Characterization of the aeolian terrain facies in Wadi Araba Desert, southwestern Jordan

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Abstract

The sand dunes in Wadi Araba Desert, southwestern Jordan, conform to the influence of two main wind systems: (1) the Shamal “ north wind”, the main determinant of dune patterns, and (2) the southerly winds caused by the Red Sea Trough and Khamasin winds. Wadi Araba is a narrow elongated morphotectonic depression bordered by the high eastern and western mountain ranges that would obstruct most of westerlies in winter and the hot dry easterlies in summer. Wadi Araba Desert is caused by the rain shadow of the eastern and western topographic highs, with an arid–hyperarid climate and moisture index between 0.1 and < 0.05. An aeolian terrain occupies about 16% of Wadi Araba Desert divided into four sand dune fields that contain different dune types, interdunes, and sand sheet facies. The development of the aeolian terrain was more likely in the interglacial periods of latest Pleistocene–Holocene during which wind deposition and fluvial erosion were prompted. The variability of wind direction, wind speed and rates of sand supply led to a variety of simple, compound and complex dune formations. Barchanoid (barchan, barchanoid ridge, transverse dunes) are very common. Linear dunes, nabkhas and climbing dunes are less common. The dunes in Wadi Araba are either mobile (active) or stabilized. Dune fixation is primarily by desert shrubs or cementation in case of aeolianites. Interdune troughs vary from dry to damp. Sand sheets are mainly unvegetated or barren. Dune sands are well sorted to moderately well sorted with more than half the sand falling in fine–medium sand fraction. Interdune areas and sheet sands are moderately–poorly sorted with grain size between 0.06–5 and 0.1–0.5 mm, respectively. Angular-subrounded grain shapes are relatively common on all sides of barchanoid dunes. The sediments of the aeolian terrain are chiefly composed of quartz, feldspar, mica and kaolinite. Calcite and dolomite are less predominant minerals.

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1. Introduction

1.1. Background

Fascinating geological and geomorphologic features occur within and around Wadi Araba, Jordan. An aeolian terrain covers about 16% (≈ 400 km²)
of the Wadi Araba area, extending between a point ≈ 35 km north of the Gulf of Aqaba to about 22 km south of southern margin of the Dead Sea basin. The aeolian terrain contains various types of dunes in association with deflation zones, sand sheets, aeolianite and interdune troughs. Coalescing alluvial fans, mud flats (Qa or “Gaa” in Arabic) and inland sabkhas are common features interrupting the aeolian terrain.

1.2. Previous studies and aim of study


1.3. Study area

The study area is a part of the Dead Sea Transform Valley (DSTV). It occurs between 29°50′N and 31°N latitudes (Fig. 1).

1.4. Geomorphology, climate, soil

The DSTV extends for 375 km along the entire length of Jordan from the Gulf of Aqaba to Lake Tiberias (Sea of Galilee) (Fig. 1). It consists of three morphotectonic depressions, the Wadi Araba in the south, the Dead Sea in the middle and the Jordan Valley in the north. The DSTV is 9–25 km wide south of the Dead Sea and narrows up to 9 km near Lake Tiberias in the north. The southern end of the DSTV lies at sea level at the Gulf of Aqaba on the shores of the Red Sea; the valley rises gradually up to 250 m asl at Jabal Ar-Risha in central Wadi Araba. From there, it gradually decreases up to 412 m bsl to the present Dead Sea shorelines. The DSTV is bordered by the Western Mountain Range (in excess of 1000 m asl) and the Eastern Mountain Range (elevation 1200–1500 m asl) (Bender, 1974). The topographic highs have a very steep escarpment that is crossed by deep gorges that drain towards the DSTV from the east and west. Large alluvial fans built out from wadi mouths at the base of the escarpment onto inland sabkhas and/or mud pans in the heart of the Wadi Araba (Fig. 1). These alluvial fans indicate periods of heavy rains with resultant debris-flow and stream-flood deposition.

Desert conditions prevail in Wadi Araba and the adjacent Negev desert to the west, which are in the rain shadow of the bordering Eastern and Western Mountain Ranges. A short dry cool winter and a very hot dry long summer characterize the modern climate. The mean air temperature is ≈ 15 °C in January and ≈ 33 °C in August. The daily maximum temperature reaches 50 °C in summer and drops to 0 °C in winter. The temperature difference between daytime and night is large enough (>30 °C) to restrict the growth of plants. The annual rainfall is usually <100 mm. Based on long-term meteorological data (1955–2002) obtained from Ghor As-Safi Meteorological Station (GSMS) and Aqaba Airport Meteorological Station (AAMS), the annual precipitation decreases from ≈ 80 mm (GSMS) to ≈ 40 mm (AAMS) from north to south of the study area. The eastern bordering highlands (Jordan Plateau), however, receive some 200 mm/year of rainfall. Localized storms occur mainly in conjunction with the presence of the active Red Sea Trough (RST) or in conjunction with a low-pressure center over southern parts of Jordan and the adjacent regions of Israel, Sinai, and northwestern Saudi Arabia (Sharon and Kutiel, 1986; Kahana et al., 2002). The mean annual potential evapotranspiration fluctuates between ≈ 2200 mm (northern Wadi Araba) and ≈ 2500 mm (southern Wadi Araba). A rapid loss of soil moisture results from high levels of evapotranspiration. In winter, the relative humidity fluctuates between 53% (AAMS) and 62% (GSMS) but in summer it diminishes to 30% (AAMS) in the southern parts and 40% (GSMS) in the northern parts (Fig. 2). Dust storms are common in Wadi Araba beginning in the spring and continuing through the early summer (March–June) (Fig. 3). The DSTV is of a typical Sudanian bioclimate.
Fig. 1. (A, B) Distribution of the aeolian terrain facies in Wadi Araba Desert, southwestern Jordan.
with entisol, enceptisol and aridisol soil types and with tropical trees and shrubs such as *Acacia savanna* and *Ziziphus spina-christi*. (Al-Qudah, 2000).

1.5. Geology

The DST fault is an active left lateral strike-slip fault that has undergone tectonic movement since
Miocene–Pliocene time and is still active (Quennell, 1958; Garfunkel et al., 1981; Galli, 1999; Atallah, 2002) The highlands flanking Wadi Araba correspond to the northern end of the Arabo–Nubian Shield and they resulted from the upheaval of the eastern Transjordan block during the Tertiary (Bender, 1968). The eastern Transjordan block consists of a complexity of igneous, volcanic and metasedimentary rocks of Precambrian age (Jarrar, 1984) overlain by Paleozoic–Tertiary sediments. Three inland sabkas (Fig. 1) located in southern Wadi Araba are chiefly composed of detrital mudstone, dolomite and gypsum (Abed and Al-Hawari, 1991; Abed, 2002). Lake varved sediments of late Pleistocene age are exposed in the central parts of the DSTV (Abed, 1982; Abed and Yaghan, 2000). These sediments have been easily eroded into a badland landscape of gullies, hills and cliffs near the Dead Sea basin. Late Pleistocene–Recent detrital sediments cover the center of Wadi Araba.

2. Materials and methods

The description of the aeolian terrain facies in Wadi Araba is based on field observations, landsat imagery (TM, scale 1:150,000) and aerial photographs (1:10,000). The grain size, roundness and mineralogy of the aeolian sediments were investigated. Grain-size analysis was determined using a Retsch AS200 Digit Analytical Sieve Shaker and Retsch-Graintest software. Powers (1953) roundness chart and binocular microscopy were used to determine the roundness of sand grains. A Phillips PW-1729 X-ray Diffractometer was used for the mineralogical analysis. To describe weather conditions in the study area, mean temperature (°C), total rainfall (mm), relative humidity (%), mean evaporation (mm), wind speed (m s⁻¹), wind direction (degrees) and the time periods of dust storms (days) were analyzed. Meteorological data (1955–2002) were collected from two stations operated by the Jordan Meteorological Department (JMD): (1) the Aqaba Airport Meteorological Station (AAMS) in the south, and (2) Ghor As-Safi Meteorological Station (GSMS) in the north. The aridity index is given by AI = P/T + 10 (De Martonne, 1925; Murai and Honda, 1991), where P = annual rainfall (mm), T = sum of monthly mean temperature divided by 12. The moisture index is given by: MI = P/PET (Pahari and Murai, 1995), where P is the annual
rainfall and PET is the potential evapotranspiration. Sand drift potential \(\text{DP} = \frac{U^2(U - U_t)}{100t}\) (Fryberger, 1979) was determined, where \(U\) is wind velocity, \(U_t\) is threshold velocity, and \(t\) is time of wind blew (%). Sand roses were plotted after simplifying wind speed data and related calculations using a visual basic application programmed by A. Saqqa.

3. Results and discussion

3.1. Aridity and moisture indices

In terms of aridity and moisture indices (AI and MI), the Wadi Araba area is classified as a desert arid zone or arid–hyperarid dryland. The AI and MI are respectively between 1.0–2.2 and < 0.05–0.1 along south–north ascending precipitation gradient. Usually, a zone of aridity is classified as desert if the aridity index (AI) is < 5 (Murai and Honda, 1991; Pahari and Murai, 1995). The terms arid and hyperarid are applied if the moisture index (MI) is respectively between 0.05 and 2.0 and less than 0.05 (UNEP, 1992; Pahari and Murai, 1995; Middleton and Thomas, 1997).

3.2. Determinant winds and sand drift

Wind speed in the Wadi Araba Desert varies throughout the year. The variation in wind speed is more likely to be caused by a large pressure difference developed between the lowland DSTV and the surrounding eastern and western high elevation mountains, in addition to the influence of desert surface roughness and local topography. The air above the DSTV is heated, expands, and rises leaving behind a low-pressure zone. Cool air flows down from a high-pressure zone and is balanced by an outward flow of the air (Krauskopf and Beiser, 2002). Fig. 4 shows that wind speed in Wadi Araba Desert reaches up to 36 knots/h \((18.5 \text{ m s}^{-1})\) or it may exceed this value during windstorm events. Monthly wind speed averages 6–15 knots/h \((3.1–7.7 \text{ m s}^{-1})\) but it diminishes to < 5 knots/h \((2.6 \text{ m s}^{-1})\) as a minimum value. The annual wind speed (Fig. 5) averages 8–13 knots/h \((4.1–6.7 \text{ m s}^{-1})\). Practically, maximum wind speed is not restricted in a certain period of year, but it is rather common at end of spring–early summer (April–June), end of summer (September), and end of winter (February–March). These periods are favorable times for the Khamasin sandstorms, monsoons, and the RST. Undoubtedly, wind speed variation has its own effect on rates of wind erosion, wind deposition, advance and reshaping of dunes. Moreover, it may serve as indicator for climatic condition variations in the whole region.

By relating the annual wind speed data to values \(\geq 12 \text{ knots/h (6.2 m s}^{-1}\), where the value 12 knots/h is taken as the threshold velocity, we found that the ratios of surface wind energy responsible for sand drift are between 56% (year 1987) and 25% (year 2001) (Fig. 6). A remarkable decrease in the annual wind energy is expected in the ninetieth, as it drops from 38% (1991) to 25% (2001) (see Fig. 6). It appears from wind speed analysis that only about half to quarter the potential wind energy is used as a motive force for wind erosion and sediment mobilization.

Wind roses (Fig. 7) suggest that four groups of surface winds blown out over the Wadi Araba Desert from NW, NE, SW and SE directions. The strongest and most frequent winds are encompassed within the first two groups: NW and NE, attributed to the northern winds, the “Shamal” (Cooke et al., 1993) drawn in by the Asian monsoon; these northern winds are the driving force for movement of sand and dune formation in the Arabian Peninsula (Memberry, 1983). The Shamal wind comprises 70% of the total wind energy (Abed, 2002). An active Red Sea Trough (RST) implies southerly winds over southern Israel, southern Jordan, Sinai, and northern Saudi Arabia and is capable of transporting moist tropical air in the middle troposphere causing some rainfall in winter over such arid regions (Itzigsohn, 1995; Dayan et al., 2001; Kahana et al., 2002). The southerly winds which compose about 8% of the total prevailing winds (Abed and Al-Hawari, 1991; Abed, 2002), may also be initiated by the Khamasin winds blowing in the early and late summer months and lasting for several days at a time before terminating abruptly as the wind direction changes and much cooler air follows. Such winds arise when cold winds from Europe converge with the hot winds from the Sahara. The resulting atmospheric depression moves eastward from Africa off the Arabian Peninsula pushing hot,
Fig. 4. Time series 1975–2001, showing monthly minimum, maximum and average wind speed (AAMS).
Fig. 4 (continued).
sand-bearing winds followed by a drop in temperature. Extreme dust storm events in the study area are simultaneous with the Khamsin (see Fig. 3). A comparison between the meteorological data from the two stations AAMS and GSMS showed that duststorms and wind speeds are more intense in the southern Wadi Araba Desert. Southwesterly winds generated by depressions across the Mediterranean and the Middle East would affect the Sahara and the Arabia as far as 30°N in winter (Breed et al., 1979; El Baz, 1986; Jones et al., 1986). Accordingly, the southwesterly winds do not blow across the whole region of Wadi Araba, but their zone of influence extends only up to 55 km north of the Gulf of Aqaba. The interaction between the westerlies and the hot dry eastern trade winds (Wilson, 1971; Fryberger, 1980) or between moist cold air of low pressure and hot dry air of high pressure (Trenberth, 1992) at the boundary between winter and summer brings a complex wind regime to the northern edges of the Sahara and the Arabian Desert. We believe that this kind of wind complexity may have little chance of operating in the Wadi Araba Desert. It is clear that the eastern and western topographic highs (elevation > 1000m asl) bordering DSTV from east and west serve as natural obstructions against most easterly and westerly winds.

Fig. 8 shows the annual sand drift potential in Wadi Araba Desert. The resultant sand drift direction (RDD) is mainly to the south, southeast or southwest. Previous studies in many localities of the Arabian Gulf showed that the season of high drift potential of dune sands is associated with the Shamal winds at

![Graph](image-url)
speeds between 12 and 26 knots/h (6.1–13.3 m s\(^{-1}\)) (ARAMCO Survey, 1974; McKee, 1979) similar to results of the present study. For example, sand drift shown in Fig. 8 is compatible with the Dhahran sand rose of the Jafurah sand sea, Saudi Arabia where the RDD implies drift to the SSE (Fryberger et al., 1983). The annual sand drift potential (DP) measured in vector units is between 140 (year 2001) and 1070 (1975), which corresponds to values between 10 and 80 m\(^3\)/meter-width (Fryberger et al., 1983). The index of directional variability RDP/DP (RDP: resultant drift potential) is between 0.5 and 0.9, which is arbitrarily classified as intermediate-high ratios (McKee, 1979). This means a narrow–wide unimodal flow pattern of less directional variability typical for barchanoid dunes (McKee, 1979).

3.3. Aeolian terrain facies

3.3.1. Sand dunes

Four sand dune fields were mapped in Wadi Araba. From north to south these are: (1) Salmani sand dune field, (2) Fidan sand dune field, (3) Qa As–Suaydiyyin sand dune field, and (4) the Gharandal–Taba sand dune field (Fig. 1). The Gharandal–Taba is the largest dune field (200 km\(^2\)). Dune types were determined according to the descriptive typology of dune and the mode of complexity of pattern displayed by the groups of dunes (McKee, 1979). Simple and compound dune types are common in the study area (Figs. 9 and 10). Complex dunes are occasional and only observed by aerial photograph analysis in the mountainous region in the eastern parts of the Gharandal–Taba- and Qa As–Suaydiyyin sand dune fields (Fig. 11). Barchanoid dunes and intergrading subtypes: barchans (Figs. 9 and 10), barchanoid ridges (Fig. 12) and transverse dunes (Figs. 12 and 13) are very common; linear dunes (Fig. 14), nabkhas and climbing dunes are less common.

The barchan dunes vary in size. They usually occur as a solitary or as a compound type mainly anchored by shrubs. Small ephemeral barchans (<3m. high) lasting a few months and mesobarchans (3–10 m high) lasting 1–30 years (Cooke et al., 1993) are present. Barchan dunes are mainly confined to the regions where winds allow for crescent forms to grow and to persist (Bagnold, 1941; Finkel, 1959; Sharp, 1964). Barchan dunes maintain their size by sands supplied from the upwind side (Hastenrath, 1987). Isolated barchans tend to develop where limited amounts of sand are available and almost unidirectional currents (McKee, 1979; Wasson and Hyde, 1983). We suggest that the compound barchan ridges occur where the large basal mound has a single proportional slip face and an upwind slope covered with smaller barchans or smaller barchanoid ridges. The small dunes are all oriented in the same direction as the main dune. An oblique sand supply accounts for some asymmetric barchans.

Barchanoid ridges and transverse dunes form in the same wind regime as barchan dunes in regions where more sands are available and winds are variable (Wasson and Hyde, 1983). Barchans merge into wave-like shapes to produce barchanoid ridges (Cooke et al., 1993) or transform into transverse dunes in regions where greater sand supply is available (Araya-Vergara, 1987). In the study area, the barchanoid ridges and the transverse dunes do occur as

\[\text{Fig. 6. Time series 1975–2001, showing the frequency % for wind speed above the threshold velocity.}\]

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groups with interdune corridors of different sizes (tens–hundreds meters). The crestlines are sinuous and the dune bodies tend to adopt linguoid–barcha-
noid shape with a gentle northern stoss side and a steep southern lee side. The average dip angle of the windward side (stoss side) is usually $<20^\circ$ and for the
leeside is >30°. As noted by McKee (1966), such dip angles are typical for barchan dunes early in their formation in the presence of unidirectional winds. Active short-lived wind ripples appear on the stoss sides and crests of dunes but are absent on the slip faces. Active short-lived ripples tend to develop on sandy surfaces that are in a state of relative equilibrium or slow deposition, while surfaces experiencing marked erosion or vigorous deposition generally do not display ripples (Sharp, 1963). Seppälä and Lindé (1978) reported that ripples in well-sorted sands of a mean size ca. 0.2 mm take less than 10 min to reach equilibrium under new wind conditions. Field study showed that the stoss sides of barchanoid dunes are usually overloaded with coarse-grained sands and the lee sides are overloaded with fine-grained sands.

Fig. 8. Annual sand roses based on wind data from AAMS station. Sand drift potential (DP) are in solid lines, resultant sand drift direction (RDD) are in dash lines. Values of DP and resultant drift potential (RDP) are in vector units. RDP/DP is index of directional variability.
Bagnold and Barndorff-Nielsen (1980) showed that the probability that a grain will be eroded or deposited is likely to be a function of the logarithm of the grain size. Small grains tend to move away from the windward side (stoss side) and accumulate in the lee side.

Despite the prevalence of barchanoid dunes created by northerly winds, with their downwind slip faces directed south, it seems that some wind reversals, i.e. southerly winds produced another group of barchan dunes with downwind slip faces directed north (e.g. Gharandal–Taba sand dune field). A similar case was mentioned by Fryberger et al. (1983) in the Jafurah sand sea, Dhahran area in Saudi Arabia.

Barchanoid dunes have tabular cross bedding preserved on slip (downwind) faces with foresets 0.5–1.0 m in length and a slope angle around 30°. Tabular cross bedding is caused by grainfall produced-strata or straight-crested dune migration (Fryberger et al., 1983). The crossbeds are truncated at the top by subhorizontal to slightly oblique reactivation surfaces as a result of scouring, possibly simultaneous with wind direction shifts. Cross bedding in dunes is occasionally disturbed or contorted by plant roots bioturbation.

Linear dunes (Gharandal–Taba sand dune field) are straight-crested to slightly sinuous sand ridges with two symmetrical slip faces. They are typically much longer than they are wide. The dunes stand a few meters above the interdune corridors with their long axes extending more than 1 km. They commonly
Fig. 9. Solitary barchan dune (Qa As–Suaydiyyin sand dune field) with two asymmetric arms directed SE. Western mountain range seen in background of photo.

Fig. 10. Aerial photograph showing compound dunes of barchanoid type (Fidan sand dune field). Arrow indicates sand drift direction.
occur as isolated ridges or sets of parallel ridges separated by regularly spaced corridors of widths < 100 m. The origin of linear dunes is confusing, with several models suggested in the literature. The barchan-to-linear dune model (Bagnold, 1941) is more likely to explain the origin of linear dunes in Wadi Araba Desert for two reasons: (1) availability of wind strength variation throughout the year, and (2) the juxtaposition of barchanoid (dominant type) and linear dunes. An asymmetric barchan is converted to a linear dune by alternating strong winds that build the dune and gentle persistent winds which elongate the dune. However, this model has little support from field evidence (e.g. Lancaster, 1980; Tsoar, 1984; Hastenrath, 1987). Fig. 14 shows pairs of linear dunes that merge at low angle Y-junctions. This means that parallel linear dunes are no longer stable in their original places, but they are reworked by wind currents. The Y-junctions are open in an east–southeast direction. They may be produced by side winds at oblique angles (Bagnold, 1941; Tsoar, 1982; Thomas, 1986). An alternative explanation is that Y-junctions have been formed by wind flowing in contra-rotating helical spirals or roll vortex pairs transporting sands obliquely (Bagnold, 1953; Folk, 1971; Glennie, 1987; Cooke et al., 1993). The downwind vortices may be produced by duststorms frequently occurring in parallel lines (e.g. Hastings, 1971; Swift et al., 1978; Middleton et al., 1986). The subdued topographic mass of Jabal Um-Nukhyla, located east of the Gharandal–Taba dune field (Fig. 1), may create natural wind passages or tunnels that would facilitate the formation of wind vortex pairs.

Nabkha dunes of various sizes develop when aeolian sands move between growing bushes and feed sand mounds between them. The nabkhas in the study area are randomly distributed because of the scattering of desert plants. Stems of bushes are still visible in small nabkhas, but in larger ones they may entirely covered by accumulated sands. Different sizes of the nabkhas are affected by changes in wind velocity and wind direction, being large when...
the winds are light, and small when they are strong (Worrall, 1974; Warren, 1988; Hesp, 1989; Cooke et al., 1993). However, some of the nabkhas from the
study area are built of cohesive, silty-fine sand when they form near stream channels and around mudflats.

Fig. 12. Aerial photograph showing barchanoid-transverse dune system (Fidan sand dune field). Arrow indicates sand drift direction.

Fig. 13. Transverse dune in front of basement igneous rock (upper right) and an interdune trough in foreground.
The anchored dunes (climbing, echo, and flank dunes) occur within the mountainous regions where drifted sands climb above the foothills. They may reach sometimes the summits of the topographic highs (e.g. Jabal Um-Nukhyla, south Gharandal). These dunes develop as a result of upwind vortices in front of the scarp (slope angle is >50°). The vortex developed by the velocity gradients above the ground (Bowen and Lindley, 1977; Cooke et al., 1993). Echo dunes develop when the vortex is capable of sweeping out a corridor in front of the scarp. When winds accelerate around the flanks of the topographic highs, they become more erosive and are capable of clearing corridors between the topographic highs and the accumulating sand mounds. Flanking dunes develop in at these locations (Cooke et al., 1993).

3.3.2. Interdune troughs

Interdune troughs which vary in size and location are mainly at three types: (1) dry depositional interdunes between barchanoid dunes, which consist of laterally continuous and flat-gently dipping thin layers or laminae showing almost a transition upward into dunes. These laminae resulted from downwind migration and deposition of current ripples. Surfaces of interdunes show current ripples and have some plant remains and mottled-roots usually around moist areas. Gravel deposits of intermittent streams occur in the dry depositional interdunes, (2) dry erosional interdunes between linear dunes, with scour surfaces developed as a result of the storage of drifting sand by the dunes immediately upwind, despite lower drift rates in interdune areas (Ahlbrandt and Fryberger, 1981; Fryberger et al., 1983), and (3) damp depositional interdunes with flourishing desert plants (≥1 m. height) between barchanoid dunes (e.g. Salmani and Fidan sand dune fields). This type of interdune has no signs of salt encrustation. Moistening of interdunes and the growth of plants above the interdune surface are
more likely to be related to a saturated water table of low salinity. The presence of vegetative cover is indicative of lower wind velocity and lower rates of sand drift. It seems that this type of interdune was built up of laterally continuous thin beds of horizontal lamination (≥ 0.3 m total thickness) considered as a subsoil layer for plant growth. Episodic stream flooding causes some local erosion in interdune troughs that has led to exposure of the horizontally laminated sediments close to the banks of dry stream channels. Horizontal lamination in sands is indicative of slow deposition rates. Gravelly sediments occur on the bottom of stream channels.

3.3.3. Sand sheets

Sand sheets in Wadi Araba are of the unvegetated or barren sand sheet type. The sand sheet surface is almost smooth, flat and covered with thin mantle of coarse sands and widely spaced sinuous-crested megaripples (wavelength >1 m). The smoothness of sand sheet surface is likely to be created by the abrasion of transported sand. Lian-You et al. (2003) showed that the abrasion capacity of transporting sand is increased logarithmically with wind velocity, and the abrasion itself is affected by sand transport rates. Megaripples formed at the sand sheet surface were more likely developed during windstorms. Sharp (1963) and Sakamoto-Arnold (1981) suggested that the megaripples in the southern San Joaquin Valley, California were produced in a short period of time, of not more than hours, under windstorm conditions. However, some megaripples may take years to develop and may last for centuries (Cooke et al., 1993). The idea that ripples tend to flatten out at higher wind strengths and greater rates of deposition, and can result in sand sheet formation (Glennie, 1987) is a possible origin for the unvegetated sand sheets in the study area. Shallow test pits (≈ 0.5 m depth) made into the sand sheets showed a horizontal lamination facies truncated at the top by gently dipping ripple cross-lamination (dip angle ≈ 15°). Glennie (1987) pointed to the possible change in lamination patterns when the mode of transport changes from traction to grainfall. Fryberger et al. (1983) assumed that sand sheets develop independently of dunes particularly when steeply dipping crossbeds are absent, and, the sand sheets grow vertically and laterally from ripples and grainfall deposition during episodes of sandstorms.

3.4. Textural attributes of the aeolian terrain facies

3.4.1. Grain size

Bagnold (1941) defined dune sands as any particle with a grain size between 0.02 and 1.0 mm. Ahlbrandt (1979) used the range between 0.1 and 1.6 mm. Despite the discrepancies in the quantitative definition of dune particle sizes, dune sands are qualitatively defined as any particle that is light enough to be moved by winds but too heavy to be held in suspension in the air. Results of the grain size analyses (Fig. 15a) showed that dune sands fall between 0.05 and 2 mm, which are very, much close to grain size of most sand sea deposits as mentioned by Cooke and Warren (1973). The mean grain size is between 0.2 and 0.35 mm. Dune sands are generally well sorted to moderately well sorted (0.35f–0.7f).

The sediments from the interdunes (Fig. 15a) are more variable in grain size and less well-sorted than the dune sands. The grain size is between 0.06 and 5 mm with as up to 40% coarser particles (>1 mm). The mean grain size varies between 0.25 and 0.6 mm.

Sands of sheet-like bodies have grain sizes between 0.1 and 0.5 mm, i.e. towards an excess of fine sand in comparison with interdune sands (Fig. 15a). The mean grain size is between 0.15 and 0.4 mm. Sheet sands are better sorted than the interdune sands.

Significant differences of grain size were observed between the sand particles from the stoss side (upwind side), crest, and the lee side (slip face) of barchanoid dunes. The results (Fig. 15b) showed that an excess of fine tailing is in the order: stoss side > crest > lee side. The mean grain size of sand particles in the stoss side, crest, and lee side is between 1.0 and 1.6 mm, 0.7 and 1.1 mm, and 0.6 and 0.9 mm, respectively.

3.4.2. Roundness

The roundness of 85–90% of sand populations from the stoss side, crest, and lee side of barchanoid dunes is commonly between angular to subrounded, with little differences in the abundance ratios of the classes (Fig. 16). The degree of angularity is more among sand grains from the stoss side, in contrast to rounding in the crest and lee side samples which tend to be more rounded, remembering that the fine tailing of grain size is toward the lee side (Fig. 15b). At first glance, these findings are incompatible with the
Fig. 15. (a) Grain size analyses of sands from barchanoid dune, interdune trough, and sand sheet. (b) Grain size analyses of sands from the stoss side, crest, and lee side of the barchan sand dune.
Fig. 15 (continued).
general idea that rounding in aeolian sands increases with grain size (e.g., Goudie and Watson, 1981; Khalaf and Gharib, 1985). Similar trends were achieved from the analysis of aeolian sands from different regions of the world, where the degree of highest rounding does not necessarily occur in coarser grains (e.g., 1–2 mm), but instead mean rounding values are relatively higher in finer sand grains (Thomas, 1984, 1987a; Khalaf and Gharib, 1985) as later discussed by Thomas (1987b). This contradiction may resolved by considering that sand grains on the upwind side (stoss side) as they moved by creep or sliding; rolling and grain impacts or collisions may be inhibited. In the crest and lee side, fine–medium sand particles are more likely to be moved by saltation, involving grain impacts and collisions. Moreover, strong winds preferentially sort out more spherical, rounded sand grains (Mazzullo et al., 1986) more likely near the crest and brink point. In terms of sand sources, the aeolian sands in the study area are derived mainly from similar nearby sources. So, it is not unlikely that the grain shape is an inherited property (Tucker, 1991). Quartz sand grains display slight elongation in the direction of their optic c-axes, a property that is more obvious in coarser grains in comparison with small grains, which will inhibit well rounding, and high sphericity of sand particles.

3.5. Mineralogy of the aeolian terrain facies

The sediments of barchanoid dunes, interdunes, and aeolianite (cemented sand sheet-like bodies) are composed mainly of quartz, feldspar, mica, calcite, and dolomite. The dominant clay mineral is kaolinite with a minor content of smectite. Sands of interdunes showed a peak at \( 2\theta \ 10.7 \ (d=8.26 \ \text{Å}) \) characteristic for the hydrated iron-sulphate mineral coquimbite \( \{\text{Fe}_2(\text{SO}_4)\}_2\cdot 9\text{H}_2\text{O}\} \) which forms instead of gypsum under conditions of low salinity and more oxidizing nearsurface groundwater conditions.

Quartz, feldspar and mica of detrital origin are derived from the nearby basement rocks and the overlying sediments. The dominance of quartz in dune sands is primarily due to its durability and resistance to chemical weathering in addition to the source availability. It is likely that the Paleozoic sandstones (mainly arkosic- and quartz arenite), exposed along the eastern shorelines of the Dead Sea, are other
suppliers for dune sands in the Salmani field, the nearest dune field to the southern tip of the Dead Sea. The nearby Mesozoic–Tertiary carbonate rocks and inland sabkha sediments in southern Wadi Araba are possible sources for detrital carbonate minerals in dune sands. In the interdunes and in the aeolianite, calcite and dolomite are presumably authigenic and act as pore-filling cement. The calcite in the aeolianite is more abundant in the uppermost horizons (peak intensity ~ 1500 counts) and tends to decrease in the intermediate and lower horizons (peak intensity ~ 1250 counts). Calcite possibly concentrates in the capillary fringe zone with the help of evaporation. Sporadic rainfall may leach calcite downward into the lower horizons. Results of XRD give unequivocal evidence of clay illuviation provided by more kaolinite in the lower horizons of the aeolianite where the peak intensity changes from 1000 to more than 1500 counts.

3.6. Expected age of the aeolian terrain

Speculations about the age of sand dunes and other associated facies in the aeolian terrain of Wadi Araba Desert are based on previous research. The paleoclimate trend in the late Pleistocene Last Glacial Maximum (LGM) (≈ 22–15 ka) showed evidence that it was cold and dry in the Sahara, Arabia and SE Asia (China) with less rainfall, and expansion of deserts (Gasse et al., 1987; Petit-Maire and Guo, 1997; El-Baz, 1998; Zhuo et al., 1998). Abed and Yaghan (2000) suggested a similar cold, dry paleoclimate trend in the Jordan Valley during the LGM (≈ 22–15 ka). Glennie (1987) considered the glacial periods of the Pleistocene as times for the major erosional features of globally stronger winds that blew with strength for a much greater part of the year in modern deserts. He believed also that fluvial erosion and wind deposition were much more effective in the interglacial periods of late Pleistocene–early Holocene “climatic optimum” around 8000–5000 years BP, when convection-induced thunderstorms were more likely to be responsible for building alluvial fans in today’s deserts (Hötzl and Zötl, 1978; Jado and Zötl, 1984). Similarly, Gasse et al. (1987), Petit-Maire and Guo (1997), El-Baz (1998) and Zhuo et al. (1998) believed that warmer interglacial periods in the late Pleistocene and the Holocene optimum were associated with high rainfall and retreat of deserts. Hence, the climatic optimum of interglacial periods of latest Pleistocene–early Holocene are more likely time periods of fluvial erosion, wind deposition and sand dune formation in the study area, despite the likelihood that deserts shrunk at these times. The eluviation of kaolinite into lower aeolianite sand-sheet-like horizons is evidence of wet pluvial episodes between dry, cold climate conditions.

4. Conclusions

The principal approaches to dune study in Jordan and probably elsewhere are practically based on the classification of dune forms from field study, landsat imagery, and aerial photographs. Textural studies for grain size and grain morphology, and understanding physical processes of wind transport, sand drift, and deposition of sands are also important. Sand rose patterns enabled us to draw conclusions on the probable relation of wind strengths and directions to dune type in the study area. The present study shows that the types of aeolian terrain facies mainly depend on the amounts of sand supply, wind variability and wind speed, and vegetation cover. Local topography (mountain scarps, inland- or marginal hills), desert plants, and desert plain roughness are responsible for obstructing wind forces, allowing sands to pile up in drifts and ultimately for different shapes and sizes of dunes to form.

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