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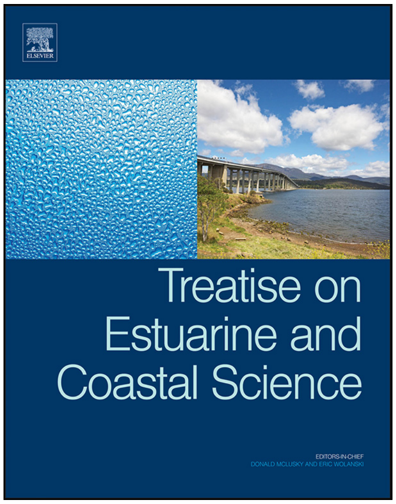

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3.08 Dune Coasts

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Glossary

Blowout A blowout is a saucer-, cup-, bowl-, or trough-shaped depression or hollow formed by wind erosion on a preexisting sand deposit. The adjoining accumulation of sand, the depositional lobe, derived from the depression and possibly other sources, is normally considered part of the blowout landform.

Established foredune Established foredunes develop from incipient foredunes and either occupy the seaward-most position at the rear of the beach or are situated behind an incipient foredune. They are commonly distinguished by the growth of intermediate, sometimes woody plant species, and by their greater morphological complexity, height, width, age, and geographical position.

Foredune (or fore-dune) A foredune is a shore-parallel dune ridge formed on the backshore by aeolian sand deposition within vegetation.

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Foredune plains Foredune plains are plains comprising two or more (typically more) former foredunes. They are formed where progradation has resulted in the multiple formations of foredunes and swales over time.
Incipient foredune An incipient foredune is a new or developing foredune that forms by aeolian sand deposition within pioneer plant communities.
Parabolic dune Parabolic dunes (also termed U-dunes, upsiloidal dunes, and hairpin dunes) are typically U- and V-shaped dunes, characterized by short to elongate trailing ridges, which terminate downwind in U- or V-shaped depositional lobes.
Protodune A bedform that has migrated away from the initial depositional or nucleation site and which does not have a slipface.
Relict foredune A relict foredune is a stable, nonaccretionary, former foredune which has become isolated from accretionary or erosional processes by seaward formation of a new foredune. In some cases, a relict foredune has been termed a 'beach ridge'.
Transgressive dunefield Transgressive dunefields are aeolian sand deposits formed by the downwind or alongshore movement of sand over vegetated to semi-vegetated terrain. Such dunefields may range from quite small (hundreds of meters in alongshore and landward extent) to very large fields that can be similar in size to some small to moderate-sized desert dunefields. Transgressive dunefields may be very large dunefields: they may be low, rolling, blowout-dominated, machair-type dunefields, or they may be low plains dominated by sand sheets, erosional knobs, and nebkha fields.

Abstract

A review of the processes of initiation, evolution, dynamics, geomorphology, and location of coastal dunes is provided. Four principal dune types are covered: foredunes, blowouts, parabolic dunes, and transgressive dunefields. The reasons why dunes first form and where they are formed are initially discussed, and it is concluded that dunes can be created on almost any coast in the world as long as some basic conditions are met and the coast is ice free for at least a small portion of time. In terms of location, foredunes can occur on almost all coasts of the world. Blowouts may occur anywhere as well as in conjunction with the other three dune types. Parabolic and transgressive dunefields are predominantly found on high-energy east and west coasts and in arid regions with a significant sediment supply. The initiation mechanisms vary for each dune type, but vegetation presence or absence is a major factor in all cases. The morphodynamics, landform units, and evolutionary patterns are described for the dune types, and the chapter concludes with a brief review of beach–dune interaction models.

3.08.1 Introduction

This chapter provides a brief review of the processes of initiation, evolution, dynamics, geomorphology, and location of coastal dunes. Four principal dune types are covered: foredunes, blowouts, parabolic dunes, and transgressive dunefields.

3.08.2 Types of Coastal Dunes

While a number of botanical and geomorphological classifications of coastal dunes exist (e.g., Pye, 1983; Tinley, 1985; Pye and Tsoar, 1990), many are cumbersome, difficult to use, and of limited utility due to a variety of factors (e.g., too local or over-complex; e.g., van Dieren, 1934; Olson and van der Maarel, 1989) or are generic and cover a wide variety of dune types (e.g., yellow, gray, and brown dunes). Herein four main coastal dune types are recognized: foredunes, blowouts, parabolic dunes, and transgressive dunefields. This review does not include dunes that have been partly or wholly built, modified, or seriously impacted by humans (see Nordstrom, 1994 for review).

3.08.3 Why Do Dunes Form?

Wherever there is a minimal sand supply to build a beach, there will exist the potential for a sand dune to form, so sand supply is the crucial first factor in determining whether dunes can form or not. The dominant force in driving initial sediment supply has been sea-level change, and for many of the Holocene coastal dunefields of the world, the postglacial marine transgression has been the fundamental driving force. Surf zones, beaches, and barriers have developed firstly where the rising sea level has driven sediment landward from a glacial low stand resulting in gradual coastal barrier and beach translation or rollover (at least in part) Dillenburg and Hesp, 2009a, 2009b). Once present sea level was reached, all the available sediment was utilized to form a surf zone, beach, and backshore. Sand-sized sediment supply then becomes vital. If only limited sand supply was available, a surf zone and beach may be all that formed, but as the sand supply increased, more and larger dunes became possible. As sea levels reached the present (at different times on diverse coasts), sand supply was initially driven by factors such as shoreface slope (Roy et al., 1980; Cowell et al., 1995), the availability of sediment to be reworked/redistributed on the shoreface (Davies, 1980; Thom, 1984; Thom et al., 1985, 1992; Carter, 1988; Orford et al., 1991; Carter and Woodroffe, 1994; Orford et al., 2002), the local and regional supply from rivers (driven by climate and regolith type; Davis, 1978, 1994; Davies, 1980), and wave energy (Short and Hesp, 1982; Miot da Silva et al., 2008; Miot da Silva and Hesp, 2010; Martinho et al., 2008; Dillenburg et al., 2009). Subsequent variations in sand-sized sediment supply to a beach may then depend on subsequent relative sea level (stable, rising, or falling), longshore drift, river supply, continued adjustments to an equilibrium condition on the

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shoreface, and *in situ* production of sediment (e.g., on a carbonate-dominated shelf; Woodroffe, 2002).

The wind energy above a minimum threshold velocity was the next factor crucial for dune development. The higher the wind energy, the greater the potential for dune development, particularly for those coasts with prevailing onshore winds (Jennings, 1957). Several other factors control the actual magnitude of sand transport from a beach (see Sherman et al., 1998; Davidson-Arnott, 2010 for reviews). The degree and type of dune development then depended on the ability of pioneer plants to colonize the backshore and maintain overall stability even during periods of storm erosion, or rapid accretion/progradation, or not. At a very general level, the dominance of plants colonizing and maintaining a presence on the backshore has led to the development of foredunes and foredune plains. Disturbance of the vegetation cover by wind, wave, or climate mechanisms led to the development of other dune types such as blowouts and parabolic dunes. In some cases, a high sediment supply, or climatic conditions (e.g., arid or semi-arid), restricted plant cover, and high wave and wind energy led to the development of mobile (transgressive) dunefields.

3.08.4 Where Do Sand Dunes Form?

Sand protodunes and dunes may arise spontaneously on a beach. Warren (1979) argued that if the wind is above threshold velocity and sand is moving, then a slight variation in the horizontal wind speed alongshore or across-shore may result in the deposition of a strip of sand. Once this initial nucleus of dry sand moves a little, Kocurek et al. (1992: 624) term it a 'protodune' (**Figure 1**); that is, "a bedform that has migrated away from the initial depositional or nucleation site" and which does not have a slipface. Warren (1979) suggested that once the sand strip or protodune is formed, there is a positive feedback between form and flow. The presence of the strip of sand will alter the local wind velocity such that the air flow will be forced to rise slightly over the depositional strip or protodune, decelerate, and deposit more sand. The protodune will therefore grow in height and may become a transverse protodune or dome dune (a dome dune is a discrete, semicircular or oblong, low mound of sand; **Figure 1**). Positive feedback results in the growth of the dune, development of a slipface, and the eventual formation of a single, discrete barchan or a transverse dune (**Figure 1**; Hesp and Arens, 1997; Nield et al., 2010).

Sand dunes may be found in all regions of the world from the southern (the dry valleys of Antarctica) and northern-most latitudes to virtually the equator. However, coastal sand dunes are not particularly common in high latitudes due to the predominance of glacially derived sediment and coarse sediment sizes (Davies, 1980; Short, 1999). Historically, the only significant debate on where coastal dune development was limited, and why, was that relating to the presence or absence of dunes in the tropics. Jennings in 1964 noted that aeolian sand dunes were either very poorly developed or largely absent in the humid tropics and sparked a debate focusing on why this should be so.

Jennings in 1964 noted that dunes were largely absent from the east and west coasts of Malaysia, the Gulf of Papua in New Guinea, and the Guiana coast of South America. In the case of Malaysia, he noted that there was a sufficient sand supply, sand grain size was not a limiting factor, and strong NE monsoon winds at least on the east coast. In general, there are indeed very few dunefields in significant portions of SE Asia. For example, very few sand dunes are present in Thailand. Foredunes are present in a few places, and in some areas beach ridges may have small aeolian caps. There are no dunes in Singapore, and there is very limited dune development in Malaysia.

Jennings questioned whether the beaches dried out sufficiently between tides for sand transport to take place, and

Figure 1 Low (~30–40 cm high) protodunes, crescentic dunes, and barchanoidal transverse dunes which formed within minutes, initially as protodunes, on a flat wet backshore following the onset of strong alongshore winds. Castlepoint, NZ.

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argued that the speed of vegetative colonization of the beach in the tropics might limit the availability of sand for aeolian transport. Subsequently, Bird (1965) argued that the rapid spread of vegetation would not hinder dune development (later accepted by Jennings in 1965) and that sand supply could be restricted by high humidity and the dampness of the beach sand. Bird (1965) then described the parabolic dunefields of the Cape Flattery region in Queensland, Australia, and stated that given the large-scale development of the dunefield, the limiting factor for dune formation in the tropics was related to a lack of sand supply. Note, however, that sand supply is not limited in many parts of Malaysia or Thailand, for example. Swan (1979) indicated that the tall back beach vegetation common to many tropical beaches may reduce wind velocities and hence sediment transport.

Jennings in 1965 analyzed wind regimes around Australia and found that there was significantly less wind energy in the Australian tropics than in adjacent regions. Trenhaile (1997) also suggested that tropical winds are weaker and less persistent than on temperate coasts. However, Wong noted that at times (particularly the NE monsoon) wind action on the east coast of Malaysia was strong enough to form ripples, some aeolian caps on beach ridges, low foredunes, and rarely small blowouts. However, it is common to observe the strongest winds in SE Asia accompanied by rain and this may be a limiting factor in dune development. Swan (1979) indicated that the best-developed dune systems in Sri Lanka, where various types (foredunes, foredune plains, and transgressive dunefields) occur along 300 km of the Sri Lanka coast lying within 10° of the equator, coincided with areas where four or more months were consecutively dry.

As Pye (1983) noted, the factors listed above do not necessarily restrict dune development in the tropics, and a variety of factors acting in combination or individually lead to the presence or absence of dunes. Such factors may be a predominance of weak, variable, or offshore winds (Swan, 1979), low onshore wind energy (Pye, 1983), high wave energy, and/or rainfall during the strongest wind events, coarse sediment and steep reflective beaches with minimum fetch (Short and Hesp, 1982), and a general lack of sand-sized sediment supply.

As will be seen below, there are large-scale mobile and stabilized transgressive dunefields in subtropical Mexico (Vera Cruz coast), and especially in tropical northern Brazil within 2°–3° of the equator (Hesp et al., 2009). Here, as elsewhere in the tropics, the seasonal climate may be the crucial factor in determining whether dunes are present or not. Both regions experience a pronounced dry season, and this will result in a much higher potential for dune development (cf. Swan, 1979) even in areas quite near the equator.

Additionally, the timing of sea-level change and its relationship to dune development may be important in some areas. Some of the NE Queensland dunefields (and others elsewhere in the tropics where fronting reef systems occur) may have been initiated and partly formed under different conditions than now present. Hopley (1983) suggested that there would be a period of up to 1500 years from when the postglacial sea level first reached around the present level (around 6500 years BP in Australia) where wave energy would be higher due to the inability of coral growth to keep up with sea level rise. During this period, beaches would be higher energy and more dissipative with larger fetches, and dune formation would have potentially been greater. This effect is, in part, demonstrated by the Sigatoka dunes in Fiji, where a local break in the reef leads to a high energy, intermediate and dissipative beach, greater sediment delivery, and relatively large-scale parabolic dune development compared to the adjacent coast where a sheltering reef results in reflective and short fetch lagoonal beaches with little dune development.

3.08.5 Foredunes

Foredunes are shore-parallel dune ridges formed on the backshore by aeolian sand deposition within vegetation (Hesp, 2002; **Figure 2**). They have commonly been termed 'embryo dunes' (when relatively new) and 'frontal dunes'.

Figure 2 An incipient foredune and landward established foredune in tropical Yucatan region, Mexico.

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They may be relatively flat terrace-like ridges, especially on rapidly prograding coasts or when forming during periods of rapid progradation. They may also occur as markedly convex ridges. Actively forming foredunes occupy the seaward-most (or foremost) position in a dune system, hence the term, fore-dune, but not all seaward-most dunes are foredunes, particularly on eroding coasts where a variety of other dune types may occupy a foremost position. In addition, foredunes do not commonly or easily form on arid coasts where vegetation is scarce or absent (e.g., Namibia, Baja California, and Mexico).

Foredunes have been classified into a wide variety of types (see e.g., Hesp, 1984a, 1988b, 2002 for review) but generally fall into two main categories, incipient and established foredunes, within which there can be wide morphological and ecological variations. Foredunes can form on any shore: beach (open ocean or semi-enclosed bay), estuary and lake/lagoon (Zenkovich, 1967; Nordstrom, 1992; Nordstrom and Jackson, 1994), and in almost any climatic region (Ruz and Allard, 1994a, 1994b).

3.08.5.1 Incipient Foredunes

Incipient foredunes are new or developing foredunes that form within pioneer plant communities (hence the term 'embryo' dunes used in some European literature). They may be initiated by sand deposition around obstacles such as stranded wrack (seaweed) or log debris and flotsam (e.g., Eamer and Walker, 2010), or within discrete or relatively discrete clumps of vegetation or individual plants (types 1a and 1b of Hesp (1989)), forming nebkha (and attending shadow dunes) (Ruz et al., 2005). In total, such development often comprises an incipient foredune zone (**Figure 3**(**c**)).

Incipient foredunes may also form on the backshore by the laterally continuous alongshore and across-shore growth of pioneer plant seedlings and/or rhizomes (types 2a and 2b of Hesp (1989)) in the wrack line, or typically around the spring high tide zone (**Figures 2** and **3**(**a**)).

The subsequent morphological development of incipient foredunes depends on plant species, density, height and cover, wind velocity, rates of sand transport, and beach progradation rates (Hesp, 2002). Secondary factors such as the rate of occurrence of swash inundation, storm-wave erosion, overwash incidence, the presence of debris and wrack, and resultant wind direction can also be important in determining subsequent dune evolution (Cowles, 1898; Ranwell, 1972; Davies, 1980; Hesp, 1983, 1989; Ruz and Meur-Ferec, 2004; Anderson and Walker, 2006; Anthony et al., 2007; Ruz and Anthony, 2008).

3.08.5.1.1 *Incipient foredune dynamics*

Variations in plant species, density, and distribution strongly influence foredune morphology by affecting the mean wind flow and turbulence within the plant canopy (Mohan and

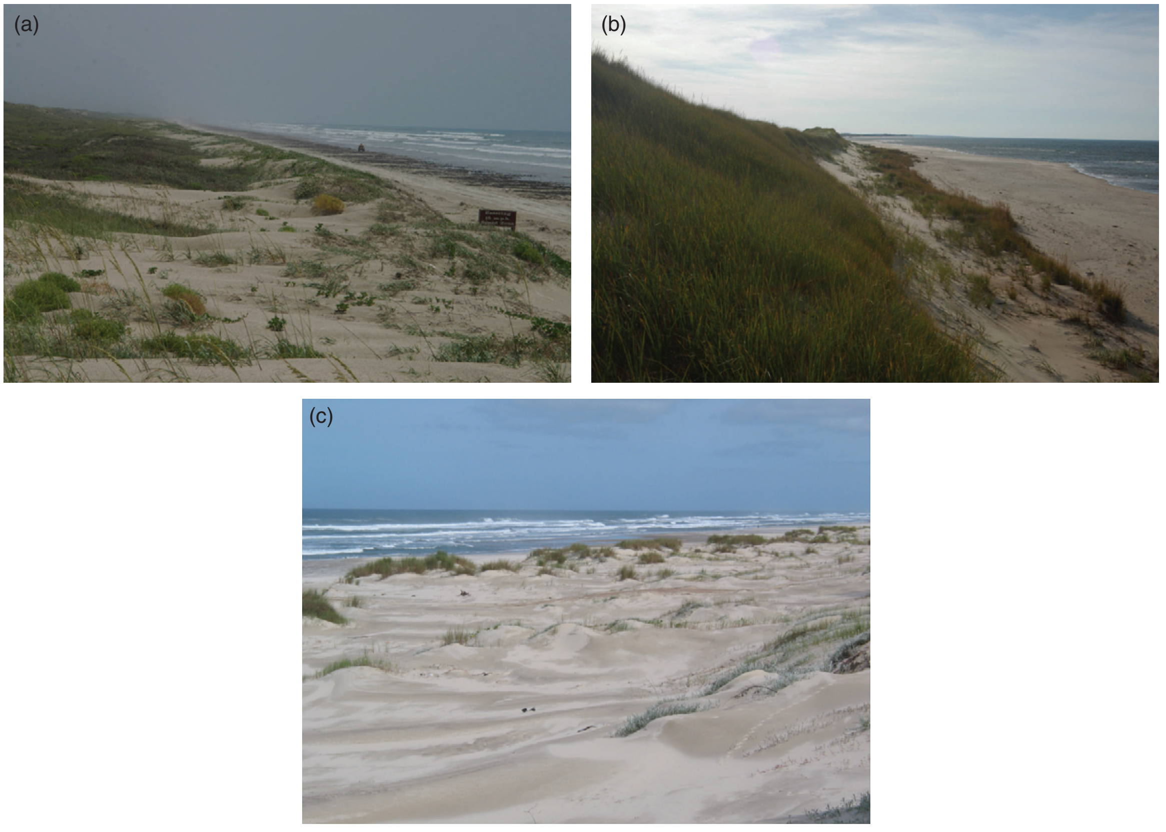

Figure 3 (a) A low incipient foredune and landward established foredune on Padre Island Texas. Multiple small blowouts along the incipient foredune provide conduits for sediment transfer into the landward swale which has developed as a limited deposition zone as the incipient foredune has formed; (b) A high (8–9 m) foredune at Prince Edward Island, Canada. The foredune is regularly scarped and retreats (or translates) landward, then revegetates to various degrees before the next storm event erodes the dune again; (c) A low foredune zone comprising scattered nebkha. Southern Brazil.

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Tiwari, 2004; Poggi et al., 2004). If plants are low prostrate species, their aerodynamic roughness is low and their effect on sand deposition is also low. Many prostrate plant species have limited or low vertical growth potential and may be virtually, or actually, buried during a significant sand transport event. Sediment bypassing of the incipient foredune zone may then result. Tall, dense plant species (e.g., *Ammophila; Uniola*) have a high aerodynamic roughness, significantly greater vertical growth capabilities following sediment deposition and, therefore, trap greater quantities of sand. All other factors being equal, low prostrate plants will tend to be associated with low, wide terrace-type incipient foredune morphologies, while tall, dense plant species will be associated with narrower, ridge morphologies (Hesp, 1983).

The lower the plant density, the less the near-surface and within-canopy drag and the lower and wider the foredune, and vice versa. As plant density increases, the flow drag increases, sand deposition takes place across a shorter downwind distance, and incipient foredunes become shorter and higher. Alongshore variations in both plant density and distribution therefore create alongshore variations in along- and across-dune morphology. Plant density and distribution also varies seasonally and, therefore, seasonal growth rates (low or absent in winter due to snow cover and high in spring) strongly influence patterns of sand transport and deposition on incipient (and established) foredunes (Hesp, 1989; Law and Davidson-Arnott, 1990; Ruz and Allard, 1994a; Trenhaile, 1997; Davidson-Arnott, 2010).

Plant density is increased as wind velocities increase since the effective density of a canopy increases as the vegetation bends and streamlines to the wind. Very high wind velocities may result in canopy changes to such a degree that, for example, wake interference flow (the wakes behind the plants do not fully develop due to interference of neighboring plants) may change to skimming flow (the flow skims across the top surface of the plants) (Carter, 1988; Nickling and Davidson-Arnott, 1990; Harwood, 1993; Wolfe and Nickling, 1993; Green et al., 1998). However, flow penetration increases with increasing velocities, so within-canopy sand transport can also be greater depending on the plant density.

3.08.5.1.2 Incipient foredune morphologies

Incipient foredunes typically adopt one of three forms: ramps, terraces, and ridges (Hesp, 2002). They may stay in this form until isolated by the formation of a new foredune to seaward, or evolve from one type to the next with time.

Ramps may form where seedlings have germinated or grow seaward on a backslope slope, and/or where plants germinate or grow on a scarp fill, or at the base of a foredune scarp which gradually accretes. Terraces form particularly where rapid plant growth takes place across the backshore, often in association with rapidly prograding beaches, where seaward plant growth roughly matches the accretion rate, or plants grow across, where scattered nebkha's form the incipient foredune zone, or on a backshore experiencing little sand accretion (e.g., such as on some tropical, or low energy shores). If the plant density is moderate, or the dominant plant species height is relatively short, then terraces may be the typical morphology. Ridges form where sand deposition preferentially takes place in the seaward portion of the plant canopy, where plant density is high, where seaward plant growth rates are slow relative to accretion rates, and/or where plant height is significant (Hesp, 1983, 1989, 2002). Ridge morphology will be more pronounced where dunes are initiated on the crest of spring tide swash bars adjacent to cut-off ponds (Hine, 1979). Wave scarping of foredunes results in the relocation of aeolian deposition to the scarp base, subsequent scarp filling, and crestal deposition (Carter, 1977; Carter et al., 1990). The latter process also leads to the development of ridge morphologies.

Swales (lee dune depressions and slacks) are typically created by seaward accretion of a foredune (**Figures 2** and **3(a)**). Swales develop as low to limited aeolian deposition zones and become deeper as seaward incipient foredunes become higher (Hesp, 1983, 1984a, 1984b, 1989; Gares and Nordstrom, 1988). They are commonly continuous alongshore where winds are predominantly onshore and the vegetation cover of the incipient foredune is moderate to high. However, where the vegetation cover is low, or more irregular, or a variety of off-, along-, and onshore winds occur, swales may be more morphologically complex alongshore and may be discontinuous. Swales may also form as linear ponds trapped behind accreting swash bars and berms where bars (or sand waves) are extending alongshore and locking onto shores some distance from the previous beach (Hine, 1979; Davidson-Arnott and Fisher, 1992; see figure 10.14 in Davidson-Arnott (2010)).

3.08.5.2 Established Foredunes

Established foredunes develop from incipient foredunes and either occupy the seaward-most position at the rear of the beach or are situated behind an incipient foredune. They are commonly distinguished by the growth of intermediate, sometimes woody plant species, and by their greater morphological complexity, height, width, age, and geographical position (Hesp, 2002; **Figure 3**). In some arid regions, there may be a single, mono-specific species present on the foredune. Foredunes may be found on virtually all coasts of the world except the very driest and/or coldest.

The morphological development and evolution of established foredunes depends on the same factors that control or drive incipient foredune development, for example, sand supply, variations in fetch, plant cover, plant species, the frequency and magnitude of wave and wind forces, the occurrence and magnitude of storm erosion, dune scarping and overwash processes, the medium to long-term beach or barrier state (stable, accreting, or eroding), surfzone-beach type, and human interference and use (e.g., Short and Hesp, 1982; Hesp, 1988a, 1988b, 1991; Thom and Hall, 1991; Davidson-Arnott and Fisher, 1992; Sherman and Bauer, 1993a; Ruz and Allard, 1994a; Davidson-Arnott and Law, 1996; Arens, 1997; McLean and Shen, 2006; Lynch et al., 2008; Ruz and Anthony, 2008; Delgado-Fernandez and Davidson-Arnott, 2009; Davidson-Arnott, 2010). Development and evolution may also be strongly dependent on the morphology and plant cover of the original incipient foredune.

The height of foredunes can range from very low scattered dunes of less than a meter or so in height on barrier islands dominated by overwash and in areas of limited sediment supply (e.g., on some estuarine shores; Nordstrom and Jackson, 1994; **Figure 3(c)**) to 10–20 m in some instances, for example, where the sediment supply is high or where the sediment

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budget is slightly negative and the foredune is translating or rolling landward (**Figure 3**(**b**)).

3.08.5.2.1 Foredune evolution and types

Hesp (1982, 1988b) classified foredunes into five morpho-ecological stages (cf. Carter, 1988), while Arens (1994) and Arens and Wiersma (1994) classified foredunes according to their longer-term temporal state of accretion, stability, or erosion. Hesp (2002) combined elements of these and other models (e.g., Hosier and Cleary, 1977; Cleary and Hosier, 1979; Godfrey et al., 1979; Psuty, 1988, 1989; McCann and Byrne, 1989, 1994; Ritchie and Penland, 1990; Saunders and Davidson-Arnott, 1990; Barrere, 1992; Ruz and Allard, 1994a, 1994b) to produce a five-stage foredune classification with additional temporal sequences.

The Hesp (2002) foredune classification provides a guide to: (1) the morphological form a foredune may take as it grows (stages 1–5) and (2) the evolutionary stages through which a foredune may progress from 1 (highly stable, well vegetated, morphologically simple) to 5 (highly erosional remnant knobs and sand sheets/blowouts).The evolutionary progression may be reversed if certain conditions occur, such as revegetation and stabilization. In the case of (1) above, a stage 1 foredune may evolve from a well-vegetated, laterally continuous, incipient foredune. A stage 5 foredune could evolve from an originally chaotic arrangement of small nebkha which formed the initial incipient foredune zone into a highly irregular collection of large nebkha and shadow dunes (**Figures 3** and **4**).

Mild to catastrophic change may be initiated by minor to severe wave scarping and/or overwash or by intense wind storms (Davidson-Arnott and Fisher, 1992; Vasseur and Hequette, 2000; Hesp and Martinez, 2007; Vespremeanu-Stroe and Preoteasa, 2007; Davidson-Arnott, 2010). The subsequent development or evolution of the foredune (and its stage) depends on the degree of foredune scarp fill and ramp development, revegetation, and re-establishment or landward migration (or translation), and the degree and spacing of subsequent storm events (e.g., Hosier and Cleary, 1977; Tsoar, 1983; Carter and Stone, 1989; McCann and Byrne, 1989; Borowka, 1990a; Carter, 1990; Carter et al., 1990; Christiansen et al., 1990; Nickling and Davidson-Arnott, 1990; Saunders and Davidson-Arnott, 1990; Morton, 2002; Christiansen and Davidson-Arnott, 2004; Claudino-Sales et al., 2008; Ruz and Anthony, 2008; Davidson-Arnott, 2010).

Figure 4 Pronounced sand suspension over an established foredune at Mason Bay, Stewart Island, NZ during a gale. The Ammophila canopy is markedly streamlined, and fine sand is transported far downwind during such conditions. Photograph courtesy of Phil Peterson and Michael Hilton.

3.08.5.2.2 Flow dynamics and sand transport

The wind flow is topographically accelerated over foredunes, particularly up stoss slopes and over crests (Arens et al., 1995; Hesp et al., 2005, 2009; Walker et al., 2009b) in similarity to many other studies of speedup over simple ridges (e.g., Jackson and Hunt, 1975; Rasmussen, 1989; Wilson et al., 1998; Kim et al., 2000; Parsons et al., 2004; Takahashi et al., 2005; Cao and Tamura, 2006). However, the variable vegetation cover of foredunes and their topographic variability leads to local deceleration and variation in roughness length (Arens et al., 1995; Hesp et al., 2005). Such variations become more pronounced as foredune morphological complexity increases and the vegetation cover becomes more variable (Hesp, 1988b). Very dense and tall vegetation results in intensely nonlogarithmic velocity profiles (Harwood, 1993). Estimation of friction velocities and roughness lengths is often difficult and may be both over- and underestimated if derived from beach measurements, depending on the changes in roughness and foredune morphology (Walker et al., 2009a, 2009b).

Patterns of sand deposition and erosion on foredunes are very strongly influenced by wind velocity and vegetation density and cover (Carter, 1977; Hesp, 1983, 1984a, 1988b; Sarre, 1989; Wolfe and Nickling, 1993; Arens et al., 1995; Arens, 1996a, 1997; Trenhaile, 1997). Under low (but above threshold) wind speeds, sediment deposition principally occurs on the lower seaward-most portion of the stoss face (or dune toe) where the toe is relatively well vegetated (Davidson-Arnott, 1988; Hesp, 1988b; Sarre, 1989; Davidson-Arnott and Law, 1990, 1996; Gares, 1992; Ruz and Allard, 1994a; Arens, 1996a). Substantial deceleration occurs upslope as the vegetation exerts greater and greater drag. As the vegetation density across a foredune decreases, sand is transported further up stoss faces, and this can increase as foredune height and/or steepness increases due to increased flow speedup or acceleration up steeper and/or higher slopes (Arens, 1996a; Hesp et al., 2009c; Walker et al., 2009a). High wind speeds combine with topographic acceleration and jet flow over dunes to significantly increase wind flow penetration of the plant canopy. The within-canopy wind flow switches from a general condition of speed-down (i.e., increasing deceleration) upslope to speedup (i.e., increasing acceleration) upslope. This enables both saltation and suspension to occur, and saltation may be enhanced by sand grains bouncing across the streamlined vegetation surface (Hesp et al., 2009). On established foredune types 4 and 5 of Hesp (1988b) or Fd and Fe of Short and Hesp (1982) and Carter (1988), the vegetation cover is low, and localized transport and deposition may occur at various locations across the dunes (Hesp, 1988b).

Figure 4 illustrates massive sand suspension taking place across a foredune at Mason Bay in the roaring forties, Stewart Island, New Zealand (NZ) under gale force winds (approximated at $+15\,\mathrm{m\,s^{-1}}$ at 1 m above the beach surface). Under these conditions, sediment is transported some distance over

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the foredunes to lee slopes and downwind to other dunes. Flow separation is common over foredune crests and lee slopes. The presence of the separation envelope, reversing flow, and significantly reduced wind velocity facilitates sand deposition leeward of the dune crest.

Offshore winds are not uncommon on many coasts and may in some regions be the dominant wind (e.g., Gares and Nordstrom, 1988). Offshore winds may also result in the closure of blowouts and an increase in topographic variability (Svasek and Terwindt, 1974; Nordstrom and Jackson, 1993; Wal and McManus, 1993; Arens et al., 1995; Lynch et al., 2009). Such winds may transport sand from lee slopes and crests where the vegetation cover is low onto the seaward foredune slope. In some cases, flow separation and steering of offshore winds over the foredune may result in beach sand being transported along the beach and into the foredune toe (Lynch et al., 2009, 2010). The regional approach wind direction is important in determining the degree of sand transport onto rather than along the foredune (Svasek and Terwindt, 1974; Rasmussen, 1989; Arens et al., 1995; Hesp, 2002; Walker et al., 2006).

On all foredunes, actual sediment transport varies widely depending on moisture levels, rainfall, and fetch, varying with perpendicular versus oblique winds (Bauer and Davidson-Arnott, 2002; Walker et al., 2006, 2009a, 2009b; Delgado-Fernandez and Davidson-Arnott, 2009; Delgado-Fernandez, 2010; Nield et al., 2010), vegetation cover and type, the presence of flotsam, debris and wrack (Hesp, 2002; Eamer and Walker, 2010), ice and snow cover, and the importance of niveo–aeolian processes (Law and Davidson-Arnott, 1990; Davidson-Arnott, 2010), wind direction, lake and sea/storm levels, and local variations in sediment supply (e.g., the presence and migration of sand waves) (Davidson-Arnott and Fisher, 1992; Wal and Mcmanus, 1993; Nordstrom and Jackson, 1994; Ruz and Allard, 1994a, 1994b; Arens, 1996b, 1997; Davidson-Arnott, 2010). In addition, the seasonal dominance of wave over wind processes may significantly alter the sediment supply to the foredune (Arens, 1996b; Delgado-Fernandez and Davidson-Arnott, 2009). The presence of wave cut scarps alters the wind flow at the dune toe and scarp fill processes then dominate (Carter et al., 1990; Isla and Espinosa, 1995; Christiansen and Davidson-Arnott, 2004; Hesp et al., 2009).

3.08.5.2.3 Medium- to long-term evolution of foredunes

In the medium to long term, foredune evolution may take several paths, including (1) gradual building *in situ*, (2) partial or complete destruction by wind and/or wave processes, followed often by reformation, (3) mild to pronounced progradation and isolation of the foredune with new foredunes forming to seaward (see below). In the first case, a single foredune may be all that is formed over the past hundred to several thousand years, depending on the length of time that the Holocene sea level has been at or near present (e.g., Nickling and Davidson-Arnott, 1990). The foredunes may therefore constitute the entire Holocene barrier and are termed ‘stationary barriers’ by Roy et al. (1994). In fact, the foredune may undergo various levels of wave scarping and wind erosion, recover to various degrees, and might comprise a foredune-blowout complex. Foredunes may be severely eroded and/or overwashed during large storms, particularly where they are low and situated on retrograding (eroding) barriers (e.g., Leatherman, 1979a, 1979b; Penland et al., 1988; Claudino-Sales et al., 2008; Pries et al., 2008). They may redevelop in place, some distance landward, be replaced by scattered nebkha fields (see below), or removed entirely (Hesp, 2002). Foredunes are perfectly capable of translating landward as sea level rises (Davidson-Arnott, 2005; **Figure 3**(**b**)).

3.08.6 Foredune Plains (and Beach Ridges)

Over time, and usually on prograding coasts, foredunes may become isolated from accretion and erosion processes by the seaward development of a new incipient foredune (e.g., McLean and Shen, 2006). The original foredune thus becomes a relict foredune (Hesp, 1999). Such relict foredunes may gradually develop into so-called gray or brown dunes over time as soil development takes place (e.g., Salisbury, 1952; Ranwell, 1972; Provoost et al., 2004). Gray dunes are defined as fixed and semifixed dunes with herbaceous vegetation (Houston, 2008). Brown dunes are more typically stable dunes with heath vegetation (e.g., Isermann, 2008). Thus, the terms gray and brown dunes are generic terms and can cover any type of dune that has been relatively stable for some time.

Beach progradation may lead to the development of foredune plains (**Figure 5**). Such individual ridge and swale couplets or plains are also commonly known as ‘beach ridges’ and ‘beach ridge plains’, respectively (Hesp, 1984a, 1984b, 2006). Such plains may develop during sea-level transgression, regression, or stability (Woodroffe, 2002; Dillenburg and Hesp, 2009a, 2009b; Nielsen and Johannessen, 2009). Firth et al. (1995) showed the rapid development of a foredune/beach ridge plain in NE Scotland to have been related to the ready availability of glacigenic sediment as sea level fell. However, they seldom develop as a function of regular sea-level rise and fall and associated scarping as preferred by Bird (1990) for his parallel dunes or due to (unproven) 20- to 60-year variations in sea level as preferred by Tanner and Stapor (1972).

Foredune-like sand ridges have commonly been termed ‘beach ridges’ regardless of genetic origin, and the term ‘beach ridge’ has been commonly applied to sand ridges when the genesis of the ridge is actually unknown. In many countries, the term beach ridge has often been used interchangeably with foredune and some definitions actually describe both foredunes and beach ridges as the same landform (e.g., Komar, 1976; Armstrong Price, 1982; Otvos, 2000). For example, Armstrong Price (1982: 160) stated that a beach ridge “may originate immediately back of the active beach as a flood-level ridge commonly of the coarser beach materials, or it may form as an aeolian accumulation caught in the vegetation immediately back of the beach proper.” He further stated that where excess amounts of sand are blown onto the ridge, it becomes a foredune.

Berms have also been described as beach ridges (King, 1972; Carter, 1986; Viles, 1988), but a berm is a shore-parallel, nonpersistent wave-built ridge or terrace formed at the limit of swash run-up (Short, 1999; Hesp, 1999, 2006). Every high tide may alter a berm to some degree, and storms usually erode or destroy fair-weather or modal surf-zone-type berms, but may build higher storm berms in some cases. In essence, it

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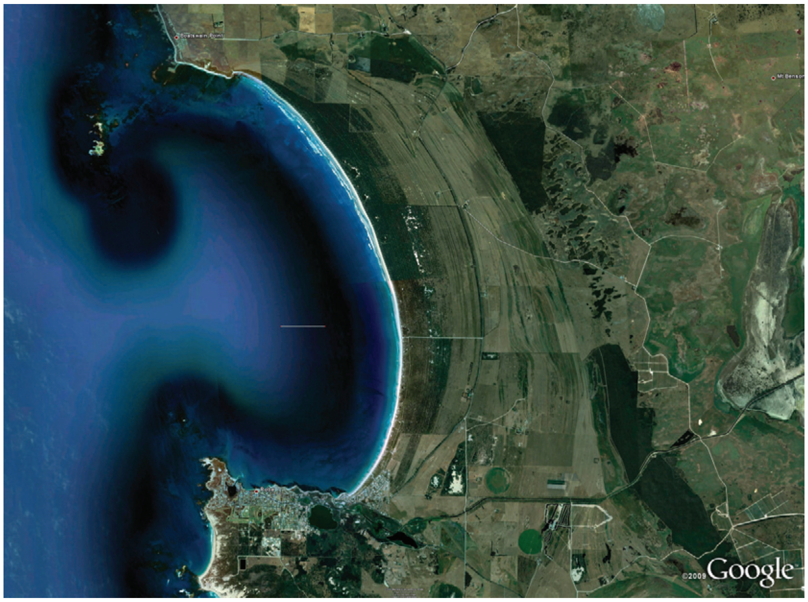

Figure 5 Aerial view of a foredune plain (beach ridge plain) at Robe, South Australia. Each dune ridge and swale couplet is formed by aeolian sediment deposition in pioneer vegetation on the backshore. The white bar (on the ocean) is 1 km long. Source: Google Earth.

could be argued that any beach ridge is a ridge formed at a beach, and, therefore, any ridge found at, or near, the beach, could be termed a beach ridge. The problem with this is that the genesis of the ridge is ignored.

Beach ridges may be formed by wave action and commonly by storm waves, particularly where the sediments are coarse sand to boulders (Redman, 1864; Johnson, 1919; Lewis, 1932; Guilcher, 1958). Chesil Beach in England is a classic example. Reineck and Singh (1975: 291) defined a beach ridge as "a continuous linear mound of rather coarser sediment near the high water line" which develops during "storms and exceptionally high waters." Beach ridges may also form where nearshore bars gradually build upward and become high spring tide berms. They may be preserved where the beach is prograding (Hine, 1979). In lakes, and modally low-energy environments, low ridges may be formed in medium to fine sand by waves. Mason (1993: 57) stated that a beach ridge "is a clastic storm-related facies, deposited beyond the influence of subsequent lower sea levels prevailing during fair-weather conditions." Hesp (2006) argued that foredunes are genetically and morphodynamically distinct from beach ridges, and that the term beach ridges should only be used to describe those ridges formed by wave processes. Foredune plains have often been interpreted as beach ridges because they appear on aerial photography to be very regularly spaced, mimic the beach plan shape (**Figure 5**), and are smooth, morphologically simple ridge forms. Some would argue that if the ridges were formed as aeolian dunes, the morphology would be less regular and they would not parallel the beach line.

Davies (1957) initially stated that beach ridges are aligned or are parallel with the beach plan shape because the beach berm follows that beach plan shape. Some foredune ridges are, in fact, formed atop spring tide berms as demonstrated by Hine (1979). However, it is not necessary to invoke a berm nucleus initiation to explain the regular, parallel nature of beach ridges or foredunes, or argue that the ridges must therefore be wave deposited. Where the foredune is initiated by seeds deposited within, and then germinating and growing within a wrack line, the seedlings are most commonly at the height of spring tide swash. They are thus aligned conformably with the beach outline (or plan shape) and the direction of prevailing swells. Where the foredune is initiated by rhizome/stolon or shoot growth, or where seeds are scattered across the backshore by winds, the seaward extent of vegetative growth is controlled by spring tide swash and storm wave inundation (Thom, 1964; Davies, 1977, 1980; Hesp, 1983). It is the occasional wave swash event, or wave scarping of the foredune on the uppermost backshore that either (1) restricts seaward growth of plants, (2) forms a shore parallel edge to the vegetation by killing any plants that grow across the spring tide or storm water line, or (3) trims or scarps the foredune and aligns it with the beach plan shape (Hesp, 2002).

While many beach-ridge systems appear to be very regularly spaced, and morphologically simple on aerial photography, detailed ground surveys usually reveal variations in ridge crest to crest spacing, swale depths, and ridge morphology (Hine, 1979; Hesp, 1984b; Shepherd, 1990). **Figure 5** clearly shows variation in ridge height and spacing across a relict foredune plain (see also Murray-Wallace et al., 2002), these variations being principally due to variations in rate of beach progradation over time and the occasional impact of storm wave scarping/erosion. Note that soil development processes, the activities of animals, and sheet and slope wash processes are important in smoothing the morphology of the ridges over time.

While some beach ridges, particularly those comprised coarse sediments, have formed via wave processes during storms or high water-level events, and some have formed via accretion of bars, many have formed in exactly the same way as foredunes form – that is, via sand accretion in pioneer plants on the backshore.

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3.08.6.1 Location

Foredune plains are common around the world including India (e.g., Cox's Bazaar), east coast, South Africa (Weisser and Backer, 1983), Ireland (Carter and Wilson, 1990), SE, E and portions of the central W coast, Australia and NZ (Hesp, 1984a; Thom et al., 1985; Shepherd, 1990; Sanderson and Eliot, 1996; Murray-Wallace et al., 2002; Shepherd and Hesp, 2003), England (Salisbury, 1952; Ranwell, 1972), Scotland (Firth et al., 1995; May and Hansom, 2003), Spain, Portugal, Po delta region and the adjacent Ravenna coast, Italy (e.g., Menez, 1977; Carter, 1990), the Viareggio region, Italy, the USA Gulf and east coasts (particularly from Florida to Georgetown, North Carolina (e.g., Nordstrom and Jackson, 1994; Otvos, 1995), the Great Lakes (e.g., Long Point, Point Pelee, Lake Erie, Canada, Sleeping Bear Dunes region, Lake Michigan), Sweden (known as 'sand hammaren'), and Denmark (Wallen, 1980; Nielsen et al., 2006), significant portions of the west, and some bays and spits on the NE coast of India, from Chur to Pasni in Pakistan, Danube Delta (Preoteasa et al., 2009), the Mersin to Seyhan River region, Turkey, the Tasucu-Kum peninsula, Turkey, Brazil central east coast (e.g., Dillenburg and Hesp, 2009a, 2009b), Vietnam, the Sea of Azov, particularly the north coast, the Black Sea (e.g., Navodari-Tulcea-Danube delta region; Kinburn Bay region), lower E margin and W coast of the Caspian Sea (e.g., Bandar Torkaman, Gilazi) (Zenkovich, 1967), and arctic and subarctic coasts (Black, 1951; Reinson et al., 1979; Martini, 1986; Rampton, 1988; Hequette and Ruz, 1991; Mason and Jordan, 1993; Ruz, 1993; Ruz and Allard, 1994a, 1994b).

3.08.7 Blowouts

A blowout is a saucer-, cup-, bowl-, or trough-shaped depression or hollow formed by wind erosion on a preexisting sand deposit. The adjoining accumulation of sand, the depositional lobe, derived from the depression and possibly other sources, is normally considered part of the blowout landform (Glen, 1979; Carter et al., 1990; Hesp, 2002).

3.08.7.1 Morphologies and Types

Blowout morphology may be highly variable. Ritchie (1972) defined four types of blowouts including cigar-shaped, V-shaped, scooped hollow, and cauldron and corridor. Smith (1960) noted that blowouts may range from pits to elongated notches, troughs, or broad basins. Cooper (1958, 1967) defined two primary types, namely saucer and trough blowouts, to which may be added the bowl blowout. Although there exists a wide range of types in aeolian environments, many blowouts can be classified into one of these three types. Saucer blowouts are semicircular or saucer-shaped and often appear as shallow dishes. Trough blowouts are generally more elongate, with deeper deflation floors and basins, and with steeper, longer erosional lateral walls or slopes. Bowl blowouts are a deeper cup- or bowl-shaped type and may often evolve from saucer blowouts (**Figure 6**).

In nature, there is a large degree of spatial and temporal variability in blowout morphologies. The initial shape, size,

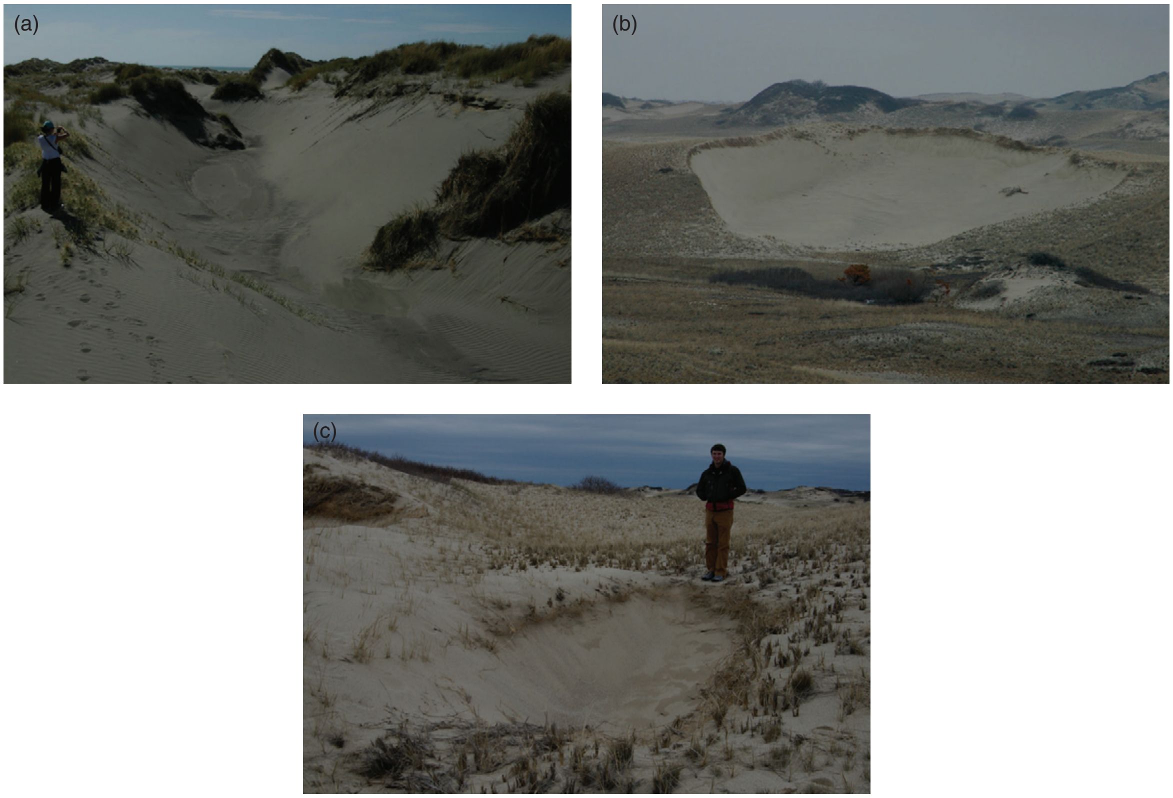

Figure 6 (a) A trough blowout in the Manawatu region, NZ. Note the steep lateral erosional walls and pronounced deflation basin. (b) A very large and deep bowl blowout at Cape Cod, USA. (c) A small saucer blowout at Cape Cod, USA.

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and location of blowouts and their subsequent development may depend on several factors. For example, Smith (1960) noted that blowouts initiated on the broad crests of foredunes are generally shallow saucer types, while those initiated on steep stoss faces are generally elongate trough types (cf. Carter et al., 1990). Once initiated, the subsequent morphological development may depend on the size of the initial constriction (Hesp, 2000), the height and width of the dune in which the blowout is developing, the degree and type of vegetation cover (Melton, 1940; Esler, 1970; de Castro Lopo, 1979), the magnitude of regional winds (Jennings, 1957; Cooper, 1958; Marta, 1958; Davies, 1980), and the degree of exposure to winds from various directions (Hogbom, 1923; Jennings, 1957; Gares et al., 1979; Gares and Nordstrom, 1987).

Blowouts occur in almost every location where there is sandy or dune terrain and are present in both coastal and desert settings (Cowles, 1898; Carter et al., 1990). Blowouts are found in deserts as wind-scoured gaps in transverse dunes (e.g., Bagnold, 1941), as preparabolic dunes (e.g., Melton, 1940; Verstappen, 1968), on linear dunes (e.g., Eriksson et al., 1989), and on many other dune types (Hack, 1941; Hugenholtz and Wolfe, 2009). They have also been described from sub-Antarctic and glacial environments (e.g., Adamson et al., 1988; Seppala, 2004).

3.08.7.2 Initiation

Blowouts are common in coastal dune environments, particularly where beaches are eroding and/or receding, but also in high-energy wind and wave environments.

Blowouts may be initiated in a variety of ways including:

1. Erosion of the dune stoss slope or toe by wave erosion followed by wind erosion of the unvegetated portion of the slope.
2. Wind erosion of poorly vegetated patches either on the foredune or along the scarp crest (Andrews, 1916; Laing, 1954; Ritchie, 1972; Carter, 1990). Erosion may occur following dieback or reduction in the vegetation cover, for example, due to nutrient depletion, climate change, a period of aridity/dryness, wave erosion, high sediment deposition, animal activity, and localized aridity (Onno, 1933; Smith, 1964; Jelgersma and van Regteren Altena, 1969; Ritchie, 1972; Jungerius et al., 1981; Martini, 1981; Ahlbrandt et al., 1983; Rutin, 1983; Jungerius and van der Meulen, 1988; Pluis, 1992; Thom et al., 1992; Thom et al., 1994).
3. Local flow acceleration between topographic highs on a dune (Hesp, 1982), over cliffs, or through arcuate concave topography and depressions (e.g., a slump in a foredune stoss face) (Esler, 1970; Ranwell, 1972; Goldsmith, 1978; Carter, 1988; Gares and Nordstrom, 1995; Hesp, 2002).
4. Wind erosion of overwash hollows and fans (Leatherman, 1976, 1979a, 1979b; Godfrey et al., 1979; Ritchie and Penland, 1990).
5. Rainwater erosion including rilling, sheet wash, and gullying followed by wind erosion (Jungerius and van der Meulen, 1988).
6. High velocity wind erosion, sand inundation and burial. Very high velocity winds and hurricanes may remove, or undermine, erode and/or bury vegetation and initiate blowouts (Marta, 1958; Bird, 1974; Hesp, 1996).
7. Activities of humans including pedestrian trampling and track creation (Parker, 1975; Quinn, 1977; Mather and Ritchie, 1978; Carter, 1988; Bate and Ferguson, 1996), forest felling (Jelgersma and van Regteren Altena, 1969), off-road 4WD activity (Brodhead and Godfrey, 1977; Godfrey et al., 1978; van der Merwe, 1988), housing and resort development (Nordstrom and McCluskey, 1984), sand extraction and military training (van der Maarel, 1979; Marston, 1986), and fires (Tinley, 1985).

3.08.7.3 Flow Dynamics

Saucer and bowl blowouts are characterized by complex flow patterns with significant flow reversal over the rims, strong topographic steering, and positive morphodynamic interactions such that flow accelerations may occur across the deflation basin and around the upper erosional walls, topographic accelerations over the depositional lobes, and flow separation over the lee slopes (Hails and Bennett, 1981; Hesp, 2002; Hugenholtz and Wolfe, 2009). Rapid deceleration downwind of either end (upwind or downwind depending on wind direction) of saucer blowouts leads to the development of short, wide, radial depositional lobes.

In deeper bowl (or cup) blowouts, complex helicoidal and swirling flow within the basins appears to be common (Hesp and Walker, in press). Saucer blowouts commonly migrate upwind due to the strong flow reversals and scarp wall slumping, and deepen to a maximum level determined by the water table, an underlying hard substrate such as beach gravel, shell, wood or coarse sand lags, or indurated layers and calcrete in carbonate-dominated environments. They have one, two, or sometimes an encircling depositional lobe depending on local wind conditions and due to the general openness of saucer and bowl blowouts (Harris, 1974; Jungerius et al., 1981; Jungerius, 1984; Lancaster, 1986; Gares and Nordstrom, 1987, 1995; Jungerius and van der Meulen, 1989; van der Meulen and Jungerius, 1989; Pluis, 1992; Hesp, 2002).

Trough blowouts are characterized by high speed and jet flows up the blowout axis and along the steep marginal walls, together with topographic accelerations and jet flows over the upper depositional lobe. Flow separation is common over the crests of the erosional walls and over the lee slope of the depositional lobe (Hesp, 2002). The topographically accelerated flow and common presence of jets in trough blowouts results in maximum erosion potential along the deflation basin floor and to a lesser extent along the blowout walls. Erosion occurs down to a base level such as the seasonally lowest water table level, a calcrete, or other cemented, indurated, or armored surface such as a gravel basement, shell, pumice, or human occupation surface (Carter, 1976; Hesp and Thom, 1990). Maximum sand transport occurs along the middle axis of the trough blowout depositional lobe and decreases radially away from the central axis. This flow structure leads to the development of arcuate or parabolic-shaped depositional lobes.

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There is very significant topographic steering of wind within all blowout types, but especially through trough blowouts (Hesp and Pringle, 2001). The degree and complexity of steering is fundamentally dependent on the blowout topography.

3.08.7.4 Blowout Evolution

Blowouts may evolve in various ways, the specific pattern depending on wind speeds, dominant wind direction, vegetation species and revegetation potential, magnitude and occurrence of beach/dune erosion and storm events, latitude, and barrier/beach status (receding, stable, prograding). The length of blowout deflation basins and blowout depositional lobes are quite strongly correlated, and blowout width is relatively well correlated with depositional lobe length, such that as width increases, depositional lobe length increases. Trough blowouts have a ratio of 1:4, while saucer blowouts have a variable ratio of approximately 1:2 to 1:3. These data reflect an obvious evolutionary trend. Trough blowouts generally start as narrow notches or slots, and as the flow is steered and directed up the trough, maximum erosion takes place along the trough base or floor. Acceleration occurs up the stoss face of the depositional lobe leading to maximum transport up the central axis of the lobe. This leads to greater rates of elongation compared to rates of widening. Nonetheless, as trough blowouts become longer, they generally become wider (Hesp, 2002).

Many blowouts become larger over time and may evolve into parabolic dunes (Carter, 1990; Carter et al., 1990; Battiau-Queney et al., 1995). Trough blowouts typically advance through evolutionary stages from erosional notches to incipient blowouts to large blowouts to decaying and revegetating blowouts (Carter et al., 1990; Gares and Nordstrom, 1995). Saucer blowouts deepen to a maximum level perhaps dictated by their morphology. Once the blowout becomes relatively wide, long, and deep, it may be that flow decelerations dominate, jet flows do not occur, and deflation is limited (Jungerius et al., 1981; Hugenholtz and Wolfe, 2006, 2009). Vegetated slump blocks falling from the rims of the blowout can re-root on the blowout walls and aid in the process of stabilization (Carter et al., 1990). If deflation basins and flats are eroded down to a level near the water table, moisture can reduce erosion and facilitate the colonization by vegetation. Incipient foredunes commonly form across the throat or entrances of blowouts eliminating through-flow of beach sand. Any subsequent dune erosion and removal of incipient foredunes can lead to re-activation of the blowouts.

The geographic location of the blowout also has some effect in morphodynamic development. Blowouts formed in latitudes above ~40° N or below ~40° S have slopes that receive significantly different degrees of solar insolation depending on the orientation of the blowout. Some blowouts will have slopes that are in shade for much, if not all, of the winter and adjacent slopes may be in the sun for much of the time. These differences in the degree of receipt of solar insolation can control slope moisture, plant growth, slope angle, type of slope failure, and slope development in blowouts (Brough, 1998; Hugenholtz and Wolfe, 2006).

3.08.8 Parabolic Dunes

Parabolic dunes (also termed U-dunes, upsiloidal dunes, or hairpin dunes) are typically U- and V-shaped dunes characterized by short to elongate trailing ridges which terminate downwind in U- or V-shaped depositional lobes. The depositional lobes may be simple, relatively featureless sand sheets, or textured with a variety of dune forms (e.g., transverse dunes, barchanoidal dunes, etc.). Deflation basins, slacks, seasonal wetlands, ponds, lagoons, and various types of low ridges (e.g., gegenwalle and dune-track ridges) occupy the low deflation area between the trailing ridges (**Figure 7**).

3.08.8.1 Location

Parabolic dunefields are common in many parts of the world's coasts, but are particularly well developed on high wind and wave energy west coasts where there is a significant sediment supply. Examples include lower west coast, North Island, NZ (Hesp, 2003), west coast, Australia, the sand islands of South Queensland, Australia (Coaldrake, 1962; Thompson, 1983), the north Queensland coast, Australia (Pye, 1982, 1983), west coast USA (Cooper, 1958, 1967), Cape Cod, USA (Motzkin et al., 2003), the Great Lakes coasts, USA/Canadian central and north coast (Olson, 1958; Arbogast, 2000; Loope and Arbogast, 2000; Arbogast et al., 2002, 2010; Hansen et al., 2009, 2010), NE Queen Charlotte Is, Canada (Wolfe et al., 2008), Natal coast of South Africa (Tinley, 1985), Sri Lanka (Swan, 1979), Mozambique,

Figure 7 Two parabolic dunes in the Manawatu region, NZ. The lower dune developed from a blowout in the foredune and has migrated downwind onto the deflation plain of the older, landward parabolic dune (in the distance). As migration occurs, trailing ridges and deflation basins form and the depositional lobe becomes larger. The white surface in the deflation basins is a shell lag.

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south coast Madagascar, Hobyo and Hafun regions in Somalia, west of Madrakah in Oman (also including sand sheets), La Paz region in the Philippines, from Chur to Pasni in Pakistan, in the Dorri to Tis region, Iran, and various bays along the southern Iran coast to Jask, the Akdeniz region in Cyprus, portions of the north and east coasts of Sardinia, Denmark (Anthonsen et al., 1996), Holland (van Der Maarel, 1979), and various parts of the United Kingdom, particularly on the east coast of Scotland (Ritchie, 1972: May and Hansom, 2003).

3.08.8.2 Terminology and Form

Parabolic dunes can be difficult to distinguish from, or define as clearly distinct and different from, blowouts. One way to do so is to separate the two on the basis of the presence or absence of the trailing ridges. Parabolic dunes typically have vegetated trailing ridges; blowouts typically do not. For example, a blowout within a foredune has lateral erosional ridges that are formed in, and of the foredune; a parabolic dune will have discrete ridges trailing upwind from the depositional lobe and formed by sand deposition in vegetation along the margins of the depositional lobe as the lobe moves downwind.

3.08.8.3 Initiation

In coastal environments, parabolic dunes often evolve from blowouts, but not always (Jennings, 1957; Bird, 1976; Pye, 1982; Cooke et al., 1993, their figure 25.9). In many cases, a blowout depositional lobe will continue to migrate downwind regardless of upwind changes. For example, where a blowout has formed within a foredune, it is common for a new incipient foredune to form across the blowout throat or entrance area (Gares and Nordstrom, 1995; Hesp, 1999), and for the deflation basin to stabilize and vegetate, while the parabolic depositional lobe continues to migrate downwind (**Figure 8**).

Blowouts and parabolic dunes may be formed on both stable (sediment supply balanced) and accreting/prograding (positive sediment supply) coasts which experience occasional or regular high-energy wind events, and sometimes also wave scarping (**Figure 9**). The lower west coast of the North Island of

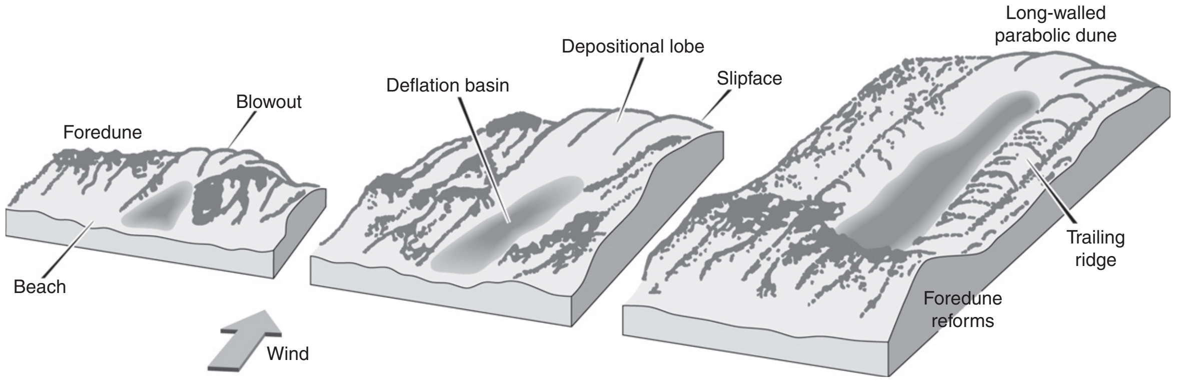

Figure 8 Schematic diagram of the evolution of a blowout in a foredune into a long-walled parabolic dune. Modified from Hesp, P.A., 2000. Coastal Dunes. Forest Research (Rotorua) and NZ Coastal Dune Vegetation Network (CDVN), 28 pp.

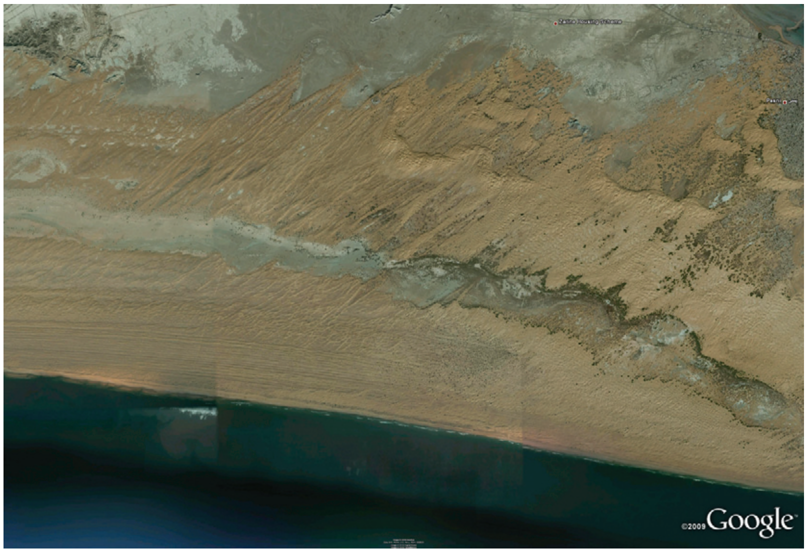

Figure 9 A complex barrier in Pakistan comprising a seaward suite of modern and relict foredunes, which break up into blowouts and parabolic dunes in the central to left side of the image and into a transgressive dunefield on the right. A deflation plain separates this portion of the dunefield from a suite of larger parabolic dunes and transgressive dunefield. The horizontal land view is 7 km across.

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NZ is prograding at around 1 m yr^{-1} (Shepherd, 1987; **Figure 7**). Frequent strong westerly winds erode small, partially vegetated, or unvegetated patches on the foredune, and blowouts are formed very rapidly, many of which evolve downwind into parabolic dunes (Hesp, 2003).

Blowouts and parabolic dunes are commonly formed on eroding coasts where the foredune stability is reduced, or indeed removed, by wave erosion, and subsequent wind erosion of the destabilized dunes or bare sand surfaces takes place (e.g., Short and Hesp, 1982; Carter et al., 1990; Psuty, 1992; McCann and Byrne, 1994; Ruz and Allard, 1994b). Psuty (1988, 1989, 1992) indicated that where the beach and foredune sediment budget is negative, blowouts and parabolic dunes are likely to form. Equally, some of the world's shorelines display inland translating foredunes where the foredune sediment budget is positive but coincident with an eroding shoreline (see **Figure 3(b)**). As such, this may represent a morphological adjustment to a changing climate and/or sea level, assuming accommodation space is available (Davidson-Arnott, 2005; Psuty and Silviera, 2010).

Transverse and barchanoidal transverse dunes may evolve into parabolic dunes as they advance into vegetation, or vegetation colonization of the dunes takes place (Anthonsen et al., 1996; Tsoar and Blumberg, 2002).

Parabolic dunes may be initiated, and evolve from the landward and downwind margins of transgressive sand sheets and dunefields, and also within those dunefields (**Figure 10**). As the mobile sands migrate into vegetation, discrete lobes may migrate ahead of the advancing dunefield to form parabolic dunes. This process occurs particularly where transgressive dunefields are stabilizing and being colonized by vegetation. Examples may be seen along the South African south coast (Tinley, 1985), USA west coast (Cooper, 1958, 1967), Denmark (Anthonsen et al., 1996), NZ west coast, and Australian coasts. Many moderately to fully vegetated transgressive dunefields (see below) have been mistakenly described as parabolic dunefields, but these have, in fact, only formed parabolic forms in the very closing stages of evolution as vegetation has taken control of transgressive dunefield evolution (see **Figure 10**).

Parabolic dunes may also be formed by aerodynamic constriction and topographic steering of the wind flow. Nested parabolic dunes are common where the sediment supply is plentiful or where phases of dune development have taken place. In some cases, subsequent migrating dunes may be forced into a parabolic dune form because of wind funneling of sediment that then drives it into a relict (vegetated) parabolic dune (**Figure 11**).

Climbing and cliff top parabolic dunes may be formed (1) during low sea levels, (2) during present (high) sea levels where steep terrain lies adjacent to the beach, and (3) where rapid erosion of unconsolidated to poorly indurated cliffed coasts is occurring (Brothers, 1954; Jennings, 1957; Short, 1988a, 1988b; May and Hansom, 2003). Parabolic dunes are often formed on the cliff tops and downwind areas as the climbing dunes migrate into vegetated terrain (e.g., west coast, northern North Island, NZ; Natal; Northern California coast).

A few authors have indicated that parabolic (and many transgressive) dunefields (now termed 'chevrons' by some of these same authors) in some regions have been formed by tsunami (Bryant, 2001: 73; Kelletat and Scheffers, 2003). Apparently, the dunes are formed by mega-tsunami racing across the coast and scalloping sandy sediments into U-shaped dune deposits in some cases up to 100 m high. The presence of chevrons on some coasts has led others to postulate the occurrence of catastrophic extraterrestrial impacts and giant

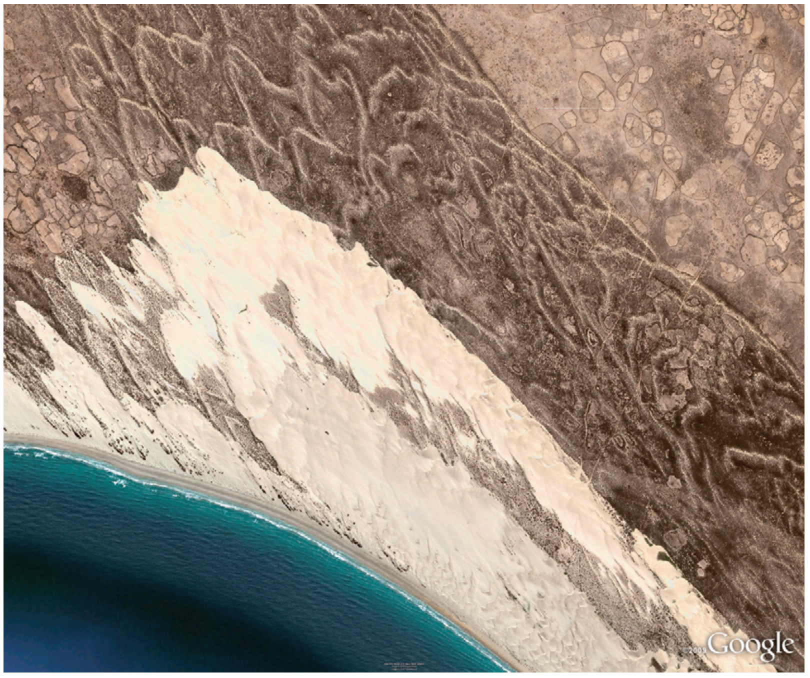

Figure 10 A trangressive dunefield and parabolic dunefield complex on the southern coast of Madagascar. It is likely that many of the parabolic dunes were formed following stabilization of transgressive dunefields as may be seen occurring on the downwind and seaward margins of the active transgressive dunefield shown here. Source: Google Earth.

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Figure 11 Nested parabolic dunes in the Manawatu region NZ. Two smaller parabolic dunes have advanced into a larger parabolic dune.

tsunami – a case of classic circular reasoning. The evidence is extremely poor to highly spurious; most is based on aerial photo mapping with little field verification (e.g., Kelletat and Scheffers, 2003). Further, there has never been any hydrodynamic explanation presented of how U-shaped dunes with marked trailing ridges and interior deflation basins can be formed by giant waves and massive overwash, and most serious research indicates that the tsunami deposits formed in sand-sized sediments are typically small- to large-scale sheets, similar to washover deposits (e.g., Dawson, 1994; Morton et al., 2007; Pinter and Ishman, 2008).

3.08.8.4 Morphology

Two principal subtypes of parabolic dune are common: long-walled types and squat, elliptical (or imbricate) types. Pye (1982) used length/width ratios to distinguish between the two. The multiple development of these leads to there being two principal subtypes of parabolic dunefields; long-walled types (e.g., Pye, 1982; Tinley, 1985) and imbricate types (Tinley, 1985), although, as with blowouts, there are significant numbers of other morphologies, sometimes very large (megadunes) and very complex types (see Davies, 1980; Pye and Tsoar, 1990, figure 6.49; Trenhaile, 1997).

Long-walled parabolic dunes display long trailing ridges and extensive deflation basins (and/or slacks) (**Figures 7**, **9**, and **10**). The trailing ridges may range from hundreds of meters long to several kilometers long. They are particularly well developed on relatively flat terrain, in regions of low heath or shrub land, high sand supply and strong winds (e.g., the Manawatu, N.Z. dunefields (Cowie, 1963; Esler, 1970); the central and northwest coast of Western Australia (Hesp and Chape, 1984; Hesp and Pelham, 1984); the north Queensland coast (Pye, 1982, 1983); the Callender area, southern and central California coast, USA (Cooper, 1967); Cape Cod, USA (Goldsmith, 1975, 1978), and southern Madagascar).

Some parabolic dunes display a squat, shorter form, often with more semicircular or elliptical deflation basins (e.g., Cooper, 1958; Goldsmith, 1975; Weise and White, 1991; McCann and Byrne, 1994). Multiple development results in the dunes overlapping each other in an imbricate fashion (e.g., Seppala, 1972, 2004; Tinley, 1985). They commonly develop in wetter areas, on flat terrain where deflation depths are limited and/or wind speeds are relatively low, on hummocky or relatively steeper terrain where significant downwind migration is impeded, in less unidirectional or multidirectional wind regimes and/or in dense, tall vegetation where the rate of advance is low and/or migration is impeded (e.g., the Monterey area, California, Cooper (1967); southern Queensland, Australia, Coaldrake (1962); Thompson (1983); Britain and Denmark, Landsberg (1956); Cape Cod, Massachusetts, Goldsmith (1978); Padre Island, Texas, USA, Weise and White (1991)). In some cases, more complex forms may develop (Filion and Morisset, 1983; Robertson-Rintoul and Ritchie, 1990).

Pye and Tsoar (1990) identified seven form variants of parabolic dunes. They state that parabolic dunes may display marked variations in morphology and orientation within a small geographical area, and that this reflects the local differences in wind climate. This is certainly true since parabolic dunes arranged along a log-spiral bay will be oriented parallel to the onshore wind resultant, which varies alongshore as different sections receive different onshore wind components (Landsberg, 1956; Jennings, 1957; Bird, 1976; Miot da Silva et al., 2008; Miot da Silva and Hesp, 2010). However, dune form is also dependent on a range of other factors, notably vegetation cover and type. Digitate parabolic dunes (Pye and Tsoar's type d) are common where the downwind vegetation is woodland and/or dense. Depositional lobe and slipface migration is impeded, and small secondary lobes advance ahead of other portions of the main lobe forming a digitate morphology (e.g., Filion and Morisset, 1983).

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3.08.8.5 Flow Dynamics

Landsberg and Riley (1943) and Olson (1958) conducted flow experiments over large blowouts, finding that marked flow compression and acceleration occurred on the upper windward slopes and over depositional lobe crests. Flow separation occurred over the lee (avalanche or slip) face, and local flow deceleration and separation occurred at sites where the topography was hummocky or more complex.

Finnigan et al. (1989) examined the flow fields around a 1:1000 scale model of a parabolic dune in a wind tunnel, principally to examine the degree of flow separation as a function of wind approach angle. They found, for winds blowing straight up the dune, that a symmetrical flow separation region formed over the dune depositional lobe and occupied an area about half the total dune width. Shear stress was substantially higher on the dune arms compared to the deflation basin. As the angle of wind approach increased ($>30^{\circ}$), the flow separation region shifted to become aligned with the approaching wind vector, not the dune axis, and there was therefore less steering of the wind by the dune. Flow over the trailing dune arms became more complex and vortex rolls were probably formed.

Robertson-Rintoul (1990) conducted field experiments of wind flow over a large parabolic dune ridge in Scotland using arrays of cup anemometers and flow visualization techniques. Her observations indicated that both upwind and leeward (downwind) flow separation occurred along the dune ridge forming both closed windward and leeward eddies. Horseshoe vortexes occurred on the upwind central dune slope and along the leeward dune margins. Crestal jets were common.

It is likely that flow conditions in parabolic dunes are similar to those in blowouts, although the scale of large parabolic dunes is such that flow constriction and acceleration is more likely to be localized to the margins and crests of the trailing ridges and the depositional lobe rather than up the deflation basins (cf. Hansen et al., 2009; Wakes et al., 2010) as in the case of blowouts.

3.08.8.6 Evolution

In many respects, parabolic dunes evolve in a similar manner to blowouts.

Parabolic dune deflation basins tend to continue to erode down to a base level such as the seasonally lowest water table level, a calcrete, or other cemented, indurated, or armored surface such as a gravel basement, shell, pumice, or human occupation surface (Carter, 1976; Hesp and Thom, 1990; **Figures 7** and **8**). (Ritchie, 1972; Carter, 1976; Pye, 1982, 1983; Carter et al., 1990; Hesp and Thom, 1990). Deflation basins may be quite flat, or they may be characterized by multiple, very low ridges termed 'gegenwalle ridges' (Paul, 1944; Pye, 1982). Wolfe et al. (2001) called these 'dune track ridges'. These ridges are formed on the downwind margin of deflation basins by sand trapping within pioneer species. Such species preferentially grow along the edge of active deflation surfaces and slacks that are commonly wet and/or have standing water in them for part of the year.

The trailing ridges of parabolic dunes primarily develop due to the presence of vegetation growing on the outside margin of the depositional lobe as it migrates downwind. As the lobe continues to migrate, the inside (deflation plain) portion of the ridge is eroded, resulting in a typically asymmetric trailing ridge (erosional on the inside, depositional on the outside). Cannibalization of older dunes is common, as younger dunes migrate into and over older relict dunes (**Figure 11**).

Depositional lobes are arcuate, hairpin, V-shaped, radial, or parabolic shaped depending on wind direction, lobe height and volume, vegetation cover and species type, and speed of migration. In the NZ west coast dunefields, the highest and shortest lobes are associated with migration into tall woodland and forest, whereas the lowest and longest lobes are associated with migration across short grasses and herbs.

3.08.8.7 Revegetation/Natural Stabilization

Cooke et al. (1993) questioned why some dunes are V-shaped while others are U-shaped. Most likely, these variant shapes result from the rate at which dunes revegetate and the size of the depositional lobe at revegetation. U-shaped dunes are more common where the lobes are extensive and rapidly revegetate in place. V-shaped dunes tend to develop where the depositional lobes are more gradually being colonized by vegetation. In the latter case, the lobe margins and associated trailing ridges gradually contract while the active lobe tapers, but continues to advance on an increasingly narrower front until only a thin V-shaped active lobe remains.

3.08.8.8 Rates of Advance or Migration

Rates of parabolic dune advance or migration vary considerably depending on the morphology, slope, and type (e.g., sandy vs. rocky) of terrain the dunes are moving across, the vegetation cover and type (e.g., woodland vs. grassland), wind velocities, and directional variability of the wind.

Typical migration rates on moderate energy coasts are around $2\,\mathrm{m\,yr^{-1}}$ (Cooper, 1958; McKee, 1979; Pye, 1982), but rates may be very low where the dunes are moving into tall, dense forest (e.g., $0.05\,\mathrm{m\,yr^{-1}}$; Story, 1982). Pye (1983) stated that parabolic dunes on the north Queensland coast migrate rapidly at rates of $5–6\,\mathrm{m\,yr^{-1}}$. However, these rates are actually moderate compared to some coasts. On the high wind energy central west coast of Western Australia, parabolic dunes migrate at rates varying from 3 to $5\,\mathrm{m\,yr^{-1}}$ where they are large, up to $13–15\,\mathrm{m\,yr^{-1}}$ where the dunes are relatively smaller, younger, and migrating over low heath (Hesp, unpublished data). Parabolic dunes on the high wind energy, west coast of Scotland migrate up to $14\,\mathrm{m\,yr^{-1}}$ at Menie in Aberdeenshire (Hansom, 2007), and on the lower North Island, NZ, display migration rates that range from low to average ($\sim 2\,\mathrm{m\,yr^{-1}}$) where the dunes are large and advancing into mature pine forest (Shepherd, 1987), high ($10\ –25\,\mathrm{m\,yr^{-1}}$) where the dunes are migrating into shrub lands, to extraordinary ($>50–70\,\mathrm{m\,yr^{-1}}$) where the dunes are low (1–3 m high) and the vegetation cover is very low shrubs to short grass (**Figure 7**).

3.08.9 Transgressive Dunefields

Transgressive dunefields are aeolian sand deposits formed by the downwind or alongshore movement of sand over vegetated to semivegetated terrain. Such dunefields may range from quite small (hundreds of meters in alongshore and

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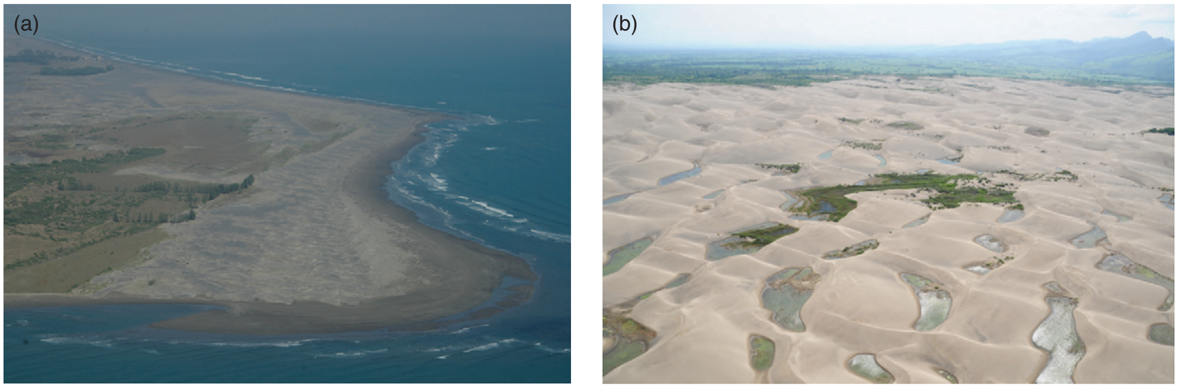

Figure 12 (a) A large, prograding transgressive dunefield on the central Veracruz coast, Mexico, displaying active transverse dunes migrating along the backbeach, precipitation ridges, trailing ridges, and deflation plains; (b) Barchanoidal transverse dunes with rainfall-filled interdunes, Dona Juana transgressive dunefield, Mexico. Some gegenwalle ridges may be seen in the central vegetated interdune.

landward extent) to very large fields that can be similar in size to some small to moderate-sized desert dunefields (Hesp and Thom, 1990; Hesp, 1999; **Figures 10, 12,** and **13**). Transgressive dunefields may be very large dunefields; they may be low, rolling, blowout-dominated, machair-type dunefields; or they may be low plains dominated by sand sheets, erosional knobs, and nebkha fields (particularly some retrogradational barriers; e.g., Hayden et al., 1995; Houser and Hamilton, 2009).

Transgressive dunefields may become completely vegetated postformation, partially vegetated, or largely unvegetated (fully active) (Hesp and Thom, 1990). Transgressive dunefields have also been termed 'mobile dunes', 'migratory dunes' (e.g., Olson and van der Maarel, 1989), and 'mendano' (e.g., Goldsmith, 1975, 1978). Some classifications (e.g., Short and Hesp, 1982; Pye, 1983) used the term 'transgressive dune sheet' rather than transgressive dunefield, but sand sheets are areas of predominantly aeolian sand where dunes with slipfaces are absent, and where the surface may be flat to undulatory (Kocurek and Nielson, 1986). Thus, both transgressive sand sheets and transgressive dunefields can occur although sand sheets may also form a landform unit or type within a transgressive dunefield.

3.08.9.1 Location

Transgressive dunefields are particularly well developed on high wind and wave energy (west and south) coasts with significant sediment supply, and also on coasts with limited vegetation growth in semiarid, arid, and cold (arctic and subarctic) regions. Thus, they are found and sometimes dominate the southwestern, southern, eastern, and northern Australian coast (Coaldrake, 1962; Thompson, 1983; Hesp and Chape, 1984; Short, 1988a, 1988b), NZ North Island west coast and Stewart Island (Brothers, 1954; Shepherd, 1987; Enright and Anderson, 1988; Muckersie and Shepherd, 1994; Shepherd and Hesp, 2003), west coast USA, SW Padre Island, Baja California Mexico (Cooper, 1958, 1967; Hunter et al., 1983; Orme and Tchakerian, 1986; Fryberger et al., 1990; Hesp and Thom, 1990; Fryberger, 1991), south and west coast of South Africa and Namibia (Tinley, 1985; McLachlan et al., 1987; Illenberger and Rust, 1988; Hesp et al., 1989; Lancaster, 1989; Hesp and Hastings, 1998), Natal coast of South Africa, west Sri Lanka (Swan, 1979), west coast South America particularly Chile and Peru (e.g., Finkel, 1959; Araya-Vergara, 1987; Avaria, 1987), the tropical West African coast, Cuba, the northern China coast, parts of the coast of

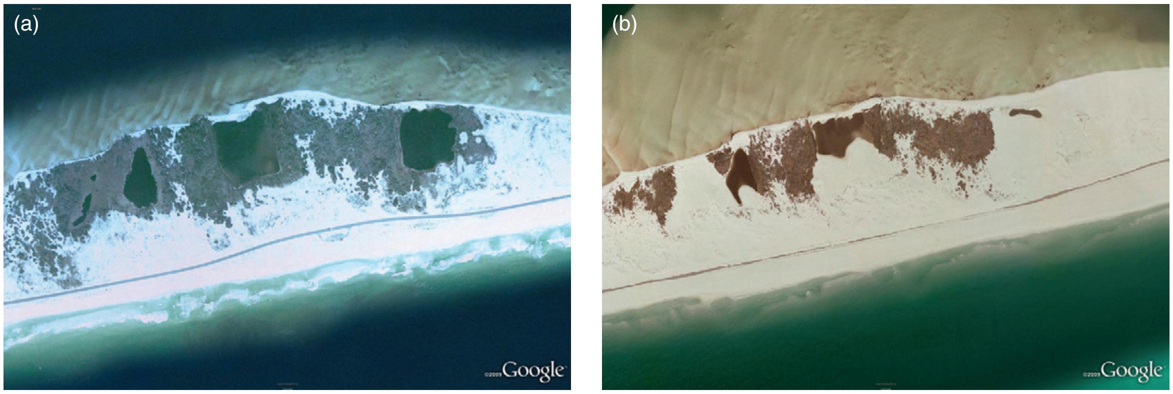

Figure 13 Pre-Hurricanes Ivan (2004), Dennis (2005) and Katrina (2005) (a) image (February 2004) and post-Hurricane Katrina (b) image (January 2008) of a portion of the barrier near Pensacola, Florida, Gulf coast, USA. During the 2004–05 period, the shoreline retreated ~44 m and recovered <10 m in the following period. A low foredune was largely destroyed in Hurricane Ivan resulting in widespread overwash, filling of back barrier ponds, and erosion of back barrier dunes by Katrina. The barrier in the 2008 image is 0.38 km across (see two-lane road for scale). Source: Google Earth.

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Vietnam from the Bay of Along (~19° N) to Cape Vung Tau (also called Cape Saint-Jacques; 10° N) in the south, the Rasiyan region in Djibouti, the Edd region in Eritrea, the coastal portion of the Wahiba Sands, Oman, the southern coasts of Pakistan and Iran, from Misratah to Ajdabiya, Libya, north coast of Egypt, the SE Mediterranean coast (Tsoar, 1990), the Karatas to Seyhan region, Turkey, the Laoag and Fort Llocandia regions in the Philippines, the Irawaddy delta region in Myanmar (Davies, 1980), the central east coast of India, particularly centered on the State of Orissa (Siddiquie, 1966), portions of arctic and subarctic environments (Black, 1951; Nichols, 1968; Filion and Morisset, 1983; Ruz and Allard, 1994a, 1994b, 1995; Seppala, 2004), Veracruz coast, Mexico (Hesp and Martinez, 2007), central west Java, Indonesia Verstappen, Somalia (e.g., Barawa and Turdho regions on the east coast; NE coast), portions of the Great Lakes, USA and Canada (e.g., Lake Michigan; Olson (1958); see Arbogast and Hansen references above; Davidson-Arnott and Pyskir (1988)), various sites in the United Kingdom (e.g., Ritchie, 1976; Hansom and Angus, 2006; Hansom, 2007), Netherlands, Belgium, Germany, Poland (Borowka, 1990a, 1990b; Piotrowska, 1991), the northern Ebro delta spit, Spain, SW Spain, Denmark, SW France (Bressolier et al., 1990), and one location on the SE coast of France (e.g., La Grau-du-Roi), and Portugal (Llamas, 1990). Some of the largest transgressive dunefields may be found on the northern and southern Brazil coasts where sediment supply over the Holocene has been significant (Bigarella, 1975; Hesp et al., 2007, 2009a, 2009b, 2009c; Martinho et al., 2008; Dillenburg et al., 2009; Dillenburg and Hesp, 2009a, 2009b).

In arid and desert regions, coastal transgressive dunefields are virtually indistinguishable from marginal continental desert dunefields and often merge imperceptibly into the deserts. Examples include the Namib Desert (Ward, 1987; Lancaster, 1989), sections of the Peruvian Coast (Finkel, 1959), Chile, the Baja California, Mexico region (Inman et al., 1966; Fryberger et al., 1990), portions of the coast of Yemen, Khaluf region in Oman, southern coasts of Pakistan and Iran, and various northwestern African coastal states (Fryberger and Dean, 1979; McKee, 1979; Seely, 1994).

3.08.9.2 Transgressive Dunefield Types

At the gross scale, transgressive dunefields may describe tabular forms, buttress forms, or they may be headland bypass dunefields, or climbing (perhaps with associated echo dunes), cliff top, and falling dunefields (e.g., Jennings, 1967; Tsoar, 1983; Tinley, 1985; Tsoar and Blumberg, 1991; Jackson and Nevin, 1992; Hellstrom and Lubke, 1993; White and Tsoar, 1998). In some cases, the transgressive dunefields that cross cuspate forelands in Victoria, Australia (Rosengren, 1981) are similar to the South African headland bypass dunefields (Tinley, 1985; McLachlan and Burns, 1992).

3.08.9.3 Initiation and Development

Transgressive dunefields develop for a variety of reasons. They may form, or have formed:

1. as a response to rising sea or lake levels and/or climatic change, particularly in the period, 10 000 to 7000 years BP (e.g., Pye, 1983; Pye and Bowman, 1984; Pye and Tsoar, 1990; Hesp, 1993; Shulmeister et al., 1993; Young et al., 1993; Hansom and Angus, 2006), but also throughout the past ~5000–6000 years (e.g., Arbogast et al., 2010);
2. in regions of high alongshore and onshore sediment supply and in sediment traps often in regions of high wind and wave energy (e.g., Short and Hesp, 1982; Illenberger and Rust, 1988; Fryberger et al., 1990);
3. on coasts experiencing erosion, both natural and human induced, and which may occur as a single, catastrophic event (Mathew et al., 2010), multiple large-scale to catastrophic events (e.g., several hurricanes over a period of years; Houser and Hamilton, 2009); or gradually over a longer period of time (e.g., Enright et al., 1988; Bressolier et al., 1990; Pye, 1990);
4. as continental shelves were exposed and/or climate changed during the Last Glacial (e.g., Lees et al., 1990, 1995; Thom et al., 1994);
5. as a response to periods of regional sea-level fall during the mid- to late Holocene (e.g., Christiansen et al., 1990); and
6. on coasts experiencing climatic extremes such as in arid (Lancaster, 1989) and arctic and subarctic environments (Ruz and Allard, 1994b), and where vegetation growth may also be limited.

3.08.9.4 Transgressive Dunefield Landforms

Active transgressive dunefields are usually characterized by a small to extensive deflation basin (or surface) or series of slacks on the seaward side, a small to extensive, mobile to partially vegetated sand sheet or dunefield, and one or more long-walled, commonly sinuous main slipface(s) or precipitation ridge(s) on the margins (Cooper, 1958, 1967; Davidson-Arnott and Pyskir, 1988; Hesp and Thom, 1990; Thom et al., 1992; Ruz and Allard, 1994b).

The surfaces of active transgressive dunefields are commonly covered with a variety of dune types ranging from simple transverse dunes and barchans, to barchanoidal and sinuous transverse and oblique dunes, to complex aklé or fish scale forms (**Figure 14**) (Cooper, 1958, 1967; Rosengren, 1981; Hunter et al., 1983; Tinley, 1985; Carson and McLean, 1986; Orme and Tchakerian, 1986; Borowka, 1990a, 1990b; Orme, 1990; Kocurek et al., 1992; Burkinshaw et al., 1993). Some simple transverse and oblique dunes, domes, and barchans may be ephemeral or seasonal forms; many are permanent and migrate downwind and landward at various rates. Star dunes are very rare in coastal dunefields, one of the few examples being west of Cocklebiddy in Western Australia (figure 6.18 in Hesp, 1999).

Deflation plains, basins, and slacks are common transgressive dunefield landforms (**Figure 12**). Ranwell (1959) stated that the term 'slack' is derived from an old Norse word 'slakki' meaning a small depression between two stretches of rising ground. It is generally taken to refer to damp or wet hollows where the groundwater table reaches or approaches the surface (Tansley, 1949), but Ranwell (1959)

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Figure 14 Four years after Hurricane Katrina impacted the Santa Rosa barrier in Pensacola, the former overwash surfaces and fans are partially covered with nebkha of various sizes. Each was formed by aeolian sand deposition within a discrete plant or group of plants colonizing the overwash areas. The margins of older, relict back barrier dunes (background) were eroded by storm surge during the hurricanes, and continue to be wind eroded into remnant knobs on this retrogradational transgressive dunefield barrier.

and others also refer to dry slacks. Geomorphologists tend to term the latter deflation hollows, basins, or plains where they are created by wind deflation, and swales where they are created by other means (e.g., by the seaward development of a new foredune; Hesp, 1991). Deflation basins, slacks, and/or interdune valleys are termed 'coralles' in Spain (Garcia-Nova et al., 1976).

Deflation plains and basins lie parallel, oblique, or transverse to the shore depending on the net direction of downwind dunefield migration (**Figure 12**). They represent the dominant erosional landform in transgressive dunefields where sediment has often been stripped down to a base level. This may be a calcrete pavement, the seasonally lowest water table, older dune or beach surfaces and palaeosols, or bedrock. Large-scale deflation areas are characteristic of arctic shores (Ruz and Allard, 1994b), where glaci-isostatic uplift surfaces are dominated by niveo–aeolian processes (Koster and Dijkmans, 1988), and many temperate dunefields, where the sediment supply is moderate or onshore winds are strong. In carbonate-dominated regions, a calcrete surface may be the most common base level. Typical desert landform units occur within, or associated with deflation areas including yardangs, ventifacts (see Laity (1994) for desert types review), nebkha, and remnant knobs. Gegenwalle ridges may also occur (Piotrowska, 1988, 1991). In Western and South Australian dunefields, the presence of calcrete, rhizotubules, cemented palaeosols, solution pipes, and pinnacles, and various levels of indurated carbonate sands and dunes may result in highly variable terrain with many remnant knobs littered with large broken clasts (Semeniuk and Meagher, 1981; Semeniuk et al., 1989). The chaotic assemblage of conical landforms found in some transgressive (and some parabolic) dunefields led Semeniuk et al. (1989) to term them 'chaots'.

Active transgressive dunefields may display a variety of generally smaller-scale dune forms and environments. These include remnant knobs, hummocks, bush pockets, nebkha, and shadow dunes (similar to many deserts; e.g., Thomas and Tsoar, 1990). A variety of shadow dunes, nebkha, hummocks, and irregular rim dunes (around the margins of washover fans), and amorphous transgressive dunefield terrains develop on barrier islands that have undergone frequent overwash and retreat (**Figures 13** and **14**) (e.g., Hosier and Cleary, 1977; Leatherman, 1979a, 1979b; Piotrowska, 1988; Ritchie and Penland, 1990; Houser and Hamilton, 2009).

As transgressive dunefields begin to revegetate or are colonized for the first time by vegetation, a variety of depositional dune forms may occur depending on the subenvironment (e.g., McLachlan et al., 1987). Bush pockets develop initially in interdune depressions in the South African dunefields, but may then develop into large hummocks or low mounds by reducing wind speeds and providing foci for sand deposition (Bate and Dobkins, 1992). Hesp and Thom (1990) recorded a temporal sequence of plant colonization of a deflation plain where isolated nebkha initially appear in the early stages, to eventually give way to a plain covered by hummocky dunes and flats. In some spectacular cases, the whole dunefield may be largely vegetated intact, and even large transverse dunes may be stabilized with little variation in form (e.g., Thom et al., 1992).

Landforms and dune units originally formed by erosion (e.g., remnant knobs (remanie dunes) and blowouts) may also be widespread in some revegetated dunefields (see **Figure 14**). Once dunefields are fully vegetated, it is often difficult to distinguish between those dunes formed by deposition and those by erosion (e.g., depositional mounds, small to large nebkha vs. remnant knobs).

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A particular form of transgressive dunefield, called machair, forms only along the exposed and low-lying mainland and island coasts of north and west Scotland, and the north and west coasts of Ireland (Ritchie, 1976; Randall, 1983; Angus and Elliott, 1992; Hansom and Angus, 2001, 2006). Machair can be described as a gently sloping coastal dune–plain formed by wind-blown calcareous sand, often incorporating a mosaic or cordon of dunes to seaward, and a species-rich grassland (managed by traditional low-intensity agriculture), slacks, wetlands, and lakes to the landward. Often the dune cordon may be missing due to frontal wave erosion, but a characteristic of machair surfaces is that they are lime rich, subject to strong, moist, oceanic winds and show detectable current or historic biotic interference from grazing, cultivation, addition of natural fertilizer such as seaweed and, sometimes, artificial drainage. Blowouts, deflation plains, and basins are common on some machair plains, and peat land often lies to the landward of machair (Angus and Elliott, 1992). Indeed, some machair could clearly be classified as an exposed version of a dune system with foredunes, blowouts, parabolic dunes, and transgressive dunefields. In other cases, the machair landscape is somewhat complex, low, and undulating hummocky topography. The development of machair is, and has been, much influenced by hundreds of years of wave and wind erosion as well as human and animal activity (Ranwell, 1980; Randall, 1983; Hansom and Angus, 2006).

Precipitation ridges (long-walled or main slipfaces) may occur along the downwind and surrounding margins of transgressive dunefields. Where the dunefields are migrating in one primary direction, they generally have one precipitation ridge. Where they are expanding or migrating landward and alongshore, they may have two ridges, and sometimes these are linked; that is, they surround the dunefield. Where the dunefields are crossing headlands (as with headland bypass dunefields) or older dune terrains, they may be bordered by two marginal (lateral) precipitation ridges and also have a downwind precipitation ridge (**Figure 12**).

3.08.9.5 Development of Dune Phases

Transgressive dunefields (and parabolic dunefields) have commonly formed in phases, quasi-phases, or episodically. Groups of dunes or individual transgressive phases overlie each other to various degrees or are separated by deflation basins and plains, slacks or wetlands (e.g., South African dunefields, Illenberger and Rust (1988); Australian dunefields, Thom (1984), Hesp and Thom (1990), and Thom et al. (1992); NZ dunefields, Brothers (1954) and Muckersie and Shepherd (1994); west coast USA dunefields, Cooper (1958, 1967), Orme and Tchakerian (1986), and Orme (1990)), and Brazilian dunefields. Sea-level changes, phases of aridity, variations in sediment supply, phases of increased coastal erosion, an increase in fire frequency, and human influences have all been cited as possible mechanisms. As yet, our understanding of the mechanisms driving the creation or initiation of dunefield phases, or pulses of dune development is exceedingly limited.

3.08.10 Models of Beach–Dune Interactions

The original generation of the wave–beach–dune model of beach and dune interactions was formulated by Hesp (1982). It followed the publication of a robust microtidal beach model with reasonably high predictability (Wright and Short, 1984; Short, 1999). The beach model enabled scientists to classify microtidal beaches into six states with characteristic morphologies, mobility, and modes of erosion and accretion. Subsequent research has resulted in other surfzone-beach states being recognized and extension of the original model to meso- and macrotidal beaches (Short, 1991; Aagaard and Masselink, 1999). An understanding of beach and backshore morphology for different microtidal surfzone-beach types allowed Hesp (1982) to develop actual and theoretical links between backshore morphology, potential aeolian transport, foredune state and morphology, and dunefield type and development. Note that there has been little development of the model for meso- and macrotidal beaches.

3.08.10.1 Surfzone-Beach State

The microtidal beach models classified beaches into six states, with the dissipative state at the high wave energy (>2.5 m) extreme and reflective state at the low wave energy (<1 m) extreme. Four intermediate beach states occur between these states (Short, 1999). Dissipative beaches are characteristically high wave energy beaches and have the highest potential onshore sediment supply (Hesp, 1988a, 1988b; Miot da Silva and Hesp, 2010). Note, however, that beaches may also be dissipative because of the presence of very fine sand (hence low gradient), or abundant sand, so some dissipative beaches may, in fact, be low wave energy beaches. They are typically wide, display flat to concave morphologies (no berms), low gradients, and minimal backshore mobility. The latter refers to the coefficient of variation of mean shoreline position, and in reality refers to the amount of volumetric and profile change the beach and backshore experiences over time and through erosion to accretion phases. Reflective beaches are characteristically low wave energy beaches with low potential onshore wave driven sediment transport. Note that they may also be moderate to high wave energy where the beach sediments are coarse sand, gravels, or boulders. They are relatively steep, narrow, linear to terraced (i.e., display a berm form) morphologies, with low backshore mobility. Intermediate beaches range from wide, relatively flat beaches with low mobility at the higher energy end of the spectrum, through moderate width with beaches with pronounced berms and high mobility, to narrow beaches and moderate to low mobility berms at the reflective end of the range. Rips dominate surfzone processes in the intermediate range.

3.08.10.2 Beach–Backshore Width and Morphology, Fetch, and Potential Aeolian Transport

Beach width is important in determining fetch which is crucial for determining the volume of sand delivered across the backshore and to dunes (Davidson-Arnott, 1988; Bauer and Davidson-Arnott, 2003). Beach morphology is important because the greater the morphological variability, the more likely that wind velocity decelerations and flow variations

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take place across the backshore. Hesp (1982, 1999) showed that the wind flow across a wide, low gradient, dissipative beach displayed minimal flow variation and gradually accelerated across the backshore, thus maximizing potential aeolian transport. The wind flow over the berm crest of an intermediate beach was accelerated but decelerated leeward of the berm crest. High narrow berms typical of some reflective beaches display significant flow disturbance and deceleration leeward of the berm crest (Short and Hesp, 1982). Sherman and Lyons (1994) modeled wind flow and potential sediment transport on a flat beach, low berm and high berm profiles, and found that sand transport from a dissipative beach was 20% higher than that from a reflective beach if just slope and grain size were taken into account. When moisture content was added, transport rates were nearly 2 orders of magnitude higher from the dissipative beach compared to the reflective beach. Note, however, that each beach had the same width (100 m wide), whereas actual reflective beaches and many intermediate beaches are considerably narrower than dissipative beaches (Short and Hesp, 1982).

Beach mobility (not just beach slope as stated by Houser and Mathew, 2011) is important because the greater the beach mobility, the greater the morphological variability (Short and Hesp, 1982; Short, 1999). The latter affects the fetch such that the beach width is at times quite narrow and at other times quite wide, particularly for intermediate beaches. It is also important because alternating episodes of cut and fill result in varying beach morphologies which then affect airflow and sediment transport as indicated above.

Thus, the link between surfzone beach state (surfzone type and wave energy, beach width, morphology, mobility and type), aeolian sediment transport, and landward dunes is that modal dissipative beaches have maximum potential aeolian sediment transport, reflective beaches minimal potential aeolian sediment transport, and intermediate beaches range from relatively high potential at the dissipative end to low potential at the reflective end. Note that a minimal sediment supply (minimal is currently undetermined) is required.

3.08.10.3 Aeolian Sediment Transport and Foredune Morphology

An examination of foredune heights and volumes on dissipative to reflective beaches allows one to examine the validity of the links above. Since established foredunes occupy a foremost backshore position, they are a medium-term indicator of beach and backshore processes. Hesp (1982, 1988) measured incipient and established foredune volumetric changes over several years at Myall Lakes National Park in NSW, Australia to find that a modal reflective beach with the same wind exposure as a modal dissipative beach received 60% less sand than the dissipative beach over the same survey period. Intermediate beach foredune volumes ranged from relatively high to relatively low between the dissipative and reflective beaches.

Surveys of established foredunes, which have potentially been present for several hundred years, provide further evidence that there is a strong link between surfzone-beach type and foredune height and volume. Hesp (1982, 1988) demonstrated that in the Myall Lakes National Park in NSW, Australia, the smallest established foredunes, with lowest sediment volumes, were found on reflective beaches, while the highest and largest foredunes occurred on dissipative beaches. Similar results are reported by Davidson-Arnott and Law (1990). Intermediate beaches followed a trend from low to high volumes on lowest to highest energy intermediate beaches respectively (see reviews in Sherman and Bauer (1993a, 1993b), Bauer and Sherman (1999), and Houser and Hamilton (2009)). Houser and Mathew (2011) also indicated that reflective and intermediate beaches have limited transport potential.

Psuty (1988, 1989, 1992) argued that the sediment budget is the primary control on foredune size and indicated that small foredunes form where the budget is high, and that the largest foredunes form where there is a slightly negative budget, apparently regardless of surfzone-beach type. Hesp (1984b) showed that very small foredunes were formed rapidly, and virtually continuously on a very low-energy reflective beach with a high sediment supply (up to ~13 m yr^{-1}). Slightly to moderately erosional low-energy reflective beaches at St. Josephs Peninsula, Florida either display high, wide, foredune-blowout complexes or arrete-type high, narrow foredunes similar to those illustrated by Psuty (1992). These examples fit Psuty's (1988) model and demonstrate that sediment budget is an important factor in foredune development. However, Short and Hesp (1982) would argue that for beaches with the same sediment supply or budget, a dissipative beach will have a far higher potential for greater foredune development than a reflective beach (cf. Houser and Mathew, 2011). Increasingly, we see that the sediment budget is not only driven by sediment availability at the close of the Holocene transgression (e.g., Roy et al., 1994; Woodroffe, 2002) and by antecedent topography (e.g., Riggs et al., 1995; Jackson et al., 2005), but also by surfzone nearshore topography and wave processes associated with alongshore topographic variability (e.g., McNinch and Luettich, 2000; Houser et al., 2008; Houser and Mathew, 2011).

3.08.10.4 Foredune Ecology

The vegetation cover, species richness, and zonation of foredunes are determined by several factors, but sediment supply and sand deposition rate, and salt spray aerosol levels are two very important factors (Hesp, 1991). Simultaneous studies carried out on adjoining reflective, intermediate, and dissipative beaches show that salt spray aerosol levels are related to surfzone-beach type. Dissipative beaches have the widest surfzones, the greatest number of breaking waves, the highest wave heights, and the highest salt aerosol levels. Reflective beaches often have only one breaking wave, narrow to very narrow surfzones, low wave heights, and the lowest salt aerosol levels. All other factors being equal, foredune species richness and zonation tends to be greatest and narrowest respectively on reflective beaches (low sediment supply and salt aerosols) and lowest and widest on dissipative beaches (highest sediment supply, high salt aerosol levels) (Hesp, 1988a, 1988b).

3.08.10.5 Foredune Stability and Type, Erosion Processes, and Dunefield Development

Foredunes bear a morphological imprint dictated, in part, by modal surfzone-beach erosion and accretion modes, and the wind often accentuates this morphological imprint. Dissipative beaches are typically eroded by swash bores and undertow

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commonly associated with elevated water levels and storm surge. Beach erosion and dune scarping is laterally continuous alongshore and at times catastrophic. Hesp (1982, 1988) and Short and Hesp (1982) theorized that such laterally continuous alongshore, large-scale foredune scarping would on occasions lead to large-scale foredune destabilization. Transgressive dunefields would most likely result from the breakdown of the large established foredune. In fact, transgressive dunefields are most commonly found on high-energy, dissipative surfzone-beach systems (e.g., Australian and South African coasts below the tropics; west coast USA; NZ North Island west coast, Brazil east coast). Indeed, many of the biggest coastal dune systems around the world (the Atlantic coasts of Europe from Portugal to Denmark; the southern hemisphere coasts of Australia, NZ, South Africa–Mozambique–Namibia, Madagascar, and Brazil) are associated with moderate to high energy, intermediate to dissipative surfzone-beach systems (e.g., Aagaard et al., 2004; Miot da Silva et al., 2008; Hesp et al., 2009a, 2009b; Dillenburg and Hesp, 2009a, 2009b).

Intermediate beaches are characterized by localized, arcuate rip embayment erosion during storms. Such arcuate erosion extends well into the foredune during extreme events resulting in large-scale, but localized foredune scarping (refs). Topographic funneling of the wind may result in the evolution of blowouts and eventually parabolic dunes at these locations. Hesp (1982, 1988) and Short and Hesp (1982) indicated that moderate to higher-energy intermediate beaches should display parabolic dune complexes as, for example, they commonly do on the west coast of the USA. There is no doubt, however, that wind energy, sediment supply, and vegetation presence, and resilience are also crucial factors especially in regard to parabolic dune development (e.g., Pye, 1990). It is common to find a mix of both parabolic dunefields and transgressive dunefields on the higher energy dissipative and intermediate USA west coast beaches. However, they are also common on the lower wave energy, often reflective beaches of the Australian west coast between Perth and Dongara, indicating that high wind energy coupled with a moderate to high sediment supply will provide the same result!

On SE Australian beach systems where overwash events are minor to absent, where sediment supply is generally not limited, and where an aggressive pioneer grass (*Spinifex* sp.) exists, relict foredune plains are common, particularly on the moderate energy intermediate beaches. Here, established foredune stability is maintained to various degrees, and progradation over the last 6000–7000 years has led to the development of pronounced foredune plains.

Reflective beaches are characterized by accentuated swash during storms and laterally continuous alongshore beach erosion. Recovery is fairly rapid. Foredunes remain relatively stable over time, and because they are typically small, with limited sediment supply, little dune transgression results. Thus, reflective beaches are typically characterized by a single foredune or a few relict foredunes. Again, and as noted below, there are plenty of places that prove the exception, and many wide foredune plains have developed on low-energy reflective beaches on the central west coast of Western Australia due to a local high sediment supply.

3.08.10.6 The Role of Sediment Supply, Sea Level, Wind Energy and Other Factors

There is no doubt that sediment supply, wind energy, sea-level state (transgressive, stable, regressive), coastal-barrier state (stable, retrograding, prograding), return interval and magnitude of extreme storm events, climate (partly dictating vegetation presence and cover), the degree of system hysteresis, tidal range, and Pleistocene inheritance factors will all, at times, and, in some places, be a controlling variable in beach–dune interactions. For example, if sediment supply is limited, sea level is rising, and coastal erosion is the general rule, the dune model as outlined above may be partly or wholly invalid.

3.08.11 Conclusion

The initiation, morphology, and basic dynamics of the four primary coastal dune types have been briefly reviewed. Foredunes may be commonly found along many coasts of the world except where there is marked aridity or prevailing snow and ice conditions, or where sand-size sediment is unavailable along with some accommodation space. Wherever we find foredunes or other dune types, blowouts may occur. Foredune plains occur where there is a moderate sediment supply coupled with shoreline vegetation growth and stability (eastern coasts and some portion of western coasts of many countries around the world). Parabolic dunefields and transgressive dunefields are more commonly found on the moderate to high wave and wind energy, moderate to high sediment supply coasts of the world, and are best developed on the south coasts of Australia, South Africa, Pakistan, Iran, and Oman, the west coasts of Australia, South Africa–Namibia, NZ, Chile, Peru, northern Mexico, and the USA, and along the Atlantic coasts of Portugal to Denmark, and along some portions of the east coasts of Australia, Brazil, South Africa–Mozambique, and India. Significantly, more comparative studies are required in order to determine the relationships between dune types and controlling variables around the world.

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