Geology of Sand Dunes

While one-quarter to one-third of the world's deserts are covered with sand, little research has taken place in ergs (sand-covered desert areas) relative to non-sandy areas. The great distances and hardships involved in reaching sandy areas, the general lack of wind data and other meteorological records, the almost total lack of human activity in ergs, and the difficulty of getting a macro-scale view of sand seas from the surface have all contributed to the lack of knowledge of the movement and accumulation of sand in deserts (McKee, 1979).

The single major work on sand dune formation was written by R. A. Bagnold and published in 1941. Bagnold used wind tunnel experiments to make quantitative predictions about sand movement and accumulation and then successfully corroborated most of those predictions in field tests in the Libyan desert. He also speculated about some of the larger-scale phenomena that he was not able to test in his wind tunnel. Despite the fact that Bagnold's work is over forty years old, it is still the most widely quoted source on eolian sand deposits. According to a modern-day student of sand dunes, Bagnold's work "with relatively few modifications, has stood the test of checking since publication" (McKee, 1979, p.5).

Since Bagnold, additional important contributions to the study of eolian sands have been made in the field by Sharp (1963; 1966; 1978) and McKee (1979), among others. Important work has also been done in the application of aerial photography (Smith, 1968) and Landsat imagery (Fryberger and Dean, 1979; Breed and Grow, 1979) in the study of erg morphology. This paper will summarize the findings of research carried out in inland (non-coastal) desert eolian sand deposits.

THE NATURE OF SAND

Sand is defined by its size, although exact quantitative ranges vary with author. Bagnold defines sand as any particle between .02 mm and 1.0 mm in diameter, while Ahlbrandt (1979) uses the range of .1 mm to 1.6 mm. Despite discrepancies in quantitative definition, which is often due to grain-size differences in the individual dunefields studied by each researcher, the authors agree that sand is qualitatively defined as any particle that is light enough to be moved by the wind but too heavy to be held in suspension in the air. Very fine particles that can be held in suspension are therefore classified as silt or dust, while heavier particles unaffected by wind are classified as pebbles or gravel (Bagnold, 1941; Cooke and Warren, 1973).

Neither mineral composition nor particle shape appears to have any significant effect on sand movement or accumulation. Any solid, non-cohesive particle, natural or human-made, that falls within the above-mentioned range is technically sand, including dry granular snow which can build up into dune-like drifts. While sand can be composed of various minerals, quartz makes up the bulk of the world's sand grains. This dominance is primarily due to the widespread distribution of quartz-containing rocks and to the chemical nature of quartz. Unlike other particles, quartz

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sand grains resist both mechanical and chemical breakup into smaller sizes (Bagnold, 1941).

While wind plays the major role in sand accumulation, the actual erosion of rock material into sand-size grains is primarily due to the action of water and ice (Bagnold, 1941). Wind does play a role in the abrading and rounding of sand grains, which results in inland dune fields having more rounded grains than coastal dunes, due to the increased distance of movement by wind (Ahlbrandt, 1979).

Sand composing inland dunefields generally varies more widely than coastal-dune sand in both grain size and degree of sorting due to the greater variety of sources for inland-dune sand and the varying distances from source to dune field. Source areas for inland sand seas include lake deposits, river deposits, alluvial fans, playas, glacial till, and sandstone bedrock (Ahlbrandt, 1979).

WIND-INDUCED SAND MOVEMENT

Individual sand grains are moved under the force of the wind in two distinct ways: saltation and surface creep. The primary method of sand movement is saltation. As wind moves over a sand deposit, it is able to pick up grains from the surface and give them a forward momentum, but the weight of the sand grains soon bring the grains back to the surface. If the surface is composed of coarse, immobile particles, such as pebbles, the sand grains will bounce directly off the hard surface and back into the air, where the wind will once again provide a forward momentum. These bouncing grains can move downwind at about half the speed of the wind. If the surface is composed of finer sand grains, however, a saltating sand grain will not bounce off the surface; rather, it will strike the sandy surface and bury itself. The impact will eject a second grain into the air to be blown downwind. This “splashing” form of saltation results in a slower rate of downwind movement than the bouncing motion on hard surfaces. Either process falls under the definition of saltation (Bagnold, 1941).

Saltation of sand grains along the surface accounts for about 75% of all sand movement by wind. However, due to the fact that sand grains average about two thousand times the weight of the atmosphere, not all winds will move sand. Wind speeds must reach what Bagnold (1941) calls a “fluid threshold,” defined as the wind speed necessary for sand to start saltating under the direct pressure of the wind. The fluid threshold varies in direct proportion to the predominant

"Grains larger than one millimeter in diameter are generally moved by a second process called surface creep."

grain size of the sand surface, generally ranging from ten to twenty miles per hour (Bagnold, 1941; Sharp, 1963).

After sand grains start moving under direct wind pressure, wind speeds lower than the fluid threshold can maintain sand movement. Once saltation has begun, direct wind pressure is no longer necessary to lift sand grains into the air. The impact of saltating grains provides enough energy to knock new grains into the air (assuming a sandy surface); thus, the wind need provide only enough energy to move the airborne grains downwind. The wind speed necessary to maintain saltation once it has begun is termed the "impact threshold" and defined by Bagnold (1941, p.32) as the velocity at which "the energy received by the average saltating grains becomes equal to that lost (by impact), so that motion is sustained." Like the fluid threshold, the impact threshold increases with increasing grain size.

Saltating sand grains usually stay close to the surface. In his wind tunnel experiments, Bagnold (1941) found the average height of windblown sand to be about ten centimeters, although both height and speed of saltating grains increased with wind speed. At Kelso dunes, a fifteen-year study indicated that 90% of saltating grains moved within sixty-four centimeters of the surface, with maximum sand-blast effect at twenty-three centimeters (Sharp and Saunders, 1978).

The overall volume of sand moved has an exponential relationship with wind velocity, i.e. as wind speeds increase, the downwind rate of sand movement increases exponentially. Even during intense "sand storms," however, at maximum wind speeds and sand movement, saltating grains rarely exceed two meters in height (Fryberger et al., 1979).

Due to an increased fluid threshold, heavier sand grains are rarely moved directly by wind pressure. Only intense storm winds can lift the heavier grains off the surface. Grains larger than one millimeter in diameter are generally moved by a second process called surface creep (Bagnold, 1941; Sharp, 1966). When saltating sand grains strike these heavy grains on the surface, they don't have enough energy to knock them into the air, but they do impart to the heavy grains a slight forward momentum along the surface. In this way, heavy sand grains up to two hundred times the mass of the saltating grains can be slowly moved downwind. Up to 25% of all wind-transported sand is moved by surface creep (Bagnold, 1941).

“Obstacles, however, are not needed for sand accumulation.”

SAND ACCUMULATION

Two primary factors are necessary for the accumulation of sand into sand sheets and dunes: 1) an adequate supply of sand, and 2) winds strong enough and persistent enough to move the sand (McKee, 1979). If these two conditions are met, large quantities of sand can be transported hundreds and even thousands of miles (Fryberger and Ahlbrandt, 1979).

What makes sand accumulate into piles rather than spread out evenly over an area? In general, sand will tend to accumulate any place “where a sufficient reduction of wind energy exists along the direction of sand drift in an active extensive system” (Fryberger and Ahlbrandt, 1979, p. 454). Any obstacle, such as a rock outcrop or a stand of vegetation, can force sand accumulation by lowering wind speeds and creating a “sand shadow” to the lee of the obstacle. Any small depression or gentle dip in an otherwise flat surface can fill with sand due to lower wind velocity within the depression (Cooke and Warren, 1973). Large areas of persistent wind deceleration, such as a basin or the base of a plateau, can spawn the creation of large ergs. In fact, most desert eolian sand seas do occur in basins (Fryberger and Ahlbrandt, 1979; Cooke and Warren, 1973).

Obstacles, however, are not needed for sand accumulation. According to Bagnold (1941, p.6), sand “alone of all artificial solids (has) the power of self-accumulation.” This self-accumulation results from two processes: 1) the differential speed of saltating sand over sandy versus non-sandy surfaces, and 2) the drag effect of saltating sand grains on wind velocity.

As mentioned in the previous section, saltation over a coarse surface generally takes the form of repeated bouncing of individual grains. In such cases, most of the wind-imparted momentum is conserved and grains move rapidly downwind. In saltation over a sandy surface, however, sand grains impact into the surface, transferring some energy to the surface (via surface creep) and some energy to dislodging other grains into the air. This process produces a slower downwind movement of sand (Bagnold, 1941).

A second factor favoring self-accumulation of sand is “saltation drag.” Friction produced by the saltating sand grains slows down wind speed near the surface. Thus, despite the smoother surface of a sandy area, “a given wind can drive sand over a hard immobile surface at a considerably greater rate than is allowed by the loose sandy surface” (Bagnold, p. 72). This negative

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feedback effect of saltating sand is so great, in fact, that it places a maximum limit on near-surface wind velocity. In effect, increasing wind velocity also increases saltation and saltation drag to the point where energy lost by increasing friction equals the energy gained by increasing velocity. At that point, further increases in wind velocity outside the sandy area do not produce any increase in velocity over the sand; in fact, increased saltation drag may actually reduce near-surface wind speed over the sand and result in vigorous deposition of grains. Hence, strong winds tend to favor sand accumulation in areas already sand-covered (Bagnold, 1941).

SMALL-SCALE SAND ACCUMULATION FEATURES

Once sand grains have accumulated into relatively large sandy patches, small-scale geomorphic features will often result, of which surface rippling is the most common. Rippling tends to develop on sandy surfaces that are in a state of relative equilibrium or slow deposition. Surfaces experiencing either marked erosion or vigorous deposition generally do not display rippling (Sharp, 1963).

Even an apparently completely flat sand sheet is inherently unstable. Due to variations in grain size a small but significant surface roughness exists, allowing for wind to pick particles. Because larger grains saltate more slowly than smaller grains, they tend to accumulate into "jams", creating more surface roughness. Also, chance unevenness on the sand surface will always be present (Sharp, 1963).

Any unevenness, either random or saltation-induced, will tend to perpetuate itself due to the sensitivity of saltating sand to slight variations in the angle at which grains impact the surface (angle of incidence). What Bagnold refers to as the "characteristic flight path" of saltating sand grains is normally at a very low angle. Since, in most natural sand surfaces, one grain size predominates with a normal distribution around the peak size, saltating sand grains are striking the surface at a relatively uniform angle (approximately ten degrees for the average grain).

When surface unevenness occurs and a small hollow is created, less saltation impacts will occur on the upwind side of the hollow than on the downwind slope (Figure 1 [page 6]). As a result, surface creep along slope AB is considerably greater than creep along slope CA, as slope CA resides in a "saltation shadow." Consequently, sand is removed from point A and deposited at point B, creating a ripple. This, in turn, produces

"As ripples increase in height, they move into levels of higher wind speeds, causing heavier grains to be blown from the ripple crests and into the troughs, filling them in."

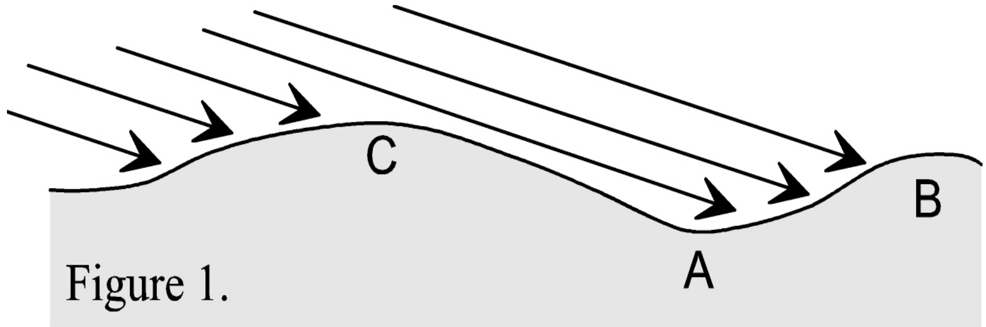

Figure 1.

a second hollow downwind of the newly-created ripple and the process repeats itself with numerous parallel ridges forming at right angles to the wind direction. The coarser sand grains will tend to collect at the crest of the ripples since they are not moved as easily by the wind and there is little surface creep down the lee side of the ripples (Bagnold,1941; Sharp, 1963).

As a rule, the "wavelength" of the ripples (the distance between crests of successive ripples) increases with increasing wind speed and reflects the increasing height of ripples and the resultant lengthening saltation shadow (Sharp, 1963). In extremely heavy winds, however, ripples flatten out completely because all grain sizes are easily moved by the wind and the differential saltation and creep rates needed for ripple formation decline (Bagnold, 1941; Sharp, 1963).

The height of an individual ripple is a function of grain sorting. The more uniform the sand surface the shallower the ripples, because of the reduced amount of differential saltation and surface creep. Due to the interference of wind speeds by the growing ripples, a maximum height limit exists. As ripples increase in height, they move into levels of higher wind speeds, causing heavier grains to be blown from the ripple crests and into the troughs, filling them in (Cooke and Warren, 1973). Bagnold (1941) claims that the ripple height is generally no more than one-tenth the wavelength of the ripples. In the Kelso dune filed, Sharp (1963) found a maximum wavelength of nineteen centimeters and a maximum height of one centimeter, with an average wavelength/height ratio of 18. However, Sharp notes that no satisfactory universal qualification of this height-wavelength relationship has been obtained.

In his studies of the Kelso dunes, Sharp (1963; 1966) also found that ripples move downwind at relatively fast rates. At a threshold velocity of 18 kph, ripples advances downwind at a rate of .9 cm per minute, with the

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rate increasing to 8 cm per minute during the strongest winds. Consequently, Sharp concluded the “adjustment in size, shape and spacing can presumably occur rapidly in response to differences in velocity” (1963, p. 631). From an initially flat surface, ripples can form a complete pattern in ten minutes (in a 48 kph wind) and can flatten out, reform or change direction as quickly. This rapid formation and movement of sand ripples also contributes to a large volume of sand movement. At one test plot, Sharp discovered that, in one hour’s time, 48 kph winds could move 6000 pounds of sand across a 32-meter line.

In areas where the sand surface has a relatively large number of coarse sand grains (greater than 1 mm in diameter), a second type of small-scale feature occurs, called a “ridge” by Ragnold and a “granule ripple” by Sharp . Granule ripples generally form in sands with a bimodal distribution of grain size - one fine and one coarse - where winds are moderate-to-strong but not strong enough to pick up coarser grains (Bagnold, 1941; Sharp, 1963).

Bagnold claimed that, like ripples, granule ripples resulted from finer saltating grains pushing coarser grains (via surface creep) into jams. Unlike sand rippling, however, these concentrated ripples of coarser grains are rarely, if ever, moved by direct wind pressure. Consequently, they are more stable and can grow to larger dimensions than sand ripples. As more and more coarse grains arrive from upwind, the granule ripple can grow quite high and the resultant saltation shadow prevents movement of large grains from the crest into the leeside trough. In contrast to sand ripples, growth and movement of granule ripples is very slow, and individual granule ripples can exist for decades and even centuries, allowing for much greater heights and wavelengths than can develop on ephemeral sand ripples (Bagnold, 1941).

Sharp (1963) tested Bagnold’s theories in the Kelso dune field. Sharp found that granule ripples were generally located in deflation hollows between dunes, where winds had removed the finer material and produced a coarse-grained surface. The granule ripples at Kelso dune field were more irregular than the sand ripples, often forming wavy chains resembling miniature barchanoid ridges. As Bagnold predicted, Sharp found that granule ripples indeed adjusted very slowly to changing wind velocities and directions. Sharp found granule ripples that were at least several months old and measured up to 12.5 cms high and over two meters in wave-

“Wind speeds are highest and saltation drag is least at the windward edge of a sandy patch.”

length. (Bagnold reported granule ripples in Africa sixty centimeters in height and six meters in wavelength.) Sharp also states that he found no gradations between sand ripples and granule ripples; rather, he found them to be distinct features even when occurring side by side, “with a sharp line of demarcation” (1963, p. 632).

LARGE-SCALE SAND ACCUMULATION FEATURES

According to Bagnold (1941), certain conditions must be met for flat sandy areas to build up into true dunes. The primary prerequisites are relatively strong winds and a fine-grained surface. Coarse sand grains and pebbles tend to stabilize a surface of sand by preventing light or moderate winds from moving the finer grain sand mixed in the coarser material. Bagnold has observed that large sand sheets (without dunes) are generally covered with coarse sand or small pebbles and are generally devoid of ripples. Only strong winds can remove the finer grains from amidst the coarse grains and carry them downwind where they can form dunes by self-accumulation. The rougher the surface of a source of sand (the more coarse material mixed with the finer sand) the stronger the winds needed for dune formation. Also, sandy areas with periodic wet-season rainfall may also be prevented from developing dunes by a light, intermittent vegetation.

Assuming a steady input of sand from an upwind source, a patch of sand that meets the above conditions and is four to six meters long can develop into a true dune. A true dune generally refers to a self-accumulating mound of sand with a distinct slip-face, as opposed to an obstacle formed pile of sand, such as a sand shadow, which will not be discussed in this paper. Saltation drag is a primary factor contributing to dune formation. Wind speeds are highest and saltation drag is least at the windward edge of a sandy patch. The further downwind on the sand, the greater the drag and the slower the winds, resulting in differential sand movement across the sandy surface. As the movement of grains slows toward the leeward side of the sandy area, accumulation of sand increases and sand begins to mound up on the leeward end (Bagnold, 1941) (Figure 2 [page 9]).

As the sand mound grows, the point of maximum sand deposition on the leeward face moves closer to the summit, causing a steepening of the leeward face relative to the windward face. The steepening and growing dune now forces wind out over the top of the dune rather than

“Any sand driven up the windward slope drops out as it hits the stagnant air above the slip face.”

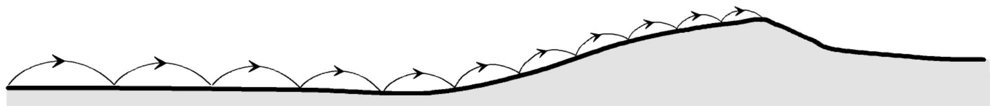


Figure 2.

down the leeward face. The saltating sand drops out at the crest and further steepens the leeward face until it reaches its angle of repose (about 32 to 34 degrees for dry sand), at which time gravity may pull sand from the crest down the leeward slope in the form of either isolated slow-flowing avalanches or the shearing and slumping of whole blocks of sand. This sand movement by slippage rather than by saltation or creep has earned the leeward face of an advancing dune the name of “slip-face” (Bagnold, 1941).

The slip-face is an effective sand trap. Bagnold’s field observations on barchans showed that even in strong winds a near-perfect wind shadow exists along the slip-face. Any sand driven up the windward slope drops out as it hits the stagnant air above the slip face, resulting in renewed steepening of the slip-face near the crest of the dune until the angle of repose is once again reached. Thus, entire dunes slowly move downwind due to avalanching and slumping along the slip-face. The rate of dune advance is directly related to the rate of sand movement over the dune crest and inversely related to the height of the slip-face, i.e. as dunes grow in height their forward movement slows (Bagnold, 1941).

Sand dunes of many different types have developed, resulting in various classification schemes and descriptive terminology as well as a considerable amount of confusion. Regardless of what terminology and classification scheme is used, most students of dune formation would agree with Bagnold’s assertion that dune accumulation “is intimately connected with the relative strength, duration and direction of alternating periods of weak and strong winds” (1941, p. 184).

Bagnold felt that strong winds tended to build up dunes in terms of height, while gentler winds generally extend the length of the dunes at the expense of height. Cooke and Warren (1973, p. 271) explain this phenomenon as follows: “a wind capable of moving sand over a sandy surface could not move it over a pebbly surface, so that a sand patch would be eroded and extended downwind by such a wind. A stronger wind on the other hand might move more sand over pebbles, because

"The primary difference among the three barchanoid subtypes is the amount of sand available for dune formation."

of the better rebound from the harder surface...and the sand patch grows."

Because it is generally agreed that the primary determinant of dune forms in any given area is the nature of the wind regime of that area, most dune classification schemes are based primarily on the direction(s) and intensity of the winds carrying the sand, although other factors, such as abundance of sand, presence or absence of vegetation, topography, the nature of surface material and the dominant grain size of the sand are also contributing factors and must be taken into consideration (Borsy, 1976; Fryberger and Ahlbrandt, 1979).

Borrowing heavily from other studies, McKee (1979) has developed a modern descriptive typology of dunes, although he also includes a generic explanation of the different types and subtypes he describes. McKee presents two related classification systems, one dealing with the basic shape and structure of individual dunes (based primarily on modifications of Bagnold's classification) and a second dealing with the complexity of pattern displayed by groups of dunes (based on field observations and remote sensing studies).

McKee classifies individual dunes in terms of the number and position of slip-faces, under the assumption that the number of slip-faces corresponds to the number of dominant wind directions in the local wind regime. Most dunes that have a single slip-face are classified in the "barchanoid" type, which consists of three intergrading subtypes: the barchan, the barchanoid ridge, and the transverse dune (Figure 3 [page 11]). All are considered to have axes perpendicular to a persistent, undirectional wind regime.

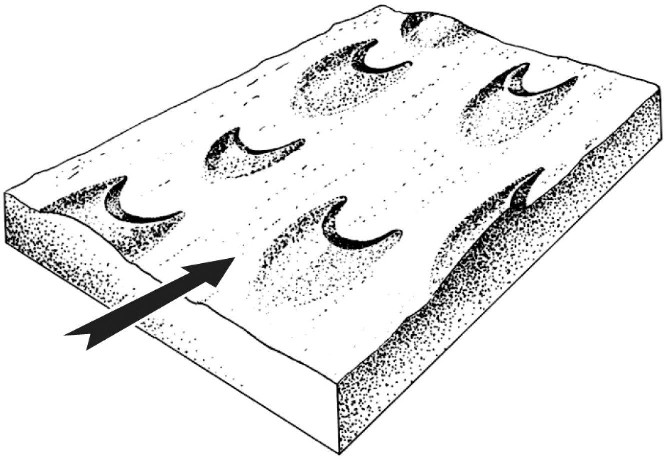
McKee feels that the primary difference among the three barchanoid subtypes is the amount of sand available for dune formation.

Barchans, despite their familiar crescentic shape, comprise only a small percentage of the world's dune areas and tend to develop where limited amounts of sand are available. For example, individual barchans have formed in the vicinity of the Salton Sea, with little or no sand found in areas adjacent to the dunes (Shelton, Papson, and Womer, 1978).

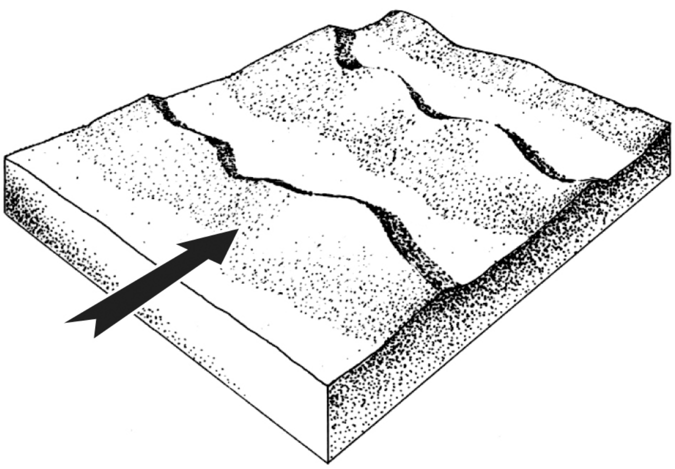
Maximum dimensions for individual barchans are approximately thirty meters in height and four hundred meters in width and length (Bagnold, 1941). Their relatively small size allows for rapid movement; individual barchans in Peru advance as much as forty-seven meters annually (Cooke and Warren, 1973).

As more sand becomes available, barchans merge into wave-like barchanoid ridges; and, if sand continues to accumulate, the barchanoid ridges grade into transverse dunes. These larger barchanoid subtypes can reach heights of over two hundred meters and extend for kilometers (Bagnold, 1941). This gradation of forms is similar to that of sand forms created by water currents. In laboratory experiments, Tyler (1979) observed “dunes” on sandy stream beds develop from barchan to barchanoid ridge to transverse forms.

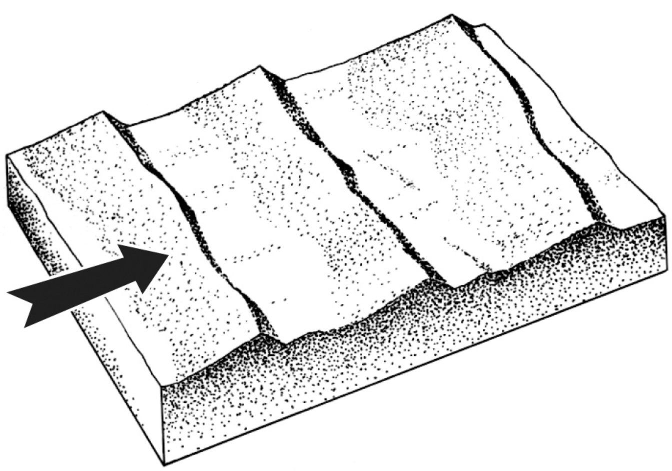
An additional type of dune with a single slip-face is the parabolic dune, with a crescentic form opposite that of the barchan (i.e. with horns pointed upwind). Parabolic dunes develop when vegetation begins to colonize the moister, less-mobile flanks of self-accumulating dunes. Vegetation has a stabilizing effect on the dune, retarding or completely stopping movement along the edges of the dune arms. The leading edge of the dune may continue to push forward and move ahead of the arms, creating the typical reverse-crescent shape. Eventually, the nose of the dune may be stabilized by vegetation as well, or it may continue to progress, break-



BARCHAN DUNES. Arrow shows prevailing wind direction.

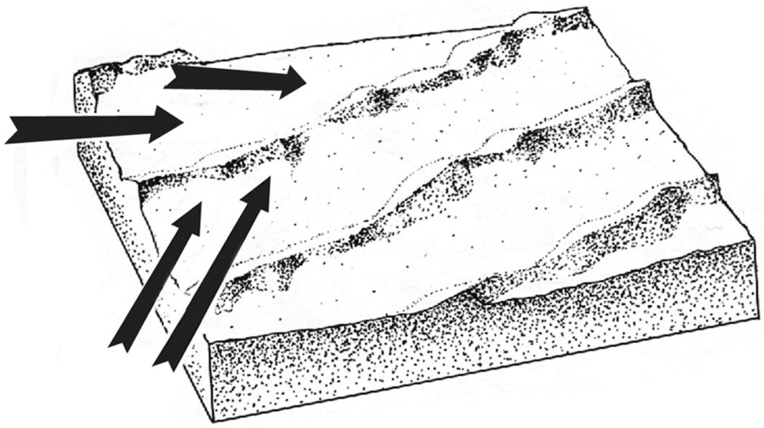


BARCHANOID RIDGE. Arrow shows prevailing wind direction.

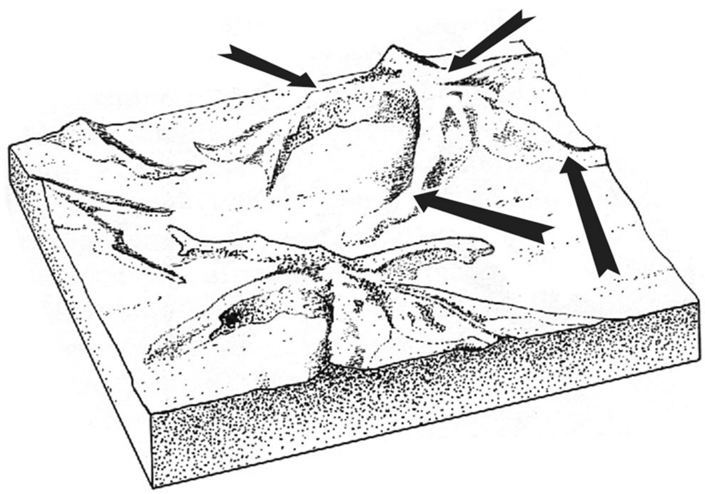


TRANSVERSE DUNE. Arrow shows prevailing wind direction.

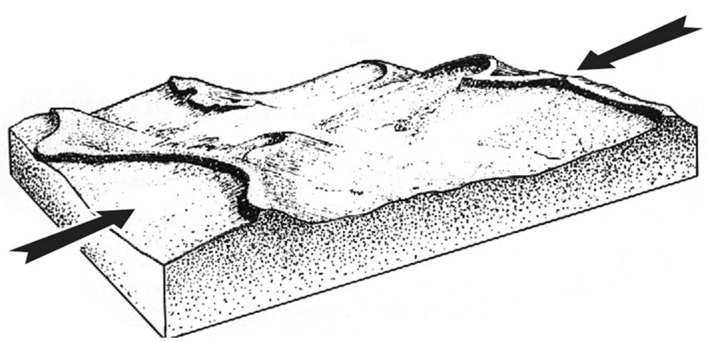
Figure 3. From McKee, 1979



LINEAR DUNES. Arrows show probable dominant winds.



STAR DUNES. Arrows show effective wind directions.



REVERSING DUNES. Arrows show wind directions.

Figure 4. From McKee, 1979

ing free of the arms and leaving two vegetated linear ridges in its wake (Cooke and Warren, 1973; McKee, 1979).

McKee recognizes two dune forms that display two slip-faces each, presumably a result of bimodal wind regimes. The linear (or seif) dune is the most common of all inland dunes and the most controversial in origin (Figure 4). These long, narrow dunes have sharp, steep crests due to the presence of a slip-face on both sides. Bagnold recorded linear dunes in Iran that were over two hundred meters high, twelve hundred meters wide and one hundred kilometers long.

While it is generally agreed that the dual slip-face structure of the linear dune does represent the work of winds from two different directions, the exact mechanisms of the wind regime is uncertain. Most researchers feel that acute bimodal winds are responsible for the creation of linear dunes (Bagnold, 1941; McKee, 1979), while others feel that linear dunes are created by helicoidal or vortex currents within a uni-directional wind regime (Borsy, 1976). Bagnold (1941) suggests that some linear dunes are formed when barchan dunes migrate from an area of unimodal winds into an area of bimodal winds, with one of the barchan's arms being extended into a linear dune.

“Reversing dunes are generally ephemeral features, occurring only immediately after reversing winds.”

The great diversity of location, surface texture and wind regimes in areas of linear dune development suggests that more than one mechanism may be responsible for the formation and growth of these dunes, and/or that a combination of two or more mechanisms may be operating at the same time. For example, some areas of linear dunes may have persistent, gentle-to-moderate unidirectional winds most of the year, during which helicoidal flow may operate; while occasional intense storm winds from a different direction (often not manifest during the brief period researchers are in the field) may also contribute to the building and lengthening of linear dunes.

A second type of dune with two slip-faces is the reversing dune (Figure 4) Better understood than linear dunes, reversing dunes are typically found in areas of bimodal winds from opposite directions. A generally persistent wind from one direction creates barchanoid-type dunes, while “reversing winds” create miniature dunes on the crest of the barchanoid dune with a slip-face in the opposite direction as the primary dune slip-face (McKee, 1979). In this respect, reversing dunes are generally ephemeral features, occurring only immediately after reversing winds. Resumption of normal wind flow from the opposite direction obliterates the smaller crestal dune and the main dune reverts to its original barchanoid form with a single slip-face.

Finally, McKee recognizes the star dune as a dune with three or more slip-faces. Star-dunes generally occur in the form of a central peak with three or more radiating arms, each with a slip-face in a different direction (Figure 4). Both star dunes and, to some extent, reversing dunes generally tend to grow vertically rather than laterally, as sand moves in and piles up from several directions (Ahlbrandt, 1979). Star dunes are sometimes found as the highest dunes amidst a field of transverse or linear dunes. The origin of star dunes in such situations is uncertain, but it is suspected that when transverse or linear dunes grow large enough their size and form may alter the local winds and create multi-directional winds and eddies persistent enough to build a star dune (McKee, 1979). Sharp (1978) noted a distinct alteration of ground-level wind direction among the transverse dunes in the Kelso dune field. As strong winds were blowing up the windward face of the dunes, winds from various directions (up to 90 degrees divergent) blew across the surface of the slip-face.

“Recent Landsat studies of the world’s ergs have allowed for a better understanding of large-scale dune-field structure and pattern.”

A second type of dune classification discussed by McKee (1979) and others before him is based upon the complexity of dune pattern: simple, compound and complex. Simple dunes are individual dune types of a unitary nature, “not divisible into clearly defined component parts” (Smith, 1968, p. 14). Any dune field with distinct individual dune types independent of each other would fall into this category, such as a field of barchans or a field of parallel linear dunes (McKee, 1979; Smith, 1968).

Compound dunes consist of two or more dunes of the same type “combined by overlapping or being superimposed” (McKee, 1979, p. 13). Examples of compound dunes are coalescing barchanoid ridges, small linear dunes forming on the top of larger linear dunes, small barchans climbing up the back of large barchans, and small parabolic dunes forming within the arms of larger parabolic dunes (McKee, 1979; Smith, 1968).

Complex dunes are those in which two different dune types coalesce or overlap. Examples of complex dunes include star dunes forming on top of linear or transverse dunes, and barchans forming in the hollows between linear dunes (McKee, 1979).

In an aerial study of North African dunes, Smith (1968) found all types of dune groupings, with complex dune patterns being the most common and most diverse. Smith believes that complex dunes have a complicated development over time, with climatic change and resultant shifts in dominant wind direction as major forcing factors. Cooke and Warren (1973) maintain that, on a global scale, complex patterns are far more common than either simple or compound forms.

Recent Landsat studies of the world’s ergs have allowed for a better understanding of large-scale dune-field structure and pattern. Breed and Grow (1979) found that dunes of similar type show the same patterns regardless of location, and concluded that “the relationships of mean length, width and wavelength are similar among dunes of each type, regardless of differences in size, form or geographic location” (p. 257). The authors also found a direct relationship between dune size and complexity.

In another important study, Fryberger (1979) used Landsat imagery and regional wind data to show global relationships between wind regimes and dune morphology. For each location under study, Fryberger developed a “sand rose” based on the traditional wind rose. The sand rose takes into account the intensity and duration of wind as

"Like barchanoids, linear dunes form in both high-energy and low-energy wind regimes."

well as wind direction, thereby providing a model of the amount of sand a particular wind regime is capable of moving in various directions.

Fryberger's findings support Bagnold's assertions that both wind direction and strength affect dune morphology. He confirmed the notion that, of all dune forms, barchanoid types are associated with the least variability of wind direction. He found most barchanoid types to form in areas with unimodal wind direction, although they also occasionally form in areas with acute bimodal winds (from two dominant directions less than ninety degrees apart) where one wind is much stronger than the other. While barchanoid dunes appear to form in both high-energy wind regimes and low-energy wind regimes, the stronger the winds, the less directional variability is required for barchanoid development.

The regional wind data used by Fryberger did not allow discrimination for each barchanoid subtype; however, Fryberger does note that all three subtypes are frequently found together in the same dune field, indicating that some other factor, presumably the amount of sand, is responsible for subtype development. Not surprisingly, Fryberger found that linear dune chains were associated with wind regimes of a greater directional variability than barchanoid dunes, ranging from wide unimodal wind regimes (with a single peak direction but less than 90% of the sand drift within a forty-five degree arc of the compass) to complex wind regimes (three or more directions of significant winds, or no clearly defined nodes). Like barchanoids, linear dunes form in both high-energy and low-energy wind regimes. The stronger the winds, the less variable the wind direction needs to be and the fewer junctures (merging of two dunes) between adjacent linear dunes. Fryberger found that fields of star dunes formed invariably in complex wind regimes---areas with winds from several directions over the course of the year. Star dunes also form in both high and low-energy wind regimes.

In recent years, desert geomorphologists have recognized a higher order of sand-accumulation form -- large, widely-spaced mega-dunes, generally known as "draas." Bagnold recognized such features in North Africa, calling them "whalebacks," and recent aerial and satellite studies of ergs have revealed that draas are not restricted to the Sahara. hile little study of draa forms has occurred, draas appear to be structurally analogous to dunes, but on a larger-scale and are considered to be slower moving

“Draas usually form a base on which fields of smaller dunes have formed.”

and older (Cooke and Warren, 1973).

Draas usually form a base on which fields of smaller dunes have formed, producing a compound pattern. The whalebacks described by Bagnold (1941) averaged fifty meters in height, were as wide as three kilometers, and some ran uninterrupted for 300 kilometers. Bagnold considered these mega-dunes to be the accumulated coarse-grained bases (“plinths”) of former linear dunes. The whalebacks were topped with smaller linear dunes running at an oblique angle to the direction of the mega-dunes, suggesting the possibility of a major change in wind regime.

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