Chapter 18

Dune Morphology and Dynamics

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Introduction

Sand dunes form part of a hierarchical self-organized system of aeolian bedforms which comprises: (i) wind ripples (spacing 0.1–1 m), (ii) individual simple dunes or superimposed dunes on mega dunes (also called draa or compound and complex dunes) (spacing 50–500 m), and (iii) mega dunes (spacing >500 m). Most dunes occur in contiguous areas of aeolian deposits called ergs or sand seas (with an area of $>100\,\text{km}^2$). Smaller areas of dunes are called dune fields. Major sand seas occur in the old world deserts of the Sahara, Arabia, central Asia, Australia, and southern Africa, where sand seas cover between 20 and 45% of the area classified as arid (Fig. 18.1). In North and South America there are no large sand seas, and dunes cover less than 1% of the arid zone. The majority of dunes are composed of quartz and feldspar grains of sand size, although dunes composed of gypsum, carbonate, and volcaniclastic sand, as well as clay pellets, also occur.

The formation of areas of dunes is determined by the production of sediment of a range of suitable particle sizes, the availability of this sediment for transport by wind, and the transport capacity of the wind (Kocurek and Lancaster, 1999). Most dunes are derived from material that has been transported by fluvial or littoral processes. Important sources include marine and lacustrine beaches, dry lake basins, river flood plains, and deltas. The availability of sediment (defined as the probability of entrainment of sand for transport) is determined by its moisture content, vegetation cover, crusting, and cohesion. The transport capacity of the wind (or the potential sand transport rate) is a cubic function of wind speed or surface shear stress above the transport threshold (see Chapter 17).

Dunes are created and modified by a series of interactions between granular material (sand) and shearing flow (the atmospheric boundary layer) as shown by Fig. 18.2. The resulting landforms are bedforms that are dynamically similar to those developed in subaqueous shearing flows (e.g. rivers, tidal currents). Their forms reflect the characteristics of the sediment (primarily its grain size) and the surface wind regime (especially its directional variability). In some areas vegetation may be a significant factor. As the bedform grows upwards into the boundary layer, the primary air flow is modified by interactions between the form and the flow which give rise to modifications of the local wind speed, shear stress, and turbulence intensity and create secondary flow circulations, especially in the lee of the dune. Many large dunes (megadunes or draa) are also characterized by superimposed bedforms that respond to changes in airflow and sediment transport on the megadune itself. Not all dunes are the products of contemporary processes and dynamics. In many areas, megadunes have a long and complex history in which the legacy of past climates and wind regimes is a significant factor in determining present-day dune morphology.

Dunes occur in self-organized patterns that develop over time as the response of sand surfaces to the wind regime (especially its directional variability) and the supply of sand (Werner, 1995). Development of these patterns is modulated by the effects of changes in climate and sea level on sediment supply, dune mobility,

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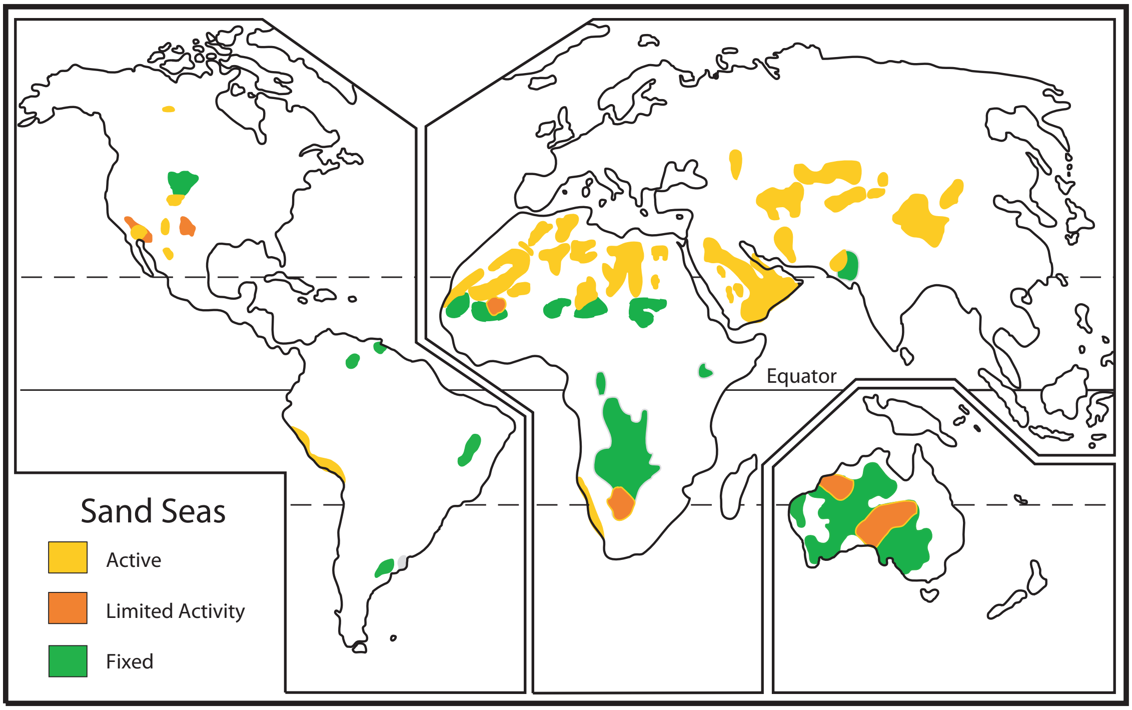

Fig. 18.1 Location of major desert sand seas and dunefields

and wind regime characteristics, often resulting in the formation of different generations of dunes. Characteristic features of dune patterns include close correlations between the height and spacing of dunes and systematic spatial variations in dune type, orientation, and sediment volume. The dune types discussed below represent the steady state attractors of the aeolian transport system and can evolve from a wide range of initial conditions. The orientation of dunes with respect to the wind regime is another aspect of the self-organizing nature of the system, in which dunes are oriented to maximize the gross sand transport normal to the crest (Rubin and Hunter, 1987).

The past two decades have seen major advances in our understanding of desert dune processes and dynamics at different temporal and spatial scales. These advances, together with parallel changes in the understanding of bedforms in subaqueous environments, have resulted in important changes in the paradigms that guide dune research. Key new paradigms include the gross bedform-normal concept for dune trends; the concept of dune generations to explain complex dune

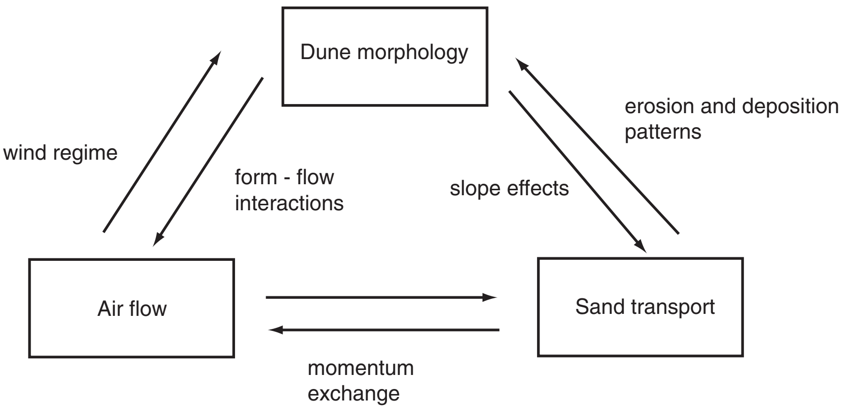

Fig. 18.2 A conceptual framework for process-form interactions on aeolian dunes

patterns; the sediment state model to explain episodic development of sand seas and dunefields; and the overarching principle of self-organization of dune systems and patterns.

Changes in the field of dune studies have been documented by new books on desert dunes and aeolian geomorphology (Lancaster, 1995a; Livingstone and Warren, 1996; Pye and Tsoar, 1990); and reviews by Wiggs (Wiggs, 2001), Walker (Walker and Nickling, 2002), and McKenna Neuman and Nickling (Nickling and McKenna Neuman, 1999) and (Livingstone et al., 2007). Progress in the field has been encouraged by the series of international conferences on aeolian research (the ICAR conferences), and the many special journal issues and books resulting from them (e.g. Goudie et al., 1999).

The following discussion of dune morphology and dynamics will proceed up the hierarchy of aeolian bedforms, with an emphasis on the fundamental processes operating at each level. Each element of the hierarchy responds to the dynamics of a component of the wind regime in an area and possesses a characteristic time period, termed the relaxation or reconstitution time (Allen, 1974), over which it will adjust to changed conditions. This increases from minutes in the case of wind ripples to millennia for draas. Change in bedforms involves the movement of sediment. Thus an increasing spatial scale is also involved at each level of the hierarchy.

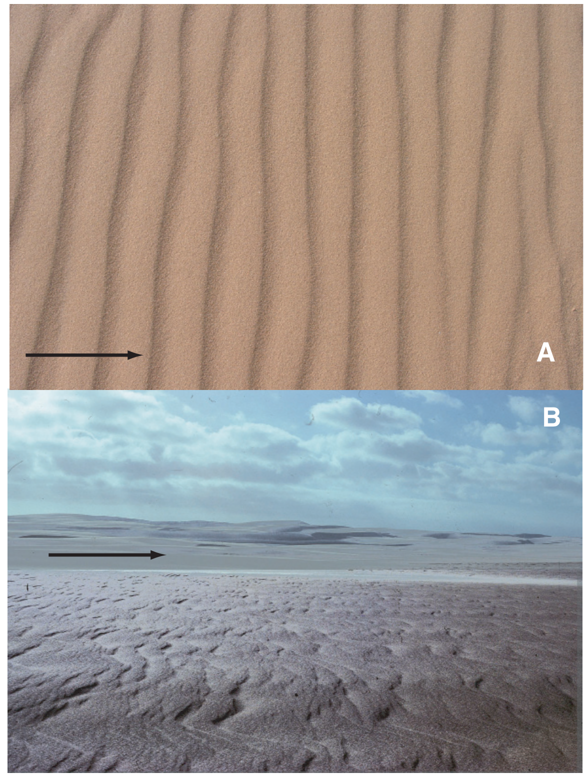

Fig. 18.3 Wind ripples: (**A**) ripples in medium–fine sand in the Gran Desierto sand sea, and (**B**) granule ripples on the Skeleton Coast, Namibia. Arrow indicates formative wind direction

Wind Ripples

Wind ripples (Fig. 18.3) are ubiquitous on all sand surfaces except those undergoing very rapid deposition and provide an excellent example of self-organization in aeolian systems (Anderson, 1990). They are the initial response of sand surfaces to sand transport by the wind, and form because flat sand surfaces over which transport by saltation and reptation takes place are dynamically unstable (Bagnold, 1941). The formation and movement of wind ripples are therefore closely linked to the processes of saltation and reptation.

Wind ripples trend perpendicular to the sand-transporting winds, although (Howard, 1977) has emphasized the effects of slope on ripple orientation. Because they can be reformed within minutes, wind ripples provide an almost instantaneous indication of local sediment transport and wind directions. Typical wind ripples have a wavelength between 13 and 300 mm (hundreds of grain diameters) and an amplitude of 0.6–14 mm (tens of grain diameters) (Anderson, 1990; Boulton, 1997). The ratio between ripple length and height is given by the ripple index, which can be used to compare ripples in different environments. Typical ripple indices for aeolian wind ripples range from 15 to 20. Much longer wavelength (0.5–2 m or more) ripples with an amplitude of 0.1 m or more are composed of coarse sand or granules (1–4 mm median grain size) and are termed 'granule ripples' (Fryberger et al., 1992; Sharp, 1963) and 'megaripples' by Greeley and Iversen (1985). Granule ripples are not distinct forms, and form one end of a continuum of wind ripple dimensions (Ellwood et al., 1975) (Fig. 18.4). Ripple wavelength is a function of both particle size and sorting and wind speed so that ripples in coarse sands have a greater spacing than those in fine sands (Sharp, 1963). For sands of a given size, ripple wavelength increases with wind

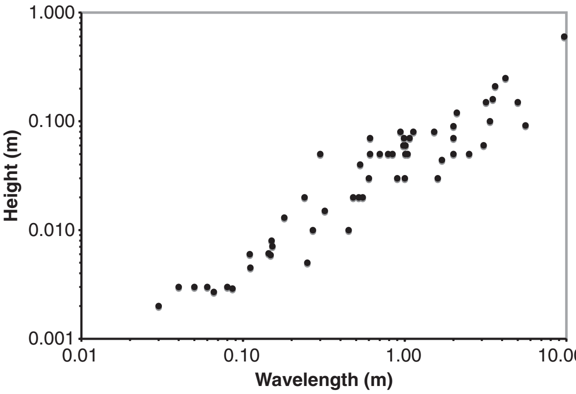

Fig. 18.4 Wind ripple morphometry. Note that there is a continuum of ripple wavelength from 'normal' wind ripples to granule ripples. Data from Sharp (1963), Walker (1981), and author's field observations

friction speed (Seppälä and Linde, 1978). Most wind ripples are asymmetric in cross-section with a slightly convex stoss slope at an angle of 1.4–7° and a lee slope that varies between 1.8 and 10°. The existence of very steep (20–30°) lee slope angles reported by Sharp (1963) is uncertain. In all cases the crest of the ripple is composed of grains that are coarse relative to the mean size of the surface sand.

Several models have been put forward to explain the formation and characteristic wavelength of ripples. Bagnold (1941) pointed to the close correspondence between calculated saltation path lengths and observed ripple wavelengths in wind tunnel experiments. He argued that chance irregularities in 'flat' sand surfaces give rise to variations in the intensity of saltation impacts (Fig. 18.5a) creating zones that would be preferentially eroded or protected. Grains from the zone of more intense impacts (A–B) would land downwind at a distance equal to the average saltation pathlength such that a zone of more intense saltation impacts would propagate downwind. In addition, variations in surface slope and saltation impact intensity cause variations in the reptation rate. Bagnold argued that interactions between the developing surface microtopography and the saltating and reptating grains would soon lead to a coincidence between the characteristic saltation pathlength and the ripple wavelength (Fig. 18.5b). Bagnold's concept of the formation of wind ripples was first challenged by Sharp (1963) who argued that grains in ripples are moved mostly by reptation or surface creep. Irregularities in the bed and interactions between grains moving at different speeds give rise to local increases in bed elevation. These 'proto-ripples' begin as short-wavelength, low-amplitude forms and grow to their steady state dimensions by the growth of larger forms at the expense of smaller. Each developing ripple creates a 'shadow zone' in its lee (Fig. 18.5c), with a width proportional to ripple wavelength and impact angle. The size of the shadow zone determines the position of the next ripple downwind. Sharp argued that the controls of ripple wavelength are impact angle and ripple amplitude, both of which are dependent on grain size and wind speed, but he could see no obvious reason why ripple wavelength should be dependent on the mean saltation pathlength. His observations on ripple development to an equilibrium size and spacing have been confirmed experimentally by wind tunnel experiments (Seppälä and Linde, 1978; Walker, 1981) and by numerical model simulations of sediment surfaces (Anderson, 1990; Werner, 1988).

Anderson has provided a rigorous model for ripple development based on experimental data and numerical simulations of sand beds (Anderson, 1987; Anderson, 1990; Anderson and Bunas, 1993). Recent experimental and theoretical work on aeolian saltation (see Chapter 17) has demonstrated that saltating sand consists of two populations: (a) long trajectory, high impact energy 'successive saltations' and (b) short trajectory, low impact energy 'reptations'. There is a wide distribution of saltation trajectories with typical pathlengths that are much longer than ripple wavelengths, and a low range (1–2°) of impact angles. This suggests that the high impact energy grains do not contribute directly to ripple formation, as Bagnold hypothesized, but drive the reptation process. Using a simplified model of aeolian saltation, Anderson (1987, 1990) was able to show that a flat bed is unstable to infinitesimal variations in bed elevation, giving rise to spatial variations in the mass flux of reptating grains. Convergence and divergence of mass flux rates result in the growth of grain-scale perturbations on the bed, with the fastest growing perturbations having a wavelength 6–10 times the mean reptation distance (Fig. 18.5d). These perturbations subsequently develop into a self-organized pattern of ripples by coalescence and convergence of bedforms moving at different speeds, with the rate of change decreasing asymptotically with time. A quasi-stable wavelength emerges that is the effect of the sharply decreasing rate of ripple mergers with increasing ripple

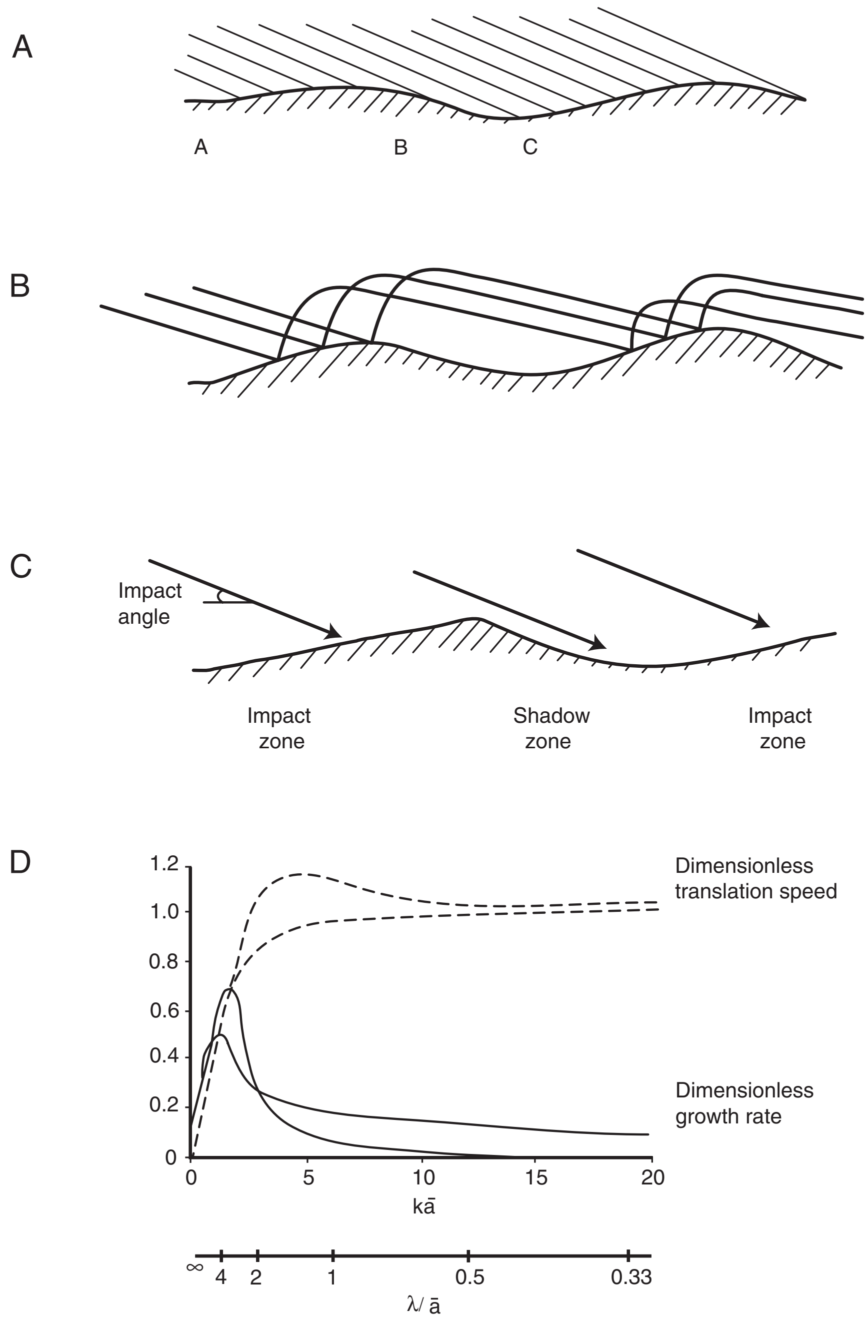

Fig. 18.5 Models for wind ripple formation. (**A, B**) Variation in impact intensity over a perturbation in the bed (after Bagnold, 1941). Note higher impact intensity in zone A–B compared with B–C. (**C**) Alternation of impact and shadow zones on a developing wind ripple (after Sharp, 1963). (**D**) Growth and movement of developing bed perturbations that evolve to wind ripples (after Anderson, 1987)

size (Werner, 1988). Numerical models of ripples suggest that their wavelength is partially dependent on the length of the shadow zone, and therefore the angle of incidence of the saltating grains, thus validating the model of Sharp (1963). Ripple height may depend in part on impact angle, but is also influenced by the higher wind shear stresses experienced on the ripple crest.

Dune Morphology and Morphometry

Dunes occur in a variety of morphologic types, each of which displays a range of sizes (height, width, and spacing). Most dune patterns are quite regular, as evidenced by distinct statistical populations of dunes defined by dune crest orientation, crest length and spacing

(Ewing et al., 2006) as well as the correlations between dune height and spacing (Fig. 18.6).

Many different classifications of dune types have been proposed (see Mainguet, 1983; 1984a, for a list of different schemes). They fall into two groups: (a) those based on the external morphology of the dunes (morphological classifications), and (b) those that imply some relationship of dune type to formative winds or sediment supply (morphodynamic classifications). Figure 18.7 provides a framework for classifying dunes based on morphology, sediment thickness, and other key parameters.

Crescentic Dunes

The simplest dune types and patterns are those that form in wind regimes characterized by a narrow range of wind directions in which dune crests are oriented tranverse to the sand transport direction. In the absence of vegetation, crescentic dunes are the dominant form. (Hunter et al., 1983) and (Tsoar, 1986) indicate that this dune type occurs where the directional variability of sand transporting winds is 15° or less about a mean value. Isolated crescentic dunes or barchans occur in areas of limited sand availability. As sand supply increases, barchans coalesce laterally to form crescentic or barchanoid ridges that consist of a series of connected crescents in plan view (McKee, 1966). Larger forms with superimposed dunes are termed compound crescentic dunes (e.g. Breed and Grow, 1979; Havholm and Kocurek, 1988; Lancaster, 1989b).

Barchans (Fig. 18.8a) are common in areas of limited sand supply and unidirectional winds (Andreotti et al., 2002), where they can migrate long distances with only minor changes of form (Gay, 1999; Hastenrath, 1987; Haynes, 1989; Long and Sharp, 1964). Recent field studies and modelling suggest that the morphological stability of these dunes may be overemphasized, and collisions leading to merging of faster-moving smaller dunes with larger dunes, as well as creation of new dunes are common (Elbelrhiti et al., 2005).

Barchans occur in two main areas: (a) on the margins of sand seas and dune fields (Sweet et al., 1988) and (b) in sand transport corridors linking sand source zones with depositional areas (Corbett, 1993; Embabi, 1982; Embabi and Ashour, 1993; Hersen

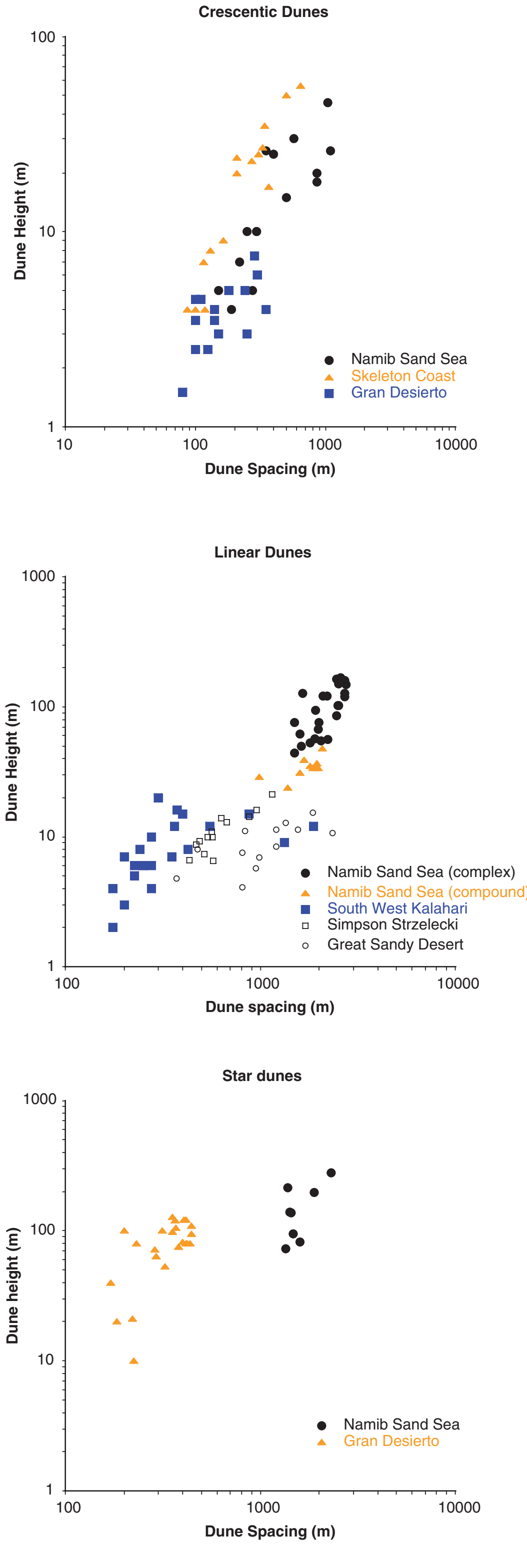

Fig. 18.6 Relations between dune height and spacing. Data from author's measurements and Wasson and Hyde (1983b)

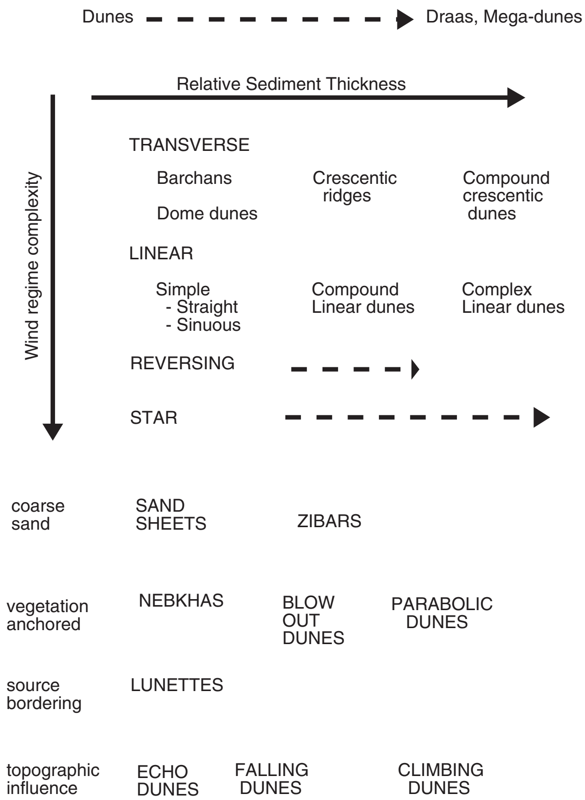

Fig. 18.7 A scheme for the classification of desert dunes. Modified from Lancaster (1995a)

et al., 2004). Barchans are characterized by an ellipsoidal shape in plan view, with a concave slip face and 'horns' extending downwind (Sauerman et al., 2000). Dune height is typically about one-tenth of the dune width (Finkel, 1959; Hastenrath, 1967). Strongly elongated horns and asymmetric development of barchan plan shapes occur in some areas (e.g. Hastenrath, 1967; Lancaster, 1982a), and have been attributed to asymmetry in the wind regime or sand supply, but may be a product of merging and reconstitution of dunes as the pattern becomes self-organized (Hersen and Douady, 2005). In some areas, barchans of this type are transitional to linear dunes (Lancaster, 1980; Tsoar, 1984).

Crescentic dunes of simple and compound varieties occupy some 10% of the area of sand seas world-wide (Fryberger and Goudie, 1981) and occur in all desert regions. The patterns of most crescentic dunes are quite regular, as indicated by the close correlations that exist between dune height and spacing (Fig. 18.6). Most simple crescentic and barchanoid ridges (Fig. 18.8b) are between 3 and 10 m high, with a spacing of 100–400 m. Typical areas of simple crescentic ridges occur at White Sands, New Mexico (McKee, 1966), Guererro Negro, Baja California (Inman et al., 1966), and in the Skeleton Coast dunefield in the northern Namib (Lancaster, 1982a).

Compound crescentic dunes (Fig. 18.8c) occupy some 20% of the area of sand seas world-wide. They usually consist of a main dune ridge 20–50 m high with a spacing of 800–1200 m. Smaller crescentic ridges 2–5 m high and 50–100 m apart are superimposed on their upper stoss slopes and crestal areas. Compound crescentic dunes are typified by those along the western edge of the Namib Sand Sea (Lancaster, 1989b), the Algodones Dunes of California (Havholm and Kocurek, 1988; Sweet et al., 1988), and the Liwa area of the United Arab Emirates (Stokes and Bray, 2005).

Linear Dunes

Linear dunes are characterized by their length (often more than 20 km) straightness, parallelism, and regular spacing, and high ratio of dune to interdune areas. Lancaster (1982b) estimated that 50% of all dunes are of linear form, with the percentage varying between 85 and 90% for areas of the Kalahari and Simpson–Strzelecki Deserts to 1–2% for the Ala Shan and Gran Desierto sand seas. Linear dunes are the dominant form in sand seas in the Southern Hemisphere and in the southern and western Sahara.

Many linear dunes consist of a lower gently sloping plinth, often partly vegetated, and an upper crestal area where sand movement is more active. Slip faces develop on the upper part of the dune, their orientation depending on the winds of the season. The average form of the dune may be symmetrical with an approximately triangular profile, but in each wind season its profile tends to an asymmetric form with a convex upper stoss slope and a well-developed lee face (Tsoar, 1985). Several varieties of linear dunes are recognized (Fig. 18.9). Simple linear dunes (Fig. 18.9a,b) are of two types: the long, narrow, straight, partly vegetated linear dunes of the Simpson (Wasson, 1983; Wasson et al., 1988) and Kalahari Deserts (Bullard and Nash, 1998; Bullard et al., 1995) (equivalent to the vegetated linear dunes of Tsoar [1989]), and the more

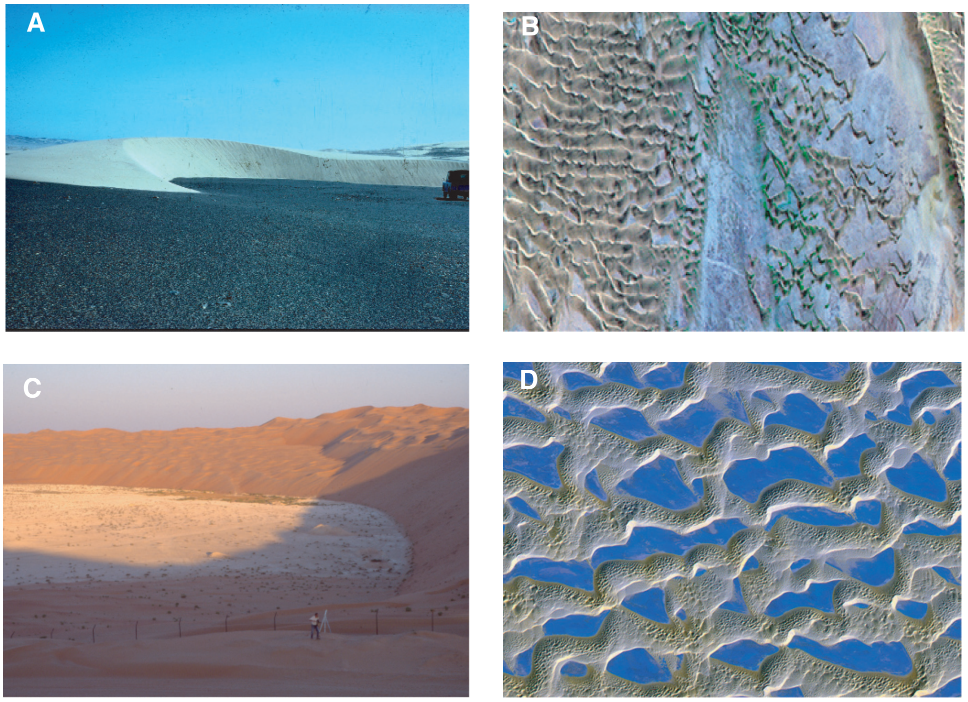

Fig. 18.8 Crescentic dunes: (**A**) barchans on the upwind margin of the Skeleton Coast dunefield, Namibia; (**B**) barchans and simple crescentic dunes, Conception Bay, Namib Sand Sea; and (**C, D**) compound crescentic dunes, Liwa area, UAE: note two orders of dune spacing, primary dunes and superimposed dunes

sinuous seif-type dunes found in Sinai [Tsoar, 1983], and in parts of the eastern Sahara. Compound linear dunes consist of two to four seif-like ridges on a broad plinth and are typified by those in the southern Namib sand sea (Lancaster, 1983). Large (50–150 m high, 1–2 km spacing) complex linear dunes (Fig. 18.9c) with a single sinuous main crestline with distinct star-form peaks and crescentic dunes on their flanks (Lancaster, 1983) occur in the Namib and parts of the Rub al Khali sand seas. Wide (1–2 km) complex linear dunes with crescentic dunes superimposed on their crests occur in the eastern Namib, parts of the Wahiba Sands (Warren, 1988; Warren and Allison, 1998), the Takla Makan (Wang et al., 2004) and the Akchar sand sea of Mauritania (Kocurek et al., 1991).

The origins of linear dunes and their relationship to formative wind directions have been the subject of considerable controversy (Lancaster, 1982b; Tsoar, 1989). A widely held view has been that linear dunes form parallel to prevailing or dominant wind directions (Folk, 1970; Glennie, 1970; Wilson, 1972). Their parallelism and straightness are believed to result from roller vortices in which helicoidal flow sweeps sand from interdune areas to dunes (Hanna, 1969; Wilson, 1972). However, there are inconsistencies between the spacing of many dunes and the reported dimensions and stability of helical roll vortices (Lancaster, 1982b; Livingstone, 1986), which would have to be positioned at exactly the same place at successive wind episodes in order to allow dunes to grow and extend (Greeley and Iversen, 1985). The only observational evidence for helical roll vortices in dune areas comes from studies of tethered kites in the Simpson Desert, Australia (Tseo, 1990). There is, however, a substantial body of empirical evidence that indicates that linear dunes form in bidirectional wind regimes. Correlations between dune types and wind regimes (e.g. Fryberger, 1979), studies of internal sedimentary structures (Bristow et al., 2000; McKee, 1982; McKee and Tibbitts, 1964) and detailed process studies on linear dunes (Livingstone, 1986, 1988, 1993; Tsoar, 1983) support such a view.

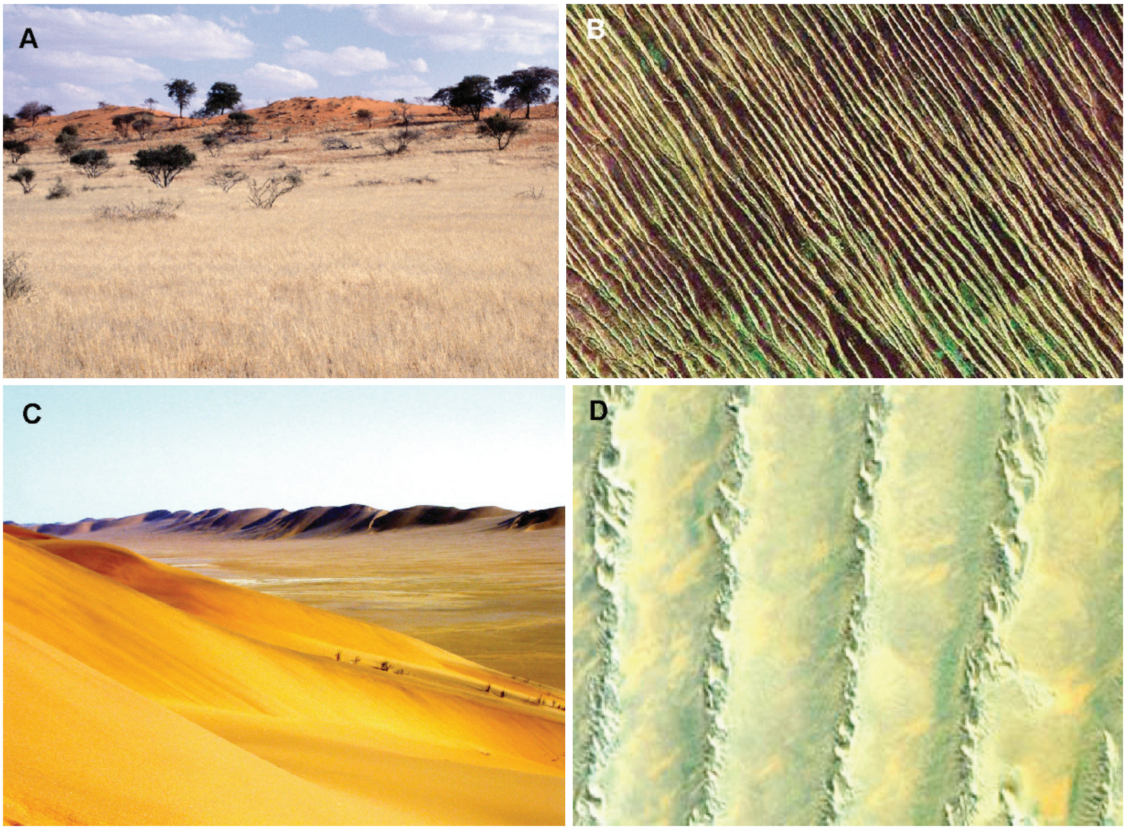

Fig. 18.9 Linear dunes: (**A, B**) vegetated simple linear dunes, south-western Kalahari; and (**C, D**) complex linear dunes, Namib sand sea

Star Dunes

Star dunes (Fig. 18.10) are the largest dunes in many sand seas and may reach heights of more than 300 m. They contain a greater volume of sand than any other dune type (Wasson and Hyde, 1983b) and appear to occur in areas that represent depositional centres in many sand seas (Dong et al., 2004; Lancaster, 1983; Mainguet and Callot, 1978). Star and related reversing dunes comprise 9–12% of all dunes in the Namib, Gran Desierto, and central Asian sand seas. They are absent from the Australian deserts, from the Kalahari, and from India, but comprise 40% of dunes in the Grand Erg Oriental (Mainguet and Chemin, 1984).

Star dunes are characterized by a pyramidal shape, with three or four arms radiating from a central peak and multiple avalanche faces. Each arm has a sharp sinuous crest, with avalanche faces that alternate in aspect as wind directions change. The arms may not all be equally developed and many star dunes have dominant or primary arms on a preferred orientation. The upper parts of many star dunes are very steep with slopes at angles of 15–30°; the lower parts consist of a broad, gently sloping (5–10°) plinth or apron. Small crescentic or reversing dunes may be superimposed on the lower flank and upper plinth areas of star dunes.

Star dunes have been hypothesized to form at the centres of convection cells, at the nodes of stationary waves in oscillating flows, above rock hills, or at the nodal points of complex dune alignment patterns created by crossing or converging sand transport paths (Lancaster, 1989a). Comparisons between the distribution of star dunes and wind regimes suggest that they form in multidirectional or complex wind regimes (Fryberger, 1979). A strong association between the occurrence of star dunes and topographic barriers has also been noted (Breed and Grow, 1979). Topography may modify regional wind regimes to increase their directional variability, as in the Erg Fachi Bilma and Namib (Mainguet and Callot, 1978; McKee, 1982;

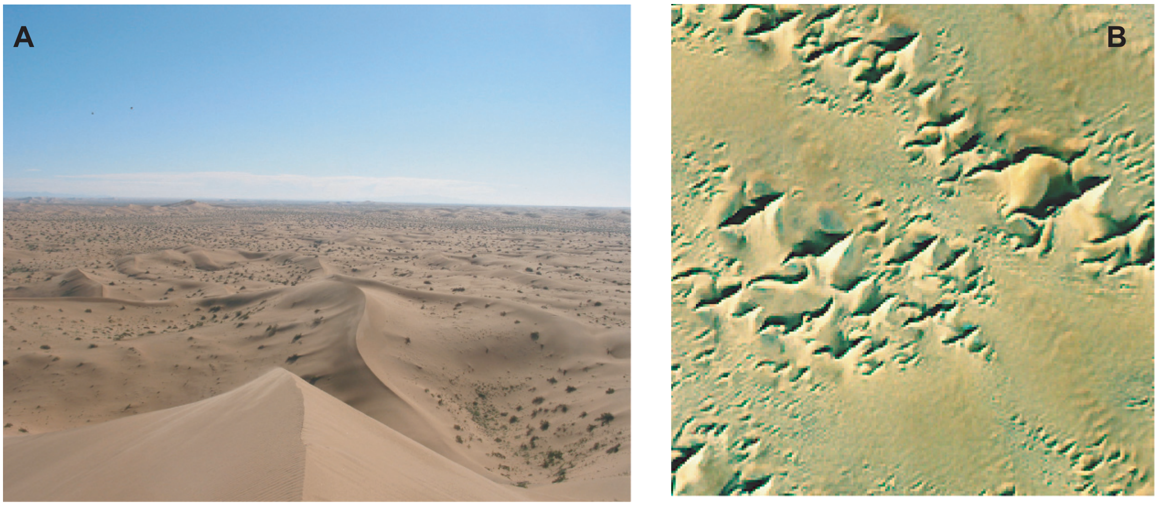

Fig. 18.10 Star dunes, Gran Desierto sand sea: (**A**) ground view; (**B**) Landsat image. Note preferred east-west orientation of major dune arms and small reversing dunes in middle distance

Lancaster, 1983), or create traps for sand transport, as at Kelso Dunes (Sharp, 1966), Dumont Dunes (Nielson and Kocurek, 1987) and Great Sand Dunes (Andrews, 1981).

Parabolic Dunes

Parabolic dunes (Fig. 18.11), common in many coastal dunefields and semi-arid to sub-humid areas, have a restricted distribution in arid region sand seas. The only major sand sea with significant areas of this dune type is in the Thar Desert of India (Kar, 1993; Singhvi and Kar, 2004; Verstappen, 1968). Small areas of parabolic dunes occur in the south-western Kalahari (Eriksson et al., 1989), Saudi Arabia (Anton and Vincent, 1986), north-east Arizona (Hack, 1941), and at White Sands (McKee, 1966).

Parabolic dunes are characterized by a U shape with trailing partly vegetated parallel arms 1–2 km long and an unvegetated active 'nose' or dune front 10–70 m high that advances by avalanching. In the Thar Desert, compound parabolic dunes with shared arms occur in some areas and result from merging or shingling of several generations of dunes with different migration rates (Wasson et al., 1983). The conditions under which parabolic dunes form are not well known. They seem to be associated with the presence of a moderately developed vegetation cover, and with unidirectional wind

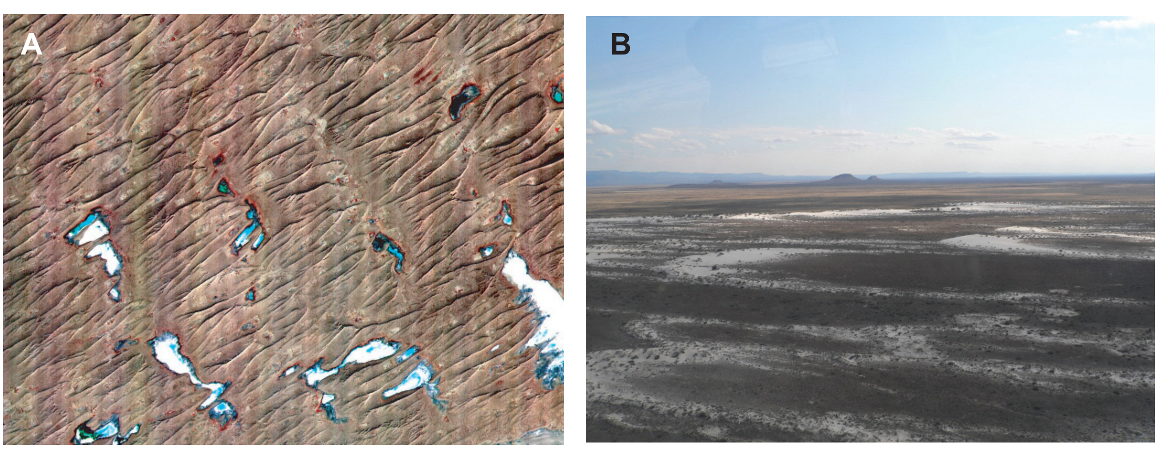

Fig. 18.11 Parabolic dunes: (**A**) Thar Desert, India (ASTER image); (**B**) White Sands, New Mexico (photograph by David Bustos, NPS)

regimes. Downwind, some parabolic dunes are transitional to crescentic dunes as vegetation cover decreases (Anton and Vincent, 1986).

Zibars and Sand Sheets

Not all aeolian sand accumulations are characterized by dunes. Low relief sand surfaces such as sand sheets are common in many sand seas and occupy from as little 5% of the area of the Namib sand sea to as much as 70% of the area of Gran Desierto (Lancaster et al., 1987). Extensive sand sheets also occur in the eastern Sahara (Breed et al., 1987; Maxwell and Haynes, 2001). Fryberger and Goudie (1981) estimate that 38% of aeolian deposits are of this type. Many sand sheets and interdune areas between linear and star dunes are organized into low rolling dunes, without slipfaces, known as zibars (Holm, 1960; Nielson and Kocurek, 1986; Warren, 1972) with a spacing of 50–400 m and a maximum relief of 10 m. Typically zibars are composed of coarse sand, and occur on the upwind margins of sand seas.

Sand sheets develop in conditions unfavourable to dune formation (Kocurek and Nielson, 1986). These may include a high water table, periodic flooding, surface cementation, coarse-grained sands, and presence of a vegetation cover all of which act to limit sand supply for dune building. Coarse grains armour the surface of sand sheets in the eastern Sahara (Breed et al., 1987). Sand sheets in the north-western Gran Desierto and the eastern Sahara are also composite features resulting from multiple generations of aeolian deposition separated by episodes of soil formation (Lancaster, 1993; Stokes et al., 1998). Those in the Gran Desierto have developed in conditions of a sparse vegetation cover which is insufficient to prevent sand transport taking place, but sufficient to cause divergence and convergence of airflow around individual plants in the manner suggested by Ash and Wasson (1983) and Fryberger et al. (1979) giving rise to localized deposition by wind ripples and shadow dunes.

Dune Processes and Dynamics

The initiation, development, and equilibrium morphology of all aeolian dunes are determined by a complex series of interactions between dune morphology, airflow, vegetation cover, and sediment transport rates. In turn, the developing bedforms exert a strong control on local transport rates through form-flow interactions and secondary flow circulations, leading to a dynamic equilibrium between dune morphology and local airflow. In multidirectional wind regimes, the nature of interactions between dune form and airflow change as winds vary direction seasonally, and lee side secondary flows become important. A conceptual framework for dune dynamics is provided by Fig. 18.2.

Dune Initiation

Dune initiation is poorly understood and little studied but is important in understanding how dunes and dune patterns develop. It involves localized deposition leading to bedform nucleation, which will then fix a pattern that can propagate downwind (Wilson, 1972). Reductions in the local sediment transport rate can occur through convergence of streamlines, by changes in surface roughness (e.g. vegetation cover, surface particle size), or by variations in microtopography (slope changes, relict bedforms).

There are few studies of the initiation and early development of dunes. Cooper, (1958), Jäkel (1980) and Kocurek et al. (1992) have described the development of barchans and transverse ridges from thin sand patches with no flow separation in their lee to small dunes with lee side flow separation, but only Kocurek et al. (1992) and Lancaster (1996) have documented the initiation of sand patches where changes in aerodynamic roughness or microtopography cause a reduction in near-surface wind speeds and transport rates.

Kocurek et al. (1992) recognized five stages of dune initiation and development (Fig. 18.12) with a progressive evolution of the lee face and bedform-induced secondary flow expansion and separation: (a) irregular patches of dry sand a few centimetres high, (b) 0.1–0.35-m-high protodunes with wind ripples on all surfaces, (c) 0.25–0.40-cm protodunes with grainfall on lee slopes, (d) 1–1.5-m-high barchans with grainflow, and (e) 1–2-m-high crescentic ridges. The developing dunes were characterized by a reverse asymmetry (steeper stoss slope) in stage (b) similar to that noted by Cooper (1958). The change from flow expansion to flow separation came as

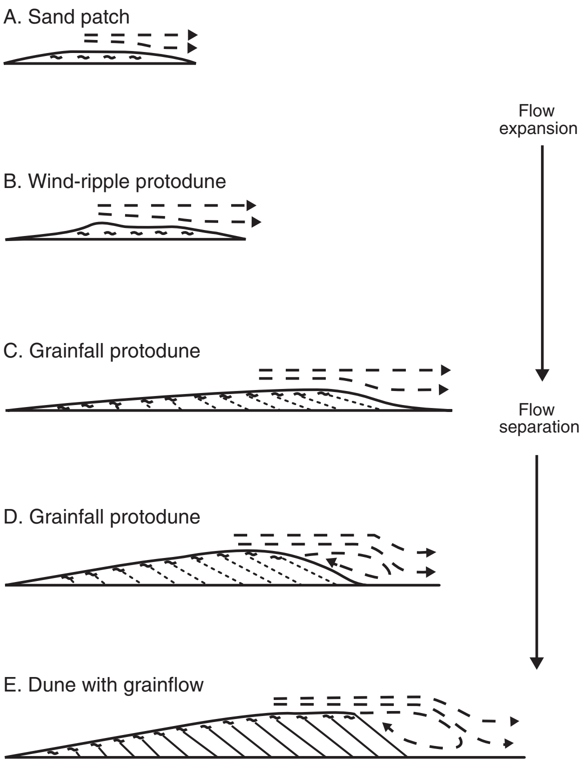

Fig. 18.12 Stages of dune development at Padre Island, Texas (after Kocurek et al., 1992)

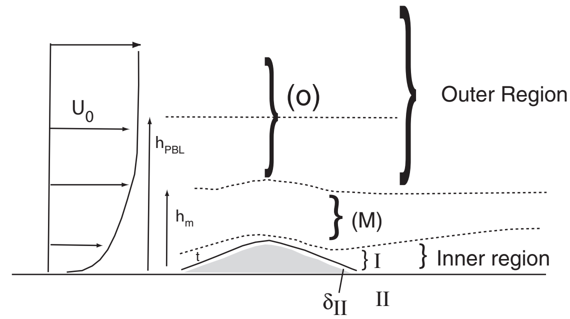

Fig. 18.13 Airflow over an isolated two-dimensional hill, showing components of the boudary layer development (after Nickling and McKenna Neuman, 1999). M – middle region

lee slopes exceeded 22°. Subsequent evolution and self-organization of the dunefield mainly takes place by repeated mergings, splittings, lateral linking, and cannibalization of dunes. Dune growth is at the expense of interdune areas and the pattern evolves so that the height: spacing ratio tends toward 1:20, a figure that is common in many areas of crescentic dunes.

Airflow Over Dunes

As dunes grow, they project into the atmospheric boundary layer so that they affect the airflow around and over them in a manner similar to isolated hills. Boundary layer theory and numerical modelling of flow over low hills provide a conceptual framework for understanding air flow over dunes, (e.g. Hunt et al., 1988; Jackson and Hunt, 1975; Jensen and Zeman, 1985; Mason and Sykes, 1979). Based on this framework, air flow over dunes can be divided into an outer inviscid region and an inner region which follows the topography of the dune (Fig. 18.13). The inner layer is further divided into two sub-layers: (1) a very thin inner surface layer, in which the shear stress is constant, with a thickness equivalent to the surface roughness (1/30 of the grain diameter for a sand surface); and (2) a shear stress layer in which shear stress effects decrease with height above the surface. It is the shear stress generated in this part of the boundary layer that is responsible for sediment transport on dunes, yet its measurement in the field presents many problems (Wiggs, 2001). Estimates of the inner layer thickness range between 0.54 m (Lancaster et al., 1996) and 0.8 m (Wiggs, 2001). The basic features of this conceptual model have been confirmed by field studies of air flow over dunes and reproduced in numerical models (Parsons et al., 2004; Weng et al., 1991).

The Stoss or Windward Slope

Winds approaching the upwind toe of a dune stagnate slightly and are reduced in velocity (Wiggs, 1993) On the stoss, or windward slope of the dune, streamlines are compressed and winds accelerate up the slope. Data from field studies (Burkinshaw et al., 1993; Lancaster, 1985; Mulligan, 1988; Tsoar, 1985) have shown that the magnitude of the velocity increase is represented by the speed-up ratio Δs or amplification factor Az represented by U_2/U_1 where U_2 is the velocity at the dune crest and U_1 is the velocity at the same height above the surface at the upwind base of the dune. Typical values for Az range between 1.1 and 2.5 and vary with dune height and aspect ratio (Fig. 18.14), in good agreement with the Jackson and Hunt model. Field studies show that the rate of

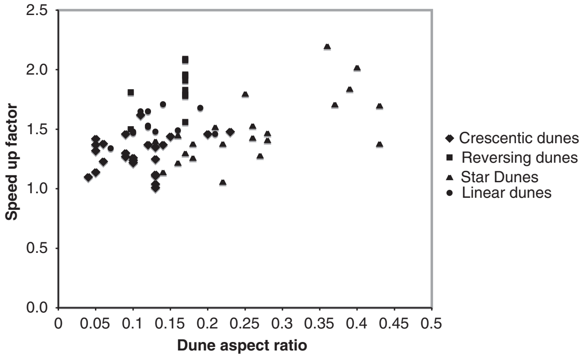

Fig. 18.14 Relations between velocity increase (amplification factor) and dune aspect ratio *h*/*L*. Data from author

velocity increase is not linear, and follows the surface of the dune closely (Fig. 18.15). Significant problems and uncertainties exist with the field measurement of wind shear velocity and shear stress on dunes in conditions of non-uniform flow and wind profiles that are not log-linear as a result of flow acceleration and development of internal boundary layers (Frank and Kocurek, 1996a). Wind-tunnel simulations and models indicate however that shear stress increases on dune slopes in a manner similar to that illustrated in Fig. 18.15. In addition to the effects of velocity

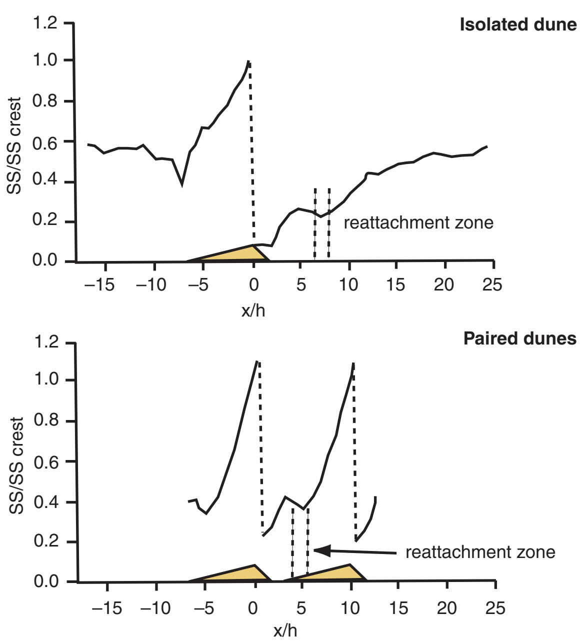

Fig. 18.15 Patterns of normalized shear stress (SS/SS crest) measured over wind tunnel models of isolated and paired transverse dunes (after Walker and Nickling, 2002)

amplification, curvature of wind streamlines may play an important role in the pattern of wind shear stress on dunes (Fig. 18.16). Concave upward curvature of streamlines at the toe of the dune enhances shear stress, whereas convex streamline curvature on the mid stoss slope and between the crest and the brink acts to decrease shear stress (Wiggs et al., 1996).

Lee Side Flow

In the lee of the crest of dunes, wind velocities and transport rates decrease rapidly as a result of flow expansion between the crest and brink of the lee or avalanche face and flow separation on the avalanche face itself. There is a complex pattern of flow separation, diversion, and re-attachment on the lee slopes of dunes, which is determined by the angle between the wind and the dune crest (angle of attack) and the dune aspect ratio (Walker and Nickling, 2002). Secondary flows, including lee-side flow diversion, are especially important where winds approach the dune obliquely, and are an important component of air flow on linear and many star dunes (Fig. 18.17). (Sweet and Kocurek, 1990) suggested that there are three types of flow in the lee of dunes: (a) separated, (b) attached, and (c) attached deflected. The nature of lee-side flow is controlled by the dune shape (aspect ratio), the incidence angle between the primary wind and the crestline, and the stability of the atmosphere (Fig. 18.18). High angles of attack on high aspect ratio (steep) dunes result in flow separation in the lee, which results in the development of an eddy in the lee of the dune. This may have the form of a roller vortex if flow is truly transverse. A conceptual model for airflow in the lee of flow-transverse dunes is provided by Walker and Nickling (Fig. 18.19) based on field and wind-tunnel studies (Frank and Kocurek, 1996b; Walker, 1999; Walker and Nickling, 2002). The model identifies a series of distinct regions of flow that vary in wind speed, shear, and turbulence intensity: a separation cell extending for 4–10 dune heights downwind (A), two wake regions (B, C) that merge at 8–10 h, and an internal boundary layer (D) that grows downwind of the point at which the separated flow reattaches to the surface. With isolated dunes, the internal boundary layer is identifiable at 8–10 h, and comes into equilibrium at around 25–30 h.

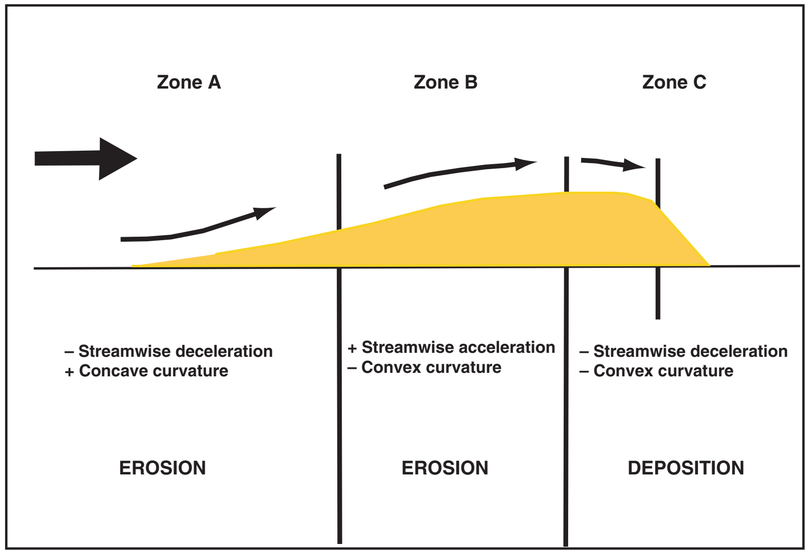

Fig. 18.16 Conceptual model of streamline convergence and divergence over a transverse dune (after Wiggs et al., 1996)

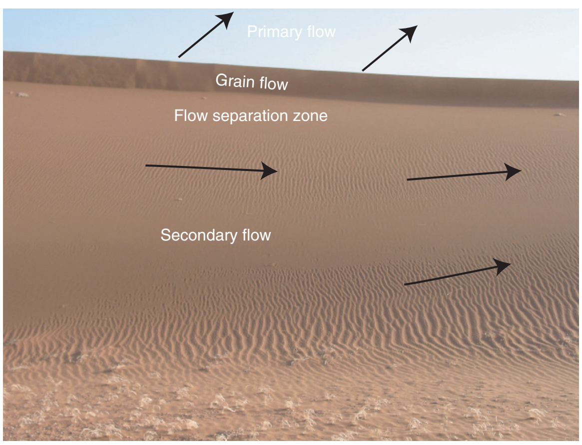

Fig. 18.17 Flow separation and diversion in the lee of a dune

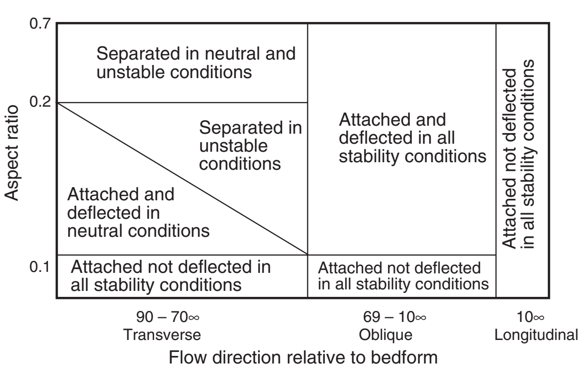

Fig. 18.18 Relations between lee-side flow speed, incidence angle, and aspect ratio for crescentic dunes (after Sweet and Kocurek, 1990)

When flow is oblique to the dune crest a helical vortex develops. The oblique flow is deflected along the lee slope parallel to the dune crest, with the degree of deflection being inversely proportional to the incidence angle between the crestline and the primary wind. Relatively high lee-side wind velocities are associated with low aspect ratio dunes and oblique primary winds. The degree of wind deflection is, in all cases, a cosine function of the incidence angle between the crestline and the primary wind.

When the angle between the dune and the wind is less than 40° the velocity of the deflected wind is greater than that at the crest and sand is transported along the lee side of the dune (Tsoar, 1983; Nielson and Kocurek, 1987; Lancaster, 1989a). Such a process is especially important on linear dunes, where it leads to dune extension, and on star dunes where it extends the arms of the dune. When winds are at more than 40° to the crestline the velocity of the deflected wind is reduced, giving rise to lee-side deposition. Changes in the local incidence angle between primary winds and a sinuous dune crest result in a spatially varying pattern of deposition and along-dune transport on the lee face. Deposition dominates where winds cross the crest line at angles approaching 90°, and erosion or along-dune transport occurs where incidence angles are <40° (Fig. 18.20).

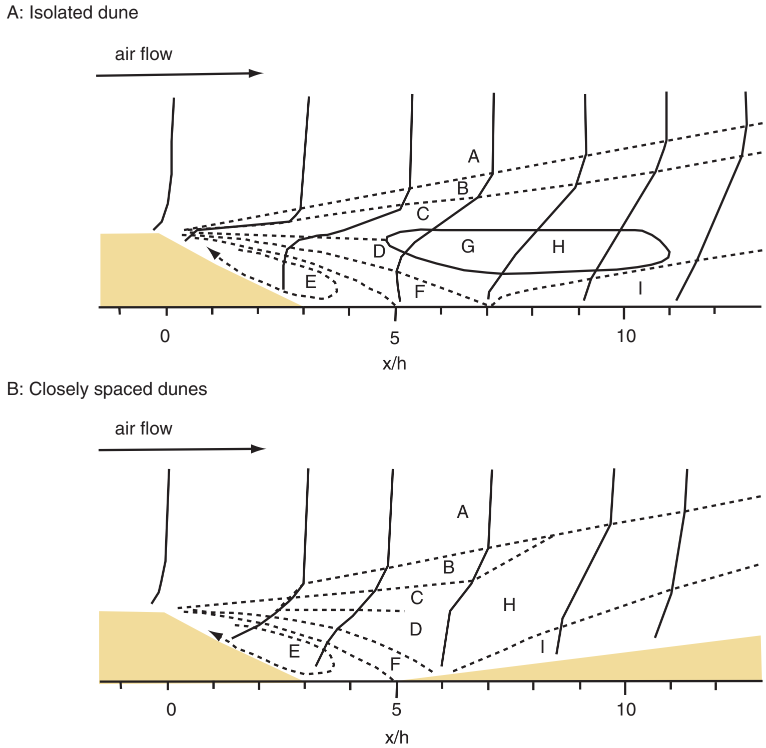

Fig. 18.19 Conceptual model for flow in the lee of a transverse dune (after Walker and Nickling, 2002). Labelled regions represent: A – outer flow; B – overflow; C – upper wake; D – lower wake; E – separation cell; F – turbulent shear layer; G – turbulent stress maximum; H – turbulent shear zone; I – internal boundary layer

Erosion and Deposition Patterns on Dunes

Flow acceleration, coupled with effects of stream line curvature, on the windward slopes of all dunes give rise to an exponential increase in sediment transport rates (Fig. 18.21) towards the dune crest (Lancaster et al., 1996; McKenna Neuman et al., 1997), resulting in erosion of the stoss slope. There are indications that the pattern of erosion and deposition varies with overall wind speed, so that areas near dune crests may experience erosion at lower wind speeds and deposition at wind speeds significantly above threshold (McKenna Neuman et al., 2000). In addition, numerical models suggest that the non-linear increase in sediment transport with height on a dune limits dune size and results in an equilibrium dune configuration (Momiji et al., 2000).

Downwind, the wind has to transport an increasing amount of sand eroded from the dune slope. This in turn requires that wind velocities and surface shear stress should change to increase transport rates proportionately. If the amount of sand in transport exceeds the capacity of the wind to transport it, deposition will occur, leading to adjustment of dune form (increasing dune steepness) and, hence, of local wind speed and transport rates. There is thus a high degree of interaction between the shape of the dune, the amount of change in wind velocities and sand transport rates and the rate and pattern of erosion or deposition.

Measurements of erosion and deposition on linear, reversing, and star dunes show that they consist of a crestal area where erosion and deposition rates are high and a plinth zone in which there is little surface change but considerable throughput of sand (Sharp, 1966; Lancaster, 1989b; Livingstone, 1989, 2003; Wiggs et al., 1995) (Fig. 18.22). On Namib linear dunes and Gran Desierto star dunes, over half the total amount of erosion and deposition takes place in the crestal zone (Lancaster, 1989a,b; Livingstone, 1989), as the position of the crestline varies seasonally over a distance of 3–15 m. The observed patterns of erosion and deposition follow changes in wind speed over the dunes and are directly related to the magnitude of velocity and transport rate amplification on dune slopes through the requirements of sediment continuity. This does not lead to a lowering of the dune crest because this zone is reworked from season to season, and there is very little net change in the position of the crestline. Time-series of erosion and

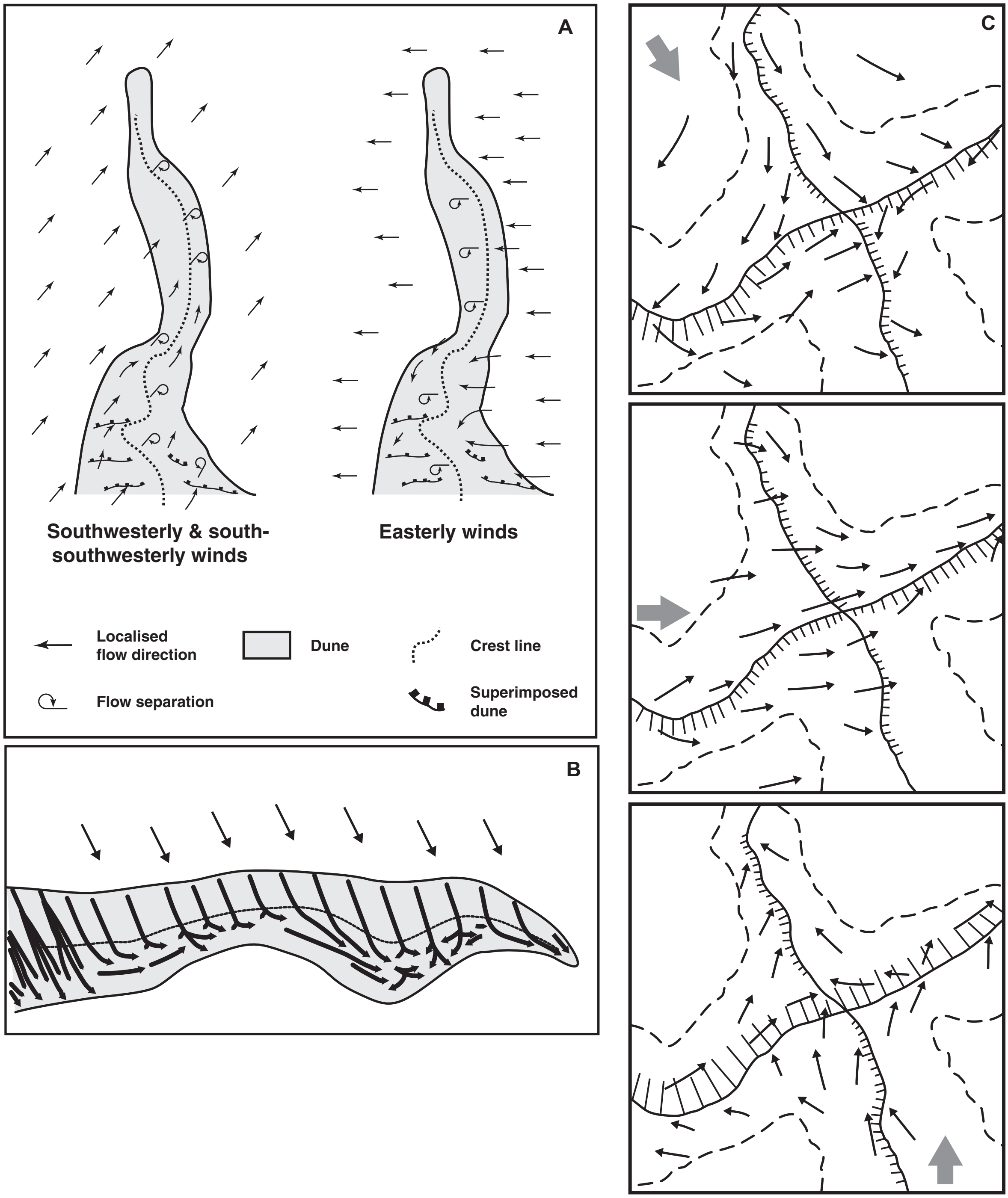

Fig. 18.20 Patterns of airflow on linear and star dunes: (**A**) Simple linear dune – after Bristow et al., 2000; (**B**) Linear dune – after Tsoar, 1983; (**C**) Star dune – after Lancaster, 1989a

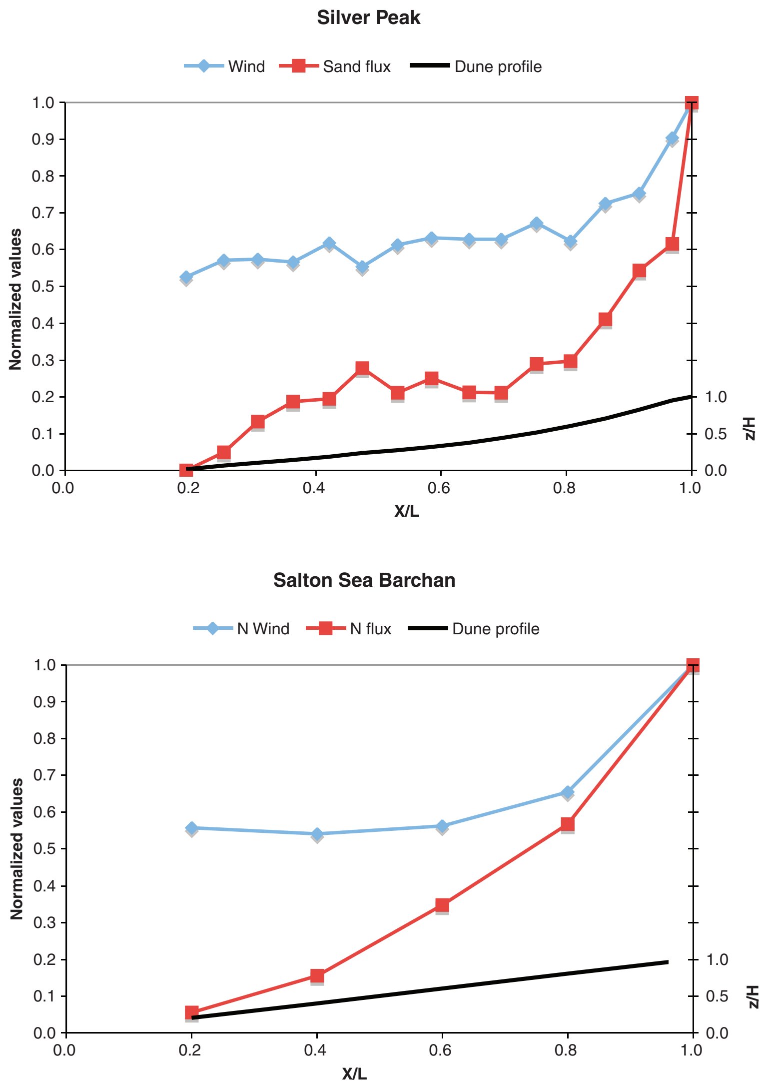

Fig. 18.21 Changes in wind speed and sediment mass flux on the stoss slopes of flow transverse dunes: Silver Peak - reversing transverse dune (McKenna Neuman et al., 1997); Salton Sea - barchan dune (Lancaster et al., 1996)

deposition show that the greatest amount of change occurs when winds change direction seasonally. This is the result of winds encountering a dune form that is out of equilibrium with a new wind direction. Field observations and models show that the crestal profiles of linear and star dunes tend toward a convex form similar to that of transverse dunes (Tsoar, 1985) and erosion and deposition rates near the crest decline as the dune comes into equilibrium with a new wind direction.

Despite their importance to the dynamics of modern dunes and the interpretation of the rock record (Howell and Mountney, 2001), there have been few studies of the depositional processes that occur on dune lee faces. Prior studies consist of two modeling efforts (Anderson, 1988; Hunter, 1985), and three known field experiments (Hunter, 1985; McDonald and Anderson, 1995; Nickling et al., 2002). Following the terminology of (Hunter, 1977), grain flows occur when sediment deposited on the lee face by grain fall from the overshoot of saltating grains transported over the brink of the lee face builds up so that the lee slope is steepened above the angle of repose and fails, initiating a grain flow or avalanche.

McDonald and Anderson (1995) proposed a numerical model to account for the pattern of grainfall deposition developed on the lee of transverse dunes that suggested an exponential decay in deposition rates, rather

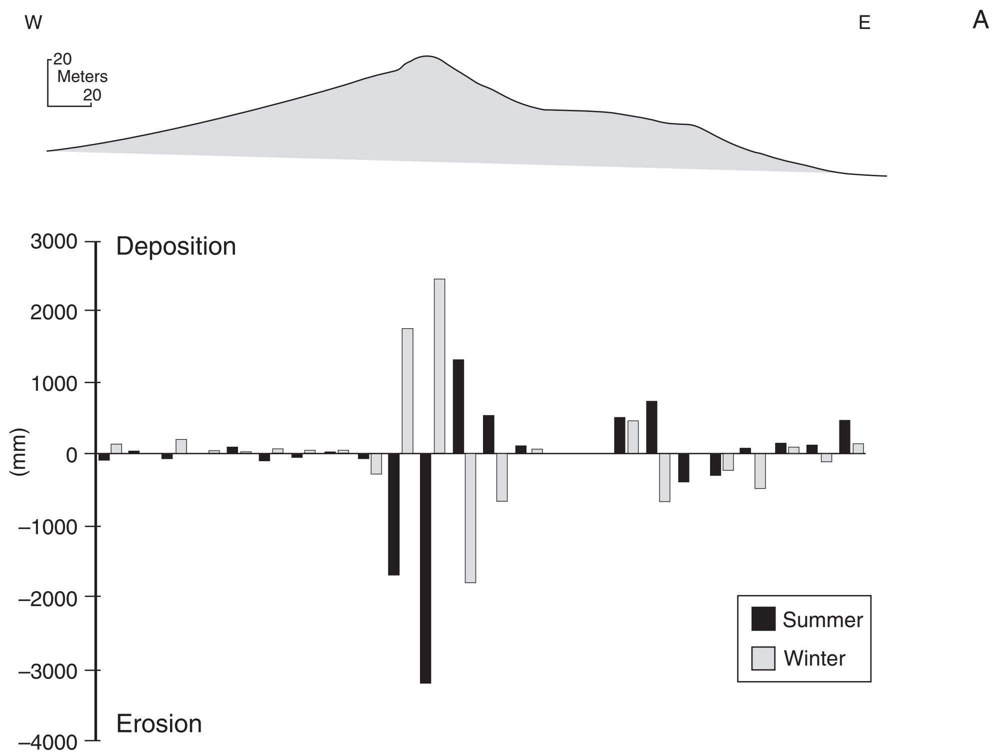

Fig. 18.22 Patterns of annual net erosion and deposition across a Namib complex linear dune. Primary winds is oblique to the dune from the south-south-west to the south-west (summer) or NE-E (winter). After Livingstone (1989)

than the power function suggested by Hunter (1985). Field testing of the model appears to confirm the exponential decay as well as the development of a depositional maximum on the upper lee slope. The experimental data, while providing a good fit to the numerical model, also indicates the importance of wake effects and suspension of grains in the lee of the dune. McDonald and Anderson also concluded that increased wind speeds resulted in more frequent, rather than larger avalanches.

Field experiments conducted by Nickling et al. (2002) clearly indicated that the decay of grainfall with horizontal distance is best described by an exponential function, with 55–95% of the grainfall being deposited within 1 m of the crest (Fig. 18.23a). However, comparison of field data with predictions from the model of Anderson (1988) showed significant lack of agreement with respect to the magnitude of deposition rates and the magnitude and location of the lee slope depositional maximum, which was an order of magnitude larger and located further down slope than indicated by the model. Avalanches appear to be initiated downslope from this area, as confirmed by field observations in Namibia and experimental wind tunnel studies.

Wind tunnel experiments using a small but true-scale artificial flow-transverse sand dune (Cupp et al., 2004) indicate that grainfall decreases approximately exponentially with distance from the brink, confirming prior field studies. The grainfall flux to the lee slope scales with overall crest wind speed and therefore incoming saltation flux (Fig. 18.23a) and shows the development of a small depositional maximum just downslope from the brink. Significant quantities of grainfall are redistributed downslope by the reptation of grains on the surface of the lee slope, a previously unrecognized process. Reptation rates scale with grainfall rates (Fig. 18.23b) and therefore decline rapidly with distance from the brink. The spatial variations in reptation rates downslope are reflected in changes in the profile of the lee slope and suggest that reptation is primarily responsible for the formation of a large sediment wedge across the upper lee slope. This sediment wedge contributes to the failure of the

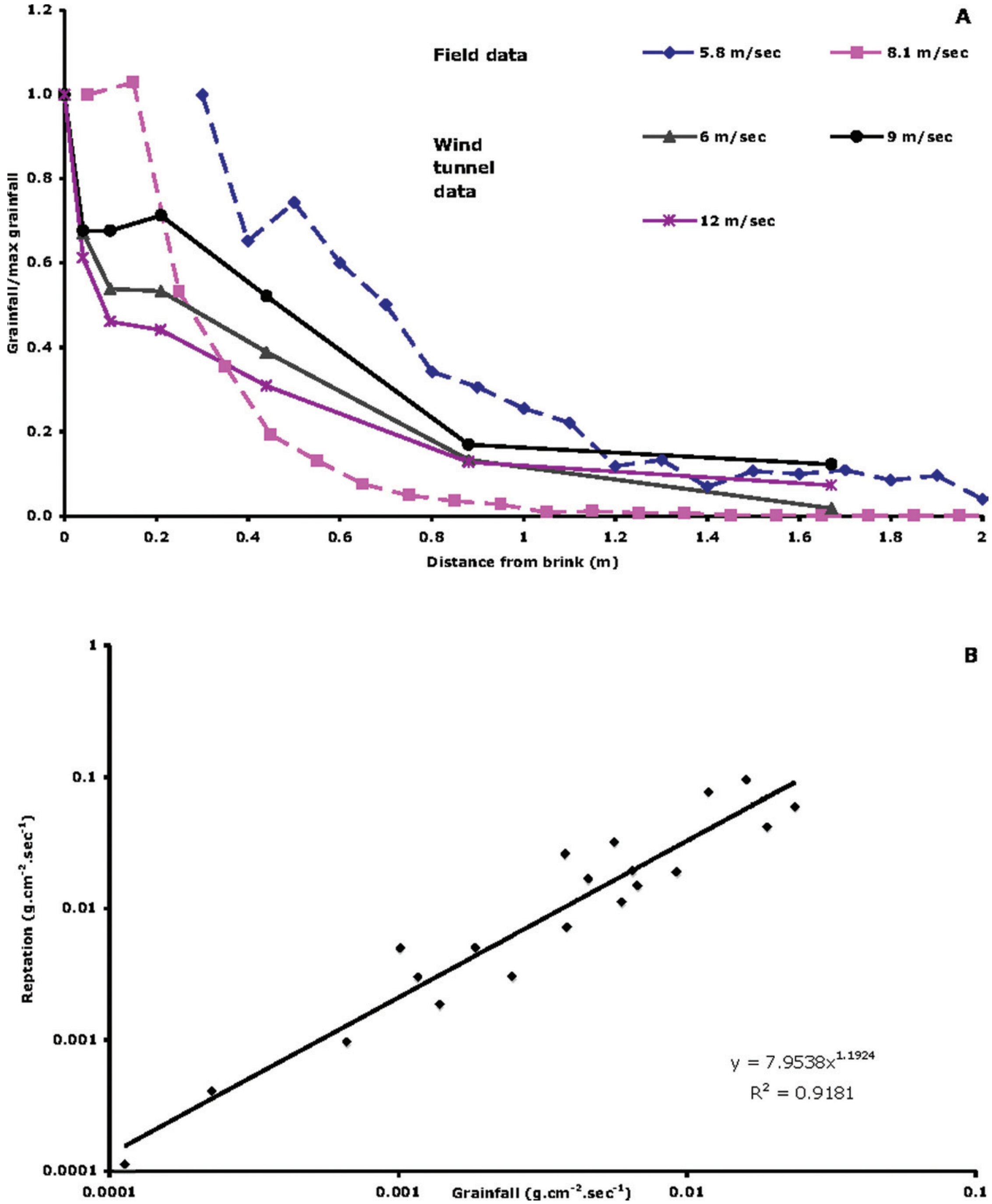

Fig. 18.23 (**A**) Depositional patterns in the lee of a transverse dune at varying incident wind speeds (field data from Nickling et al., 2002; and wind tunnel data from author); and (**B**) relations between rate of grainfall and rate of downslope reptation on experimental wind tunnel dune (author unpublished data)

upper lee slope, and the resulting redistribution of sand by grainflow. The experiments show that grainflow magnitude remained relatively constant with wind speed whereas grainflow frequency increased with wind speed.

Long Term Dune Dynamics

The majority of the process studies discussed above have a duration of days or weeks, and the measured rates of erosion and deposition are difficult to extrapolate to the longer time scales over which dunes and dune fields form. One approach is to model dune behaviour over time using information on the fundamentals of dune dynamics. This field is still in its infancy but recent studies show that this approach can produce realistic dune shapes and patterns using a number of different approaches (Partelli et al., 2006; Schwämmle and Herrmann, 2004; Werner, 1995).

Long term monitoring can provide insights into dune dynamics over annual to decadal timescales. Comparison of the position of dunes on time-series of aerial photographs or satellite images has provided a large data set on dune migration rates. An inverse relationship between barchan dune height and migration rate (Fig. 18.24) has been determined by using this approach by numerous investigators (Haff and Presti, 1995; Long and Sharp, 1964; Sweet et al., 1988; Finkel, 1959; Hastenrath, 1967; Marîn et al., 2005; Slattery, 1990). Such investigations are now facilitated by GIS and GPS methods which also enable changes in dune volume (including dissipation of dunes) to be determined (Bristow and Lancaster, 2004; Stokes et al., 1999).

Studies of the long-term dynamics of individual dunes are rare. Monitoring of surface topographic change on a large linear dune in the central Namib Desert since 1980 (Livingstone, 1989, 1993, 2003) has shown that the crest region of the dunes is the most active as it is reworked by winds from different directions according to season. The crestlines migrate over a lateral distance of as much as 14 m over a 12-month period but with little net change over periods of years. The crest also changed from a relatively high single crest in the 1980s to a slightly lower double crest form in the 1990s, and then back to single crest form by 2001, regaining much of its original height (Fig. 18.25). The lower, or plinth, areas of the dune showed little change over the period of study. Livingstone (2003) attributed the changes in dune crest characteristics to changes in the relative magnitude and frequency of strong easterly winter winds, which increased in the late 1980s. The studies also suggested that these large dunes are not migrating laterally.

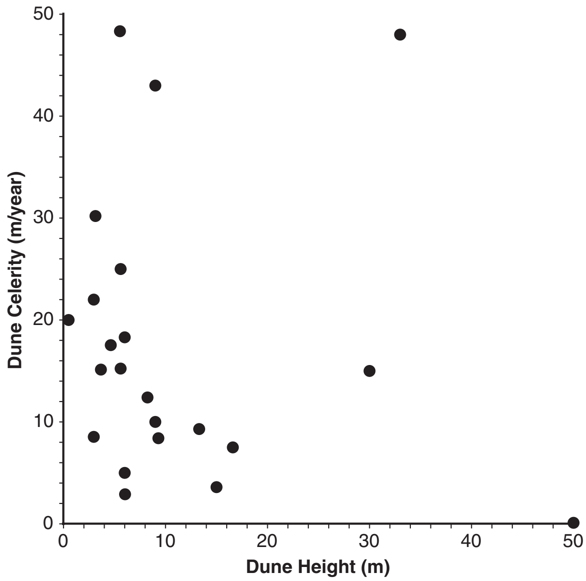

Fig. 18.24 Relations between crescentic dune migration rate (celerity) and dune height. Scatter in points in *y* direction is a function of the overall wind energy at the different localities. Data are mean values calculated from Finkel (1959), Long and Sharp (1964), Hastenrath (1967), Tsoar (1974), Embabi (1982), Endrody-Younga (1982), Haynes (1989), and Slattery (1990)

A longer-term (centuries to millennia) perspective on dune dynamics can now be gained through the application of high-resolution OSL dating of dunes, especially when used in combination with GPR studies of dune structures. Recent studies in the Namib Desert show that crescentic dunes superimposed on the southern edge of a N-S oriented linear dune (Fig. 18.26) have migrated to the west at an average rate of 0.12 m/yr over the past 1570 years (Bristow et al., 2005), while a large linear dune shows evidence of episodic development and lateral migration at rates of up to 0.13 m/yr over the past 6000 years (Bristow et al., 2007). Bray and Stokes adopted a similar approach to show that a 20 m high transverse dune in the Liwa area had migrated 250 m to the south over the past 320 years at an average rate of 0.78 m/yr, with more rapid migration of 0.91 m/year between 220 and 110 years ago (Stokes and Bray, 2005).

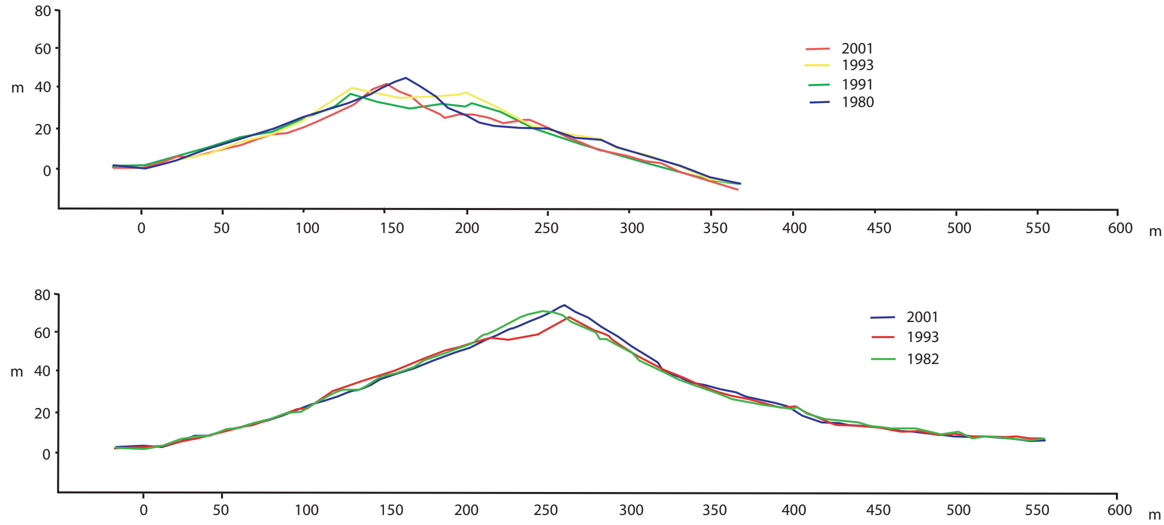

Fig. 18.25 Changes in the profile form of a Namib linear dune over a 20 year period (after Livingstone, 2003)

GPR studies can also provide information on the development and growth of dunes through information on sedimentary structures. Bristow et al. (2000) showed how linear dunes in the Namib Sand Sea go through a series of stages in which secondary flows and form-flow interactions become more important as the dune increases in size, leading ultimately to the establishment of superimposed bedforms (Fig. 18.27).

Controls of Dune Morphology

Recent work has recognized that the directional variability of the wind regime is a major determinant of dune type. Wind speed, grain size, and vegetation play subordinate roles. The effects of sand supply are uncertain (Rubin, 1984; Wasson and Hyde, 1983b).

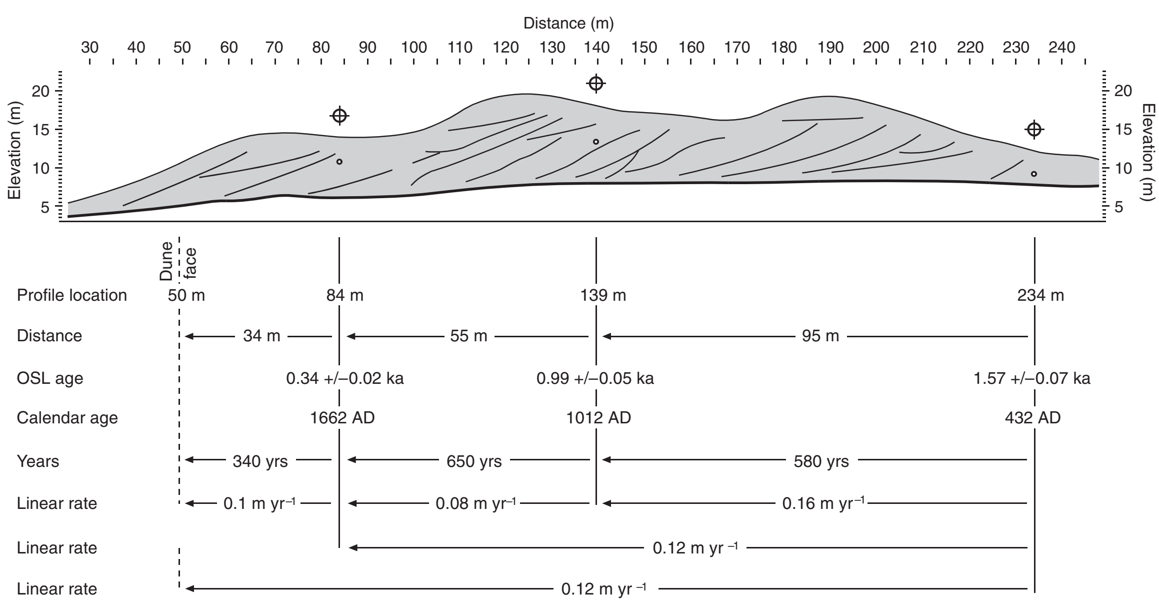

Fig. 18.26 Long term migration of dunes determined using GPR and OSL (after Bristow et al., 2005)

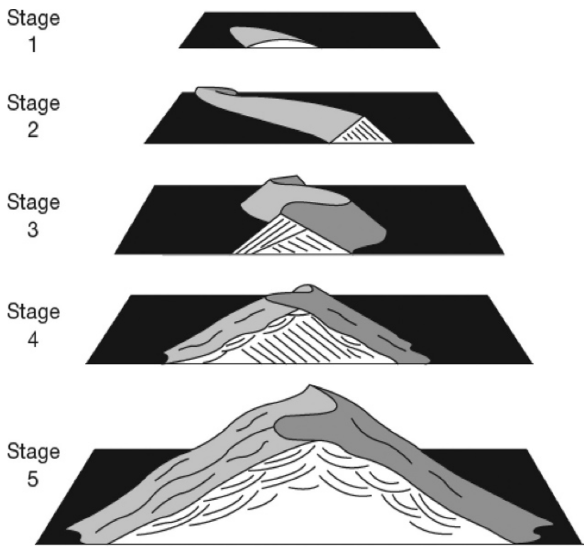

Fig. 18.27 Growth and extension of a linear dune (after Bristow et al., 2000)

Sediment Characteristics

There appears to be no evidence for a genetic relationship between the grain size and sorting character of sands and dune type, except that sand sheets and zibar are often composed of coarse, poorly sorted, often bimodal, sands (Bagnold, 1941; Kocurek and Nielson, 1986; Lancaster, 1982a; Maxwell and Haynes, 1989; Nielson and Kocurek, 1986; Warren, 1972). The hypothesis of Wilson (1972) that dune spacing is related to sand particle size is not supported by empirical data (Wasson and Hyde, 1983a).

Wind Regimes

The association of dunes of different morphological types with wind regimes which have different characteristics, especially of directional variability, has been noted by many investigators. Fryberger (1979) compared the occurrence of each major dune type (crescentic, linear, star) on Landsat images of sand seas with data on local wind regimes, using the ratio between total (DP) and resultant sand flow (RDP) as an index of directional variability. High RDP/DP ratios characterize near unimodal wind regimes, whereas low ratios indicate complex wind regimes. Fryberger (1979) found that the directional variability or complexity of the wind regime increases from environments in which crescentic dunes are found to those where star dunes occur (Fig. 18.28). Crescentic dunes occur in areas where RDP/DP ratios exceed 0.50 (mean RDP/DP ratio 0.68) and frequently occur in unimodal wind regimes, often of high or moderate energy. Linear dunes develop in wind regimes with a much greater degree of directional variability and

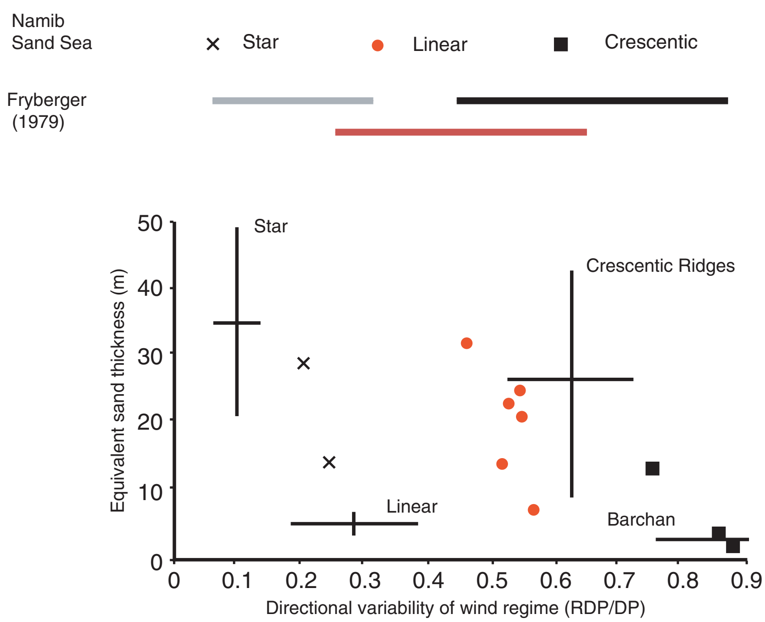

Fig. 18.28 Relations between dune type, wind regime variability (RDP/DP ratio), and sand thickness (RDP/DP data from Fryberger, 1979; Wasson and Hyde, 1983a; and Lancaster, 1989b). Note that linear dunes in the Namib occur in less variable wind regimes than those in Wasson and Hyde's sample

commonly form in wide unimodal or bimodal wind regimes with mean RDP/DP ratios of 0.45. Star dunes occur in areas of complex wind regimes with RDP/DP ratios less than 0.35 (mean = 0.19).

Comparison of the distribution of dunes of different morphological types in the Namib sand sea with the information on sand-moving wind regimes (Lancaster, 1989b) tends to confirm Fryberger's hypothesis that there is an increase in the directional variability of the sand-moving wind regime from crescentic to star dunes. However, the overall directional variability of wind regimes in the Namib sand sea is less than that in Fryberger's global sample (Fig. 18.28). Computer simulations of dune patterns using a cellular automaton approach further demonstrate that each of the major morphological types is independent of initial conditions and forms an attractor in a complex system (Werner, 1995; Werner and Kocurek, 1997; Werner and Kocurek, 1999), in which wind regime variability (equivalent to variability in transport directions) is the main determinant of dune type and orientation (Fig. 18.29).

Process studies give some indications as to why dunes of different types should occur in different wind regimes. The primary response of sand surfaces to the wind is to form an asymmetric transverse dune with a convex stoss slope. This is the most common dune form in unidirectional wind regimes. In multidirectional wind regimes, profiles of the crestal areas of linear and star dunes tend toward this form in each wind season. As dunes grow, the cross-sectional area increases exponentially and their reconstitution time

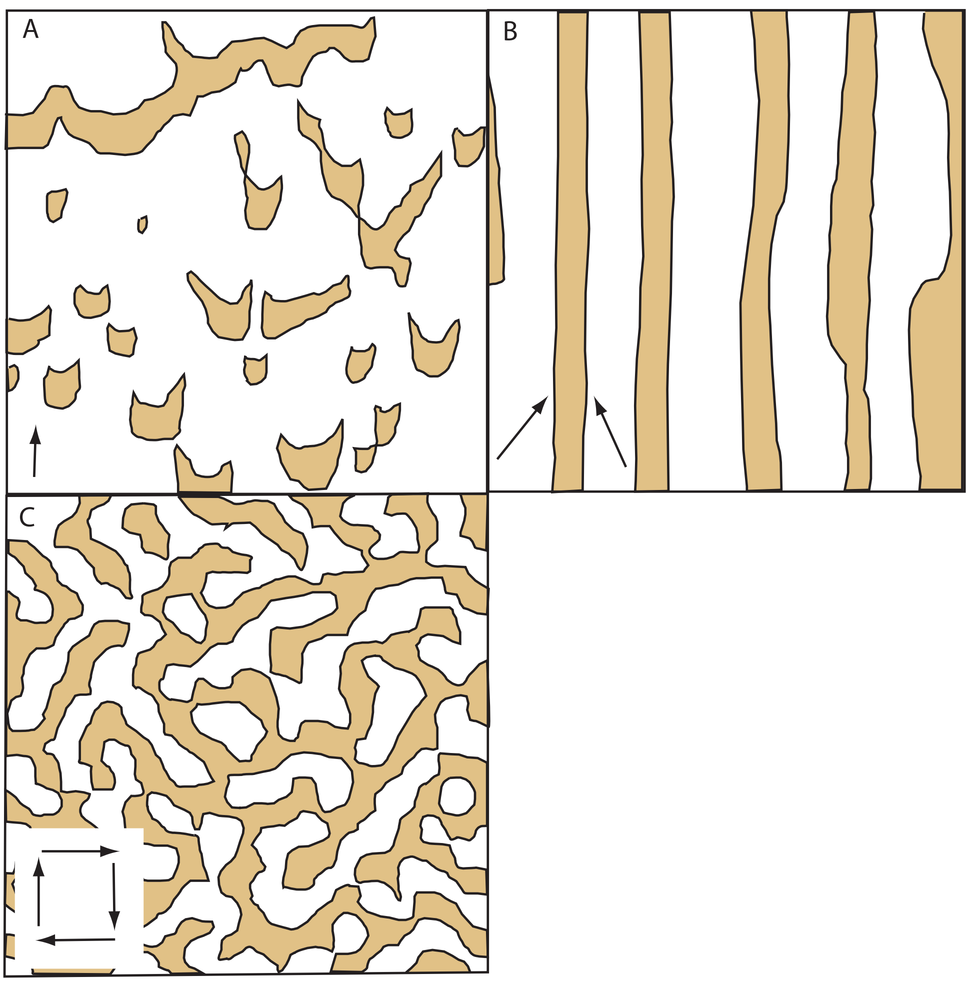

Fig. 18.29 Numerical simulation of dune morphological types and patterns (after Werner, 1995). Arrows show principal sand transporting wind directions used in simulations

increases by one or two orders of magnitude. Because of their size, they can no longer be remodelled in each season, so form–flow interactions become significant, and they develop a morphology that is controlled by more than one wind direction. There are many examples of dunes of different types occurring together in the same wind regime: the small dunes are almost always crescentic forms because they can be reformed entirely in each wind season; larger dunes often tend to be linear or star types.

The essential mechanism for linear dune formation is the deflection of winds that approach at an oblique angle to the crest to flow parallel to the lee side and transport sand along the dune. Thus any winds from a 180° sector centred on the dune will be diverted in this manner and cause the dune to elongate downwind. Linear dunes are not stable in a unidirectional wind regime because erosion and deposition are concentrated at the same localities, resulting in their eventual break-up (Tsoar, 1983). Thus a seasonal change in wind direction is necessary to maintain the dune and its triangular profile.

In bimodal wind regimes, deflection of oblique winds on the lee side will tend to elongate the form, producing a linear dune. The effects of each wind direction will be controlled by their direction relative to the dune. If a high percentage of winds are at optimal angles for lee-side deflection, then the dune will extend strongly, and such dunes will tend to be long and relatively low. If winds blow at higher angles to the dune, then longitudinal movements of sand will be replaced by deposition on lee-side avalanche faces. Sand will tend to stay on the dune and increase its height. A limiting case will occur when winds are perpendicular to the crest, producing a reversing dune. Such dunes tend to accrete vertically and develop major lee-side secondary flow cells that move sand toward the centre of the dune, leading to the formation of star dunes.

The development of star dunes is strongly influenced by the high degree of form–flow interaction, which occurs as a result of seasonal changes in wind direction, and the existence of a major lee-side secondary circulation. The pattern of winds on star dunes indicates that most of the resultant erosion and deposition involves the reworking of deposits laid down in the previous wind season. Sand, once transported to the dune, tends to stay there and add to its bulk. This is a result of the high deposition rates on lee faces and the fact that wind velocity and sand transport rates decrease away from the central part of the dune. The prominent lee-side secondary flows also tend to move sand towards the centre of the dune and promote the development of the arms.

Sand Supply

The availability of sand for dune building has long been considered a factor influencing dune morphology (Hack, 1941; Wilson, 1972). Using a sample of dunes of different types from sand seas in all major desert areas, Wasson and Hyde (1983b) established that the mean equivalent or spread-out thickness (EST) of dune sand in a given area for all dune types was statistically identical and concluded that although sediment availability was a significant variable determining dune type, it was not the only one.

However, by plotting EST against Fryberger's RDP/DP ratio, a clear discrimination of dune types was achieved (Fig. 18.28), leading to the conclusion that barchans occur where there is very little sand and almost unidirectional winds; transverse dunes are located where sand is abundant and winds variable; linear dunes develop where sand supply is small, but winds are more variable; and star dunes form in complex wind regimes with abundant sand supply. Similar relationships are evident in the Namib sand sea, although the range of directional variability is less (Fig. 18.28).

However, EST is not a measure of sand supply, but of the volume of sand contained in the dunes and may be a reflection of dune type with the dune type being influenced by the other factors, especially the wind regime (Rubin, 1984). In the Namib sand sea it is possible to clearly discriminate between dune types on the basis of their relationships with wind regimes. The EST data merely suggest that there is more sand in complex linear and star dunes than in compound linear and all types of crescentic dunes.

Vegetation

The effects of vegetation on sand transport, which include direct protection of the surface, absorption of momentum, and partitioning of shear stress between the surface and plants are discussed in depth in Chapter 17. The effects of vegetation on dune morphology are, however less well known. Hack (1941) suggested that in north-eastern Arizona there was a transition from crescentic to parabolic dunes with increased vegetation cover, and that linear dunes occurred in areas with less sand and vegetation than parabolic dunes (Fig. 18.30). In the Negev Desert, Tsoar and Møller (1986) documented changes in the morphology of linear dunes as vegetation was removed, including development of sharp sinuous 'seif' dune crests and braided patterns.

Many dunes, even in hyperarid regions like the Namib, are vegetated to some extent, mostly in the plinth areas of linear and star dunes where relatively little surface change occurs (Thomas and Tsoar, 1990). In other areas, such as the Kalahari and Australian deserts, dune crests may be sparsely vegetated, while vegetation cover is sufficient to restrict sand transport on dune flanks. Partially vegetated linear dunes are widespread in the Australia deserts and in the Kalahri of southern Africa. Stratigraphic evidence and OSL dating (e.g. Stokes et al., 1997) shows that many linear dunes of this type were originally formed during the late Pleistocene and have since become stabilized by vegetation in more humid and/or less windy conditions, although localized Holocene activity has been documented. Studies of dune dynamics in relation to vegetation cover show that the crests of these dunes can be very active, with relatively large amounts of erosion and deposition being recorded, especially in periods of drought or after disturbance e.g. by fire (Bullard et al., 1997; Hesse and Simpson, 2006; Thomas and Leason, 2005; Wiggs et al., 1994, 1995). As pointed out by Bullard and others, inter-annual and decadal scale changes in rainfall, temperature, and wind strength give rise to significant temporal changes in the amount of vegetation cover and dune surface activity. Empirical studies (Lancaster and Baas, 1998; Wiggs et al., 1995) suggest that a vegetation cover of 14% is sufficient to restrict sand transport on dune surfaces. Thomas and Leason (2005) used Landsat image data to show that the proportion of the area of the SW Kalahari dune field with less than 14% vegetation cover ranged between 10 and 16% for dry years but was only 3–6% for wet years, suggesting that dune activity is likely much more extensive after periods of extended drought.

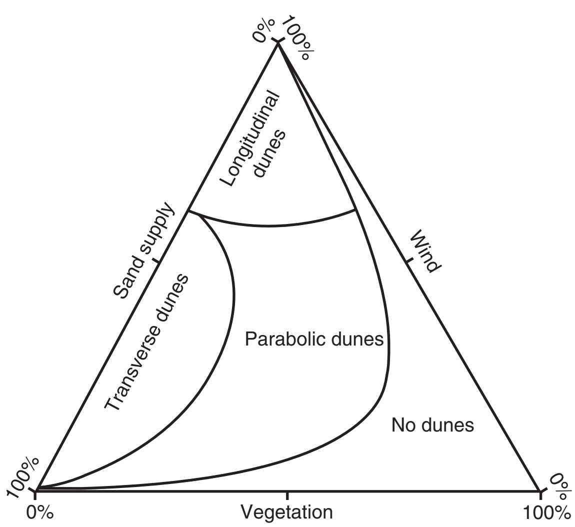

Fig. 18.30 Relations between dune type, sand supply, wind strength, and vegetation cover (after Hack, 1941)

Controls on Dune Orientation

The controls on the orientation or alignment trend of dunes have long been a subject for speculation. Many workers have concluded that dunes are oriented relative to the resultant or vector sum of sand transport (Fryberger, 1979). Thus dunes can be classified as transverse (strike of crestline approximately normal to resultant), longitudinal (crest parallel to resultant), or oblique (15–75° to resultant direction) (Hunter et al., 1983; Mainguet and Callot, 1978). Others believe that dunes are oriented parallel to or normal to prevailing winds (Bagnold, 1953; Glennie, 1970), or that the trend of dunes is influenced by secondary flows (Tsoar, 1983). As discussed above, numerous explanations of why dunes are longitudinal or transverse to transport directions have been advanced, including wind regime characteristics, existence of helical vortices, and sand availability.

Field and laboratory experiments with wind ripples and sub-aqueous dunes (Rubin and Hunter, 1987; Rubin and Ikeda, 1990) suggest that all types of bedforms are oriented in the direction subject to the maximum gross bedform-normal transport across the crest (Fig. 18.31) leading to the conclusion that a wide variety of wind regimes can produce the same dune trend.

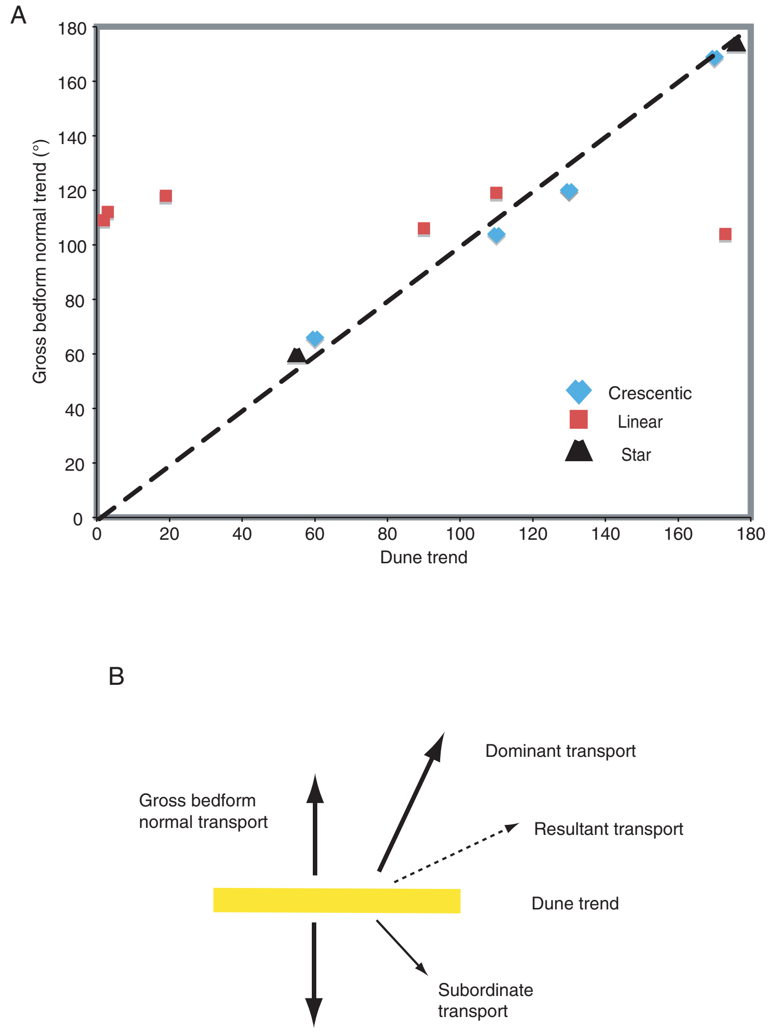

Fig. 18.31 Gross bedform normal (GBN) transport direction and dune trends; (**A**) relations between dune trend and GBN direction (after Lancaster, 1991, with additional data); (**B**) Explanation of Gross Bedform Normal (after Rubin and Ikeda, 1990)

In this approach, transport from all directions contributes to bedform genesis and growth. By contrast, use of the resultant direction of transport as a control of bedform orientation implies that transport vectors from opposing directions cancel out each other and thus do not contribute to bedform growth. The type of dune that occurs is determined by the divergence angle between the dominant and subordinate transport vectors and the ratio between the two transport directions (Transport ratio). A trend parallel to or normal to the resultant direction of sand transport is purely coincidental.

Using a sample of dunes from sand seas in Namibia and the southwestern United States, Lancaster (1991) determined that there was a close agreement between observed and predicted gross-bedform normal orientations in the case of many barchans, crescentic dunes, simple linear dunes, and star dunes (Fig. 18.31), suggesting that all major dune types are oriented to maximize gross bedform-normal sediment transport and therefore are dynamically similar. Differences between the form and orientation of dunes therefore result from variations in the directional characteristics of the flow, especially the angle(s) between the major transport directions and the ratio(s) between the magnitudes of flows from different directions.

This approach can also be used to identify dune trends that are not in equilibrium with the modern sand-transporting wind regime, as demonstrated for Mauritania (Lancaster et al., 2002), and to suggest wind regimes that could have produced the trends of dunes constructed in the past.

Development of Compound and Complex Dunes

Many large dunes (complex and compound dunes, draas, mega dunes) exhibit superimposed dunes. There has been much debate over the orgins of the superimposed bedforms. Lancaster (1988) showed that crescentic dunes superimposed on linear dunes were clearly in equilibrium with secondary airflow patterns on the larger dune form. Similar observations were made for crescentic dunes (Havholm and Kocurek, 1988; Rubin and Hunter, 1982). The slopes of large dunes present an effectively planar surface on which sand transport takes place. Therefore, variations in sand transport rates on mega dunes in time or space will lead to the formation of superimposed dunes if the major dune is sufficiently large. This suggests that there is a minimum size for compound and complex dunes. Although the sample size is small, data on the width and spacing of simple, compound, and complex crescentic and linear dunes (Breed and Grow, 1979) indicate that the mean sizes of simple, compound, and complex dunes are statistically significantly different from one another, suggesting that a minimum dune size must be reached before superimposed dunes can develop. In the Namib sand sea, simple crescentic dunes have a spacing of less than 500 m, whereas compound dunes are all more than 500 m apart. There is, however, a continuum of the height and spacing of the major bedform from simple to compound and complex types. This suggests that, given sufficient sand supply and time, simple dunes may grow into compound and complex forms. This seems to be confirmed by the sedimentary structures of linear dunes (e.g. Bristow et al., 2000), which show the dominance of superimposed dunes once the dune reaches a certain size.

In some areas, however, the larger primary dunes are clearly a product of past wind regime and sediment supply conditions as demonstrated by OSL dating of the major form. Large dunes have considerable inertia and require significant periods of time to adjust to changed conditions of wind regime and sediment supply (Warren and Allison, 1998). The inertia of a dune can be represented by their reconstitution time, or the time required for the bedform to migrate its length in the direction of net transport. In the Namib, typical complex linear dunes in the northern parts of the sand sea have a width of approximately 600 m and migrate laterally at a rate of $0.1\,\mathrm{m\,y^{-1}}$ (Bristow et al., 2007). The minimum reconstitution time for these dunes is therefore 6000 years. Crescentic dunes superimposed on the flanks of the linear dunes have a mean spacing of 90 m and migrate at a rate of $3\,\mathrm{m\,y^{-1}}$, giving a reconstitution time of 30 years. Reconstitution time therefore increases by several orders of magnitude from simple to complex dunes. This implies that the morphology of simple dunes and superimposed dunes is governed principally by annual or seasonal patterns of wind speed and direction and by spatial changes in wind speed over draas. The lifespan of these dunes is about 10–100 years. Compound and complex dunes are relatively insensitive to seasonal changes in local air flow conditions and may persist for 1000–10000 years, as confirmed by OSL ages obtained from many large dunes.

Sand Seas

Sand seas (also called ergs) are dynamic sedimentary bodies that form part of regional-scale sand transport systems in which sand is moved by the wind from source zones to depositional sinks. Major desert sand seas occur in the Saharan and Arabian deserts, the Thar Desert of India, interior Australia, and southern Africa (Namib and Kalahari deserts) (Breed et al., 1979; Wilson, 1973). The source of sediment for sand sea accumulation is commonly external to the sand body, although internal sources, e.g. alluvial deposits in interdune areas, may be important (e.g. in Australia). Sediment for transport by wind may be derived from deflation of bedrock (Besler, 1980), but the primary source of sand-sized grains is sediments deposited by fluvial/alluvial, coastal, or lacustrine systems (e.g. Muhs, 2004; Muhs et al., 1995). In those sand seas for which the source is known, fluvial systems provide the major input of sand for wind transport, either directly through deflation of alluvial sediments, or indirectly through transfer of sediment in coastal sediment transport systems, as in the Namib, Gran Desierto, Sinai, Atacama, and Arabian sand seas (Lancaster, 1999).

Deposition of sand and the accumulation of sand seas occur downwind of source zones wherever sand transport rates are reduced as a result of changes in climate or topography. Compilations of wind data together with information from aerial photographs and satellite images (Fryberger and Ahlbrandt, 1979; Mainguet, 1984b; Wilson, 1971) show that long-distance transport of sand by the wind occurs in the Namib, Sahara, and Arabian Deserts (Fig. 18.32). However, Australian sand seas and many North American dunefields receive sand from local sources (Blount and Lancaster, 1990; Muhs, 2004; Muhs et al., 2003; Ramsey et al., 1999; Wasson et al., 1988). Some sand seas accumulate where sand transport pathways converge in the lee of topographic obstacles (e.g. Fachi-Bilma Erg) or in areas of low elevation (Wilson, 1971; Mainguet and Callot, 1978; Wasson et al., 1988). Others occur where winds are checked by topographic barriers (e.g. Grand Erg Oriental, Kelso Dunes, Great Sand Dunes). Decelerating winds and reduced potential rates of sand transport may be the result of changes in regional circulation patterns that decrease wind speeds and/or increase their directional variability. Sand seas and dunefields in the western and southern Sahara, Saudi Arabia, the eastern Mojave and Sonoran Deserts, and the Namib occur in areas of low total or net sand transport compared with areas without sand sea development (Fryberger and Ahlbrandt, 1979; Fryberger et al., 1984; Lancaster, 1989b; Mainguet, 1984b). Many sand seas (e.g. the Akchar and Makteir sand seas in Mauritania

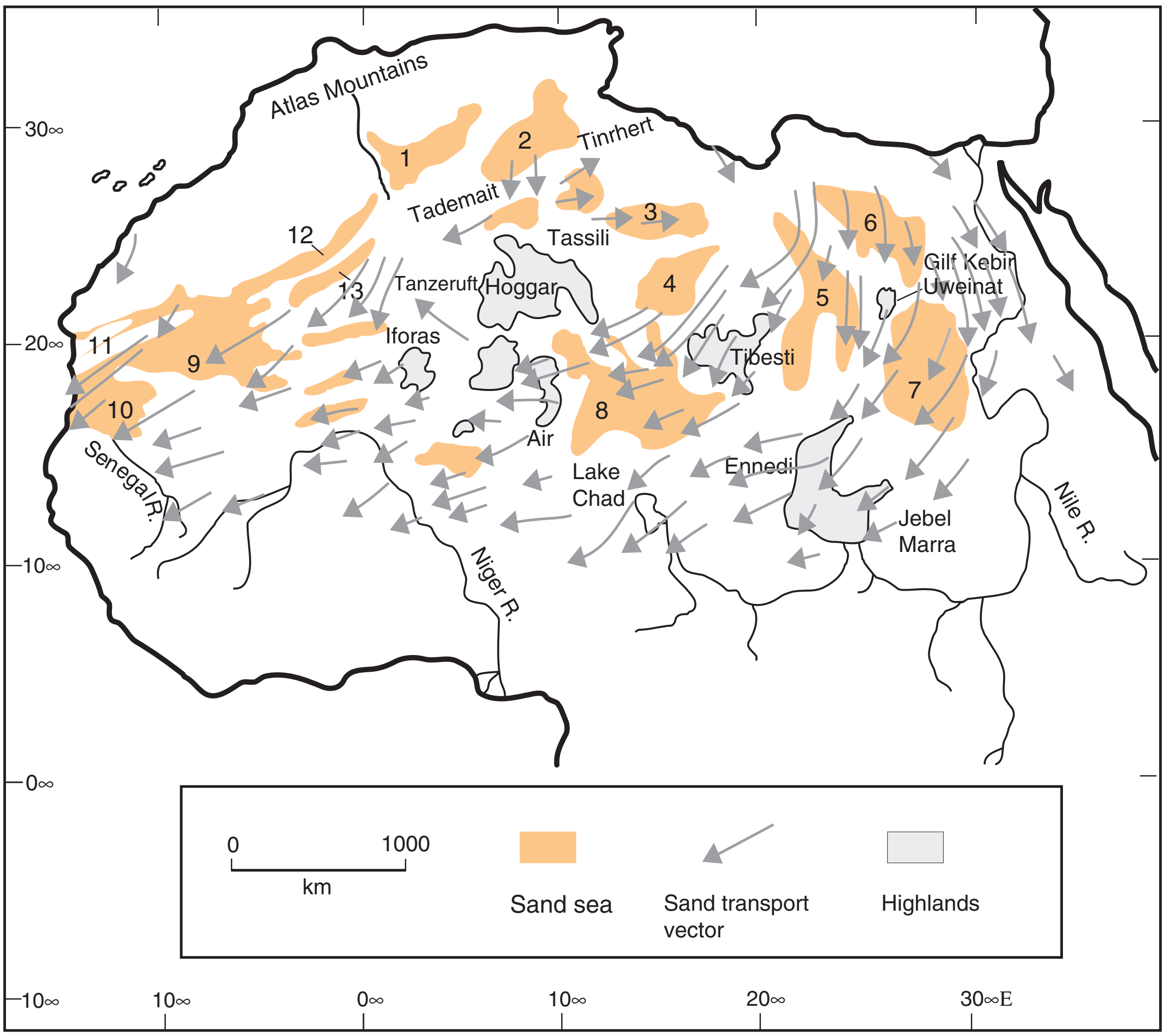

Fig. 18.32 Long distance sand transport pathways in the Sahara (after Mainguet and Chemin, 1983)

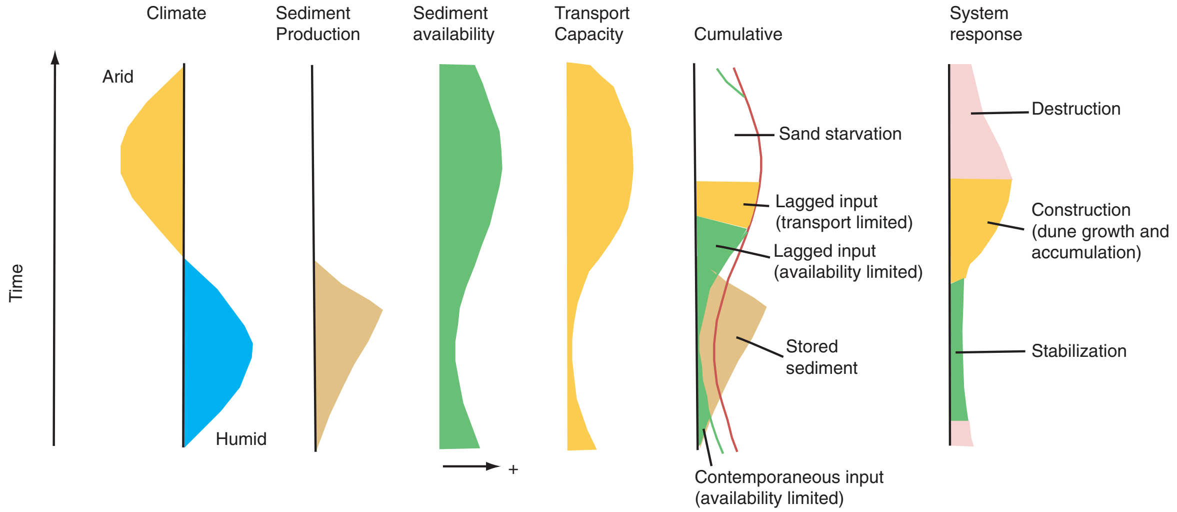

Fig. 18.33 Variation in sediment states and system response over time in a hypothetical example (after Kocurek, 1998)

and the Nafud of Saudi Arabia) are crossed by sand transport pathways. The same winds that transport sand to the sand sea can also remove it at the downwind end. These are the 'flow crossed' sand seas of Wilson (1971).

Sand seas are important depositional landforms and have accumulated over periods of 10^3 to 10^4 years, during which changes in climate and sea level have affected sediment supply, availability, and mobility. The relations between sediment supply, availability, and mobility at any point in time, as well as their variation through time, define a series of states of the system (Kocurek, 1998; Kocurek and Lancaster, 1999), as summarized in Fig. 18.33. They also determine the response of the system to external forcing factors so that a sand sea may be described as:

(1) Transport limited, in which actual rate of sand transport (Qa) equals the potential rate (Qp) and the system is limited only by the capacity of the wind to move sediment from source zones. Examples are the Namib sand seas and many coastal dune fields.
(2) Availability limited, in which Qa $<$ Qp and the system response is controlled by vegetation cover, for example as in the Kalahari and Australia.
(3) Supply limited, in which Qa $<<$ Qp and the system is starved of sediment, as in many central and northern sand Saharan sand seas, some of which are apparently eroding in present conditions.

Dune Patterns in Sand Seas

Many sand seas show a clear spatial patterning of dune types (Fig. 18.34), dune size and spacing, crest orientation, and sediment thickness (Breed et al., 1979; Ewing et al., 2006; Lancaster, 1999). Once thought to be largely the product of regional changes in wind regimes (e.g. Lancaster, 1983), sharp transitions between dune types and differences in their sedimentary characteristics (e.g. in the Simpson-Strzelecki, the Wahiba Sands, and the Gran Desierto) suggest that many sand seas are composed of different generations of dunes, each with a distinct morphology and sediment source (Lancaster, 1999). Dune generations (Fig. 18.35) can be recogized in many ways including: differences in dune morphology giving rise to statistically distinct populations of dunes with different crest orientation, spacing, and defect density (Kocurek and Ewing, 2005); variations in dune sediment composition, color, and particle size (Teller et al., 2002; Wasson, 1983); variations in vegetation cover and dune activity (Lancaster, 1992); and geomorphic relations between dunes of different types (e.g. crossing patterns, superposition of dunes) (Lancaster et al., 2002).

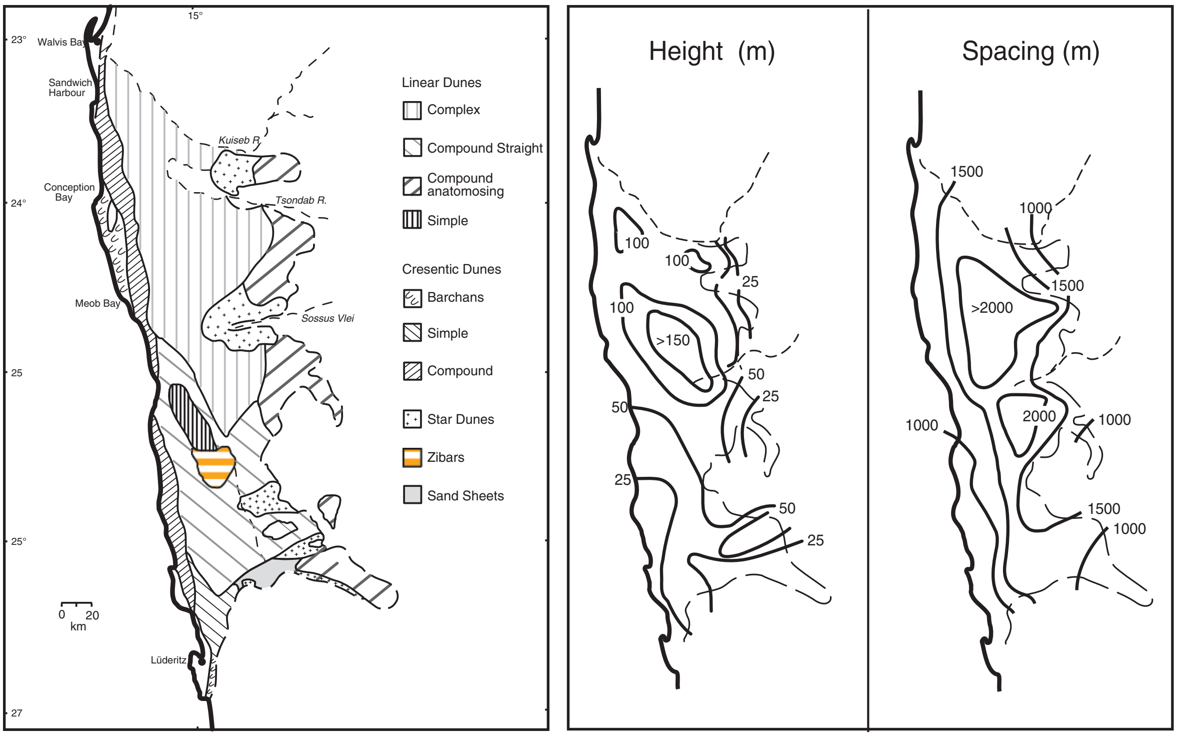

Fig. 18.34 Spatial variation dune types and morphology in the Namib Sand Sea (after Lancaster, 1989b)

Concepts of self-organization of complex systems (Werner, 2003; Werner and Kocurek, 1999), indicate that dune patterns should evolve through time as a result of interactions between dunes so that spacing increases asymptotically and defect density decreases. Further, each generation of dunes represents the product of a set of initial conditions of wind regime and sediment supply, together with evolution of the pattern

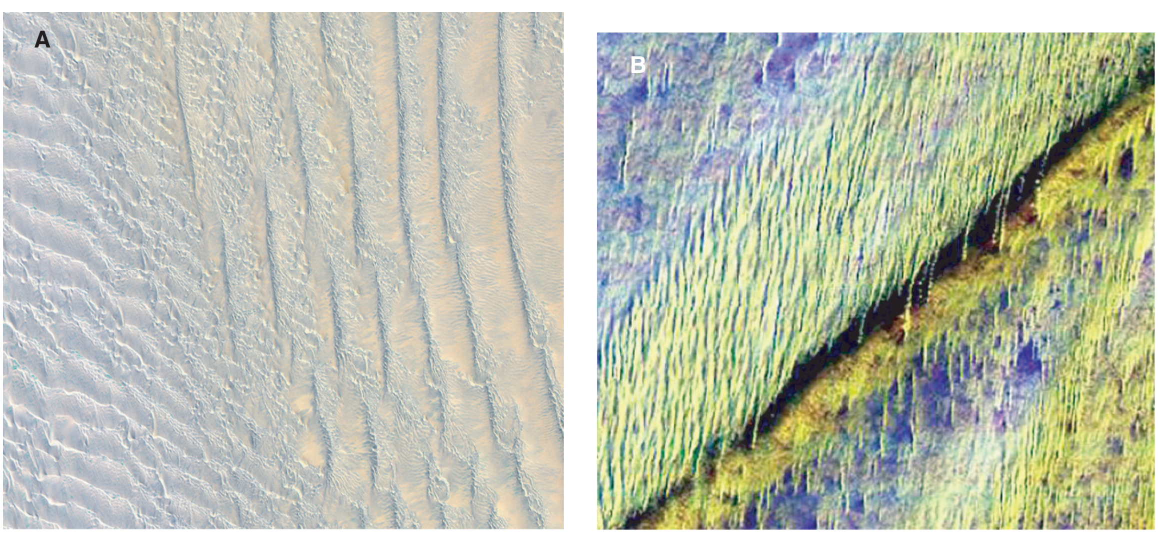

Fig. 18.35 Examples of dune generations: (**A**) juxtaposition and superposition of pale compound crescentic and redder complex linear dunes, Namib Sand Sea; and (**B**) superimposition of modern N-S oriented linear dunes on NE-SW oriented late Pleistocene red brown linear dunes, Azefal Sand Sea, Mauritania

over time by merging, lateral linking, migration of defects, and creation of terminations (Fig. 18.36) (Kocurek and Ewing, 2005). As a result, dune patterns (especially those of crescentic dunes) should evolve downwind, as demonstrated by dunes at White Sands and on the Skeleton Coast of Namibia. Change in dune patterns can only occur at terminations, so that dunes with few terminations per unit crest length (e.g. linear dunes) will tend to be more stable than those with large numbers of terminations (e.g. many crescentic dunes). These concepts are broadly supported by empirical data from several sand seas (Ewing et al., 2006), as discussed below.

The Azefal, Agneitir, and Akchar Sand Seas of western Mauritania show three distinct crossing trends of linear dunes (Fig. 18.37). Analyses of Landsat images, in conjunction with geomorphic and stratigraphic studies, and OSL (optically stimulated luminescence) dating of dunes demonstrate the existence of three main generations of dunes that were formed during the periods 25–15 ka (centered around the Last Glacial Maximum), 10–13 ka (spanning the Younger Dryas event), and after 5 ka (Lancaster et al., 2002). Modelling of the wind regimes that produced these dunes using the gross bedform normal approach shows that the wind regimes that occurred during each of these periods were significantly different, leading to the formation of dunes on three distinct superimposed trends-northeast, north-northeast, and north. Satellite images of the Azefal Sand Sea show clear evidence for reorientation of the termini of older dunes, supporting the model of Werner (1995) and Werner and Kocurek (1997) for change in dune patterns at termination sites. In this model, because linear dunes have few terminations per crest length, these dunes are remarkably stable features. The stability of linear dunes is further illustrated by the superposition of progressive generations of linear dunes at an angle to older trends, arguing for a rate of dune formation that is greater than the rate at which the older, larger dunes can be reoriented.

The Gran Desierto Sand Sea of northern Mexico exhibits a spatially diverse and complex pattern of dunes of several morphological types (Figs. 18.38) (Blount and Lancaster, 1990; Lancaster, 1995b). Analyses of dune patterns, coupled with OSL dating, shows that the sand sea has evolved by coalescence of several generations of dunes, each created during relatively short-lived periods of aeolian construction in the late Pleistocene and Holocene (Beveridge et al., 2006). The characteristic star dunes represent a late Holocene modification (<3 ka) of linear dunes originally constructed between 26 and 12 ka; degraded crescentic dunes formed around 12 ka; while compound crescentic dunes on the easteren margin of the sand sea formed around 7 ka.

Conclusions and Future Research Directions

Major advances in our knowledge of the dynamics of dunes and sand seas have occurred in the past three decades. There is now a general understanding of the processes that form or at least maintain most major dune types. The importance of scale effects is recognized, so that the processes that form wind ripples or simple dunes need to be considered at different temporal and spatial scales from those that form compound and complex dunes and sand seas. The importance of past conditions of sediment supply, availability, and mobility, determined by climatic and sea level changes, in the formation of mega dunes and sand seas and dunefields is now well recognized.

This progress has come about as a result of several important methodological and conceptual advances. Increasing sophistication of instrumentation has enabled more precise measurement of airflow and sediment transport on dunes, although the limitations of this reductionist approach are being increasingly recognized (Livingstone et al., 2007). The application of increasingly precise optical dating techniques has provided new data on the age of dunes. When coupled with ground penetrating radar studies of dune sedimentary structures, this technique can provide unique information on dune dynamics on timescales of centuries to millennia. Numerical modelling of dunes and dune systems has been facilitated by vastly increased computing power, as well as better understanding of fundamental dune processes, although this area of study is still evolving very rapidly. Remote sensing images and GIS techniques now make it possible to analyze dune patterns in detail and trace patterns of sediment movement from source areas to depositional sinks.

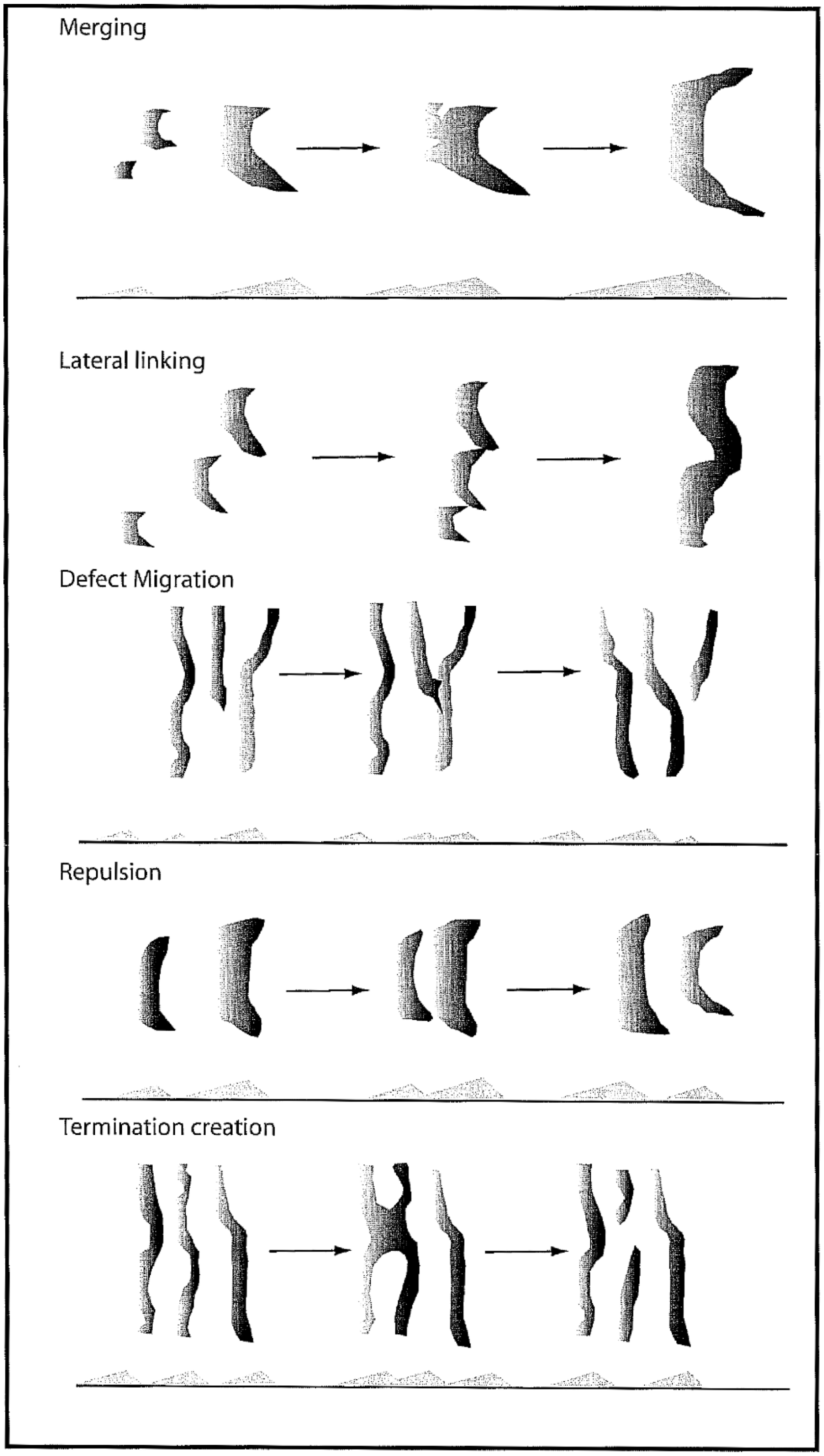

Fig. 18.36 Evolution of dune patterns (after Kocurek and Ewing, 2005)

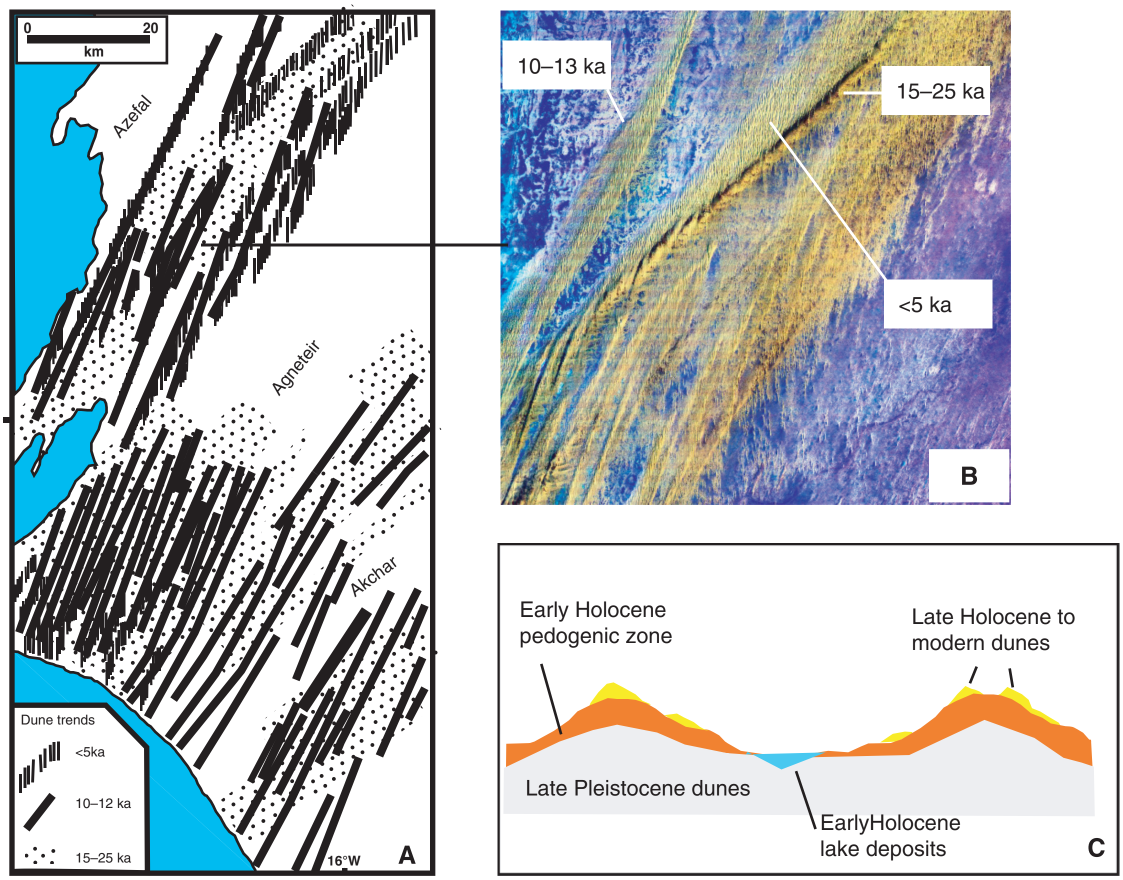

Fig. 18.37 Dune patterns and dune generations in Mauritania (after Lancaster et al., 2002 and Kocurek et al., 1991)

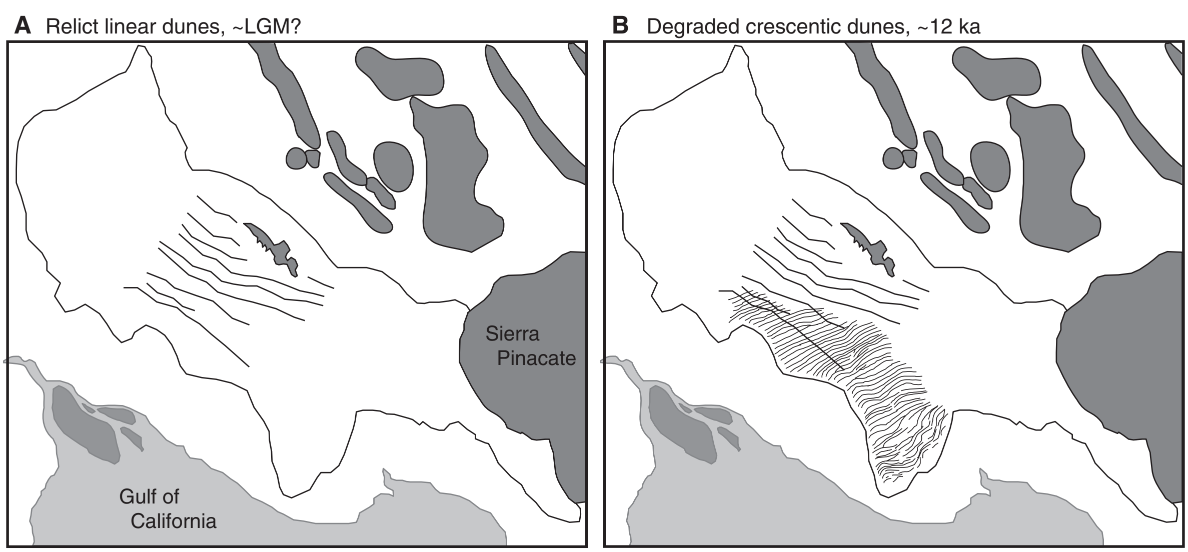

Fig. 18.38 Development of dune patterns in the Gran Desierto, Mexico by assembly and modification of different dune generations (after Beveridge et al., 2006)

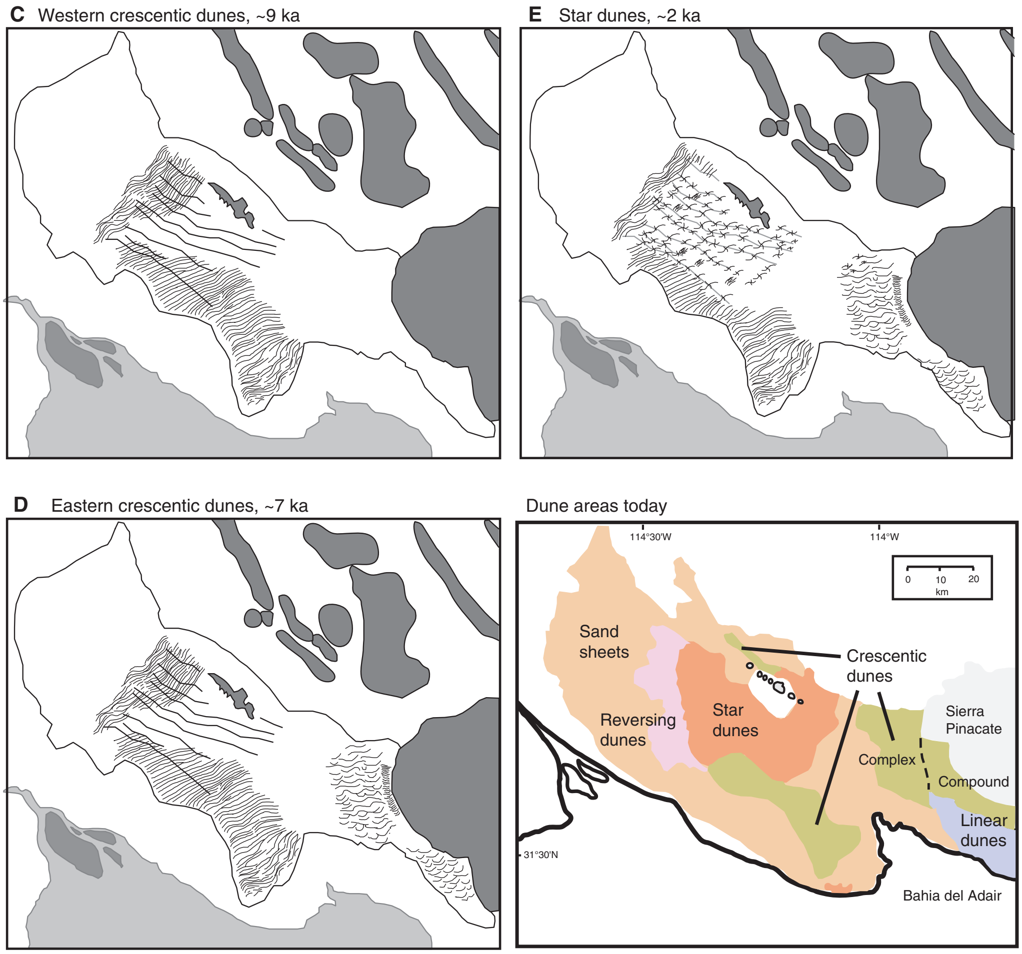

Fig. 18.38 (continued)

References

Allen, J.R.L., 1974. Reaction, relaxation and lag in natural sedimentary systems: general principles, examples and lessons. Earth Science Reviews, 10: 263–342.

Anderson, R.S., 1987. A theoretical model for aeolian impact ripples. Sedimentology, 34: 943–956.

Anderson, R.S., 1988. The pattern of grainfall deposition in the lee of aeolian dunes. Sedimentology, 35(2): 175–188.

Anderson, R.S., 1990. Eolian ripples as examples of self-organization in geomorphological systems. Earth Science Reviews, 29: 77–96.

Anderson, R.S. and Bunas, K.L., 1993. Grain size segregation and stratigraphy in aeolian ripples modeled with a cellular automaton. Nature, 365: 740–743.

Andreotti, B., Claudin, P., and Douady, S., 2002. Selection of dune shapes and velocities. Part I: Dynamics of sand, wind, and barchans. European Physics Journal, 28 B.

Andrews, S., 1981. Sedimentology of Great Sand Dunes, Colorado. In: F.P. Ethridge and R.M. Flores (Eds.), Recent and Ancient Non Marine Depositional Environments: models for exploration. The Society of Economic Paleontologists and Mineralogists, Tulsa, OK, pp. 279–291.

Anton, D. and Vincent, P., 1986. Parabolic dunes of the Jafurah Desert, Eastern Province, Saudi Arabia. Journal of Arid Environments, 11: 187–198.

Ash, J.E. and Wasson, R.J., 1983. Vegetation and sand mobility in the Australian desert dunefield. Zeitschrift fur Geomorphologie Supplement, 45: 7–25.

Bagnold, R.A., 1941. The Physics of Blown Sand and Desert Dunes. Chapman and Hall, London, 265pp.

Bagnold, R.A., 1953. The surface movement of blown sand in relation to meteorology, Desert Research, Proceedings of the International Symposium. Research Council of Israel, Jerusalem, pp. 89–93.

Besler, H., 1980. Die Dunen-Namib: entstehung und dynamik eines ergs. Stuttgarter Geographische Studien, 96: 241.

Beveridge, C., Kocurek, G., Ewing, R., Lancaster, N., Morthekai, P., Singhvi, A. and Mahan, S., 2006. Development of spatially diverse and complex dune-field patterns: Gran Desierto Dune Field, Sonora, Mexico: Sedimentology, 53: 1391–1409.

Blount, G. and Lancaster, N., 1990. Development of the Gran Desierto Sand Sea. Geology, 18: 724–728.

Boulton, J.W., 1997. Quantifying the morphology of aeolian impact ripples formed in a natural dune setting, University of Guelph, Guelph, Ontario, Canada, 172pp.

Breed, C.S., Fryberger, S.G., Andrews, S., McCauley, C., Lennartz, F., Geber, D. and Horstman, K., 1979. Regional studies of sand seas using LANDSAT (ERTS) imagery, In: McKee, E.D. (Ed.), A Study of Global Sand Seas: USGS Professional Paper, 1052, pp. 305–398.

Breed, C.S. and Grow, T., 1979. Morphology and distribution of dunes in sand seas observed by remote sensing. In: E.D. McKee (Editor), A Study of Global Sand Seas. Professional Paper. United States Geological Survey Professional Paper 1052, pp. 253-304.

Breed, C.S., McCauley, J.F. and Davis, P.A., 1987. Sand sheets of the eastern Sahara and ripple blankets on Mars. In: L.E. Frostick and I. Reid (Eds.), Desert Sediments: Ancient and Modern. Blackwell Scientific Publications, Oxford, London, Edinburgh, Boston, Palo Alto, Melbourne, pp. 337–359.

Bristow, C.S., Bailey, S.D. and Lancaster, N., 2000. Sedimentary structure of linear sand dunes. Nature, 406: 56–59.

Bristow, C.S., Duller, G.A.T. and Lancaster, N., 2005. Combining ground penetrating radar surveys and optical dating to determine dune migration in Namibia. Journal of the Geological Society (London), 162(2): 315–321.

Bristow, C.S., Duller, G.A.T. and Lancaster, N., 2007. Age and dynamics of linear dunes in the Namib Desert. Geology, 35:555–558.

Bristow, C.S. and Lancaster, N., 2004. Movement of a small slipfaceless dome dune in the Namib Sand Sea, Namibia. Geomorphology, 59: 189–196.

Bullard, J.E. and Nash, D.J., 1998. Linear dune pattern variability in the vicinity of dry valleys in the southwest Kalahari. Geomorphology, 23: 35–54.

Bullard, J.E., Thomas, D.S.G., Livingstone, I. and Wiggs, G.F.S., 1995. Analysis of linear sand dune morphological variability, southwestern Kalahari Desert. Geomorphology, 11: 189–203.

Bullard, J.E., Thomas, D.S.G., Livingstone, I. and Wiggs, G.F.S., 1997. Dunefield activity and interactions with climatic variability in the southwest Kalahari Desert. Earth Surface Processes and Landforms, 22(2): 165–174.

Burkinshaw, J.R., Illenberger, W.K. and Rust, I.C., 1993. Wind speed profiles over a reversing transverse dune. In: K. Pye (Ed.), The Dynamics and Environmental Context of Aeolian Sedimentary Systems. Geological Society, London, pp. 25–36.

Cooper, W.S., 1958. Coastal Sand Dunes of Oregon and Washington. Geological Society of America Memoir, 72: 167.

Corbett, I., 1993. The modern and ancient pattern of sandflow through the southern Namib deflation basin. International Association of Sedimentologists Special Publication, 16: 45–60.

Cupp, K., Lancaster, N. and Nickling, W.G., 2004. A Dune Simulation Wind Tunnel for Studies of Lee Face Processes. Eos, Transactions American Geophysical Union, 85(47, Fall Meeting Supplement): Abstract P31B-0989.

Dong, Z., Wang, T. and Wang, X., 2004. Geomorphology of megadunes in the Badain Jaran Desert. Geomorphology, 60(1–2): 191–204.

Elbelrhiti, H., Claudin, P. and Andreotti, B., 2005. Field evidence for surface-wave-instability of sand dunes. Nature, 437: 720–723.

Ellwood, J.M., Evans, P.D. and Wilson, I.G., 1975. Small scale aeolian bedforms. Journal of Sedimentary Petrology, 45: 554–561.

Embabi, N.S., 1982. Barchans of the Kharga Depression. In: F. El Baz and T.A. Maxwell (Eds.), Desert Landfroms of Egypt: a basis for comparison with Mars. NASA, Washington D.C., pp. 141–156.

Embabi, N.S. and Ashour, M.M., 1993. Barchan dunes in Qatar. Journal of Arid Environments, 25; 49–69.

Endrody-Younga, S., 1982. Dispersion and translocation of dune specialist tenebrionids in the Namib area. Cimbebasia (A), 5: 257-271.

Eriksson, P.G., Nixon, N., Snyman, C.P. and Bothma, J.d.P., 1989. Ellipsoidal parabolic dune patches in the southern Kalahari Desert. Journal of Arid Environments, 16: 111–124.

Ewing, R.C., Kocurek, G. and Lake.L.W., 2006. Pattern analysis of dune-field parameters. Earth Surface Processes and Landforms, 31(9): 1176–1191.

Finkel, H.J., 1959. The barchans of Southern Peru. Journal of Geology, 67: 614–647.

Folk, R., 1970. Longitudinal dunes of the northwestern edge of the Simpson Desert, Northern Territory, Australia, 1. geomorphology and grain size relationships. Sedimentology, 16: 5–54.

Frank, A. and Kocurek, G., 1996a. Airflow up the stoss slope of sand dunes: limitations of current understanding. Geomorphology, 17(1–3): 47–54.

Frank, A. and Kocurek, G., 1996b. Towards a model for airflow on the lee side of aeolian dunes. Sedimentology, 43(3): 451–458.

Fryberger, S., Ahlbrandt, T. and Andrews, S., 1979. Origin, sedimentary features, and significance of low-angle eolian "sand sheet" deposits, Great Sand Dunes National Monument and vicinity, Colorado. Journal of Sedimentary Petrology, 49(3): 733–746.

Fryberger, S. and Goudie, A.S., 1981. Arid Geomorphology. Progress in Physical Geography, 5(3): 420–428.

Fryberger, S.G., 1979. Dune forms and wind regimes. In: E.D. McKee (Ed.), A Study of Global Sand Seas: United States Geological Survey, Professional Paper. U.S.G.S. Professional Paper, pp. 137–140.

Fryberger, S.G. and Ahlbrandt, T.S., 1979. Mechanisms for the formation of aeolian sand seas. Zeitschrift für Geomorphologie, 23: 440–460.

Fryberger, S.G., Al-Sari, A.M., Clisham, T.J., Rizoi, S.A.R. and Al-Hinai, K.G., 1984. Wind sedimentation in the Jafarah sand sea, Saudi Arabia. Sedimentology, 31(3): 413–431.

Fryberger, S.G., Hesp, P. and Hastings, K., 1992. Aeolian granule ripple deposits, Namibia. Sedimentology, 39: 319–331.

Gay, S.P., Jr., 1999. Observations regarding the movement of barchan sand dunes in the Nazca to Tanaca area of southern Peru. Geomorphology, 27(3–4): 279–294.

Glennie, K.W., 1970. Desert Sedimentary Environments. Developments in Sedimentology, 14. Elsevier, Amsterdam, 222pp.
Goudie, A.S., Livingstone, I. and Stokes, S. (Eds.), 1999. Aeolian Environments, Sediments, and Landforms. John Wiley & Sons, Chichester, 325pp.
Greeley, R. and Iversen, J.D., 1985. Wind as a Geological Process. Cambridge University Press, Cambridge, 333pp.
Hack, J.T., 1941. Dunes of the Western Navajo County. Geographical Review, 31(2): 240–263.
Haff, P.K. and Presti, D.E., 1995. Barchan dunes of the Salton Sea region, California. In: V.P. Tchakerian (Ed.), Desert Aeolian Processes. Chapman and Hall, New York,pp. 153–178.
Hanna, S.R., 1969. The formation of longitudinal sand dunes by large helical eddies in the atmosphere. Journal of Applied Meteorology, 8: 874–883.
Hastenrath, S., 1987. The barchan dunes of Southern Peru revisited. Zeitschrift fur Geomorphologie, 31(2): 167–178.
Hastenrath, S.L., 1967. The barchans of the Arequipa region, Southern Peru. Zeitschrift fur Geomorphologie, 11: 300–311.
Havholm, K.G. and Kocurek, G., 1988. A preliminary study of the dynamics of a modern draa, Algodones, southeastern California, USA. Sedimentology, 35: 649–669.
Haynes, C.V.J., 1989. Bagnold's barchan: a 57-yr record of dune movement in the eastern Sahara and implications for dune origin and palaeoclimate since Neolithic times. Quaternary Research, 32(2): 153–167.
Hersen, P., Anderson, K.H., Elbelrhiti, B., Andreotti, B., Claudin, P. and Douady, S., 2004. Corridors of barchan dunes: stability and size selection. Physics Review, E, 69: 011304.
Hersen, P. and Douady, S., 2005. Collision of barchan dunes as a mechanism of size regulation. Geophysical Research Letters, 34(21): L21403.
Hesse, P.P. and Simpson, R.L., 2006. Variable vegetation cover and episodic sand movement on longitudinal desert dunes. Geomorphology, 81: 276–291.
Holm, D.A., 1960. Desert geomorphology in the Arabian Peninsula. Science, 123: 1369–1379.
Howard, A.D., 1977. Effect of slope on the threshold of motion and its application to orientation of wind ripples. Geological Society of America Bulletin, 88: 853–856.
Howell, J. and Mountney, N., 2001. Aeoian grain flow architecture: hard data for reservoir models and implications for red bed sequence stratigraphy. Petroleum Geoscience, 7: 51–56.
Hunt, J.C.R., Leibovich, S. and Richards, K.J., 1988. Turbulent shear flows over low hills. Quarterly Journal of the Royal Meteorological Society, 114: 1435–1470.
Hunter, R.E., 1977. Basic types of stratification in small eolian dunes. Sedimentology, 24: 361–388.
Hunter, R.E., 1985. A kinematic model for the structure of lee-side deposits. Sedimentology, 32: 409–422.
Hunter, R.E., Richmond, B.M. and Alpha, T.R., 1983. Storm-controlled oblique dunes of the Oregon Coast. Geological Society of America Bulletin, 94: 1450–1465.
Inman, D.L., Ewing, G.C. and Corliss, J.B., 1966. Coastal sand dunes of Guerrero Negro, Baja California, Mexico. Geological Society of America, Bulletin, 77: 787–802.
Jackson, P.S. and Hunt, J.C.R., 1975. Turbulent wind flow over a low hill. Quarterly Journal of the Royal Meteorological Society, 101: 929–955.
Jäkel, D., 1980. Die bildung von barchanen in Faya-Largeau/Rep. du Tchad. Zeitschrift für Geomorphologie, N.F., 24: 141–159.
Jensen, N.O. and Zeman, O., 1985. Perturbations to mean wind and turbulence in flow over topographic forms. In: O.E. Barndorff-Nielsen, J.T. Møller, K.R. Rasmussen and B.B. Willetts (Eds.), Proceedings of International Workshop on the Physics of Blown Sand. University of Aarhus, Aarhus, pp. 351–368.
Kar, A., 1993. Aeolian processes and bedforms in the Thar Desert. Journal of Arid Environments, 25: 83–96.
Kocurek, G., 1998. Aeolian System Response to External Forcing Factors – A Sequence Stratigraphic View of the Saharan Region. In: A.S. Alsharan, K.W. Glennie, G.L. Whittle and C.G.S.C. Kendall (Eds.), Quaternary Deserts and Climatic Change. Balkema, Rotterdam/Brookfield, pp. 327–338.
Kocurek, G. and Ewing, R.C., 2005. Aeolian dune field self-organization – implications for the formation of simple versus complex dune field patterns. Geomorphology, 72: 94–105.
Kocurek, G., Havholm, K.G., Deynoux, M. and Blakey, R.C., 1991. Amalgamated accumulations resulting from climatic and eustatic changes, Akchar Erg, Mauritania. Sedimentology, 38(4): 751–772.
Kocurek, G. and Lancaster, N., 1999. Aeolian Sediment States: Theory and Mojave Desert Kelso Dunefield example. Sedimentology, 46(3): 505–516.
Kocurek, G. and Nielson, J., 1986. Conditions favourable for the formation of warm-climate aeolian sand sheets. Sedimentology, 33: 795–816.
Kocurek, G., Townsley, M., Yeh, E., Havholm, K. and Sweet, M.L., 1992. Dune and dunefield development on Padre Island, Texas, with implications for interdune deposition and water-table-controlled accumulation. Journal of Sedimentary Petrology, 62(4): 622–635.
Lancaster, N., 1980. The formation of seif dunes from barchans - supporting evidence for Bagnold's hypothesis from the Namib Desert. Zeitschrift fur Geomorphologie, 24: 160–167.
Lancaster, N., 1982a. Dunes on the Skeleton Coast, SWA/Namibia: geomorphology and grain size relationships. Earth Surface Processes and Landforms, 7: 575–587.
Lancaster, N., 1982b. Linear dunes. Progress in Physical Geography, 6: 476–504.
Lancaster, N., 1983. Controls of dune morphology in the Namib sand sea. In: T.S. Ahlbrandt and M.E. Brookfield (Eds.), Eolian Sediments and Processes. Developments in Sedimentology. Elsevier, Amsterdam, pp. 261–289.
Lancaster, N., 1985. Variations in wind velocity and sand transport rates on the windward flanks of desert sand dunes. Sedimentology, 32: 581–593.
Lancaster, N., 1988. Controls of eolian dune size and spacing. Geology, 16: 972–975.
Lancaster, N., 1989a. Star Dunes. Progress in Physical Geography, 13(1): 67–92.
Lancaster, N., 1989b. The Namib Sand Sea: Dune forms, processes, and sediments. A.A. Balkema, Rotterdam, 200pp.
Lancaster, N., 1991. The orientation of dunes with respect to sand-transporting winds: a test of Rubin and Hunter's gross bedform-normal rule, NATO Advanced Research Workshop

on sand, dust, and soil in their relation to aeolian and littoral processes. University of Aarhus, Sandbjerg, Denmark, pp. 47-49.
Lancaster, N., 1992. Relations between dune generations in the Gran Desierto, Mexico. Sedimentology, 39: 631–644.
Lancaster, N., 1993. Origins and sedimentary features of supersurfaces in the northwestern Gran Desierto Sand Sea. IAS Special Publication, 16: 71–86.
Lancaster, N., 1995a. Geomorphology of Desert Dunes. Routledge, London, 290pp.
Lancaster, N., 1995b. Origin of the Gran Desierto Sand Sea: Sonora, Mexico: Evidence from dune morphology and sediments. In: V.P. Tchakerian (Ed.), Desert Aeolian Processes. Chapman and Hall, New York, pp. 11–36.
Lancaster, N., 1996. Field studies of proto-dune initiation on the northern margin of the Namib Sand Sea. Earth Surface Processes and Landforms, 21: 947–954.
Lancaster, N., 1999. Sand Seas. In: A.S. Goudie, I. Livingstone and S. Stokes (Eds.), Aeolian Environments, Sediments, and Landforms. Wiley, Chichester, New York, pp. 49–70.
Lancaster, N. and Baas, A., 1998. Influence of vegetation cover on sand transport by wind: field studies at Owens Lake, California. Earth Surface Processes and Landforms, 23(1): 69–82.
Lancaster, N., Greeley, R. and Christensen, P.R., 1987. Dunes of the Gran Desierto Sand Sea, Sonora, Mexico. Earth Surface Processes and Landforms, 12: 277–288.
Lancaster, N., Kocurek, G., Singhvi, A.K., Pandey, V., Deynoux, M., Ghienne, J.-P. and Lo, K., 2002. Late Pleistocene and Holocene dune activity and wind regimes in the western Sahara of Mauritania. Geology, 30: 991–994.
Lancaster, N., Nickling, W.G., McKenna Neuman, C.K. and Wyatt, V.E., 1996. Sediment flux and airflow on the stoss slope of a barchan dune. Geomorphology, 17(1–3): 55–62.
Livingstone, I., 1986. Geomorphological significance of wind flow patterns over a Namib linear dune. In: W.G. Nickling (Ed.), Aeolian Geomorphology. Boston, Allen and Unwin, pp. 97–112.
Livingstone, I., 1988. New models for the formation of linear sand dunes. Geography, 73: 105–115.
Livingstone, I., 1989. Monitoring surface change on a Namib linear dune. Earth Surface Processes and Landforms, 14: 317–332.
Livingstone, I., 1993. A decade of surface change on a Namib linear dune. Earth Surface Processes and Landforms, 18(7): 661–664.
Livingstone, I., 2003. A twenty-one-year record of surface change on a Namib linear dune. Earth Surface Processes and Landforms, 28(9): 1025–1032.
Livingstone, I. and Warren, A., 1996. Aeolian Geomorphology: an introduction. Addison Wesley Longman, Harlow, 211pp.
Livingstone, I., Wiggs, G.F.S. and Weaver, C.M., 2007. Geomorphology of desert sand dunes: A review of recent progress. Earth Science Reviews, 80(3–4): 239–257.
Long, J.T. and Sharp, R.P., 1964. Barchan dune movement in Imperial Valley, California. Geological Society of America Bulletin, 75: 149–156.
Mainguet, M., 1983. Dunes vives, dunes fixées, dunes vêtues: une classification selon le bilan d'alimentation, le régime éolien et la dynamique des édifices sableux. Zeitschrift für Geomorphologie, Suppl. Bd. 45: 265–285.
Mainguet, M., 1984a. A classification of dunes based on aeolian dynamics and the sand budget. In: F. El-Baz (Ed.), Deserts and arid lands. Martinus Nijhoff, The Haguw, pp. 31–58.
Mainguet, M., 1984b. Space observations of Saharan aeolian dynamics. In: F. El Baz (Ed.), Deserts and Arid Lands. Nyhoff, The Hague, pp. 59–77.
Mainguet, M. and Callot, Y., 1978. L'erg de Fachi-Bilma (Tchad-Niger). Mémoires et Documents CNRS, 18: 178.
Mainguet, M. and Chemin, M.-C., 1984. Les dunes pyramidales du Grand Erg Oriental. Travaux de l'Institut de Géographie de Reims, 59–60: 49–60.
Marîn, L., Forman, S.L., Valdez, A. and Bunch, F., 2005. Twentieth century dune migration at the Great Sand Dunes National Park and Preserve, Colorado, relation to drought variability. Geomorphology, 70: 163–183.
Mason, P.J. and Sykes, R.I., 1979. Flow over an isolated hill of moderate slope. Quarterly Journal of the Royal Meteorological Society, 105: 383–395.
Maxwell, T. and Haynes, C., 2001. Sand sheet dynamics and Quaternary landscape evolution of the Selima Sand Sheet, southern Egypt. Quaternary Science Reviews, 20: 1623–1647.
Maxwell, T.A. and Haynes, C.V., Jr., 1989. Large-scale, low-amplitude bedfroms (chevrons) in the Selima sand sheet, Egypt. Science, 243: 1179–1182.
McDonald, R.R. and Anderson, R.S., 1995. Experimental verification of aeolian saltation and lee side deposition models. Sedimentology, 42(1): 39–56.
McKee, E., 1982. Sedimentary structures in dunes of the Namib Desert, South West Africa. Geological Society of America Special paper, 188: 60.
McKee, E. and Tibbitts, G.C., Jr., 1964. Primary structures of a seif dune and associated deposits in Libya. Journal of Sedimentary Petrology, 34(1): 5–17.
McKee, E.D., 1966. Structures of dunes at White Sands National Monument, New Mexico (and a comparison with structures of dunes from other selected areas). Sedimentology, 7(1): 1–69.
McKenna Neuman, C., Lancaster, N. and Nickling, W.G., 1997. Relations between dune morphology, airflow, and sediment flux on reversing dunes, Silver Peak, Nevada. Sedimentology, 44: 1103-1114.
McKenna Neuman, C., Lancaster, N. and Nickling, W.G., 2000. The effect of unsteady winds on sediment transport on the stoss slope of a transverse dune, Silver Peak, Nevada. Sedimentology, 47(1): 211–226.
Momiji, H., Carretero-Gonzalez, R., Bishop, S.R. and Warren, A., 2000. Simulation of the effect of wind speedup in the formation of transverse dune fields. Earth Surface Processes and Landforms, 25: 905–918.
Muhs, D.R., 2004. Mineralogical maturity in dunefields of North America, Africa, and Austrlia. Geomorphology, 59(1–2): 247–269.
Muhs, D.R., Bush, C.A., Cowherd, S.D. and Mahan, S., 1995. Source of sand for the Algodones Dunes. In: V.P. Tchakerian (Ed.), Desert Aeolian Processes. Chapman and Hall, New York, pp. 37–74.
Muhs, D.R., Reynolds, R.R., Been, J. and Skipp, G., 2003. Eolian sand transport pathways in the southwestern United States: importance of the Colorado River and local sources. Quaternary International, 104: 3–18.

Mulligan, K.R., 1988. Velocity Profiles measured on the windward slope of a transverse dune. Earth Surface Processes and Landforms, 13(7): 573–582.

Nickling, W.G. and McKenna Neuman, C., 1999. Recent investigations of airflow and sediment transport over desert dunes. In: A.S. Goudie, I. Livingstone and S. Stokes (Eds.), Aeolian Environments, Sediments and Landforms. Chichester, John Wiley & Sons.

Nickling, W.G., McKenna Neuman, C. and Lancaster, N., 2002. Grainfall Processes in the Lee of Transverse Dunes, Silver Peak, Nevada. Sedimentology, 49(1): 191–211.

Nielson, J. and Kocurek, G., 1986. Climbing zibars of the Algodones. Sedimentary Geology, 48: 1–15.

Nielson, J. and Kocurek, G., 1987. Surface processes, deposits, and development of star dunes: Dumont dune field, California. Geological Society of America Bulletin, 99:177–186.

Parsons, D.R., Walker, I.J. and Wiggs, G.F.S., 2004. Numerical modelling of flow structures over an idealised transverse dunes of varying geometry. Geomorphology, 59: 149–164.

Partelli, E.J.R., Schwämmle, V., Herrman, H.J., Monteiro, L.H.U. and Maia, L.P., 2006. Profile measurement and simulation of a transverse dune field in the Lencois Maranhenses. Geomorphology, 81: 29–42.

Pye, K. and Tsoar, H., 1990. Aeolian Sand and Sand Dunes. Unwin Hyman, London, 396pp.

Ramsey, M.S., Christensen, P.R., Lancaster, N. and Howard, D.A., 1999. Identification of sand sources and transport pathways at the Kelso Dunes, California using thermal infrared remote sensing. Geological Society of America Bulletin, 111: 646–662.

Rubin, D.M., 1984. Factors determining desert dune type (discussion). Nature, 309: 91–92.

Rubin, D.M. and Hunter, R.E., 1982. Bedform climbing in theory and nature. Sedimentology, 29: 121–138.

Rubin, D.M. and Hunter, R.E., 1987. Bedform alignment in directionally varying flows. Science, 237: 276–278.

Rubin, D.M. and Ikeda, H., 1990. Flume experiments on the alignment of transverse, oblique and longitudinal dunes in directionally varying flows. Sedimentology, 37(4): 673–684.

Sauerman, G., Rognon, P., Poliakov, A. and Herrmann, H.J., 2000. The shape of the barchan dunes of Southern Morocco. Geomorphology, 36(1–2): 47–62.

Schwämmle, V. and Herrmann, H., 2004. Modelling transverse dunes. Earth Surface Processes and Landforms, 29(6): 769–784.

Seppälä, M. and Linde, K., 1978. Wind tunnel studies of ripple formation. Geografiska Annaler, 60(Series A): 29–42.

Sharp, R.P., 1963. Wind Ripples. Journal of Geology, 71: 617–636.

Sharp, R.P., 1966. Kelso Dunes, Mohave Desert, California. Geological Society of America Bulletin, 77: 1045–1074.

Singhvi, A.K. and Kar, A., 2004. The aeolian sedimentation record of the Thar Desert. Proceedings of the Indian Academy of Sciences (Earth Sciences), 113(3): 371–401.

Slattery, M.C., 1990. Barchan migration on the Kuiseb River Delta, Namibia. South African Geographical Journal, 72: 5–10.

Stokes, S. and Bray, H.E., 2005. Late Pleistocene eolian history of the Liwa region, Arabian Peninsula. Geological Society of America Bulletin, 117(11/12): 1466–1480.

Stokes, S., Goudie, A.S., Ballard, J., Gifford, C., Samieh, S., Embabi, N. and El-Rashidi, O.A., 1999. Accurate dune displacement and morphometric data using kinematic GPS. Zeistchrift für Geomorphologie Supplementbände, 11: 195– 214.

Stokes, S., Maxwell, T.A., Haynes, C.V. and Horrocks, J., 1998. Latest Pleistocene and Holocene sand sheet construction in the Selima Sand Sheet, Eastern Sahara. In: A.S. Alsharan, K.W. Glennie, G.L. Whittle and C.G.S.C. Kendall (Eds.), Quaternary Deserts and Climatic Change. Balkema, Rotterdam/Brookfield, pp. 175–184.

Stokes, S., Thomas, D.S.G. and Shaw, P.A., 1997. New chronological evidence for the nature and timing of linear dune development in the southwest Kalahari Desert. Geomorphology, 20(1–2): 81–94.

Sweet, M.L. and Kocurek, G., 1990. An empirical model of aeolian dune lee-face airflow. Sedimentology, 37(6): 1023–1038.

Sweet, M.L., Nielson, J., Havholm, K. and Farralley, J., 1988. Algodones dune field of southeastern California: case history of a migrating modern dune field. Sedimentology, 35(6): 939–952.

Teller, J.T., Glennie, K.W., Lancaster, N. and Singhvi, A.K., 2002. Calcareous dunes of the United Arab emirates and Noah's Flood: the postglacial reflooding of the Persion (Arabian) Gulf. Quaternary International, 68–71: 297–308.

Thomas, D.S.G. and Leason, H.C., 2005. Dunefield activity response to climate variability in the southwest Kalahari. Geomorphology, 64(1–2): 117–132.

Thomas, D.S.G. and Tsoar, H., 1990. The geomorphological role of vegetation in desert dune systems. In: J.B. Thornes (Editor), Vegetation and Erosion. John Wiley & Sons Ltd., Chichester, pp. 471-489.

Tseo, G., 1990. Reconnaissance of the dynamic characteristics of an active Strzelecki Desert longitudinal dune, southcentral Australia. Zeitschrift für Geomorphologie N.F., 34(1): 19–35.

Tsoar, H., 1974. Desert dunes morphology and dynamics, El Arish (northern Sinai). Zeitschrift für Geomorphologie Supplementband, 20: 41–61.

Tsoar, H., 1983. Dynamic processes acting on a longitudinal (seif) dune. Sedimentology, 30: 567–578.

Tsoar, H., 1984. The formation of seif dunes from barchans – a discussion. Zeitschrift fur Geomorphologie, 28(1): 99–103.

Tsoar, H., 1985. Profile analysis of sand dunes and their steady state significance. Geografiska Annaler, 67A: 47–59.

Tsoar, H., 1986. Two-dimensional analysis of dune profile and the effect of grain size on sand dune morphology. In: F. El-Baz and M.H.A. Hassan (Eds.), Physics of Desertification. Martinus Nyhoff, Dordrecht, pp. 94–108.

Tsoar, H., 1989. Linear dunes – forms and formation. Progress in Physical Geography, 13(4): 507–528.

Tsoar, H. and Møller, J.T., 1986. The role of vegetation in the formation of linear sand dunes. In: W.G. Nickling (Ed.), Aeolian Geomorphology. Allen and Unwin, Boston, London, Sydney, pp. 75–95.

Verstappen, H.T., 1968. On the origin of longitudinal (seif) dunes. Zeitschrift für Geomorphologie N.F., 12: 200–220.

Walker, D.J., 1981. An experimental study of wind ripples. MSc Thesis, Massachusetts Institute of Technology.

Walker, I.J., 1999. Secondary airflow and sediment trasport in the lee of a reversing dune. Earth Surface Processes and Landforms, 24: 437–448.
Walker, I.J. and Nickling, W.G., 2002. Dynamics of secondary airflow and sediment transport over and the lee of transverse dunes. Progress in Physical Geography, 26(1): 47–75.
Wang, X., Dong, Z., Zhang, J. and Qu, J., 2004. Formation of the complex linear dunes of the central Taklimakan sand sea. Earth Surface Processes and Landforms, 29(6): 677–686.
Warren, A., 1972. Observations on dunes and bimodal sands in the Tenere desert. Sedimentology, 19: 37–44.
Warren, A., 1988. The dunes of the Wahiba Sands. In: R.W. Dutton (Ed.), Scientific Results of the Royal Geographical Society's Oman Wahiba Sands Project 1985–1987. Journal of Oman Studies, Special Report 3, Muscat, Oman, pp. 131–160.
Warren, A. and Allison, D., 1998. The palaeoenvironmental significance of dune size hierarchies. Palaeogeography, Palaeoeclimatology, Palaeocology, 137: 289–303.
Wasson, R.J., 1983. Dune sediment types, sand colour, sediment provenance and hydrology in the Strzelecki-Simpson Dunefield, Australia. In: M.E. Brookfield and T.S. Ahlbrandt (Eds.), Eolian Sediments and Processes. Developments in Sedimentology. Elsevier, Amsterdam, Oxford, New York, Tokyo, pp. 165–195.
Wasson, R.J., Fitchett, K., Mackey, B. and Hyde, R., 1988. Large-scale patterns of dune type, spacing, and orientation in the Australian continental dunefield. Australian Geographer, 19: 89–104.
Wasson, R.J. and Hyde, R., 1983a. A test of granulometric control of desert dune geometry. Earth Surface Processes and Landforms, 8: 301–312.
Wasson, R.J. and Hyde, R., 1983b. Factors determining desert dune type. Nature, 304: 337–339.
Wasson, R.J, Rajaguru, S.N. Misra, V.N. Agrawal, D.P. Dhir, R.P., Singhvi, A.K., Kameswara Rao, K.., 1983. Geomorphology, late Quaternary stratigraphy and paleoclimatology of the Thar dunefield. Zeitschrift für Geomorphologie, Supplementband, 45: 117-151.
Weng, W.S., Hunt, J.C.R., Carruthers, D.J., Warren, A., Wiggs, G.F.S., Livingstone, A. and Castro, I., 1991. Air flow and sand transport over sand dunes. Acta Mechanica Supplement, 2: 1–22.
Werner, B.T., 1988. A steady-state model of wind-blown sand transport. Journal of Geology, 98(1): 1–17.
Werner, B.T., 1995. Eolian dunes: computer simulations and attractor interpretation. Geology, 23(12): 1107–1110.
Werner, B.T., 2003. Modeling Landforms as Self-Organized, Hierarchical Dynamical Systems. Predictions in Geomorphology, Geophysical Monograph, (135): 133–150.
Werner, B.T. and Kocurek, G., 1997. Bed-form dynamics: Does the tail wag the dog? Geology, 25(9): 771–774.
Werner, B.T. and Kocurek, G., 1999. Bedform spacing from defect dynamics. Geology, 27(8): 727–730.
Wiggs, G.F.S., 1993. Desert dune dynamics and the evaluation of shear velocity: an integrated approach. In: K. Pye (Ed.), The Dynamics and Environmental Context of Aeolian Sedimentary Systems. Geological Society, London, pp. 37–48.
Wiggs, G.F.S., 2001. Desert dune processes and dynamics. Progress in Physical Geography, 25(1): 53–79.
Wiggs, G.F.S., Livingstone, I., Thomas, D.S.G. and Bullard, J.E., 1994. Effect of vegetation removal on airflow patterns and dune dynamics in the southwestern Kalahari Desert. Land Degradation and Rehabilitation, 5: 13–24.
Wiggs, G.F.S., Livingstone, I. and Warren, A., 1996. The role of streamline curvature in sand dune dynamics:evidence from field and wind tunnel measurements. Geomorphology, 17(1–3): 29–46.
Wiggs, G.F.S., Thomas, D.S.G., Bullard, J.E. and Livingstone, I., 1995. Dune mobility and vegetation cover in the southwest Kalahari Desert. Earth Surface Processes and Landforms, 20(6): 515–530.
Wilson, I.G., 1971. Desert sandflow basins and a model for the development of ergs. Geographical Journal, 137(2): 180–199.
Wilson, I.G., 1972. Aeolian bedforms – their development and origins. Sedimentology, 19: 173–210.
Wilson, I.G., 1973. Ergs. Sedimentary Geology, 10: 77–106.