncating surfaces only occur locally. Volcanic features are of minor significance, but ring structures, in some cases falsely interpreted as meteorite impact craters, are quite common in the field as well as in satellite imagery. Other conspicuous features are inverted wadi channels figuring as linear dams which consist of chalcedony gravels and cobbles and maroon mudstones protected by an armour of chalcedony nodules (HAYNES 1982).

Roughly half of the surface of the Western Desert is covered by deposits of Quaternary age which unconformably overlie the Pliocene or older strata (cf. Lithology sheet; WYCISK this vol.). Second in surface area to the bare rocky platforms are aeolian sediments which build up the sand shields and longitudinal dunes of the Great Sand Sea, the Abu Muhariq barchan train totalling more than 600 km, the sand covered slopes ("Hangschleppen" in German) of the Gilf Kebir valleys, and the vast sheet sands and barchan fields of the Selima Sand Sheet. Fluvial and colluvial sediments such as gravelly serir or wadi deposits and sandy alluvial plains ("Sandschwemmebenen") cover a significant portion of the surface of the plains (VAHRSON 1987). Large, often coalescent alluvial fans and slope debris extend for some distance in front of the high escarpments and hills.

So-called wind streaks are a common feature on all satellite imagery of the area and are a characteristic trait of a desert relief controlled by aeolian processes. They mainly occur below escarpments and in the leeward side of mountains and hills and are oriented in the direction of the northerly to northnortheasterly trade winds (cf. Topography sheet). Bright streaks are usually composed of sand dunes and dune belts, sand sheets, and lag deposits of light coloured bedrock, whereas darktoned streaks are predominantly local lag fragments and desert pavement (MAXWELL & EL-BAZ 1981).

Within this general physiographical framework, there are several distinct relief units figuring on the maps which will be explained in more detail in the following. These major landforms are the Egyptian Plateau with the Abu Tartur promontory, the Dakhla and Kharga depressions, the Abu Ballas zeugenberg plains and escarpments, the Gilf Kebir Plateau, the Great Sand Sea, and the vast plains of Atmur el Kibeish and the Selima Sand Sheet.

3.1 Egyptian Plateau and Abu Tartur

The central part of Egypt is occupied by the large limestone plateau lying at 450–550 m a.s.l. Its cuesta landscape is founded on a gently north-dipping sequence of sandstone, shale, and limestone of Cenomanian to Paleocene age which forms the resistant caprock of the gradually northward sloping plateau. At some places such as north of Mut oasis, a gradually dropping scarpland featuring an echelon of parallel, east-west striking minor scarps is developed. In other areas, particularly in the southern, almost level part of the plateau, the upper strata have been eroded to irregular, round or amphitheatre-like depressions exposing a concentrically stepped relief. There are also signs of former drainage systems occurring as dendritic or radial patterns.

The shale-faced main escarpment rises steeply 300–400 m above the Dakhla-Kharga-depressions and extends for more than 200 km on their northern edge. Fluvial dissection has caused an irregular contour of the limestone plateau with several spurs and smaller plateaus projecting into the adjacent depressions. The largest of these subplateaus is the Abu Tartur Plateau which is known for its phosphate mines. Only a few outliers stand as isolated table hills several kilometres apart from the escarpment. Large-scale rotational landslides have occurred along most of the length of the main escarpment of the limestone plateau; pronounced examples are to be seen along the Budkhulu promontory (Gebel Gifata) or at the slopes of the Abu Tartur Plateau (photo 2).

Much of the plateau surface is a very rough and almost untrafficable hamada known as "Kharafish", meaning in Arabic light-coloured hummocky country with wind-polished and furrowed rocks (photo 3). In its most developed form it is made up of numerous elongated hillocks with every portion of the exposed rock-surfaces being deeply scored (BEADNELL 1909). The hillocks are separated by deep troughs half-buried in drift-sand and are lying parallel in the direction of the prevailing winds. In some places trade wind induced troughs ("Windgassen" in German) are superposed upon the former fluvial relief to an extent that west-east-trending valleys are crossed perpendicularly creating a grid pattern. In other places the original drainage direction has even been inverted by wind action. Other curious surface features are boulders called "El Botikh" ("water-melon"), which are up to a metre in length and result from the weathering and aeolian erosion of lower Eocene formations containing numbers of resistant, globular stromatolite concretions.

Karst phenomena are widespread on the limestone plateau and Abu Tartur. They are thought to be responsible for the formation of some of the major and minor depressions which are dotting the plateau. Karstic processes also led to the formation of large caves already reported by ROHLFS (1875) featuring metre-thick stalactites and stalagmites. Pipe-like fillings of terra rossa in solution cavities are the remnants of a former thick and continuous soil over the limestone plateau. The presently highly irregular surface therefore appears to be the result of solution and red soil development under karstic conditions during a warm humid climate possibly in Pliocene time, subsequently truncated and modified severely by aeolian

Photo 2: *Escarpment of the Egyptian Plateau, 300 m high, with landslide features and part of Dakhla depression.*

Photo 3: *Wind-scarved surface of the limestone plateau (Kharafish area).*

processes (HAYNES 1981).

3.2 Dakhla and Kharga Depressions

Extending 70 km WNW to ESE and up to 20 km N to S, Dakhla Oasis occupies a basin below the 140-m-contour between 25° and 26° N. Its edges are conspicuous only in the north with the huge steep escarpment of the limestone plateau, while the south is completely open. The western boundaries are difficult to trace because of the dunes of the Great Sand Sea. In the east there is a dune-covered lowland which reaches as far as Kharga Oasis. Great rocky tongues of the limestone plateau such as Gebel Gifata (or Budkhulu promontory) stick into the depression creating "bays" of bottomland which penetrate northward and carry some wadis (ABU AL-IZZ 1971). The bottom of the Dakhla depression is higher than that of the Kharga depression with its lowest point at about 90 m a.s.l. and is devoid of mountain blocks except for the Gebel Edmonstone, an outlier plateau named after the first European to visit the depression in 1822 (see Topography sheet). The two major physiographic units of the depression are a southern sandstone plain and cuesta at 140–160 m a.s.l. and a central shale-floored lowland at 90–140 m a.s.l. (BROOKES 1989).

The Kharga depression is located about 120 km east of the Dakhla depression and some 150 km west of the Nile. It stretches in a north-south direction between latitudes 24° and 26°N along the abruptly rising western edge of the Egyptian Plateau. The 220 km long and 15-80 km wide depression is only 1 m above sea level at its lowest point and encloses at least 3000 km^2. However, it is difficult to establish the precise total area since the western and southern boundaries are not as clearly defined as the steep escarpments forming its eastern and northern borders.

The western edge consists of level sand with sand dunes and a high wall in the northwestern part which is strongly dissected by deep wadis descending from the plateau. To the south, the depression is open without any distinctive features. The northern boundary consists of the steep limestone scarp ranging from 370 m in the west to 355 m in the east where a steep valley emerges from the northeastern corner of the depression. In this area, the Abu Mohariq barchan dunes descend from the limestone plateau to form a continuous, 200 km long dune strip which covers a large portion of the floor of the depression. The 400 m high escarpment forming the eastern border of the depression is clearly the highest. It is cut by steep, deep wadis which serve as passes leading to the most important ancient caravan road crossing the eastern Sahara, the Darb el Arba'in[1].

Along a north-south axis between Kharga and Baris, parallel to the plateau scarp, extends a series of oases and wells (SCHWEINFURTH 1875). During the more humid periods of the early Holocene, discharge from springs, coupled with runoff

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[1] Influenced by this route, HAYNES (1981) has proposed the name "Arba'in Desert" to designate the particularly dry area (formerly called Libyan Desert) lying south of the Eocene Limestone Plateau, west of the Nile, north of Wadi Howar in Sudan and east of the border with Libya and the Ennedi Mountains in Chad, and covering about 400000 km^2.

from the surrounding plateau, was sufficient to form extensive but shallow lakes which are evidenced by aerodynamically shaped remnants (yardangs) of extended playa sediments and wind-scoured spring vents in various parts of the depression (CATON-THOMPSON 1952, WALTHER 1912). These flat siliceous mud deposits are typical features of the southern half of the Western Desert but are most prevalent in the southern part of the depression. Though by far the largest in extent, the Kharga playa deposits have received only little attention (EMBABI 1968/69, WENDORF & SCHILD 1980). Artesian springs are known to have occurred in Kharga up to Acheulean times (CATON-THOMPSON & GARDNER 1932), but the ground water table has dropped for about 10 m between Roman times and the middle of this century (MURRAY 1952), with a significant acceleration during the past few decades (HEINL & THORWEIHE this vol.).

3.3 Abu Ballas area

The Abu Ballas area derives its name from a conical hill (Pottery Hill or Abu Ballas meaning "father of jars" in Arabic) at which more than a hundred large pottery jars, probably dating from the New Kingdom around 1200 BC, have been found (BALL 1927, GABRIEL 1986). It is typical of the geomorphology of the Western Desert east of the Gilf Kebir and consists of nearly flat, largely sand-covered plains with isolated conical and flat-topped sandstone hills separated by minor escarpments. This landscape of sand sheets and erosional remnant hills (nusab), on an average 50 m high, may best be termed a zeugenberg plain. Structural control is obvious in only a few places and many structures are truncated. Deflational hollows, mostly occurring on the leeward side of the hills, range in size from some square metres to many square kilometres and frequently bear playa deposits. Alluvial fans are relatively small and essentially inactive. All exposed rock surfaces and stones have been polished, pitted, grooved, or frosted by persistent unidirectional sand-blasting producing a variety of abrasion features and ventifacts (windkanter) in proportion to their time of exposure (HAYNES 1981).

The generally west-east trending Abu Ballas escarpment sensu strictu is about 50 km long and 100 m high. Following its projecting and receding course, however, it can be traced almost to the foreland of the Abu Tartur Plateau in the east, and to the Gilf Kebir Plateau in the west. The whole length thus amounts to about 500 km. The scarp is relatively well defined only southwest of Abu Ballas; elsewhere it is in advanced stages of disintegration. Along the foot of the escarpment is a chain of discontinuous subsequent basins, into which open short, steeply walled valleys from the north and flat-bottomed wadis from the south. Rocky sandstone promontories, outliers and swells, as well as kilometre-long falling dunes and barchan trains, divide the depressions into numerous subbasins which are occupied by extensive, early- to mid-Holocene playa deposits. The yardang fields below the western margin of the Abu Ballas scarp are among the most conspicuous; they exhibit advanced deflation of originally at least 8 m thick pelitic and psammitic accumulations formed by episodic playa-lakes during the first third of the Holocene (photo 4; PACHUR et al. 1987, VAHRSON 1980).

Photo 4: *Abu Ballas escarpment and relict playa deposits aerodynamically shaped into yardangs, 2–5 m high, with blunt side facing northerly trade-winds.*

Photo 5: *Valley in the eastern Gilf Kebir Plateau (Wadi el Bakht). Width of valley is 650 m, depth 100–130 m. Trucks for scale.*

The formation of fine-grained deposits described as "a chain of ancient lakes or mud-pans" (BAGNOLD et al. 1939) in the eastern depressions seems to be older and more complex than that of their western counterparts. Only the decimetre-thick top layers which bear abundant neolithic artifacts (KUPER 1989) appear to have been deposited during the first half of the Holocene. They are largely composed of detrital Tertiary shale and show features of repeated soaking and weathering (KRÖPELIN, unpubl. data). Excavations along the littoral slopes yielded an exceptionally rich spectrum of early and mid-Holocene faunal remains (VAN NEER & UERPMANN 1989) and charcoal of savanna vegetation (NEUMANN 1989; cf. chapter 5.3). Neighbouring outlier hills also feature some of the rare occurrences of rock engravings in the Western Desert.

3.4 Gilf Kebir Plateau

The Gilf Kebir is a huge flat-topped sandstone plateau with occasional basaltic intrusions located in the remotest and most arid part of the Western Desert (KLITZSCH 1979). With its northwestern part, the so-called Abu Ras Plateau, it is 12 000 km^2 in size, i.e. one third larger than the island of Corsica. The height of the plateau declines northwards from 1050 m in the south to about 900 m a.s.l. in the north until it merges into the plains of the Great Sand Sea. Bounded by steep to almost vertical cliffs with an estimated total frontage of more than 3000 km (BAGNOLD et al. 1939), the plateau rises more than 300 m above the surrounding desert floor. Fronting the irregularly dissected plateau scarp are numerous residual interfluves which consist of flat-topped and conical buttes of varying sizes evidencing significant gradual scarp retreat (McCAULEY et al. 1981). Prominent examples are the hills of El-Aqab el-Qadim and Gebel el-Ailam in the eastern foreland (cf. Topography sheet).

The predominantly west-east trending valleys of the Gilf Kebir show rather angular courses and are devoid of significant catchments. In cross-section the wadis are typically flat-floored with remnants of drainage channels in their alluvial fill. They are bounded by invariably steep cliff walls with an average slope angle of 35° which become steeper to the top and culminate in a vertical or nearly vertical lip (photo 5). The often amphitheatre-like heads of the wadis are very abrupt and often consist in a sheer cliff, frequently with signs of former sapping at the cliff foot (MAXWELL 1981). In contrast to other Saharan mountains, more elevated wadi terraces are almost absent in the valleys of the Gilf Kebir Plateau with one prominent exception at a wadi junction in Wadi el Maftuh where there is a more than 100 m wide and about 3 km long accumulation terrace rising 10-15 m above the wadi floor (KRÖPELIN 1989).

Due to the constant trade winds blowing roughly north-south and a source of sand in the north (Great Sand Sea), dune barriers are partly or completely blocking the upper courses of the west-east trending valleys at several locations. During the early and middle Holocene, the combination of these factors has enabled the accumulation and preservation of fine-grained playa deposits, which do not occur in the adjacent Abu Ras Plateau with its north-south-trending valleys, where only undifferentiated wadi alluvium has been found. The most significant playa accumulations outcrop in Wadi el Bakht and cover an area of some 65000 m^2 with a thickness of more than 8 m. They are composed of thin alternating layers of siliceous mud and partly cemented sands which point to episodic pools exclusively fed by surficial run-off and containing water for weeks or months at most (KRÖPELIN 1987, 1989; PACHUR & RÖPER 1984).

3.5 Great Sand Sea

The Great Sand Sea of Egypt covers a surface of more than 70 000 km^2 and is roughly delimited by the 400-m-contour. A basin position is also indicated by seismic surveys and a test well in the northern part showing sand thicknesses of 120-300 m (HAYNES 1982). This suggests a giant wedge of sand in the north thinning out in the south. The edges of the Sand Sea vary between an abrupt limit with the virtually sand-free rocky plains in the southwest and a gradual feathering out in the southeast, a result of regional topography and aerodynamics.

The most characteristic features of this erg are parallel longitudinal dunes with steep-sided ridges rising up to 100 m and lengths sometimes exceeding 100 km. These dunes are called "seif" after the Arabic name for sword. They are superimposed upon so-called "whalebacks" (BAGNOLD 1941) which are continuous dome-like dunes 1-3 km wide and 50 m high, running in straight lines for distances of the order of 300 km in a general direction of southsoutheast to due south at the southern end of the Sand Sea (photo 6). These plinth-like sand bodies with overlying seif dune chains running along their edges are separated by regularly spaced interdune corridors 1–10 km wide, called "gassi" in Arabic. In the southern part these streets expose wind-abraded bedrock, partly shales with silcrete-like pavement, and gravelly alluvium, but northward they disappear under a continuous sand shield consisting of medium to high rolling dunes with occasional crests. Consequently, only a minor part of the surface of the erg (totalling only 7 % after STEFFAN 1983) is actually covered by fine-sandy dunes.

While the weakly consolidated whalebacks with their very pale yellow surface armour of finer pebbles and coarse sand have to be considered as quasi-stable and pre-Holocene because of onlying Paleolithic artifacts, the steeper sided, yellow to reddish-yellow and loose seif dunes are active. Their slip faces which are low in relation to the total height of the dune, occupy only the upper part of the slope; running roughly parallel with the prevailing wind and away from the cross-winds, they are therefore a temporary phenomenon. Seif dunes occur when the wind regime is such that the strong winds blow from a quarter other than that of the general drift of sand caused by the more consistent gentle winds (BAGNOLD 1941). The bi-directional hypothesis of seif dune formation consequently assumes that cross-winds are essential while barchans occur where the wind is nearly uni-directional. Crucial for seif formation is therefore not the average wind direction (i.e. the northerly trade winds) but the wind force of occasional winds from mostly easterly directions, as significant sand movement only sets in at wind speeds of about 40 km/h.

Photo 6: *Great Sand Sea. Longitudinal (or 'seif') dune, up to 100 m high and more than 100 km long, resting on whaleback.*

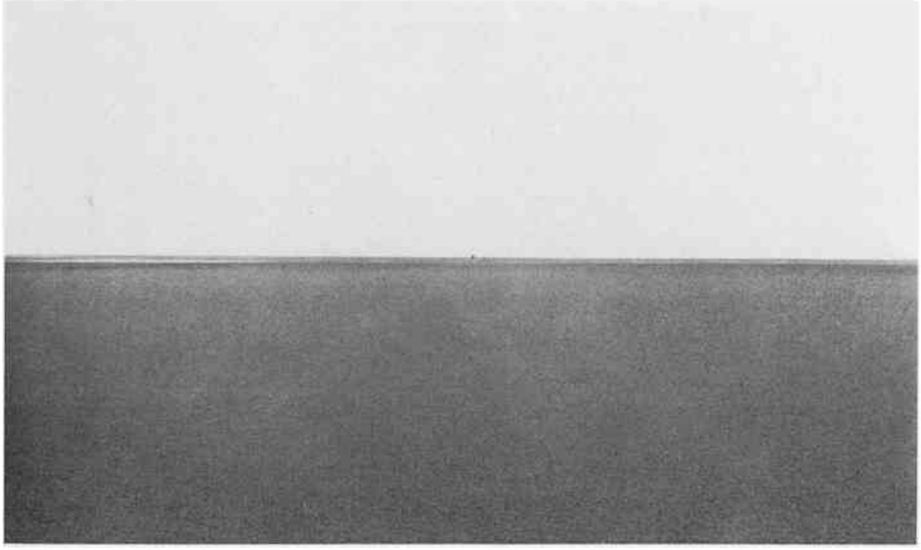

Photo 7: *Featureless flat of the Selima Sand Sheet. Trucks for scale.*

Ancient drainage lines extend from the Gilf Kebir highlands into the Sand Sea following the regional gradient. Some appear as inverted channels which mainly consist of quartzite pebbles and cobbles and raise up several metres above the floor of the dune corridors. Siliceous playa deposits, sometimes overlain by dunes, indicate ephemeral ponds in the Sand Sea during the early and mid-Holocene (HAYNES 1982, JUX 1983). Widespread calcretes, extending as far as 26°30′N, show the existence of lakes and a near-surface groundwater table during late Pleistocene wet phases (PACHUR et al. 1987).

A crater-like circular feature about 4 km in diameter and centering at 24°12′N–26°39′E occurs among the dunes in the sandstone. A complex structure with a flat floor, a terraced wall and a crenulated rim, it was thought to be an impact crater (EL-BAZ 1981) but is in fact volcanic in origin. Noteworthy is also the occurrence of the so-called Libyan Desert Silica Glass in the southwestern part of the Great Sand Sea. This material consists almost entirely (98%) of amorphous silica and is not known from any other location. Inclusions of neogene microfossils, mainly plant tissues and sporomorphs, and its in-situ occurrence in desiccation cracks of neogene lacustrine deposits imply a terrestrial origin from colloidal mobilisation and final gelation of silica derived from biogene siliceous skeletons during a warm arid climate (JUX 1983). Fission track dating yielded ages of 28.5 ± 2.3 Ma (GENTNER et al. 1969).

3.6 Atmur el Kibeish and Selima Sand Sheet

Nearly flat or gently undulating sandy plains cover extensive areas of Southwest Egypt and Northwest Sudan. Atmur el Kibeish ("plain of the sheep") is a nearly level plain south of the Kharga depression with elevations between 200 m and 300 m a.s.l. It is surrounded by more broken ground with elevations from 300 m to 400 m to the north and to the south where it is bordered by the north-northeast to south-southwest trending Kiseiba escarpment. The plain is blanketed by horizontally-laminated deposits of sand, silt, granules, and pebbles, which are typically armoured by a lag surface layer of the coarsest particles (BREED et al. 1987).

The "Great Selima Sand Sheet" (BAGNOLD 1933) is a virtually featureless desert plain covering more than 60 000 km^2 of southwestern Egypt and northwestern Sudan. The gently northward sloping sand sheet consists of a compacted layer of wind-sorted sand generally not thicker than 50 cm. While in some places there is a gently undulating, subtle low relief of about 10 m (HAYNES 1982), the Egyptian part of the sand sheet is almost perfectly flat except for a few isolated rock outcrops and occasional barchan dune trains (Photo 7). A 57-year record of dune movement over the sand sheet shows an average advance rate of 7.5 m/yr for a 200 m long, 275 m wide and 20 m high barchan with an estimated weight of 760 000 t (HAYNES 1989).

Late Pleistocene calcretes occur in isolated patches within many shallow depressions below a thin drift sand layer in the northern Selima Sand Sheet between Bir Tarfawi and the Gilf Kebir (PACHUR et al. 1987) and are not confined to the areas around Bir Sahara, Bir Tarfawi and Bir Dibis previously outlined by ISSAWI

(1978) and SAID (1980). Up to several 100 m^2 in size and 5–30 cm thick, the strongly consolidated, monogenetic formations which lie upon loose aeolian sands are of the groundwater to lacustrine type, and different from the complex, more than 1.5 m thick calcretes in positions with near-surface groundwater such as Bir Tarfawi and Bir Dibis (RÖPER 1988).

Field observations and geophysical work in the Selima Sand Sheet have shown that the surficial deposits of aeolian origin, typically only a few centimetres to metres thick, with a maximum of about 10 m, disconformably overlie either wind-truncated sediments of alluvial, colluvial, and mixed origin, or shallow bedrock (McCAULEY et al. 1986). The sand sheet, although the dominant landform today, is therefore only an aeolian overprint on a much earlier fluvial landscape (cf. chapter 4).

4 Landscape evolution

The present morphology of the Western Desert, like that of most desert landscapes, is a palimpsest of forms inherited from earlier, contrasted morphogenetic systems and responses to climatic changes. With most relief features removed by erosion, the large rocky platforms of the Western Desert may be regarded as remnants of ancient planation surfaces.

Isolated flat-topped and conical hills on the pediplain east and south of the Gilf Kebir Plateau which are interfluve remnants of wadi walls, as well as the recent evidence of a subjacent fluvial topography beneath the Selima Sand Sheet (the so-called "radar rivers"; McCAULEY et al. 1982, 1986) corroborate earlier views that the physiographic framework of the Western Desert is largely the legacy of pre-Quaternary fluvial erosion. Streams of probably middle to late Tertiary age have drained the Gilf Kebir highlands in all directions, the plateau acting as a major drainage divide in southwestern Egypt (McCAULEY et al. 1981).

The succeeding drainage system was related to the late Miocene Messinian event about 5 million years ago, when the Mediterranean became desiccated and the modern Nile incised. Remains of this drainage pattern are obscured by presumably later Pliocene alluvial deposits which were bevelled by aeolian erosion during the extremely arid episodes of the Pleistocene (SAID 1983).

In the Quaternary, running water has played a comparatively minor role in the large-scale shaping of landforms. During the numerous pluvials (chapter 5), however, episodic wadi erosion and deposition occurred in the foreland of plateaus and escarpments. Even a minor, short-range functioning of the "radar rivers" may have to be taken into account (McHUGH et al. 1988b, 1989). Sheetfloods also contributed to a large extent to the present morphology of the desert surface (SAID 1990a). The wind-striated terrain of the plains displays successive and alternating imprints of running water, mass wasting and wind action. The main factors in the land-climate interrelationships during the Quaternary thus seem to be the alternations from moist to dry and the attendant changes in processes acting upon rocks with differential resistance caused by silification and ferruginization possibly partly resulting from former spring outlets (HAYNES 1981).

The origin of the oasis depressions is a long disputed and still controversial issue (cf. ABU AL-IZZ 1971, WALTHER 1912). Some authors have proposed a more or less purely aeolian origin (BALL 1927, BEADNELL 1909), while others have postulated structural preconditions (KNETSCH & YALLOUZE 1955, PFANNENSTIEL 1953). A more recent hypothesis is a fluvial formation of the only semi-enclosed Dakhla and Kharga depressions by a hypothetical Pliocene river system formerly draining in a southwesterly direction (SAID 1983). As it is obvious, however, that (apart from ice) wind is the only medium able to remove material from depressions, the differences in views mostly concern the remote late Tertiary origin of the oasis depressions and become less important in respect of the recent geological epoch. According to HAYNES (1981), most, if not all, of the depressions are the result of aeolian corrasion and deflation of beds weakened by leaching of cement and salt efflorescence which is a self-enhancing mechanism, once started by the development of an initial area of internal drainage. The driest part of a climatic cycle would not be as effective as the transitional parts from wet to dry, when it would take less time for vegetation and soil to be destroyed than to be reestablished. The occurrence of faults or fractures could augment corrasion not only because of physical weaknesses, but by bringing groundwater to the surface where evaporation and efflorescence would aid the comminution of the rocks.

As can be observed along the edge of the present head scarp above Dakhla and Kharga, the main factors causing scarp retreat and thus the expansion of the Egyptian oasis depressions, after the tectonic premises which account for the primary alignment of faults and the lithological structure controlled by the resistant limestone cap rock, are gravitational landsliding processes due to shearing stress and the lack of abutment (cf. photo 2). Fluvial action results in wadi incision at the crown, collection and infiltration of run-off into transverse cracks and slip planes, and downward transport of eroded slump material to the lowest parts of the depression. Deflation removes the detrital sediments against the gradient. Supplementary factors include sapping and piping along the scarp, and salt weathering at the bottom of the depressions. Thus a complex combination of processes has been involved in the formation of the depressions. To account for the decakilometres of scarp retreat, all these processes must have been effective with varying rates under changing climates from the upper Tertiary onwards. Active transverse cracks along the Dakhla-Kharga escarpments clearly indicate that landsliding is a continuing process even under the extant hyperarid regime.

The origin of the nearly level Atmur El Kibeish plain may be due to the depth of wind scour and deflation being limited by the depth of the water table, which is generally relatively flat in porous sandstone (HAYNES 1981). The surface of these extensive peneplains is not necessarily parallel to the slightly undulatory bedding, but truncates it as well as shallow structures. Lesser irregularities in the surface of the sandstone are filled and thus smoothed by the overlying sand sheets.

Pleistocene alluvial and lacustrine deposits, reworked and transported by aeolian processes probably make up the bulk of the sand sea, sand sheet and sand shield sediments. This implies export of the sands of the Sahara and thus a negative balance during the dry periods of the Quaternary (STEFFAN 1983). Another primary source of the sands in the Great Sand Sea is aeolian abrasion of bedrock sandstone, in particular in the great depressions.

The origin and distribution of the widespread aeolian sand sheet deposits in the Selima Sand Sheet over tens of thousands of square kilometres remain a problem. The depositional processes responsible for these widespread but relatively thin tabular sand bodies are poorly understood, but are thought to be primarily accretion, resulting from surface creep of the coarser particles, and sedimentation, due to grain fall of the finer particles from saltation and suspension (McCAULEY et al. 1986). Key factors encouraging accretion in other regions such as vegetation, high water tables and the presence of dissolved salts promoting cementation of sand ripple laminae, do not apply to the active Selima Sand Sheet. Perhaps the extreme horizontality of structures is a type of planar lamination, which occurs where particles are driven across a smooth surface under high-speed winds (BREED et al. 1987).

The wadis of the Gilf Kebir show features which cannot be explained by a theory of stream erosion alone. To account for the abrupt wadi heads, the hypothesis has been suggested that the bulk of rainfall upon the plateau did not run off over the surface and so erode directly, but sank by piping through the permeable sandstone to reappear at the base of the cliffs as springs (PEEL 1939, 1941). Such groundwater springs would contribute much to basal sapping and account for the lack of water erosion on the lips of the cliffs, which would retreat owing to undercutting and wasting. The same processes may apply along the open edge of the Gilf Kebir plateau as well as its broad "bays". Resistant beds of silicified sandstone over wide areas of the plateau forming a protective capping, the delayed retreat of the upper part of the cliffs contributes toward maintaining the steep profile. There is little doubt that these processes have been much more active at periods of more abundant and more frequent rainfall during the Quaternary pluvials and the late Tertiary than under hyperarid conditions such as the present ones.

A circular depression in the upper Wadi Ard el Akhdar in the southern Gilf Kebir Plateau provides insight into some of the morphodynamic processes occurring in the Western Desert during the first half of the Holocene with its more humid conditions. The sandy to silty strata in the periphery of the episodic paleoplaya lake consist of seven single fanglomeratic deposits which have been washed from the marginal slopes by sheetfloods after exceptionally heavy rainfall (KRÖPELIN 1989). Radiometric ages indicate that these few depositional events occurred between about 9400 and 5000 bp [2] (cf. chapter 5.1), while the intervening long periods only witnessed slight soil formation and mass wasting. Stratified early Neolithic pottery indicates long-term occupation surfaces. Thus comparatively stable conditions with an apparently minor importance of aeolian erosion and deposition have prevailed, while short, but intensive depositional events account for the stratigraphic legacy.

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[2] "bp" refers to conventional uncalibrated radiocarbon years (half-life time 5568 yrs) or U/Th years before present (1950), "BP" for tree-ring- or otherwise calibrated radiocarbon years before present, "BC" to both calendric years and tree-ring-calibrated radiocarbon years before Chr.

Major linear erosion and channelling of the playa deposits only occurred as a direct consequence of the abrupt breaching of the dune blocking the valley. Later erosion by running water during the second half of the Holocene obviously had no more significant effect on the degradation of the playa deposits and the downvalley wadi floor. Noteworthy is also that coarse slope debris mostly occurs on the surface of the paleoplayas while it is scarcely or not at all interstratified with the playa deposits. In conclusion, there seems to have been no general, but only a gradual difference to the present morphodynamics.

Among the active processes deflation is of prime importance, as has been the case for most of the Quaternary, though deflation rates may vary from millimetres to metres per millennium depending on resistance, position and exposure. Wind dismantles scarps, deepens hollows and erodes rocks. The products of erosion are either hurled in the atmosphere as dust, accumulated as sand sheets and dunes, or are left behind as coarse lag deposits (EL-BAZ & WOLFE 1981). Rare desert storm rainfalls like those on December 16/17, 1977 in the Kiseiba-Dungul depression, only result in fresh runnels and mud curls on playa deposits with the water infiltrating through large shrinkage cracks (HAYNES et al. 1987).

5 Quaternary paleoclimatology

There is ample evidence of pre-Quaternary climates fundamentally different from today's, ranging from tillites in the Abu Ras Plateau pointing to a middle Paleozoic glaciation of the Gilf Kebir area (KLITZSCH 1983b) to various Mesozoic tropical-wet soil formations (WYCISK 1984). Presumably late Tertiary paleosoils such as the relict *terra rossa* found in solution cavities on the Egyptian Plateau east of Kharga suggest a pre-Pleistocene karst terrain (HAYNES 1981). Other evidence is the formation of apparently lacustrine silica glass during the Neogene in the south-western part of the Great Sand Sea (JUX 1983). All of these were a consequence of largely different latitudinal positions due to plate drift whereby, for example, the phanerozoic paleolatitudes of Cairo shifted from 70°S to 30°N according to apparent polar wander paths (MORGAN 1990). The Quaternary climatic cycles, however, have been entirely independent of plate dynamics and were controlled exclusively by global climatic change.

In accordance with the first glaciation of the subarctic continents (SHACKLETON et al. 1984), the beginning of the Quaternary has to be put back to at least 2.4 Ma bp; in discussion is a triggering effect of the isostatic rise of the Himalayas at that time. As there is no generally accepted subdivision of the Pleistocene, in the following the early Pleistocene is understood as lasting from 2.5 to 0.7 Ma bp (a period roughly corresponding to the Matuyama reverse magnetic epoch), the middle Pleistocene from 700 to 300 ka bp (from the beginning of Brunhes normal magnetic epoch to the approximate limit of U-series dating-techniques) and the late Pleistocene from 300 000 to about 10 000 bp, with the terminal Pleistocene starting at the last glacial maximum (LGM) around 18 000 bp.

5.1 Early and Mid-Pleistocene

Little is known about the early and mid-Pleistocene environments and climatic conditions in the Western Desert. With respect to world-wide climatic events well documented in the chronology of isotopic stages (EMILIANI & SHACKLETON 1974, FINK & KUKLA 1977, and others), however, the hypothetical concept can be adopted that, in SW Egypt too, moist and dry alternations like those in the better known last glacial cycle have occurred at least twenty times during the past 2 million years. The average interval between them was 100 000 years, according to changing insolation rates controlled by astronomical parameters (BERGER & TRICOT 1986). There is little doubt that aridity has been the dominating factor for the whole Saharan belt for most of the Quaternary. During the long dry periods, most, if not all, evidence of the preceding, comparatively short wet phases has repeatedly been removed by aeolian erosion under extreme desert conditions.

In a recent classification for the Egyptian Quaternary mainly based on the Nile valley, SAID (1983, 1990b) proposes two major units belonging to the early Pleistocene. These two oldest Quaternary pluvials, the Idfuan and the Armantian, are not preceded or succeeded by interpluvial deposits and therefore hold a questionable position. The Idfuan corresponds to the Proto-Nile stage around 1.8 Ma when the source of the Nile river was within the borders of Egypt. Northern Sudan and the source lands were subjected to intense chemical disintegration, leaving behind surfaces with siliceous cobbles and pebbles, which are thought to indicate a vegetative cover and a wetter climate throughout the year at that time. The Armant formation is put at around 1.2 Ma, thus directly preceding the Pre-Nile stage, and is recognized in the Nile valley by travertines interbedded with gravel and locally-derived detritus and overlain by massive breccias; near Luxor it is associated with early human pebble tools.

It is obvious that these humid episodes, as well as the others postulated above, must have caused responses in surface hydrography, groundwater recharge, landforming processes, pedogenesis, vegetation and fauna. It is, however, not clear to what extent relict landforms caused by ephemeral streams - such as ancient drainage patterns or now inverted channel deposits - have been formed during these wet periods by down-cutting, erosion and accumulation. At present it is not possible to distinguish or date these stages or events in the Western Desert with the methodical spectrum available to terrestrial paleoclimatology as even rough chronologic indicators such as early paleolithic artifacts (pebble tools) have not yet been reported.

The advent of the middle Pleistocene at around 700 ka bp was marked by the forming of the Pre-Nile. With its headwaters in the Ethiopian highlands, it was the first river with an African connection since the Oligocene. The middle Pleistocene was terminated by a pluvial with intense rains (Abbassian I) resulting in thick, locally derived gravel deposits lying unconformably over Pre-Nile sediments (SAID 1990b).

A few Uranium-series dates on carbonate-cemented alluvium in the Wadi Arid and at Bir Tarfawi (SZABO et al. 1989, WENDORF et al. 1990) indicate humid conditions and shallow groundwater-tables allowing calcrete formation before 300 ka. As these dates are approaching the limit of the dating technique, they can only provide minimum ages. Nevertheless, they are the first firm evidence of mid-Pleistocene pluvial phases and groundwater recharge in the very core of the eastern Sahara. There is no further information on their paleoenvironmental characteristics.

5.2 Late and Terminal Pleistocene

Late Acheulean artifacts are the oldest material indicators of climatic conditions permitting early human occupation of SW Egypt during the late Pleistocene. Acheulean handaxes found in situ in carbonate-impregnated sediments in Wadi Arid and Wadi Safsaf south of Bir Tarfawi indicate a minimum age of 212 ka for the late Acheulean (SZABO et al. 1989), but they may be as old as 500 ka (WENDORF et al. 1985). Concentrations of Acheulean artifacts in the Bir Tarfawi area (WENDORF & SCHILD 1980), along the paleochannels of the "radar rivers" (McHUGH et al. 1988a,b), in the valleys and at the piedmonts of the Gilf Kebir (MYERS 1939) or near the spring deposits of Kharga and Dakhla (CATON-THOMPSON & GARDNER 1932) point to occupation sites near groundwater outlets or temporary ponds. This does not necessarily imply equally favorable ecological conditions for the open plains during the early Paleolithic wet phases. The post-Acheulean aridity must have been long and intense because a deflational hollow has been cut down to about 3 m below the modern water table at Bir Sahara East which later was filled by sand sheet deposits older than 175 ka (MILLER et al. 1991).

Uranium-series dating indicates that widespread carbonate deposition occurred in SW Egypt about 212 ka, 141 ka and 121 ka ago even though there are considerable age discrepancies within single layers (SZABO et al. 1989). Most of the carbonate appears to have been precipitated from groundwater suggesting that these three episodes of deposition may be related to late Pleistocene humid climates that facilitated human settlement. Formation of spring travertine in the Kharga depression east of Bulaq, first reported by CATON-THOMPSON & GARDNER (1932), also occurred around a U/Th-date of 149 ka bp (PACHUR et al. 1987). According to the meanwhile widely accepted isotopic scale established on ocean cores, the last interglacial (stage 5) lasted from 130 to 75 ka bp with two short, but pronounced interruptions (cold substages 5d and 5b; EMILIANI & SHACKLETON 1974). Substage 5e (about 127–115 ka), called Eem in the strict sense in the terrestrial nomenclature, appears not only to have been the warmest climatic phase of the last interglacial but of the last 700 ka. Therefore it would have been surprising if such worldwide climatic changes had produced no environmental and human responses in the Western Desert.

In fact, extensive archaeological research into the Middle Paleolithic of Bir Tarfawi has shown that six episodes of lake development occurred from about 175 to 70 ka ago (WENDORF et al. 1990). These contrast with only two pluvials, Saharan I and II, in SAID's (1990b) classification. The exact chronological position of these lake phases is still complicated by the large range of the respective U-series-, TL- and ESR-dates, but is attributed to 175 ka, 160 ka, 130 ka (with three phases within 10 000 years) and between 104–70 ka (CLOSE et al. 1990). The apparently permanent lakes were primarily fed by groundwater, but the increases in available moisture are thought to have resulted from northward shifts of the monsoonal belt. Both the faunal spectrum which includes fish, amphibians, reptiles, birds and small mammals (KOWALSKI et al. 1989), and the levels of lakes and groundwater are unequalled in the Holocene wet phase. Nevertheless, paleorainfall estimates of 600 mm per annum postulated for the wet optimum around 130 ka bp on the basis of specific rodent species seem high on general hydrogeologic, ecological and climatic grounds (KRÖPELIN & PACHUR 1991).

A more controversial problem is the on-going discussion of possibly uninterrupted aridity in the Sahara from 70-10 ka bp which arose from new dating methods that cast doubt on most of the radiocarbon dates older than 20 000 bp (CAUSSE et al. 1988, SCHILD & WENDORF 1981). This leads to the perplexing situation that late Pleistocene paleoclimatic chronologies have to be re-examined even before they have been appropriately established. There is, however, various evidence pointing to more humid conditions in the Western Desert of Egypt during this 60 000-year period of the early, middle and late glacial. Calcrete formation in the Wadi Arid, Wadi Safsaf and Bir Tarfawi areas is clearly indicated at 45 ka by ten U/Th-dates (SZABO et al. 1989). This humid episode corresponds to the Kubbaniyan in SAID's (1990b) classification. The question must remain open whether some of the ten radiocarbon dates of calcretes from the Great Sand Sea, the Bir Tarfawi area and the southern edge of the Selima Sand Sheet ranging from 38 to 19 ka bp (PACHUR et al. 1987) may be attributed to this humid stage, or whether they indicate another or even several wet intervals between 35 ka and 25 ka bp. The organic matter of a mud underlying a calcareous precipitate in the "Waterhole depression" near Bir Tarfawi, radiocarbon-dated at 26 ka bp, confirms the carbonate dates; so does the 26 ka bp date from the organic matter of a mud layer in a yardang in the southern Bahariya depression (PACHUR et al. 1987). Spring-deposited travertine at the slope of the Abu Tartur plateau yielded a similar radiocarbon age of 29 ka bp (PACHUR & RÖPER 1984). All these dates are corroborated by pronounced phases of groundwater formation between 34 and 28 ka, and 24 and 20 ka bp (THORWEIHE 1986). Further evidence of more humid late Pleistocene environments allowing human occupation are the rich Aterian sites in Kharga (CATON-THOMPSON 1952) and several other parts of the Western Desert.

The last glacial maximum (LGM) has been radiocarbon-dated at 18 ka bp, which according to U/Th-calibration is equivalent to about 21 500 BP or 19 500 BC (BARD et al. 1990). For this time, no indicators of increased humidity whatever have been reported from the entire Western Desert. Therefore it can be considered certain that the core of the Eastern Sahara around the Tropic of Cancer witnessed extremely arid conditions during the peak of the last pleniglacial in the northern hemisphere.

There is also a general consensus on the terminal Pleistocene hyperarid phase lasting from at least 22 ka to 11 ka BP. This period, named "Ogolien" in the

Western and Central Sahara, can be clearly deduced from the notable absence of radiocarbon-dates and artifacts, but also from morphological indicators such as corraded sandstone surfaces at the base of early Holocene playa deposits, or from the deflational origin of the lake basins (PACHUR et al. 1987).

5.3 Early to Mid-Holocene

During the Alleröd interstadial (12 000–10 800 bp [3]) when NW Europe witnessed temporarily waning ice-sheets, as well as during the 500-year-long Younger Dryas, the final cold phase of the Pleistocene, the Western Desert was still hyperarid. Only with the beginning Preboreal, i.e. the onset of post-glacial conditions in the mid-latitudes, is a sudden return of humidity evidenced by an overall appearance of playa lake deposits and other paleoenvironmental indicators. It occurred around 9500 bp, which corresponds to about 8800 BC according to BECKER & KROMER (1986), and apparently affected not only the whole eastern Sahara (KRÖPELIN 1993), but likewise the northern and southern sides of the entire Saharan belt (GASSE et al. 1990). Even taking into account the recently established, possibly up to 400 yrs long plateau phase of constant radiocarbon age at that time (LOTTER 1991), this dramatic climatic and environmental change must have taken place within a few centuries, a clear contrast to the slow and progressive southward shifting of the isohyets in the middle Holocene (chapter 5.4).

This last of the Quaternary pluvials, called Nabtian in SAID's (1983) classification, is represented in the Nile valley by the deposits of the recessional episode between the gamma- and delta-stages of the Neo-Nile. It is well documented in the playas of the Western Desert even if there are only few radiocarbon-dated stratigraphies covering the entire Holocene wet phase. "Nabtian" derives from Nabta playa (22°30'N/30'45'E; about 15 km east of the map sheets), where one of the most complete sequences has been found. Three phases of playa formation occurred between 9000 and 5800 bp (HAYNES 1980); intervening arid intervals, only a few centuries long, had to be revised after subsequent investigations (WENDORF et al. 1984). The sections of Bir Kiseiba lying 100 km further west, parallel that of Nabta Playa but playa formation began earlier, around 10 000 bp at the latest, and also ended earlier (BANKS 1984, ISSAWI & EL HINNAWI 1984).

The Dakhla Oasis region in south-central Egypt records variable sedimentary responses to the early and mid-Holocene pluvial. Twenty samples of artifactual ostrich eggshell and hearth charcoal associated with basinal lacustrine, playa and sandsheet deposits yield radiocarbon ages between 8800 and 4700 bp (BROOKES 1989). Differences in hydrogeology and morphometry among and within the basins, however, complicate the detailed paleoclimatic interpretation of the aggregate data.

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[3] As at present, continuous tree-ring calibration does not cover the entire Holocene, most early Holocene dates are given in uncalibrated radiocarbon years for reasons of uniformity and differing correction factors applying to different materials. It must, however, be kept in mind that, according to natural ^{14}C variations, conventional radiocarbon dates from 8000-5000 bp correspond to calendar ages roughly 1000 years older (KROMER et al. 1986). Similar time scale differences apparently also apply to the period between 13 000 and 9000 bp (LOTTER 1991).

Several radiocarbon dates of charcoal from fine-grained playa deposits indicate extensive ephemeral lakes filling the depressions below the Abu Ballas escarpment in the period between 10 100 and 6600 bp (PACHUR et al. 1987). Archaeological excavations of large Neolithic sites in the Mudpan area below the more dissected escarpment further northeast, point to a possibly continuous occupation between approximately 8300 and 6300 bp (KUPER 1989). The spectrum of mammalia identified from associated bone remains is unequalled by any other Holocene site within the maps outside the oasis depressions and includes savanna to semi-desert dwellers such as elephant, giraffe, oryx or caracal (VAN NEER & UERPMANN 1989). Species identifications of large concentrations of charcoal remains from the prehistoric sites show a maximum northward expansion of tropical savannas as contracted outposts between 7000 and 6500 bp, while desert formations consisting of the same vegetation elements as today, though with a wider distribution, prevailed in SW Egypt during the preceding and subsequent wet periods (NEUMANN 1989).

The playa in Wadi el Bakht, a valley in the southeastern part of the Gilf Kebir Plateau (cf. chapter 3.4), has been singled out for its outstanding potential for reconstructing late prehistoric paleoenvironments (McHUGH 1980). Owing to the complete blocking of the valley by a dune, the absence of unconformities, and the independence of groundwater influence, every single major surface runoff from local rainfall in the intake area during more than 5000 calendric years has been recorded in the sedimentary sequence. The non-calcareous deposits consist of thin alternating layers of siliceous mud and medium- to coarse-grained sand and are virtually sterile, thus fully contrasting with the calcareous, highly fossiliferous lake sediments typical of the more southerly parts of the Eastern Sahara (KRÖPELIN & SOULIE-MÄRSCHE 1991). The microstratigraphic sequence indicates their event-controlled character as well as their mainly short duration of sedimentation (KRÖPELIN 1987). The pelitic layers recur 89 times within the top 8 m of the section and have an average thickness of 14 mm only, disregarding the singular, more than one metre thick top layer (photo 8).

The stratigraphical and sedimentological evidence in combination with several charcoal radiocarbon dates suggests the following climatic development. Up to about 9500 years bp a hyperarid climate, similar to the present one, prevailed. The period between 9500 and approximately 6000 bp had an arid climate with rare heavy rainfalls (on an average four events per 100 years) enabling meagre soil formation and sparse plant growth in the surroundings of the temporary rainpools which contained water for days or weeks at most. Between 6000 and 5000 bp conditions tended toward moderate aridity (KRÖPELIN 1989) with the occurrence of slightly more demanding tropical plant species (NEUMANN 1987). This millennium also seems to represent the main phase of Neolithic settlement (KUPER 1988). Possibly a west-wind-induced type of occasional winter rainfall with steady rains has alternated with a regime of secular monsoonal-convective summer rains triggered by exceptional northward surges of the surface intertropical convergence zone. However, there was still no appreciable biological activity in or around the ephemeral lake, even though, for geomorphological reasons, Wadi el Bakht would have been the most appropriate location for any potential occurrence.

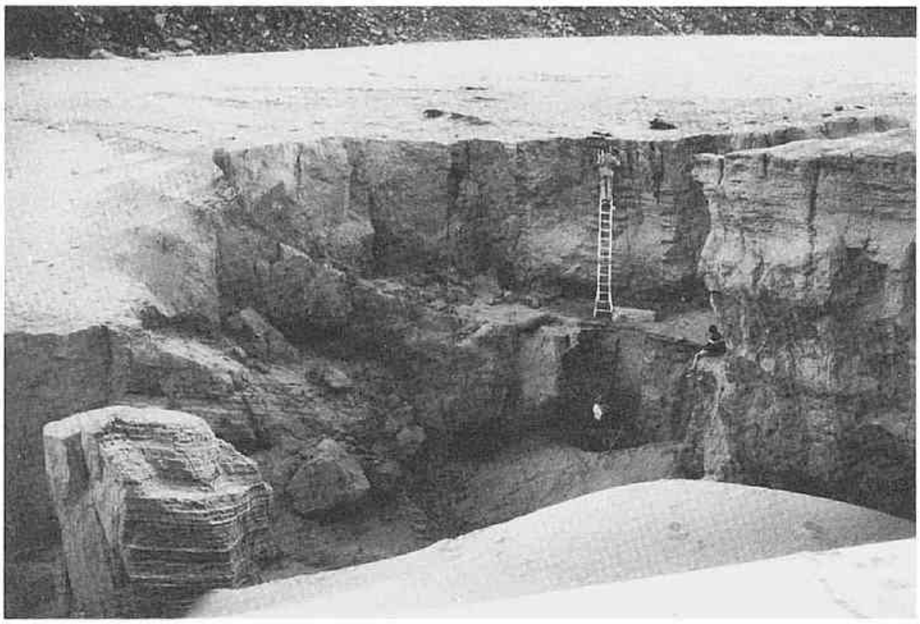

Photo 8: *Wadi el Bakht, Gilf Kebir. Early Holocene playa deposits, dipping upvalley and more than 8 m thick, on right side. Residual terminal Pleistocene blocking dune on left side (cf. photo 5).*

At about 4800 bp the deposition of playa sediments stopped, suggesting that the dune blocking Wadi el Bakht was breached at that time owing to an unprecedented, exceptionally high water pressure or level, pointing to a climatic optimum and/or a unique millennial rainfall event.

In synthesis, the past decade of field research into the stratigraphy of playa deposits and the paleobotany of Southwest Egypt has resulted in growing evidence that the climate in this central part of the eastern Sahara around the Tropic of Cancer was relatively arid even during the early- and mid-Holocene wet phase with an estimated maximum precipitation of 100 mm per annum (KRÖPELIN 1987, 1989; NEUMANN 1987, 1989), and that the earlier notion of a pronounced Holocene pluvial with rainfall up to 600-800 mm in the Western Desert of Egypt adopted by some authors (McHUGH 1974) may have been an overstatement. This conclusion supports the general view that "maximalist" estimates of rainfall in the range of 200 mm to 600 mm, or even more, are not necessary to explain all of the known facts established for the Holocene wet phases in the entire Sahara (MUZZOLINI 1985).

Nevertheless, even these low figures of paleorainfall during the first half of the Holocene equal more than 50 times the present rates of rainfall (<2 mm of mean

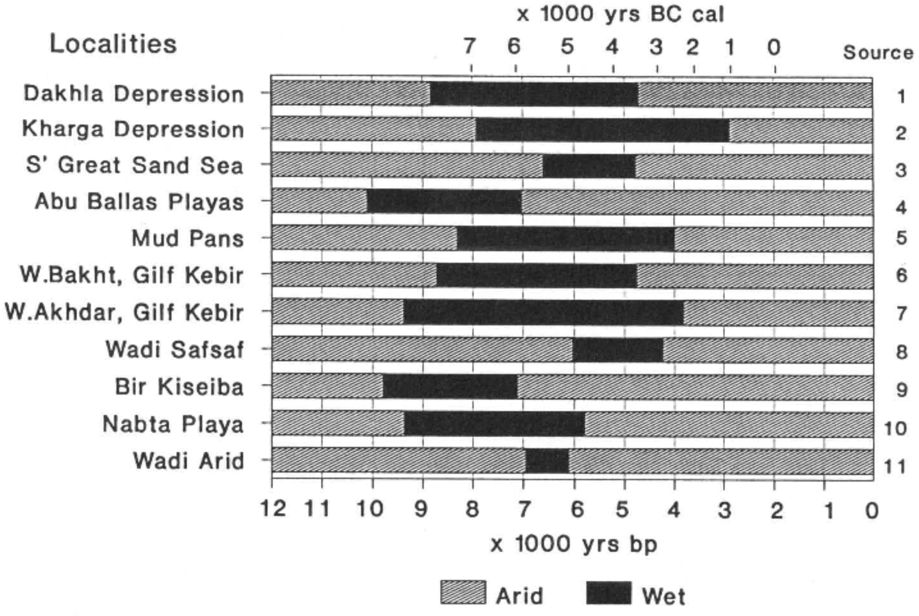

Fig. 2: *Holocene lacustrine phases in Southwest Egypt (according to oldest and youngest radiocarbon dates; carbonate and aberrant dates neglected).*

Sources: 1: BROOKES 1989, 2: HAAS & HAYNES 1980, 3: HAYNES 1982, 4: PACHUR et al. 1987 (top layers deflated), 5: NEUMANN 1989, 6,7: KRÖPELIN 1987, 8: McHUGH et al. 1989, 9: BANKS 1984, 10: HAYNES 1980, 11: McHUGH et al. 1989. See text and footnote 3 for further explanations.

annual rainfall) in the hyperarid core of the Eastern Sahara. This is in line with the ratio established for the whole Saharan belt along the Tropic of Cancer (PETIT-MAIRE & KRÖPELIN 1991). These more favorable climatic conditions were of eminent importance for faunal migrations, nomadic cattle-keeping, gathering activities and the origins of Egyptian agriculture (HASSAN 1986), and allowed for cultural contacts of prehistoric man from East to West and North to South. However, local differences are observed in the climatic scenario along the Tropic of Cancer with the optimum occurring sooner (8300-6700 bp) in the central Western Sahara than in Egypt (6000-5000 bp), the latter always remaining more arid than the former. The duration of the wet phase and of its optimum were, as is logical, shorter along the tropic (9200 to 4500 bp) than in the northern and southeastern continental Sahara (FONTES et al. 1985; PACHUR & KRÖPELIN 1987, 1989), or Atlantic Sahara (PETIT-MAIRE ed. 1979) and the Sahel (SERVANT 1983). While the overall beginning and end of the early and mid-Holocene wet phase in SW Egypt can be traced to a satisfactory degree (fig. 2), estimations of the relative degree of humidity for each time segment or a subdivision of these periods by "arid intervals" vary considerably and are far from being consistent. The causes for local differences among the stratigraphies and impacts of increased moisture upon the environment lie in differing topography, drainage patterns, lithologies, groundwater control or other boundary conditions. Therefore, a high-resolution subdivision of the Holocene climatic evolution of the Western Desert of Egypt in the scale of centuries requires more detailed research, in spite of some 400 radiocarbon dates available and the multitude of paleoenvironmental data collected during the last 15 years in this formerly virtually unknown region.

5.4 Middle and Recent Holocene

There is general agreement that arid to hyperarid conditions in the Western Desert set in around 4500 bp, i.e. about 3500 years BC (HAYNES 1987, KRÖPELIN 1987, PACHUR & KRÖPELIN 1989, RITCHIE et al. 1985). The aridification of the Western Desert was the consequence of a continuous southward shift of the desert boundary (roughly corresponding to the 100-mm-isohyet) progressing with an average rate of about 35 km/100 yr or 1° lat./300 yr (HAYNES 1987, KRÖPELIN 1993). The aridification of southwestern Egypt, the high continental centre of the eastern Sahara, has been an irreversible process, in contrast to northern Sudan at latitudes 15°–17° N (KRÖPELIN 1993, MAWSON & WILLIAMS 1984) or the foreland of high mountains such as the Serir Tibesti (PACHUR 1974), where episodes of more humid conditions reappeared around 2000 bp. There was no ecologically significant revival of the rains over the desert after the end of the Neolithic wet phase, and the Nile valley was settled during the Predynastic of Egypt, an interval between ca. 4000 and 3050 BC, by refugees from the eastern Sahara fleeing from mid-Holocene droughts; this was the very time when the identity of Egyptian society was forged (HASSAN 1988). The coincidence of the desert lands becoming uninhabitable, and the forming of the Pharaonic Nile culture, now confirmed by geoarchaeological field evidence, has been a much-discussed historical and philosophical issue (cf. BUTZER 1959).

In the first dynasties of the Old Kingdom (3050-2050 BC), even before the Pyramid age, the specific savanna fauna of Egypt had died out even along the Nile valley. A review of textual and architectural evidence bearing on rainfall suggests that the Middle Kingdom (2050-1570 BC) had conditions similar to the present ones with heavy rainfalls being somewhat less rare (BELL 1975). In the New Kingdom (1570–715 BC) desert neighbours played no part whatever for Upper Egypt (BUTZER 1959).

Apart from a few radiocarbon dates from hearth charcoal pointing to nomadic camps, there is virtually no sedimentological or archaeological record of the deteriorating environmental conditions within the plains and plateaus of the Western Desert during the whole Dynastic period. This indicates extreme desert conditions dominated by deflation. Even from the "Great Oasis" of Kharga, little is known of positive value until the advent of the Persians at 525 BC when the oasis was rendered habitable by the opening of deep wells and the introduction of foggara irrigation (CATON-THOMPSON & GARDNER 1932). In the subsequent period, Roman wells fully exploited the artesian water initiating the lowering of the near-surface groundwater table (HEINL & THORWEIHE this vol.).

Despite the persisting extreme desert conditions, a few faunal and floral retreat areas have remained until recently in the northern valleys of the Gilf Kebir and the Abu Ras Plateau underlining the paramount effect of topography and relief in a hyperarid area. Noteworthy is also the evidence of short-term standing water more than 2 m deep in the upper part of Wadi el Maftuh after secular heavy rainfall even under the present climatic regime (KRÖPELIN 1989). Nevertheless, the very last survivors of highly adapted animals such as addax (*Addax nasomaculatus*) or mufflon (*Ammotragus lervia*), still observed in some numbers in the twenties of this century, are now extinct. In danger of extinction are also relict trees (*Acacia raddiana*) in the last vegetated areas in the so-called Contrast Wadi in the northern Gilf Kebir and in the wadis Abd el Malik and Hamra in the Abu Ras Plateau, to which, at the beginning of this century, cattle were still being driven from Kufra to graze after exceptional summer rainfall. There are also indications of an increasing southward extension of the dunes of the Great Sand Sea on the northern side of the Gilf Kebir during the last decades. Apparently, the process of aridification of the Western Desert is approaching its climax.

6 Closing remarks

Investigations into the Quaternary geology of the Eastern Sahara are not only essential for the understanding of past climatic, landscape, biological and cultural evolution but also provide a long-term scenario for future changes in this now sparsely populated part of the earth. Excluding man's influence, it provides a paleoanalogue that could anticipate the slow natural evolution of northeastern Africa in the next millennia. In the more probable case of a more man-made scenario related to the greenhouse temperature increase resulting from atmospheric pollution (cf. FLOHN 1985), the examples of the last glacial-interglacial transition and of the last warm climatic optimum still provide the best and most concrete model available for the Saharan and Sahelian belts (PETIT-MAIRE 1990).

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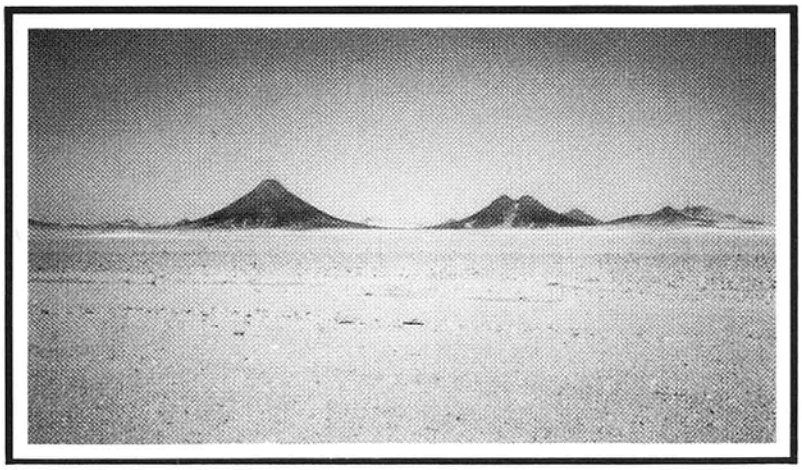

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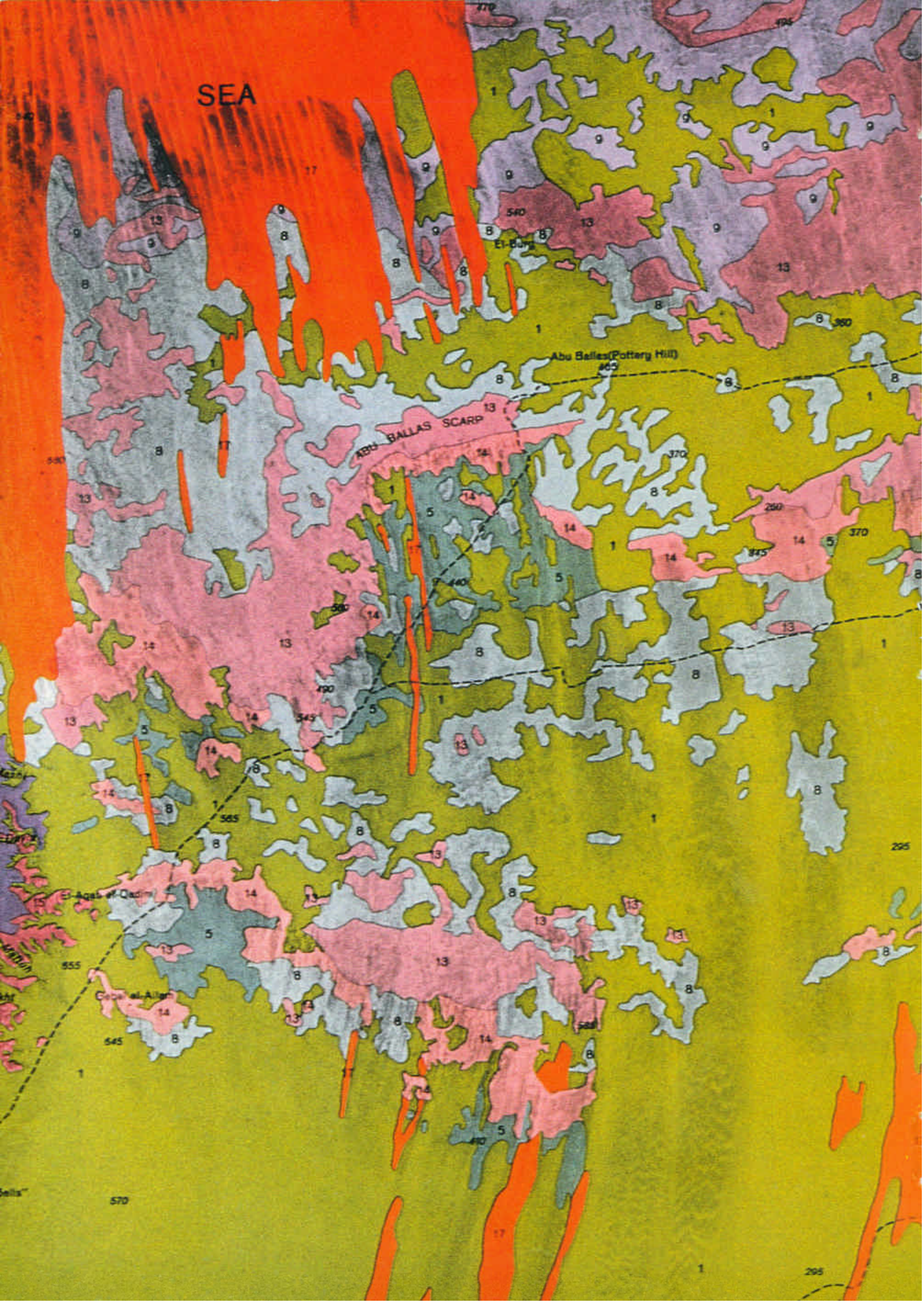
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