31

Chapter Eleven

Erosional landforms in coastal dunes

R. W. G. CARTER,
Department of Environmental Studies, University of Ulster

PATRICK A. HESP,
Carlingford, New South Wales

AND

KARL F. NORDSTROM
Institute of Marine and Coastal Sciences, Rutgers University

1. INTRODUCTION

Set against the extensive literature on depositional landforms in coastal dunes, there are few publications focussing on their erosional counterparts. However, erosional landforms constitute a significant proportion of many dune systems, and form an integral part of many postulated dune morphological cycles (Sokolow, 1894; Aufrère, 1931; Melton, 1940; Davies, 1972). Moreover, erosional forms add a significant aesthetic dimension to dunescapes (Engel, 1981, 1983) and require special management (Ranwell and Boar, 1986; Nordstrom and Lotstein, 1989).

Erosional dune forms include a range of slope failure and deflation structures from small isolated features to more complex terrains. Most dune landscapes tend to be indicative of negative sediment budgets and are likely to include erosional forms. Many other dune systems are mixtures of erosional and depositional morphologies, often with the former controlling the latter and giving rise to a suite of second-phase eolian landforms. The aim of this chapter is to provide a brief review of the erosional landforms encountered in coastal dunes, including wave-eroded features, large-scale deflation surfaces, and blowouts.

Coastal Dunes: Form and Process. Edited by K. F. Nordstrom, N. P. Psuty and R. W. G. Carter

Small-scale erosional and depositional forms associated with these features are also discussed. Examples of dune systems in Ireland, the USA and Australia are used to supplement results of previous investigations. Blowouts are examined in detail to illustrate the feedback between dune shape and wind flow as well as the reversals of net sediment transport caused by alterations in the location of deposition and scour. The complex interaction of processes with dune form leads to a reorganization of topography and creates complex mosaics of topography with vegetation. One of the principal goals of the chapter is to highlight the significance of erosional processes in the interpretation of these seemingly irregular dune landscapes.

2. EROSION OF DUNES BY WAVES

The reworking and remobilization of dunes by wave attack (Fig. 1) is a common process that has been discussed by numerous authors including Bremontier (1833), Dolan (1972), Parker (1975), Leatherman (1979), Vellinga (1983, 1984), van de Graaf (1986), Carter and Stone (1989) and several others in this volume.

Fig. 1 Undercutting and erosion of coastal dunes by storm waves (photograph by John Greer)

Because of the potential dangers, there have been a number of attempts to define conditions leading to dune erosion (Edelman, 1968, 1972; van der Meulen and Gourlay, 1968; van de Graaff, 1977, 1986; Hughes and Chui, 1981; Vellinga, 1983, 1984). These studies, including theoretical, empirical and experimental approaches, have highlighted the critical parameters leading to erosion, including morphological form (beach slope, dune height), and sediment texture (grain size, shape and packing) as well as hydrostatic (water level) and hydrodynamic (wave height, period and type) factors. Van de Graaff's (1986) work would suggest that surge height is by far the most important variable (82.8% variance in his studies), followed by particle size (7.3%) surge duration (2.6%) and initial profile (1.3%). Research indicates that the rate of erosion decays exponentially through a storm as the dune to beach sediment exchange re-establishes an equilibrium to the changed conditions. In fact, Hughes and Chui (1981, p. 186) state that between 70 and 90% of dune erosion is accomplished *before* the surge peak. Both Edelman (1972) and Hughes and Chui (1981) point out that the speed of storms is often the key factor in predicting the amount of erosion at any point on the shore. Van de Graaff (1986) recognizes two types of erosion: (i) gradual, perhaps involving 50 $m^3 m^{-1}$ of shoreline each year (approximately 2 m of retreat) and (ii) storm, when as much as 400 $m^3 m^{-1}$ (equivalent to 15 to 20 m of recession) may occur in 5 to 10 hours.

Erosion is caused by basal undercutting due to periodic wave attack (Fig. 2), although inherent slope stability factors, like soil moisture, increases in overburden and dynamic loading due to vegetation may also play a part. The main focus of attack is the foot of the foredune, which may be directly undercut by waves or trimmed by swash. Water in the sediment interstices usually provides sufficient cohesion to allow a vertical scarp to form. Scarp height (H) is related to repose (residual) angle (ϕ), slope angle (i), cohesion (c), unit weight of sediment (γ) in the form (Lohnes and Handy, 1968):

$$H = \frac{4c}{\gamma} \frac{\sin i \cos \phi}{[1 - \cos(i - \phi)]}$$

If the sediment was cohesionless ($c = 0$) then $H = 0$ and no scarp would form. However, undercutting leads to a state of tension on the upper slopes, often visible as parallel cracks (Carter and Stone, 1989, Fig. 4C), several tens of centimetres deep. Scarping alters the balance of forces on the dune slope, leading eventually to slab wedge-type or steeply inclined rotational failure, which may leave a vertical mid-slope scar and so cause further progressive slumping (Fig. 2). Assuming $\gamma = c$. 18 kNm^{-3} (a reasonable figure for moist, loose sand of mixed grain size) and $\phi = 34°$, then a vertical scarp ($i = 90$) of between 0.2 m and 2 m will be maintained as cohesion, c, increases from 0.5 kPa (kilo Pascals) to 10 kPa.

Rapid oversteepening of the lower slope leads to the collapse of the vertical scarp and eventually secondary failure of the entire cliff. Following undercutting, the time to failure of the dune face is associated with the cohesiveness of the

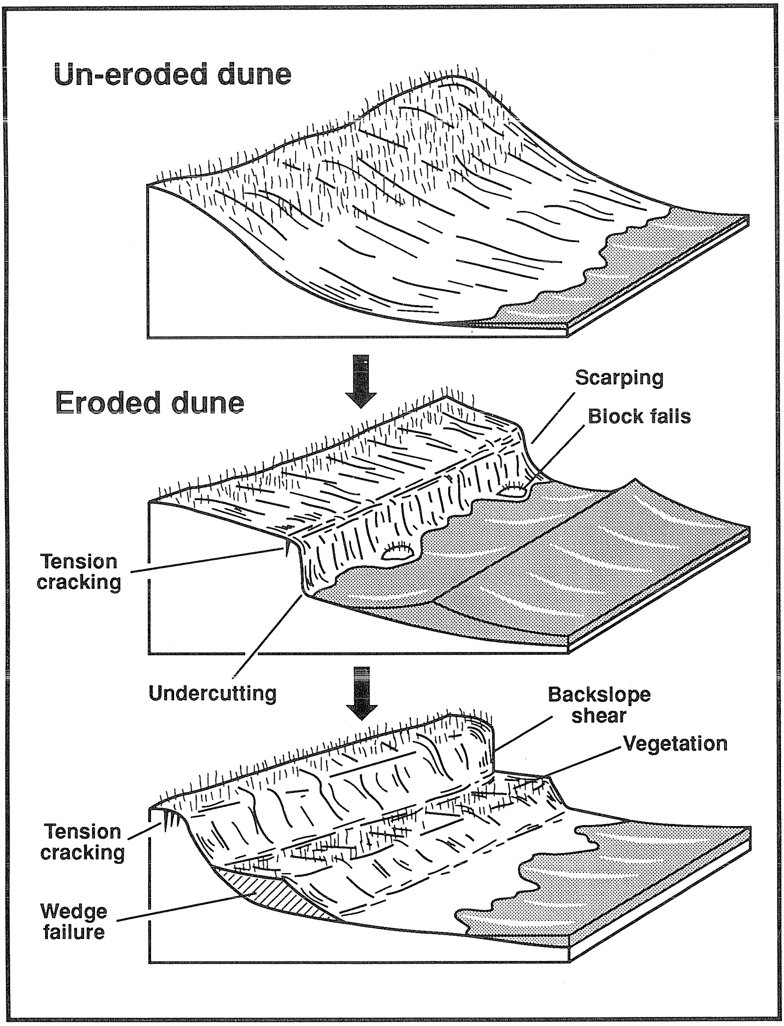


Fig. 2 Sequence of undercutting, failure and post-failure slope development on coastal dunes

dune, which in turn is related to the density of vegetation roots and rhizomes, soil moisture and chemical cementation. Slope failure may take place continuously, intermittently or only once. A single failure may protect the dune foot from further collapse, especially where a vegetation and soil mat slumps to form a protective sheath at the base of the dune. On granular, non-cohesive dune slopes above the vertical scarp, failure following undercutting is usually continuous and occurs as shallow surface slides or avalanches with slope angles

Fig. 3A Retrogressive slope failure of a dune scarp following basal undercutting

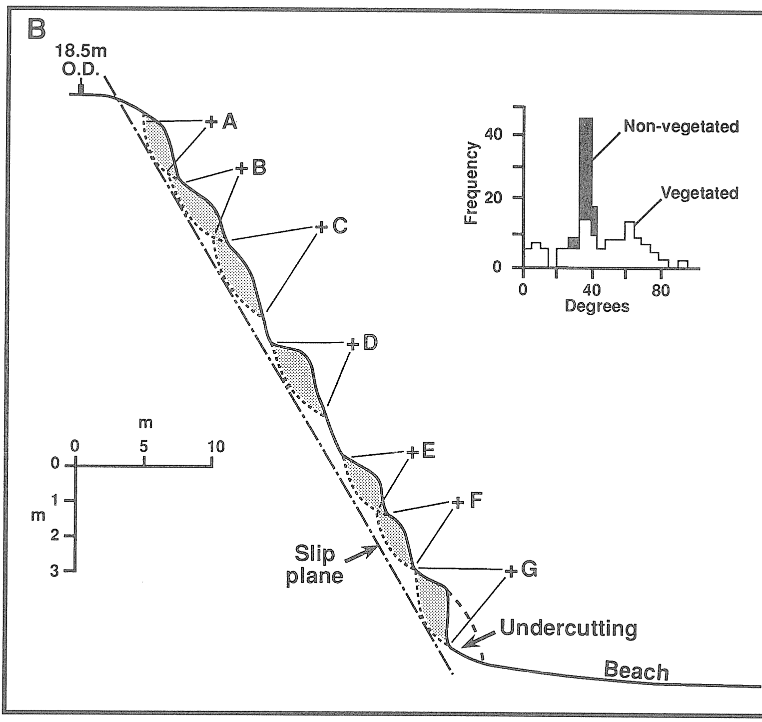


Fig. 3B Slope profile of same site illustrating the relationship between the shallow rotational slumps and the overall failure surface. The inset on the upper right shows the distribution of slope angles immediately after failure. The vegetation slopes (white) show a complex distribution of slope angles while the unvegetated (black) display only one mode correlated with the angle of repose remaining around 32° to 34°, the repose angle for loose sand grains. Presence of vegetation roots increases cohesion from zero (cohesionless) to 5 to 15 kPa, and allows steeper (40° to 43°), stable slopes (Carter, 1980; Greenway, 1987). Mode of failure also changes to larger, less frequent collapses. Vegetated or loosely cemented slopes fail intermittently, perhaps two or three times during each undercutting episode, usually as discrete slumps or slides (Fig. 3A). A wave-undercut dune slope may also fail as a series of shallow retrogressive translational slides with rotational elements. Fig. 3B illustrates a dune slope failure with seven small rotational slumps, apparently moving on a shallow (*c.* 1.8 m) slip plane coincident with the root vegetation depth. Higher slumps overlie lower ones (E over F, B over C, A over B on Fig. 3) indicating concatenated failure.

The slope failures associated with marine undercutting may be considered as first-phase adjustments. Second-phase activity includes both erosion and deposition. Drying-out of the slope often triggers small avalanches and block falls, occasionally associated with gravity tunnelling, where dry sand underlies a moist cohesive layer. Desiccation and slumping may form breccias (Vortisch and Lindstrom, 1980) and earth and root falls, which may be rolled by swash action into balls (Foss, 1985; Carter and Stone, 1989). In some cases, whole slump blocks, held together by vegetation (commonly a single species), litter

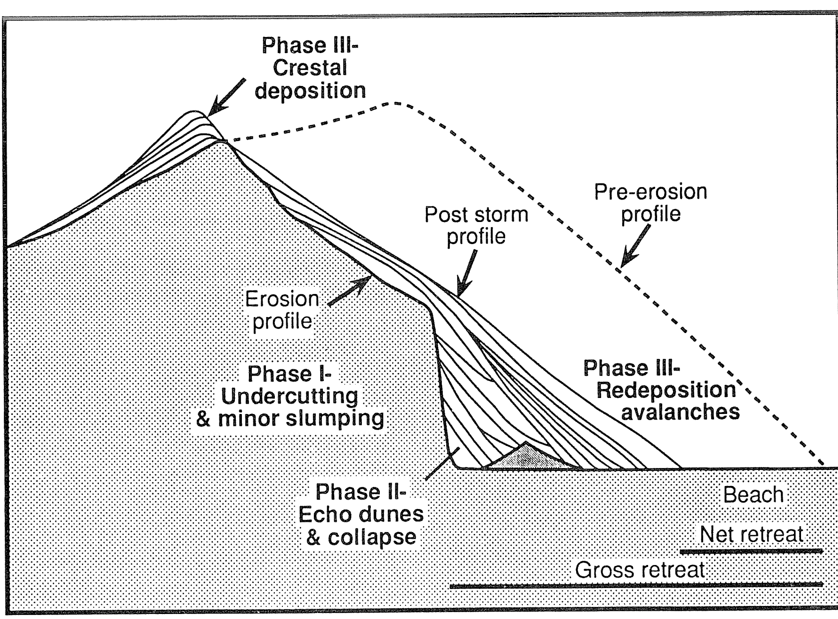


Fig. 4 Three phases associated with the undercutting, slumping and reforming of the dune slope

the avalanche slope. Where the vegetation has multiple roots (e.g. *Ammophila* spp.), it is not uncommon for the plant to survive and establish new roots within the surrounding and underlying avalanche material. Scarp slopes are revegetated by this means. In the examples shown in Fig. 3, the Phase I (Fig. 4) was followed by a number of small avalanches and chutes, which together with fresh accumulations of blown sand, formed a rectilinear facet at 38° within 3 months (Phase II—Fig. 4) (Carter, 1980).

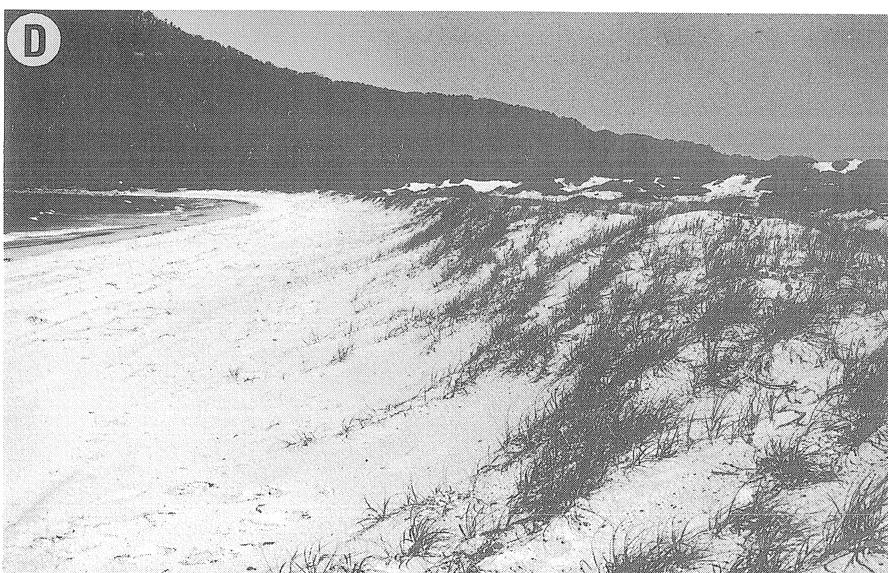
Sediment removed from the dunes during scarping is usually returned to the slope face as part of the beach/dune recovery cycle (Phase III on Fig. 4). The initial accumulations are commonly echo dunes (Tsoar, 1983) forming between the HWM and the scarp (Fig. 5A). Later, windblown material may accumulate at the crest (Fig. 5B) and/or on the mid-slope (Fig. 5C), leading to overloading, instability and further failure, with the material sliding down to mask the lower slope and upper beach (Fig. 5D). Once the entire scarp is covered by aeolian sediment, material may accrete at or beyond the crest (Hesp, 1988), often in distinct lobes. The scarp may be filled by marine sediments thus 'armouring' the dunes and facilitating wave overtopping (Orford and Carter, 1985). Once the scarp slope is filled, and lying below the angle of repose, vegetation growth by vegetative expansion from slump blocks, propagation of seed material, or colonization by rhizomatous species extending from the scarp crest gradually results in partial or full recovery. The internal structure of dunes subjected to scarp/recovery cycles is distinctive, with dominant large-scale, seaward-dipping cross-beds (Fig. 6), often overlain by landward-dipping sets representing an advancing sand-sheet. Some shoreline dunes show a strongly bipolar, shore-normal azimuth structure (Goldsmith, 1973). As the dune retreats, the rebuilding of seaward faces and crests tends to maintain the dune form (Figs 4 and 5D), particularly on well-vegetated slopes.

3. DEFLATION SURFACES

Eolian processes remove the finer material on flat or gently inclined surfaces, leaving an immobile, often coarser, residue (Fig. 7A) (Segerstrom, 1962; Cooke, 1970; Carter, 1976).

The formation of an immobile surface marks a limit to eolian entrainment and thus sediment supply, and introduces a distinctive aerodynamic domain through the creation of a new boundary layer. Limits to deflation may arise via the aggregation of coarse particles at the surface (Carter and Rihan, 1978), the formation of chemical crusts (Carter, 1978; Pye, 1983), presence of algal mats (Van der Ancker, Jungerius and Mur, 1985); the exhumation of buried soils (Cooper, 1958), or on encountering the water table (Ahlbrandt and Fryberger, 1981; Wiedemann, 1984). Where a salt crust forms or there are temporary fluctuations in the water table deflation may cease for a period before resuming. The presence of an immobile shell or gravel pavement may form a semi-permanent deflation limit, which is occasionally re-exposed (Carter and Rihan, 1978).

Fig. 5A–D A scarp fill sequence on eroding dunes from Fens Embayment, Myall Lakes National Park, Australia. A: Post-scarp slumping. *Spinifex sericeus* rhizomes are being undermined and exposed (Phase I in Fig. 4). B and C: Scarp filling takes place via slumping and later eolian ramp formation (Phases II and III in Fig. 4). Sand blown into



the original scarp crest vegetation forms asymmetric ridges and lobes. In C small-scale dry sand flows are visible. These are being fed from within and above the vegetation. D: Subsequent regrowth of *Spinifex* results in stabilization of the scarp fill

Fig. 6 Undercut slopes during Phase III commonly display a range of erosional and depositional forms. In this photograph seaward-dipping cross-beds indicating an earlier erosion/accretion cycle have been exposed. The lower slopes are being covered by dry sand avalanches, tongues often fed by upper slope rills and chutes. Vegetation block falls are also visible as are soil 'balls' at the foot of the slope

Formation of a deflation surface is often rapid. Carter (1976) describes shell lags 'emerging' in only a few days when conditions are favourable. During emergence, a variety of minor structures may appear including adhesion ripples and tears (Berry, 1973), saltation impact depressions, pedestals (Carter, 1978; Vortisch and Lindstrom, 1980) and tilted stones (Mattson, 1976). Most of these structures are destroyed as deflation proceeds. For example, laboratory experiments on 5 mm thick crusts suggest that the time taken both for pitting (impact craters) and penetration (crust collapse) is inversely related to an exponent of wind velocity above 4.4 m s^{-1} (Franzen, 1989). In a wind speed of 14 m s^{-1}, crusts were destroyed within 40 minutes. In contrast to erosional forms, Ahlbrandt and Fryberger (1981) note that where water tables are rising, aggradation may take place, and adhesion ripples can provide the basis for depositional features as high as a metre.

The formation of a deflation surface radically alters the near-surface air flow (Carter, 1976). Coarse lag deposits often allow particle overpassing (Everts, 1973) as single grains or as discrete bedforms ranging in size from ripples to small dunes (Fig. 7B). Transport is often within strong turbulent airflow, characterized by streaking, bursting and kinematic waves, so that movement is spatially and temporally intermittent (Allen, 1985). Some deflation surfaces may accumulate sediment in interparticle voids during periods of light winds. As wind velocity increases, this stored material is rapidly re-entrained and transported downwind. The ability of coarse-grained deflation surfaces to trap and then release material enables a transport flux to be maintained.

Deflation surfaces linked to water tables may become occupied by dune slacks, ponds (lagoons) or sabkhas (evaporite interdunes) and they may attain considerable size. The deflation plain of the Oregon dunes on the Pacific northwest coast of the USA is up to 2 km wide, and some deflation plain surfaces have increased 0.8 km in width the past 50 years. The rapid expansion is attributed to the elimination of sediment input from the beach caused by high, linear foredunes (Fig. 8) that built up following the spread of exotic European beach grass, *Ammophila arenaria* (Pinto *et al*, 1972; Wiedemann, 1984).

Some lagoons that form on deflation surfaces are large enough to form distinct shorelines, so that significant reworking of marginal dunes may occur (Ahlbrandt and Fryberger, 1981). On occasions, almost complete deflation may occur, leaving broad terrace-like structures (Segerstrom, 1962). The coastal dunes may reform downwind as coherent bedforms, such as transverse ridges, parabolics or sand seas (Hesp *et al.*, 1989). Alternatively the sand may disperse into cover sands (e.g. Kocurek and Nielson, 1986) engulfing wide areas. One especially advanced deflation form is the 'machair' (Fig. 9A), a widespread, vegetated dune plain found in western Scotland and northwest Ireland (Ritchie, 1976; Bassett and Curtis, 1985). This flat or slightly landward-dipping feature probably originates from a mixture of factors including slow sea-level change, strong onshore winds, failing sediment supply and anthropogenic intervention. Ritchie

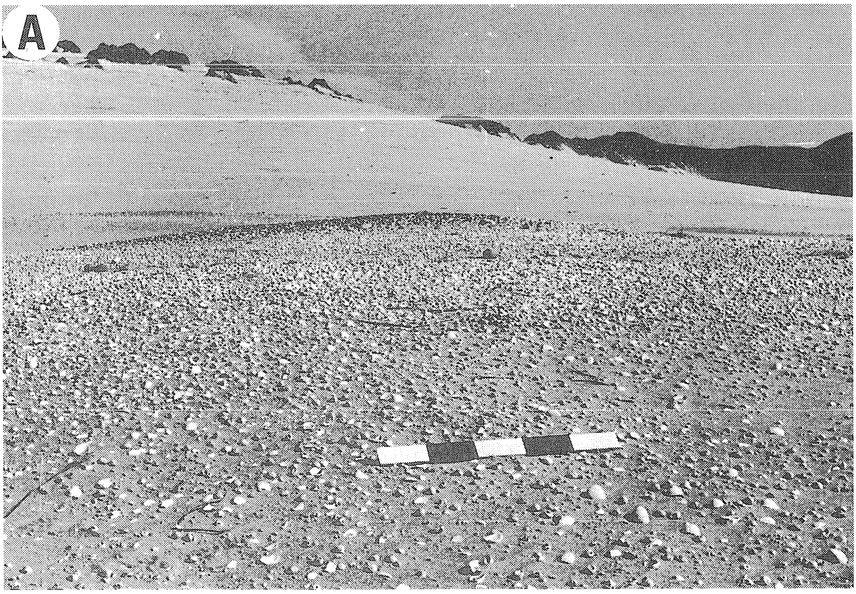


Fig. 7A A shell and gravel lag deposit forming an extensive interdune surface

Fig. 7B Small barchan dunes moving across a shell deflation surface. The foreground barchan is 0.8 m high and moving right and away from the camera exposing the internal structure. Note the scoured deflation surface with numerous 'tear' marks and shell pedestals

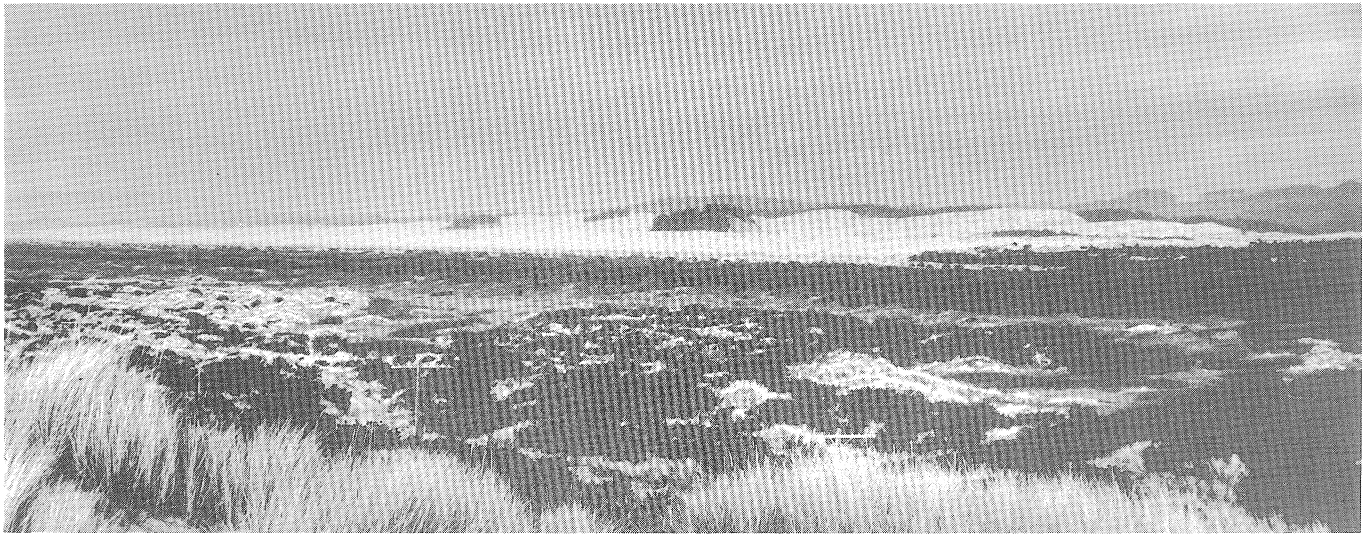


Fig. 8 The deflation plain in Oregon Dunes National Recreation Area showing the high foredune upwind (foreground), the vegetated deflation plain and the active migrating dunes downwind (background)

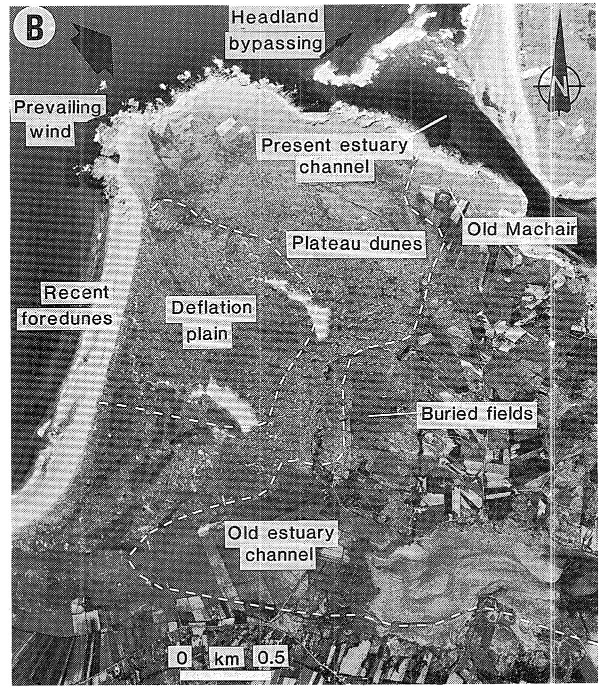


Fig. 9A Dune machair in northwest Ireland. Machair appears to be a strongly erosional landform, dominated by deflation to the water table

Fig. 9B The dune machair at Trawenagh Bay in Co. Donegal was formed in the late eighteenth century following destabilization of high dunes by agricultural impact. Blowing sand engulfed numerous buildings and farms leaving an extensive machair plain at the water table

(1977) proposes two models of machair development, both involving erosion and subsequent landward dispersal of a coastal foredune and deflation to a base imposed by the water table. Many machairs in Ireland rise steeply at the seaward margin, which is often cut by semi-circular blowouts with rim dunes, marking active landward sediment movement (Carter, 1990). In Co. Donegal, Ireland, nineteenth-century destabilization of the dunes at Trawenagh Bay (Fig. 9B) produced a modern machair plain. The sediment deflated from this site was blown over a wide area, much of it infilling and closing a tidal channel.

4. BLOWOUTS

4.1. Terminology

The generic term blowout is usually employed to describe an erosional hollow, depression, trough, or swale within a dune complex. These landforms were identified by early workers as important morphological elements (the 'trough-shaped wind sweeps' of Cowles, 1898). The actual word 'blowout' appears to have gained scientific acceptance in Melton's (1940) paper on the semi-arid dunelands of the southern High Plains and Bagnold's (1941) monograph on desert landforms, although the term was well known before (Kurz, 1942; Oosting and Billings, 1942). Melton used the term to describe parabolic dunes arising from the deflation of sand surfaces, whereas Bagnold's blowouts referred to wind-scoured gaps in an otherwise continuous transverse dune. Despite Brothers' (1954) and Landsberg's (1956) use of Melton's definition to explain parabolic coastal landforms in New Zealand and Scotland, Bagnold's definition has gained common currency, through the works of Laing (1954), Cooper (1958), Olson (1958) and Ranwell (1972).

4.2. Initiation of blowouts

Blowouts form readily in vegetated dunes (Fig. 10), where stable and unstable morphologies may co-exist. There are a number of ways that blowouts are initiated. Most of these involve the acceleration of wind where deflation potential has increased due to shoreline erosion and/or washover (Laing, 1954; Godfrey, Leatherman and Zaremba, 1979), vegetation die-back and soil nutrient deficiency (Jungerius, Verheggen and Wiggers, 1981), destruction of vegetation by animals (rabbits—Ritchie, 1972, Ranwell and Boar, 1986; bears—Martini, 1981), overland flow (Jungerius and van der Meulen, 1988) and diverse human activities including recreation (Mather and Ritchie, 1978) and fencing and house-building (Nordstrom and McCluskey, 1984). Perhaps the most intriguing of these initiating mechanisms is the cyclic model of Jungerius, Verheggen and Wiggers (1981). These authors suggest that periodic, spatially impersistent die-back of vegetation due to natural nutrient depletion leads to disintegration of the soil

Fig. 10A Small coastal blowouts forming along the shoreline. Note the 'healed' blowout to the right, and the 'closed' blowout topography farther inland

Fig. 10B Elongate trough blowout comprising a deep deflation basin, lateral erosion walls, trailing ridges and a depositional lobe. The blowout has been migrating at 18 m a^{-1} and the houses are in imminent danger of engulfment (photograph by S. Chape)

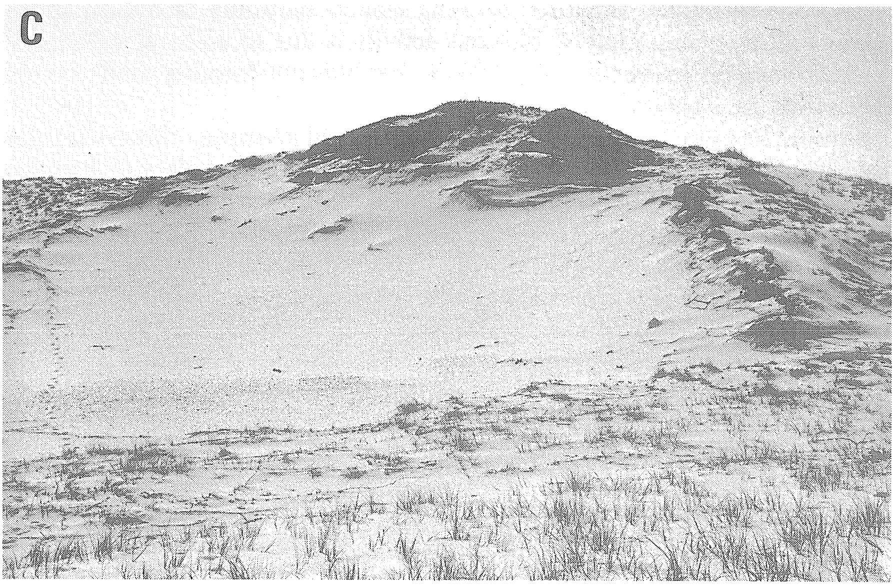


Fig. 10C Shallow saucer blowout developing on the stoss face of an established foredune. A coarse deflation lag has formed at the base of the blowout. Steep, low erosion rims mark the lateral margins of the blowout, and sand is deposited in small asymmetric bedforms immediately downwind

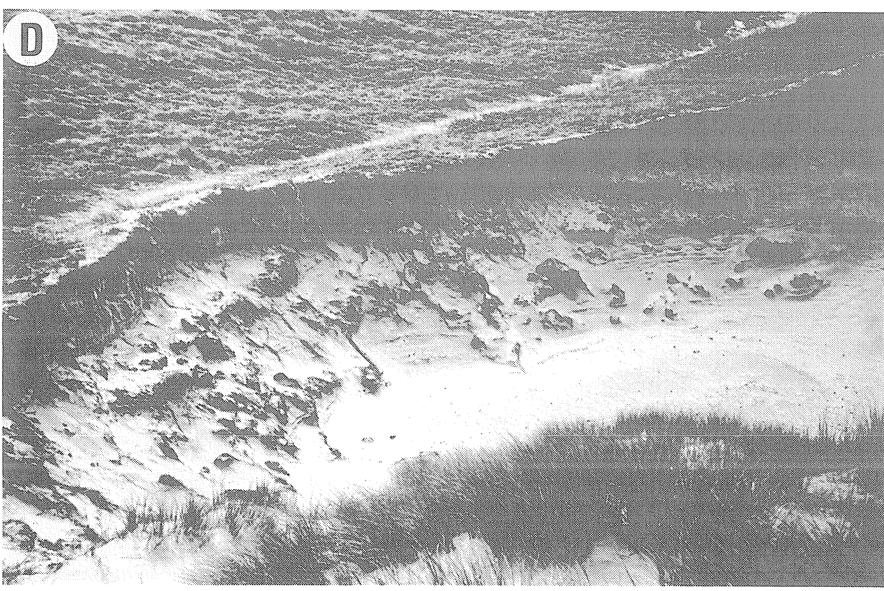


Fig. 10D A trough blowout dissecting stable dune grassland. The cross-sectional form of the blowout is highly asymmetrical with erosion (farthest from camera) and accretion (nearest) surface and eventually deflation, forming shallow blowouts. In almost all the other cited cases initiation of blowout activity is due to external influences. Jungerius and his co-workers also remark that blowout-forming winds are not necessarily the strongest.

Blowout topography need not arise from erosional processes. 'Blowouts' may develop as areas of non-deposition between mobile dune ridges (Gares and Nordstrom, 1988) or as gaps in incipient foredunes that remain open as the dune grows around them (Hesp, 1984; Carter and Wilson, 1988). Blowouts of non-erosional origin often assume the incised side wall characteristics of their erosional counterparts, and discrimination is not always straightforward. Blowout orientation appears to depend on antecedent topography and patterns of external disturbance as well as prevailing winds. Most authors describe a preferred orientation broadly commensurate with prevailing winds (Landsberg, 1956; Jungerius, Verheggen and Wiggers, 1981), but Gares and Nordstrom (1988) note three clear orientations on the New Jersey coast associated with storm wind direction (44% total blowouts), dominant winds (18%) and pedestrians (33%). Many blowout orientations on the Irish coast are shore-normal (Carter, 1990) (Fig. 10A), and probably reflect local micro-climates with wind blowing parallel or sub-parallel to the beach, and veering obliquely into the dunes.

Although there is a large variety of blowout morphologies (see Ritchie, 1972), two basic types have often been identified (Cooper, 1958), the saucer blowout (Figs 10A and C) and the trough blowout (Figs 10B and D). Saucer blowouts are shallow, ovoid, dish-shaped hollows with a steep marginal rim and commonly a flat-to-convex downwind depositional lobe. Trough blowouts are relatively deep, narrow, steep-sided topographies with more pronounced downwind depositional lobes, and marked deflation basins. Along the Australian east coast, saucer blowouts tend to develop on low gradient slopes on the windward faces of large foredunes, and on low, rolling dune topography where the vegetation cover has been locally removed. Trough blowouts are particularly well-developed where they cut through high dunes (e.g. foredunes), and they commonly evolve into parabolic dunes.

4.3. Blowout dynamics

Once initiated, the dune blowout will enlarge through a combination of deflation and slope or side wall failure. Most active blowouts enlarge laterally by wind scour that oversteepens side walls and leads to slumping and avalanching. They enlarge vertically by deflation of the blowout floor and extend downwind by deflation of the original sand surface and migration of the depositional lobe.

The wind flow in blowouts is topographically accelerated and altered as it moves through the landform. Fig. 11 illustrates wind velocity profiles through a narrow, relatively deep (8 m) trough blowout (Fig. 12). Fig. 11 indicates that jet flow is common within the blowout. Wind speeds are significantly accelerated

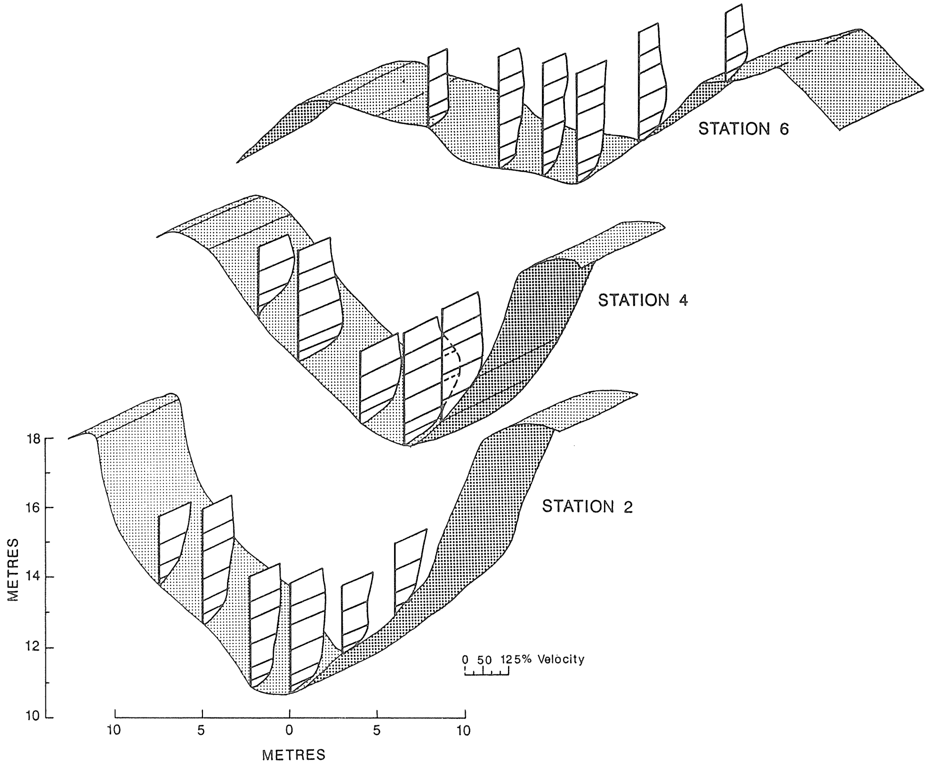


Fig. 11 Wind velocity profiles in a narrow blowout from South Australia. The profiles are expressed as a function of a permanent station sited in the middle of the blowout. The three-dimensional one metre topographic slices are taken from the throat entrance region (Station 2), mid-deflation zone (Station 4) and lower stoss slope of the depositional lobe (Station 6)

up the axis of the deflation basin (see centre profile, station 4). The steepness of the lateral erosion walls accelerates the wind flow and high jet velocities occur along the walls. Maximum shear stresses and sediment transport occur along the base of the trough and along the steepest part of the wall. As the blowout trough expands, for example, where it has cut through to the lee side of a large foredune (Fig. 11), jet flows expand and decelerate. This deceleration occurs in a radial concentric pattern across the blowout depositional lobe, whenever winds blow directly up the blowout.

When winds approach the blowout directly, these flows maximize sand transport and erosion along the deflation basin and the side walls. Flows are generally strongest up the centreline axis and decrease away from it. Shallow basins result. Erosion of the deflation basin removes support for the sides of

Fig 12A Entrance/throat region of the studied blowout illustrating narrow deflation floor (partially revegetated in upwind region), steep lateral walls and skewed orientation

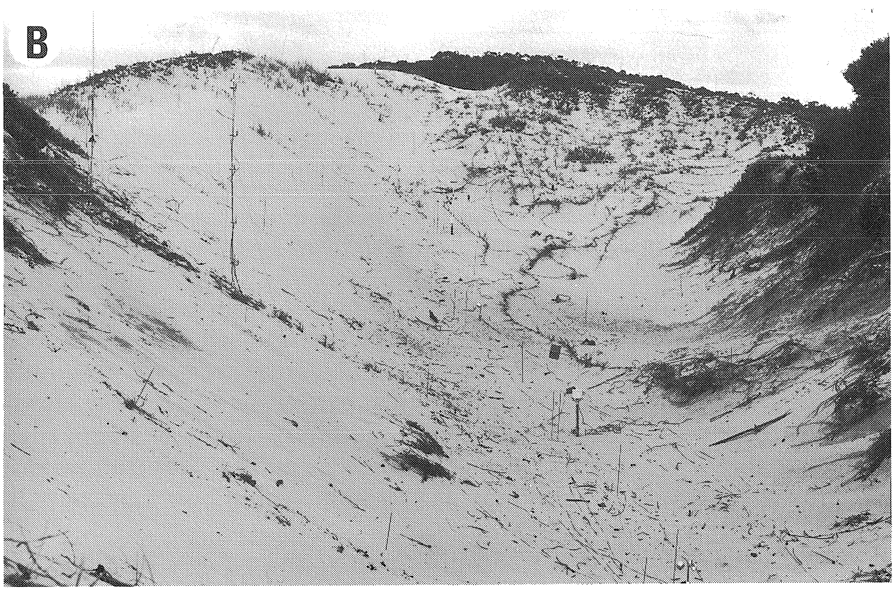


Fig 12B Upper blowout illustrating upwind face of the depositional lobe, narrow deflation floor and erosional walls (with instruments)

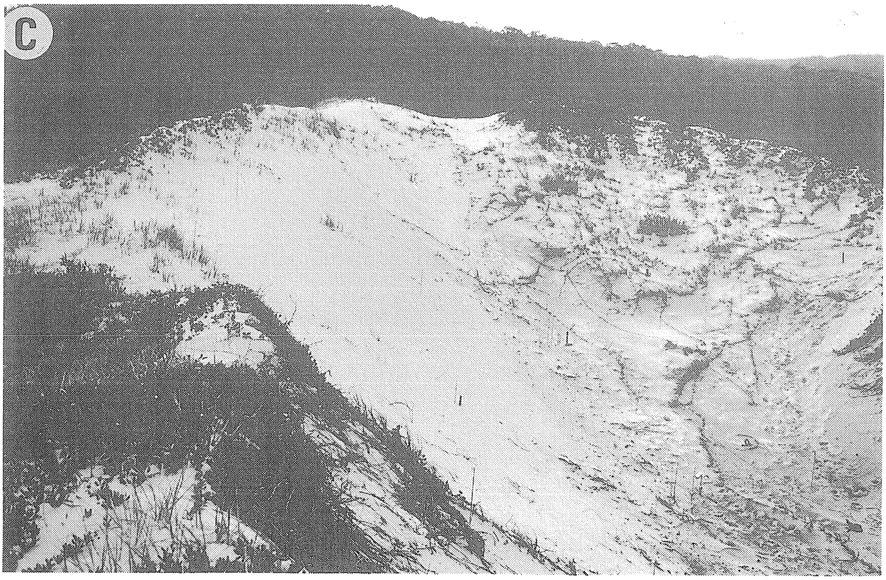


Fig. 12C Similar to B except for inch-view of lateral wall 'overtop' deposition in foreground on left

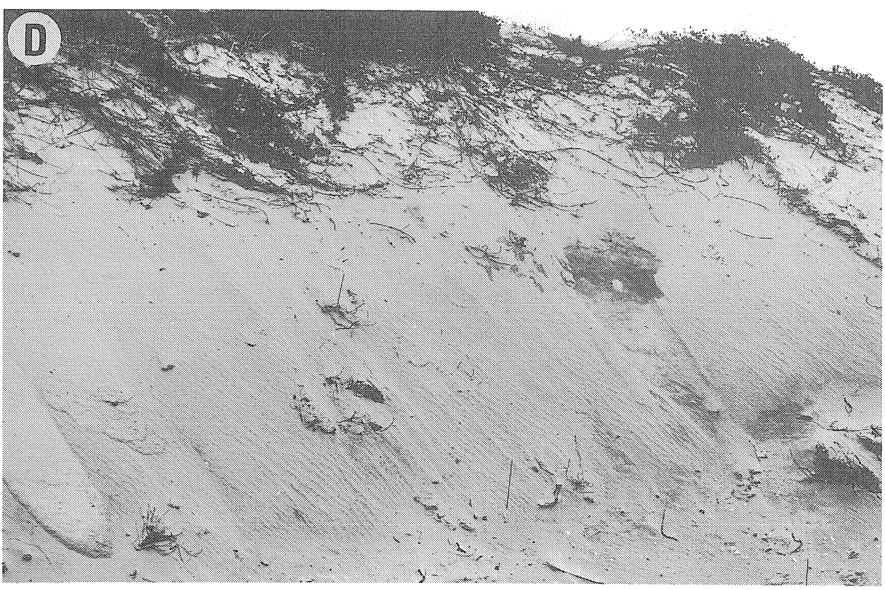


Fig. 12D Southwestern lateral wall indicating steep upper slope held by vegetation and roots, mid-lower slope dominated by avalanche cones, and deflation floor (right-hand bottom). The original lee slope of the foredune may be seen as indicated by the soil profile in the mid right-hand portion of the avalanche material

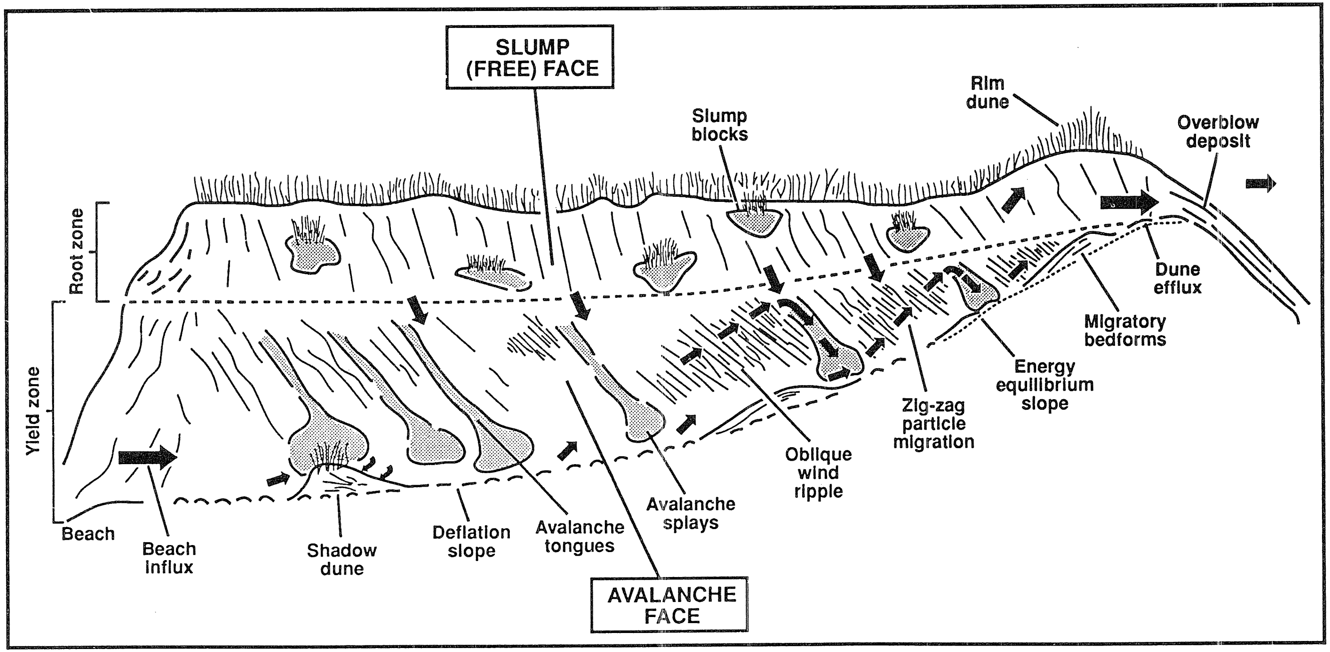


Fig. 13 Schematic long profile of a dune blowout indicating the slope facets and deposition forms. Arrows indicate the likely transport direction if wind is blowing from the beach

the blowout. Side walls are generally composite slopes with an upper free face (>40°) formed in the vegetation root zone overlying a lower, loose avalanche incline at or near the angle of initial yield (30 to 34°) (Fig. 12D). The transition between the free face and the avalanche slope is often sharp. Sediment tends to accumulate along this transition before falling back under gravity to the blowout floor, from where it is removed. Sediment is also transported directly along the blowout walls by oblique ripple migration, often across very steep slopes.

Regular exchanges of sediment between side wall and basin often occur in a zig-zag migration pattern combining ripple patches and small avalanches (Fig. 13). The blowout floor is often inclined, with the lower, flatter segment provided by a hard deflation surface (often shell or gravel) and the upper by an energy equilibrium facet reflecting the increasing wind velocity through the blowout. Spiralling helicoidal flows moving along the side walls transport sediment from the upper wall over the crest and into marginal vegetation forming rim dunes along the lateral blowout margins. Rapidly decelerating flows result in maximum sediment transport up the blowout axis with decreasing movement towards the lateral margins (Fig. 11), producing a parabolic-shaped depositional lobe. This is formed of two elements: (i) a delta-like 'lobe', often semi-circular in shape dominated by largely unvegetated foreset deposition; and (ii) beyond the lobe, a semi-circular or elliptical zone of slow deposition, perhaps marked by more vigorous sand-binding vegetation.

Where the primary wind approaches a blowout obliquely it may be directed by the topography. Many blowouts are highly asymmetrical with one side wall displaying under cutting while the other is dominated by accretion or is in the process of revegetating. Such occurrences are common on the east Australian coast where many blowouts are initiated by occasional SE storms but experience frequent NE to S prevailing winds. Here the wind enters the blowouts obliquely, attacking the facing (exposed) lateral wall, but not the sheltered wall. Blowout depositional lobes are also skewed obliquely as the wind is blowing preferentially out one side of the blowout.

4.4. Evolution of blowouts

The blowout enlarges as the side walls recede and the deflation area extends downwind. Almost all blowouts are limited in terms of depth, with erosion either arrested through the formation of a lag deposit, or controlled by the presence of a fluctuating water table (Ritchie, 1972).

There appears to be a fundamental distinction between 'open' and 'closed' blowouts, with the former having clearly defined wind gaps into and/or out of the hollow. Sediment flux is greater through open blowouts, and many act as transport corridors.

Jungerius, Verheggen and Wiggers (1981) point to a relative constancy in width/depth ratios (between 3 and 6), although Wilcock (1976), working in a more open dune system, suggested that blowout width eventually became independent of blowout depth once downward erosion ceased.

The length of blowouts usually depends on the available relief. The potential length of the small blowouts studied by Jungerius, Verheggen and Wiggers (1981) was largely unrestricted, yet most failed to develop beyond 30 m along the axis

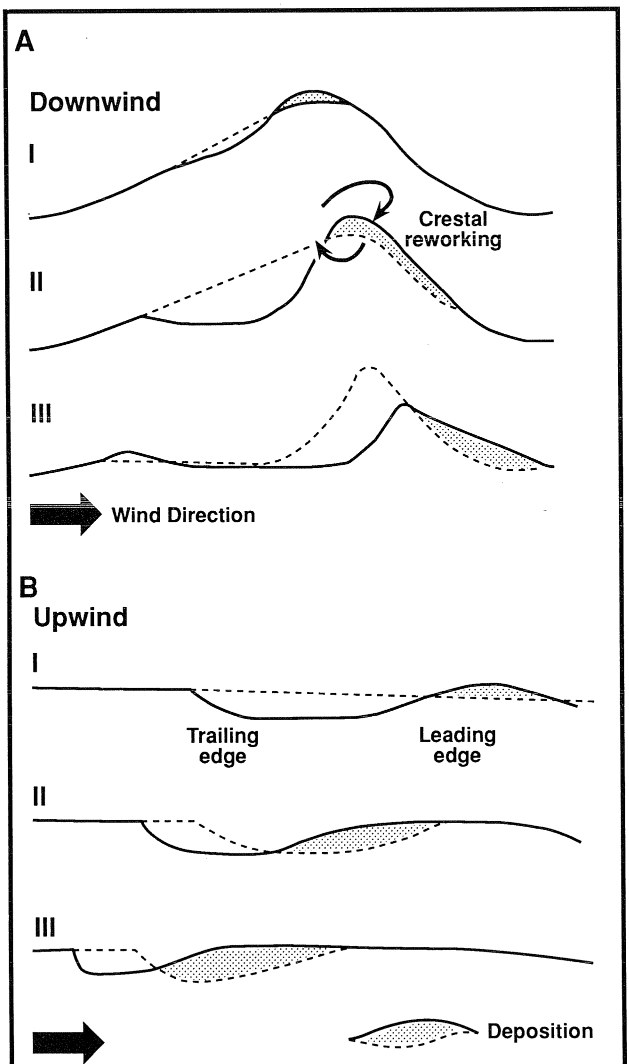


Fig. 14 Schematic views of downwind (A) and upwind (B) migration of blowouts. In A sediment efflux exceeds blowout storage and the form enlarges. In B upwind erosion is more than balanced by downwind infilling, with the result that the form appears to move into the prevailing wind direction

of the prevailing wind, perhaps because the wind-run over a 30 m length scale is accelerated insufficiently to continue sand transport downwind, so that the blowout becomes dormant and eventually revegetates. Blowouts may grow large enough to breach their host dune (Landsberg,1956; Ritchie, 1972; Gares and Nordstrom, 1987). The resulting breach may function more as a transport corridor, moving sand through the dune ridge, rather than directly from it. Breaching usually involves the flattening of the axial slope, presumably indicating a relative deceleration of airflow within the blowout, commensurate with a reduction, if not a cessation, of erosive activity.

Blowouts may migrate downwind or upwind. The downwind migration (Fig. 14) has the leading erosional edge advancing slowly with the wind, often infilling leeward of the deflation zone (Landsberg, 1956; Ritchie, 1972; Martini, 1981). This type of blowout leads to extensive reworking of the dune sediments. Blowouts may migrate upwind, eroding at the upwind edge and accumulating downwind (Jungerius, Verheggen and Wiggers, 1981; Jungerius and van der Meulen, 1989) (Fig. 14).

Significant topographic changes can often be measured over periods of a few years (Ritchie, 1972; Wilcock, 1976; Gares and Nordstrom, 1987). The rate of development varies between sites, depending on the direction, frequency and force of sand-transporting winds, exposure and pre-existing relief, the degree of vegetation cover and the characteristics of both the sand and the sand body.

Fig. 15 shows the typical arrangement of sediments within a single blowout at Portrush, Northern Ireland. Whereas the total deflation from the blowout is about 7200 m^3, almost twice this amount (13 600 m^3) is deposited in the rim dunes and the depositional plumes, showing the utility of the blowout to transfer beach material inland. Carter (1980) estimated that almost 0.25 m of an annual shoreline erosion rate of 0.3 m might be ascribed to this process.

Gares and Nordstrom (1987) measured 3.03 m of deflation in 5 years at the location where a gap formed in the foredune (Fig. 16A) while 2.02 m of deposition occurred adjacent to this location, in the direction of the resultant of the northeast storm winds and dominant northwest winds. Sediment trap data gathered at this blowout over a 5-month period in 1981 prior to breaching revealed that the highest transport rates were at the side wall on the south, which was eroded by the northwesterly winds, and at the saddle south of the blowout. The rate of transport through the saddle was higher than at any other location, including the open beach, and documents the effects of topographic channelling.

Sediment entering these oblique blowouts often collects in the throat (Fig. 16A and B) or on the outer side wall. The height of the (shadow) dune landward of the gap in Fig. 16B grew 1.8 m between 1982 and 1988. The most rapid growth followed creation of the gap by the dominant northwest winds. Sediment transported by northeast and southeast storm winds created the new zone of deposition northwest of the gap. More sediment accumulated south of the gap,

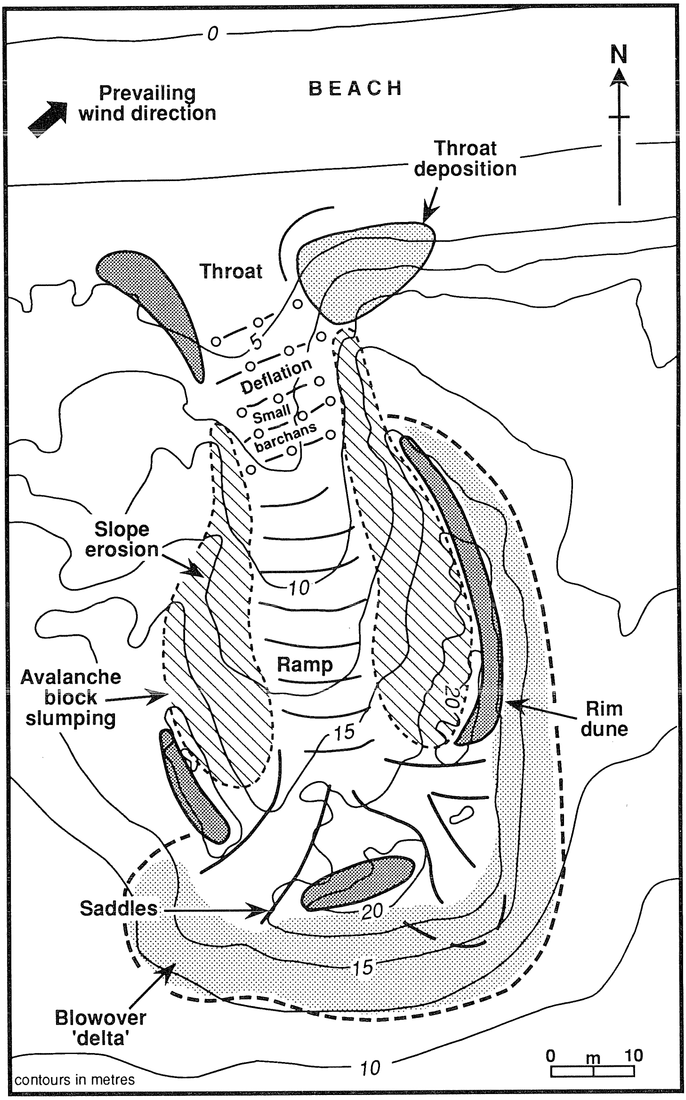


Fig. 15 Landforms and sedimentary features in a small shore-normal blowout at Portrush, Northern Ireland. The blowout comprises a series of slope facets and depositional zones commensurate with its role in the erosion and transfer of coastal sediment landward

where sand delivered both from the beach and from the blowout (by northwest winds) was blown over the crest by northeast winds. The gap thus provides a conduit for delivery of sand to the dune crest and to former low ground within the blowout as well as a conduit facilitating removal of sand from other portions of the blowout.

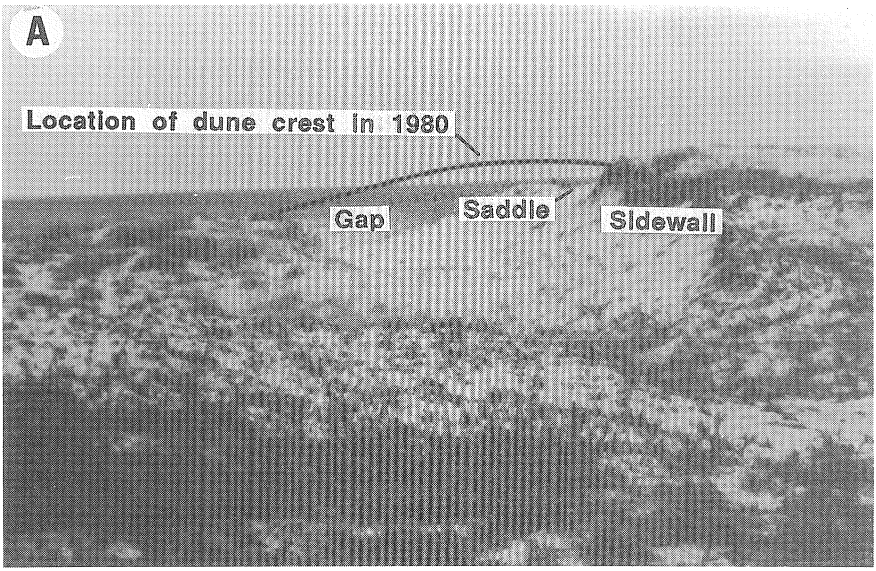


Fig. 16 Two views of deflation (A) (1982) and subsequent accretion (B) (1989) in a blowout at Island Beach State Park, New Jersey

5. EROSIONAL DUNE COMPLEXES

Complex blowouts may arise where there is great initial topographic variability (e.g. massive dunefields) or where local zones of deposition and scour alter wind and sedimentation through complex feedback mechanisms. Laing (1954) identifies small blowouts forming within large ones, while David (1977) and Filion and Morisset (1983) record a variety of dune blowout forms (imbricate, *en échelon*, digitate and hemicycle). A common complex form found in Ireland is a double or stacked blowout (Fig. 17), in which the formation of a blowout on the lower dune slopes initiates another on the upper slope. These blowouts have two erosional/depositional slope facets separated by an intermediate crest. Visual observations indicate that deposition on the intermediate crest disturbs airflow and vegetation on the upper slope, creating another erosional slope. The bottom part of Fig. 17 indicates the sediment transport pathways up this

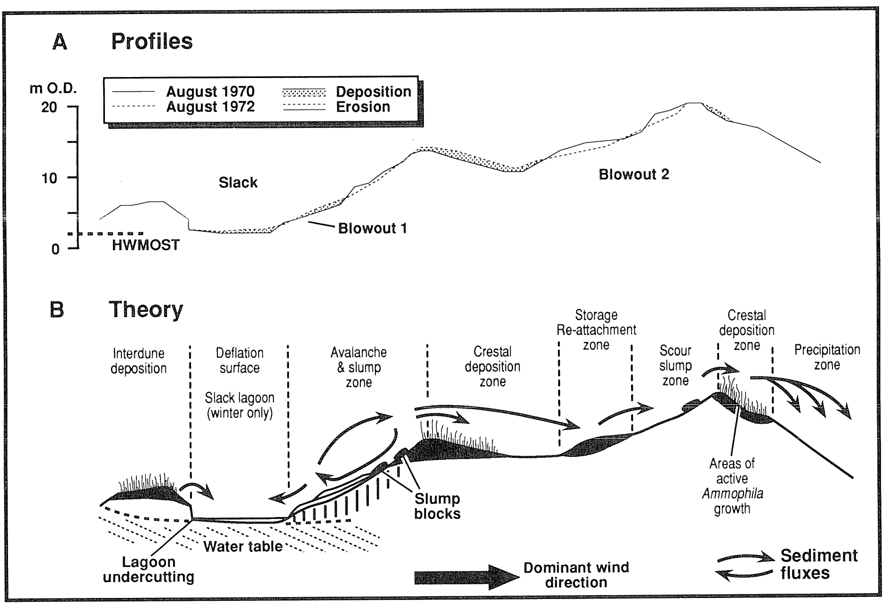
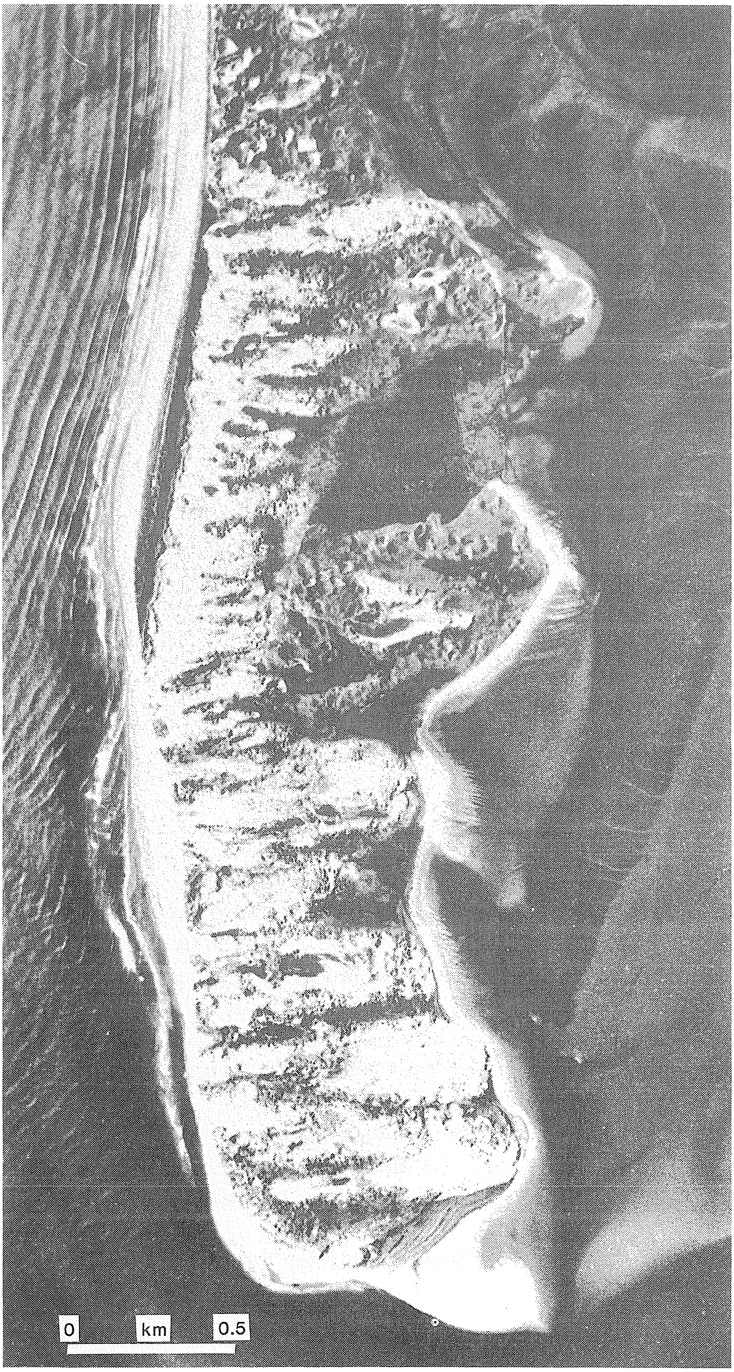


Fig. 17 A cross-profile (A) of a stacked pair of blowouts at Portstewart, Northern Ireland. Erosion of the lower blowout (1) leads to the formation of the upper blowout (2). An anatomical description of this situation is given in B

Fig. 18 (opposite) The erosional dune complex at Inch, Co. Kerry, Ireland. This unstable dunefield has been hugely dissected by onshore winds. The photograph shows numerous linear trough blowouts allowing sediment to move landward. Much of the material is stored in depositional lobes several hundred square metres in area, before being transported onto the estuarine flats and rejoining the marine transport system via the tidal channel to the south


0
km
0.5

slope. The sequence of events at the blowout portrayed in Fig. 15 also demonstrates how topography redirects wind and creates new locations of accretion and scour, resulting in a dynamic and complex topography. Constant reworking (or cannibalization) of depositional landforms often leads to complete topographic reorganization.

A dominantly beach-aligned system can become dissected by secondary forms, breaking up the simple dune-interdune pattern, and creating a more complex aerodynamic environment, which leads to more irregular topography. In time, the dune complex may become almost divorced from its associated beach system, with internal reorganization subsuming any external forcing factors. Garcia-Nova, Ramirez Diaz and Torres Martinez (1975) provide an example of this from the Donaña dunes in southern Spain.

Erosional complexes are competitive (Cooper, 1958). Major forms subsume minor forms; minor forms are superimposed on major forms; and locations of accretion and scour are transposed (Fig. 18). The resulting diverse and dynamic geomorphological landscape supports a varied vegetation succession with steep ecological gradients and species richness. The varied vegetation, in turn, contributes to further differences in sedimentation rates. Such a landscape is almost fractal in nature as process-form interactions are replicated at many scales, and the gradual evolution into chaotic form is often evident. The interpretation of these landscapes is a challenging task for geomorphologists.

6. CONCLUSION

Development of erosional landforms (especially blowouts) through time requires more extensive investigation. A considerable proportion of the literature on dune management is aimed at prescribing remedies for erosional landforms, although there is a paucity of information to determine the extent to which such forms are self-healing. Long-term studies of the origin and history of blowouts would be particularly valuable given the probability of natural cycles of stability and instability associated with both internal (nutrient cycles, plant successions, animal population fluctuations) and external (climatic changes, human alterations) factors. The net result may be a relatively balanced system, with no long-term losses or gains of sediment, species or productivity. However, the role of blowouts in dune system dynamics has yet to be defined sharply.

Given the great ecological value of erosional dunescapes, effective collaborative research between geomorphologists and ecologists would seem to be an essential prerequisite for future progress. The sheer complexity of erosional dunes offers a major research challenge to coastal scientists.

ACKNOWLEDGEMENTS

Many thanks to Paul Gares, Norb Psuty and Peter Wilson for finding time to comment on this chapter, and to Mary McCamphill, Kilian McDaid and Nigel McDowell for typing, drafting and photography.

REFERENCES

Ahlbrandt, T. S. and Fryberger, S. G. (1981) Sedimentary features and the significance of interdune deposits. In Ethridge, R. G. and Flores, R. M. (eds), *Recent and Nonmarine Depositional Environments: Models for Exploration*. SEPM, Tulsa, Okla., pp. 293–314.

Allen, J. R. L. (1985) *The Principles of Sedimentology*. Allen & Unwin, London, 272pp.

Aufrère, L. (1931) Le cycle morphologique des dunes. *Ann. Geogr.*, **40**, 34–49, 362–385.

Bagnold, R. A. (1941) *The Physics of Blown Sand and Desert Dunes*. Methuen, London, 256pp.

Bassett, J. A. and Curtis, T. G. F. (1985) The nature and occurrence of sand-dune machair in Ireland. *Proc. R. Ir. Acad.*, **85B**, 1–20.

Berry, R. W. (1973) A note on asymetrical structures caused by differential wind erosion of a damp, sandy forebeach. *J. sediment. Petrol.*, **42**, 205–206.

Bremontier, N. T. (1833) Mémoire sur les dunes. *Ann. Pont. Chaussé.* **69**, 145–224.

Brothers, R. N. (1954) A physiographic study of the recent sand dunes on the Auckland west coast. *N. Z. Geogr.*, **10**, 47–59.

Carter, R. W. G. (1976) Formation, maintenance and geomorphological significance of an eolian shell pavement. *J. sediment. Petrol.*, **46**, 418–429.

Carter, R. W. G. (1978) Ephemeral sedimentary structures formed during Aeolian deflation of beaches. *Geol. Mag.*, **115**, 379–382.

Carter, R. W. G. (1980) Vegetation stabilisation and slope failure on eroding sand dunes. *Biol. Conserv.*, **18**, 117–122.

Carter, R. W. G. (1990) Geomorphology of the Irish coastal dunes. *Catena Suppl.* **18**, 31–39.

Carter, R. W. G. and Rihan, C. L. (1978) Shell and pebble pavements on beaches; examples from the north coast of Ireland. *Catena*, **5**, 365–374.

Carter, R. W. G. and Stone, G. W. (1989) Mechanisms associated with failure of eroding sand dunes, Magilligan, Northern Ireland, *Earth Surf. Proc. Landf.*, **14**, 1–10.

Carter, R. W. G. and Wilson, P. (1988) Geomorphological, sedimentological and pedological influences on coastal dune development in Ireland. In Psuty, N. P. (ed) *Dune/Beach Interaction, J. Coast. Res. Spec. Issue No. 3*, pp. 27–31.

Cooke, R. U. (1970) Stone pavements in deserts. *Ann. Assoc. Amer. Geogr.*, **60**, 560–577.

Cooper, W. S. (1958) *The Coastal Sand Dunes of Oregon and Washington*. Geol. Soc. Am. Mem. **72**, 169pp.

Cowles, H. C. (1898) The ecological relations of the vegetation on the sand dunes of Lake Michigan. *Bot. Gaz.*, **27**, 97–117.

David, P. P. (1977) *Sand dune occurrences of Canada*. Dept. Indian Affairs/Geol. Surv. Canada Rept. 183pp.

Davies, J. L. (1972) *Geographical Variation in Coastal Development*. Oliver & Boyd, Edinburgh, 204pp.

Dolan, R. (1972) Barrier dune systems along the Outer Banks of North Carolina: a reappraisal. *Science*, **176**, 286–288.

Edelman, T. (1968) Dune erosion during storm conditions. *Proc. 11th Conf. Coast. Eng.*, 719–722.

Edelman, T. (1972) Dune erosion during storm conditions. *Proc. 13th Conf. Coast. Eng.*, 1305–1311.

Engel, J. R. (1981) Sacred sands: the civil religion of the Indiana Dunes. *Landscape*, **25**, 1–10.

Engel, J. R. (1983) *Sacred Sands*. Wesleyan University Press, Middleton, Conn., 352pp.

Everts, C. H. (1973) Particle overpassing on flat granular boundaries. *J. Waterway Harb. Coast. Eng. Div. ASCE*, **99**, 425–438.

Filion, L. and Morisset, P. (1983) Eolian landforms along the eastern coast of Hudson Bay, Northern Quebec. *Nordicana*, **47**, 73–94.

Foss, P. J. (1985) Some observations on 'sea balls' discovered in West Donegal. *Ir. Nat. J.*, **21**, 526–528.

Franzen, L. G. (1989) Experimental studies of eolian erosion on a dune sand surface, protected by an artificial crust. *Zeit. Geomorph. NF*, **33**, 355–360.

Garcia Nova, F., Ramirez Diaz, L. and Torres Martinez, A. (1975) *El sistema de duñas de Doñana*. Publ. No. 5 ICONA, Ministerio di Agricultura, Madrid.

Gares, P. A. and Nordstrom, K. F. (1987) Dynamics of a coastal foredune blowout at Island Beach State Park, N. J. *Proc. Coast. Sed.*, '87, *ASCE*, 213–221.

Gares, P. A. and Nordstrom, K. F. (1988) Creation of dune depressions by foredune accretion. *Geogr. Rev.*, **78**, 194–204.

Godfrey, P. J., Leatherman, S. P. and Zaremba, R. (1979) A geobotanical approach to classification of barrier beach systems. In Leatherman, S. P. (ed.), *Barrier Islands*. Academic Press, New York, pp. 99–126.

Goldsmith, V. (1973) Internal geometry and origin of vegetated coastal sand dunes. *J. sediment. Petrol.*, **43**, 1128–1142.

Greenway, D. R. (1987) Vegetation and slope stability. In Anderson, M. G. and Richards, K. S. (eds), *Slope Stability*, Wiley, Chichester, pp. 187–230.

Hesp, P. A. (1984) Foredune formation in southeast Australia. In Thom, B. G. (ed.), *Coastal Geomorphology in Australia*. Academic Press, Sydney, pp. 69–77.

Hesp, P. A. (1988) Morphology, dynamics and internal stratification of some established foredunes in southeast Australia. *Sediment. Geol.*, **55**, 17–41.

Hesp, P. A., Illenberger, W., Rust, I., McLachlan, A. and Hyde, R. (1989) Some aspects of transgressive dunefield and transverse dune geomorphology and dynamics, south coast South Africa. *Zeit. Geomorph. Suppl.-Bd*, **73**, 111–123.

Hughes, S. A. and Chui, T.-Y. (1981) *Beach and dune erosion during severe storms*. University of Florida, Dept. Coastal and Oceanographic Engineering, Report UFL/COEL-TR/043, 290pp.

Jungerius, P. D. and van der Meulen, F. (1988) Erosion processes in a dune landscape along the Dutch coast. *Catena*, **15**, 217–228.

Jungerius, P. D. and van der Meulen, F. (1989) The development of dune blowouts, as measured with erosion pins and sequential air photos. *Catena*, **16**, 369–376.

Jungerius, P. D., Verheggen, J. T. and Wiggers, A. J. (1981) The development of blowouts in 'de Blink' a coastal dune area near Noordwijkerhout, The Netherlands. *Earth Surf. Proc. Landf.*, **6**, 375–396.

Kocurek, G. and J. Nielson (1986) Conditions favourable for the formation of warm-climate eolian sand sheets. *Sedimentol.*, **33**, 795–816.

Kurz, H. (1942) *Florida dunes and scrub*. State of Florida Dept. Conservation, Geol. Bull. No. 23, 117pp.

Van der Meulen, T. and Gourlay, M. R. (1968) Beach and dune erosion tests. *Proc. 11th Conf. Coast. Eng.*, 701–707.

Vellinga, P. (1983) Predictive computational model for beach and dune erosion during storm surges. *Proc. Coast. Zone '83, ASCE*, 806–819.

Vellinga, P. (1984) *Movable-bed modelling law for coastal dune erosion*. Water Port Coast Ocean Eng. Div. *ASCE*, **110**, pp. 495–504.

Vortisch, W. and Lindstrom, M. (1980) Surface structures formed by wind activity on a sandy beach. *Geol. Mag.*, **117**, 491–496.

Wiedemann, A. M. (1984) *The Ecology of Pacific Northwest Coastal Sand Dunes: A Community Profile*. US Dept. Interior, Fish and Wildlife Serv. Washington, DC, 130p.

Wilcock, F. A. (1976) Dune physiography and the Impact of Recreation on the North Coast of Ireland. Unpublished DPhil thesis, The New University of Ulster, Coleraine, 169pp.

Laing, C. C. (1954) The ecological life-history of the *Ammophila breviligulata* community on the Lake Michigan Dunes. Unpublished PhD thesis, Univ. Chicago, 108pp.
Landsberg, S. Y. (1956) The orientation of dunes in Britain and Denmark in relation to the wind. *Geogr. J.*, **122**, 176–189.
Leatherman, S. P. (1979) Barrier dunes—a reassessment. *Sediment. Geol.*, **24**, 1–16.
Lohnes, R. A. and Handy, R. L. (1968) Slope angles in friable loess. *J. Geol.*, **76**, 247–258.
Martini, I. P. (1981) Coastal dunes of Ontario: distribution and geomorphology. *Geogr. Phys. Quat.*, **35**, 219–229.
Mather, A. S. and Ritchie, W. (1978) The Beaches of the Highlands and Islands of Scotland. Countryside Commission for Scotland, Redgorton, Perth.
Mattson, J. O. (1976) Wind tilted pebbles in sand—some field observations and simple experiments. *Nordic Hydrol.*, **7**, 181–208.
Melton, F. A. (1940) A tentative classification of sand dunes: its application to dune history in the southern High Plains. *J. Geol.*, **48**, 113–174.
Nordstrom, K. F. and Lotstein, E. L. (1989) Perspectives on resource use of dynamic coastal dunes. *Geogr. Rev.*, **79**, 1–12.
Nordstrom, K. F. and McCluskey, J. M. (1984) Considerations for control of house construction in coastal dunes. *Coast. Zone Mngmt J.*, **12**, 385–402.
Olson, J. S. (1958) Lake Michigan dune development. *J. Geol.*, **66**, 345–351.
Oosting, H. H. and Billings, W. D. (1942) Factors affecting vegetational zonation on coastal dunes. *Ecology*, **23**, 137–139.
Orford, J. D. and Carter, R. W. G. (1985) Storm generated dune armouring on a sand gravel barrier system, southeastern Ireland. *Sediment. Geol.*, **42**, 55–82.
Parker, W. R. (1975) Sediment mobility and erosion on a multibarred foreshore (Southwest Lancashire, UK). In Hails, J. and Carr, A. P. (eds), *Nearshore Sediment Dynamics and Sedimentation*. Wiley-Interscience, London, pp. 151–179.
Pinto, C., Silovsky, E., Henley, F., Rich, L., Parcell, J., and Boyer, D. (1972) *The Oregon Dunes NRA Resource Inventory*. US Dept. Agriculture, Forest Serv., Pacific Northwest Region, Portland, Oregon, 294p.
Pye, K. (1983) Coastal dunes. *Prog. Phys. Geogr.*, **7**, 531–557.
Ranwell, D. (1972) *Ecology of Salt Marshes and Sand Dunes*. Chapman & Hall, London, 258pp.
Ranwell, D. and Boar, J. (1986) *Coast Dune Management Guide*. Inst. Terrestrial Ecol./NERC, 105pp.
Ritchie, W. (1972) The evolution of coastal sand dunes. *Scott. Geogr. Mag.*, **88**, 19–35.
Ritchie, W. (1976) The meaning and definition of machair. *Trans. Proc. Bot. Soc. Edinb.*, **42**, 431–440.
Ritchie, W. (1977) Machair development and chronology in the Uists and adjacent islands. *Proc. R. Soc. Edinb.*, **77B**, 107–122.
Segerstrom, K. (1962) *Deflated marine terrace as a source of dune chains, Atacama Province, Chile*. US Geol. Surv. Prof. Paper 450C, C91–C93.
Sokolow, N. A. (1894) *Die Dünen, Bildung, Entwickelung und Innerer Bau*. Berlin.
Tsoar, H. (1983) Wind tunnel modelling of echo and climbing dunes. In Brookfield, M. E. and Ahlbrandt, T. S. (eds), *Eolian Processes and Sediments*. Elsevier, Amsterdam, pp. 247–260.
Van de Graaff, J. (1977) Dune erosion during a storm surge. *Coast. Eng.*, **1**, 99–134.
Van de Graaff, J. (1986) Probalistic design of dunes; an example from the Netherlands. *Coast. Eng.* **9**, 479–500.
Van der Ancker, J., Jungerius, P. and Mur, L. (1985) The role of algae in the stabilization of coastal dune blowouts, *Earth Surf. Proc. Landf.*, **10**, 189–192.