Dune slope, not wind speed, best predicts bare sand in vegetated coastal dunes

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Abstract

Globally vegetation cover on coastal sand dunes has increased since at least the 1950s. With the aim of restoring or increasing biodiversity, land managers in several countries have removed vegetation and/or reprofiled dune slopes to reinvigorate geomorphic activity. However, the longevity of these interventions can be relatively short (on the order of 5 to 10 years), and further active management is required. Hypotheses for controls on geomorphic activity on dunes have frequently suggested that wind speedis the most important controlling factor. Here we show dune slope to be the best predictor of bare sand at four predominantly vegetated coastal sand dunes in England and Wales. We suggest that bare sand on steep dune slopes is maintained by three important factors: (1) Wind erosion, due to topographic acceleration (2) Granular avalanches of unconsolidated sediment and (3) Rotational slumping of unstable slopes. Our results indicate that where land managers wish to 'rejuvenate' areas of bare sand, efforts should focus on steep windward dune slopes and reprofiling of the dune slope should mimic the concave profiles of active slope faces on active parabolic dunes with an overall slope angle of between 18° and 23° from the dune toe to the crest.

Keywords Coastal Dune Management · Sand Dunes · Remobilisation · Bare Sand

Introduction

Coastal and desert dunefields experience varying amounts of aeolian activity depending on windiness,annual precipitation, and annual potential evapotranspiration (Lancaster 1988; Tsoar 2005). Since at least the 1950's, coastal sand dunes across the globe have demonstrated a trend of reduced mobility and increased vegetation growth (Pye et al. 2014; Delgado-Fernandez et al. 2019; Jackson et al. 2019; Gao et al. 2020). This change can be attributed to a reduction in wind power (Delgado-Fernandez et al. 2019; Zeng et al. 2019; Pye et al. 2020), increase in temperature (Delgado-Fernandez et al. 2019; Pye et al. 2020), a decline in grazing pressure (Levin and Ben-Dor 2004; Provoost et al. 2011)

and changes in human activity (Tsoar and Blumberg 2002; Provoost et al. 2011).

However, even in dunefields that are predominantly stabilised by vegetation, localised areas of geomorphic activity persist, often in the form of dynamic landforms such as blowouts, parabolic dunes and sand sheets (Arens et al. 2004; Smyth et al. 2020a). Conversely, where vegetation has been purposely removed and dune slopes reprofiled to reinvigorate geomorphic activity with the aim of restoring on increasing biodiversity, e.g., The Netherlands (Arens et al. 2004), Wales (Rhind and Jones 2009), Canada (Darke et al., 2013), New Zealand (Konlecher et al. 2015); Israel (Bar 2013) and France (Laporte-Fauret et al. 2021), substantial vegetation re-establishment is often rapid e.g. between two and eight years (Arens and Geelen 2006; Bar 2013; Barchyn and Hugenholtz 2013).

Abiotic factors and dune mobility

Among the abiotic factors that may control geomorphic activity on partially vegetated sand dunes, wind strength is frequently suggested to be the most important controlling factor (Arens et al. 2004; Wiggs et al. 1995; Pye 1982; Rust

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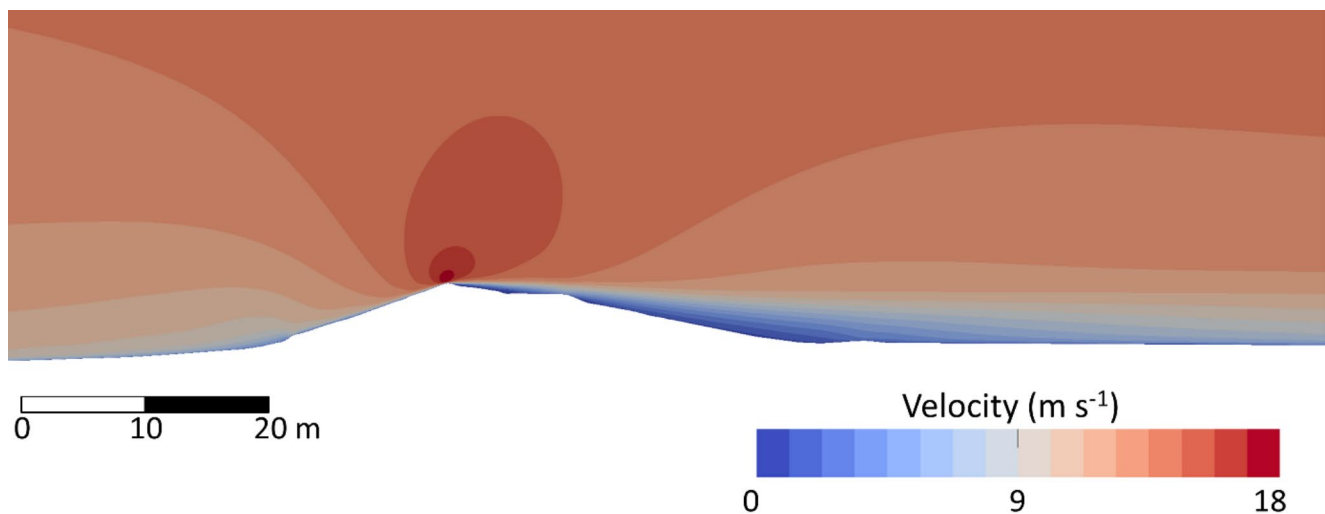[3] School of Science, Engineering and Environment, University of Salford, Manchester, UK

Fig. 1 Isovels of wind speed above a two-dimensional surface of the foredune at Greenwich Dunes, Prince Edward Island, Canada. Incident wind direction is from left to right. Figure adapted from Hesp and Smyth (2016)

Fig. 2 Map of western England and Wales denoting the four study sites investigated

1990). Levin et al. (2008a) suggest that sand dune mobility may be controlled by wind speed because not many vegetation species can tolerate the erosion and relatively high amounts of sediment transport caused by high wind speeds. This hypothesis is supported by Li et al. (2009) who found that wind erosion reduced the size of the germinable seedbank on sand dune grassland, in eastern Inner Mongolia, China. Local wind speeds over sand dunes vary,as near surface wind speed is modulated by local dune topography. Where the incident wind direction is perpendicular to the dune crest, wind speeds slow as they approach the foot of the dune and accelerate toward a maximum velocity immediately before the crest (Hesp and Smyth 2016; Fig. 1), with maximum wind speed increasing with dune height (Parsons et al. 2004). In the lee of the dune crest, wind speed dramatically decreases, and a flow separation cell may form where the angle of the dune at the lee face brink is sufficiently steep. Measured and modelled data from a range of dune morphologies indicate that the separation cell typically extends 4 to 10 dune heights beyond the crest of the dune, before flow becomes reattached and oriented in the same direction as the incident wind flow (Walker et al. 2022). Where a flat homogenous surface extends downwind of a dune, wind velocity profiles require at least 25 dune heights to recover to incident wind conditions that were present upwind of a dune (Parsons et al. 2004; Walker et al. 2022).

As well as the relatively simple dune topography depicted in Fig. 1, most coastal dunes contain a range of dune and slope types over which near surface wind flow patterns increase in complexity (Walker et al. 2022). For example, where foredunes have been subject to wave erosion, cliffing of the foredune toe may occur, resulting in a steep, near vertical foredune stoss slope scarp (Piscioneri et

al., 2019; Hesp and Smyth 2021; Bauer and Wakes 2022). The presence of a scarp substantially reduces wind speed at the foot of the dune and produces a zone of flow separation and reversal upwind of the dune, that decreases in extent with incident wind flow obliquity to the dune crest (Hesp and Smyth 2021; Bauer and Wakes 2022). As well aeolian deposits of sediment, erosional hollows, known as blow-outs, are common in vegetated coastal dunes. Blowouts are chiefly formed by aeolian erosion of sediment and are typically described as saucer, bowl, and trough in form (Hesp and Smyth 2019). Generally, saucer blowouts are described as shallow and circular, trough blowouts as long and narrow, and bowl blowouts as deep and circular. Wind flow within blowouts is substantially modified compared to the dunes around them. Within circular saucer and bowl blow-outs, a turbulent separation zone that is reduced in wind speed may develop in the lee of the upwind erosional wall (Smyth et al. 2019). As wind exits over the blowout rim, it increases in speed due to topographic acceleration and becomes realigned with the incident wind direction (Smyth et al. 2019). Within trough blowouts, winds are steered between steep erosional walls (Hesp and Pringle 2001) and where a windward slope is present along the long-axis of the landform, wind flow acceleration occurs (Smyth and Hesp 2016). As blowouts increase in size and extend downwind, they may evolve into parabolic dunes which are differentiated from blowouts by their trailing ridges, which may long-walled or relatively squat (Hesp and Smyth 2019). Like within trough blowouts, winds within the deflation basin of a parabolic dune becomes increasingly 'funnelled' and steered along the long-axis of the deflation basin as it increases in width and depth (Smyth et al., 2020b). However, only modest increases in near surface wind speeds within the deflation basin of parabolic dunes has been measured (Hansen et al. 2009; Delgado-Fernandez et al. 2018).

Aspect may also play an important role in sand dune mobility. In cold climates and regions where precipitation is common, those slopes which receive greater insolation may be drier and more frequently thawed, thus resulting in greater opportunities for sediment transport to occur (Hugenholtz and Wolfe 2006). Bradley et al. (2019) also noted that steeper slopes may inhibit vegetation stabilisation as loose sediment is prone to avalanching and thus smothering vegetation growth. The hypothesis that dune slope steepness may maintain patches of mobile sand in vegetated coastal sand dunes is shared by Ranwell (1958) who summarised that the steep windward slopes of dunes eroded "more or less continuously" (Ranwell 1958, p.98) in a three-year period between August 1952 and September 1955, surmising that both orientation to the prevailing wind and dune slope are important factors in governing geomorphic activity in dunes.

The overall purpose of this study was to examine any potential relationships between localised areas of sand dune mobility and abiotic landscape scale factors, specifically wind speed, aspect and slope angle. These relationships were tested at four coastal dune sites in England and Wales. By understanding the abiotic factors which best correlate with existing areas of bare sand, it is hoped this research will help guide the location of future dynamic dune interventions, increasing the sustainability of aeolian dynamism and thus reducing the extent and/or frequency of management interventions.

Methods

Study sites

The abiotic landscape scale drivers of coastal dune mobility were investigated at four coastal dune fields in England and Wales, namely Ainsdale, Aberffraw, Morfa Dyffryn and Penhale (Fig. 2). All four sites are designated Special Areas of Conservation (SAC) primarily due to the presence of shifting dunes, fixed dune with herbaceous vegetation, dunes with creeping willow (Salix repens ssp. argenta) and humid dune slacks. Substantial areas of dune at Ainsdale and Morfa Dyffryn are also designated as National Nature Reserves (NNRs) (Fig. 3). Topography and aerial imagery of each dune site was collected between 2014 and 2017 (Table 1), before the commencement of largescale dynamic restoration works as part of the Sands of LIFE (Wales) and DuneLIFE (England) projects at the sites (Creer et al. 2020). The sites represent a diversity of environmental factors and anthropogenic pressures that are common on many coastal dunes.

Ainsdale

Ainsdale sand dunes are located on the Sefton Coast in northwest England (Fig. 2). The area investigated forms part of the Ainsdale Sand Dunes National Nature Reserve and Ainsdale and Birkdale Sandhills Local Nature Reserve (Fig. 3a). The study area contains several active blowouts (Smyth et al. 2020a) and relatively small discrete patches of bare sand are also present on the crests of fixed parabolic dunes in the northwest of the study area (Fig. 3a). Fenced areas of the hinterdune within the National Nature Reserve have been grazed by sheep and cattle since the 1990s. Vegetation predominantly consists of a short sward of grasses, herbaceous perennials, areas of scrub, some of which are periodically cut, and extensive dune slacks between dune ridges. Much of the study site was also a Corsican Pine (Pinus nigra ssp. laricio) planation. The plantation was

Fig. 3 Digital Terrain Models (DTMs) and 2 m contours of the dunes investigated at (**a**) Ainsdale (**b**) Aberffraw (**c**) Morfa Dyffryn (**d**) Penhale Dunes. Information regarding the LiDAR tiles used is available in Table 1 and Appendix 1
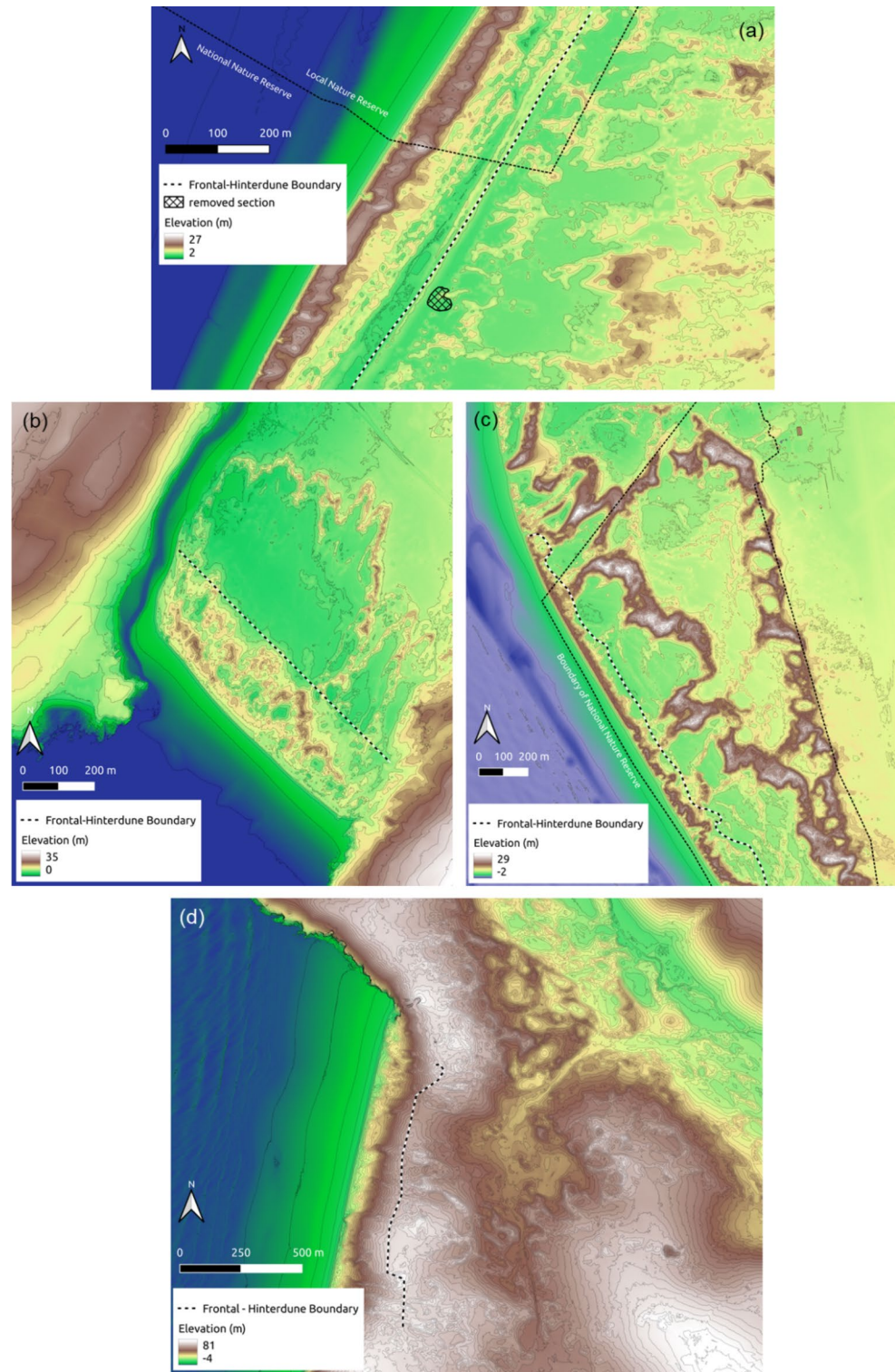
removed in the early 1990s as part of a dune restoration project and stumps remain throughout the dune landscape. A dune slack artificially created at the site, has been removed from the data analysed in this study as it generated a relatively large area of bare sand in the aerial imagery (hatched area in Fig. 3a). Visitor pressure is mostly concentrated on the foredune and clearly marked paths and roads in the hinterdune, both of which are not included in the area of dune being investigated (Sect. 2.2. and Fig. 3a).

Aberffraw

Aberffraw is a coastal dune field on the southwest coast of Anglesey (Fig. 2). The dunes cover approximately 2.9 km² and although the percentage of bare sand at the site has dramatically declined since the 1940s (Bailey and Bristow 2004; Pye et al. 2014), the erosional walls of a clearly defined ridge of parabolic dunes, approximately 500 m inland on the foredune crest (Fig. 3b), remain active. The seaward limits of the dunefield are defined by a beach, Traeth Mawr, which faces southwest. The dunes at Aberffraw are registered Common Land and remain extensively grazed by livestock. The site is a popular area for recreation, however most of the visitor pressure is restricted to a small number of access paths to the beach and along the foredunes (Pye et al. 2014).

Morfa dyffryn

Morfa Dyffryn islocated in Gwynedd on the northeast coast of Cardigan Bay, in north Wales (Fig. 2).The beach and dunes at Morfa Dyffryn form part of an eroding dune system that extends 3 km inland at its widest point though much of this area has been converted from sand dunes to agricultural land and an airfield (formerly Royal Aircraft Establishment Llanbedr). At present, the site contains 2.6 km² of coastal dunes and are designated as a Special Area of Conservation (SAC) due to the presence of dunes with creeping willow (Salix repens ssp. argenta) and humid dune slacks. A high proportion of the dunes at Morfa Dyffryn remain highly mobile (20% bare sand in 2009 (Pye et al. 2014)) and several migrating parabolic dunes extend inland from the shoreline (Fig. 3c). A large campsite is located to the north of Morfa Dyffryn National Nature Reserve resulting in considerable visitor disturbance to the dunes, particularly where visitors cross the dunes to access the beach. Visitor pressure is also compounded by easy access to the site by train. This high level of anthropogenic pressure is reflected in Williams and Davies (2001) assessment of the vulnerability and protection of coastal dunes in Wales, who gave Morfa Dyffrynthe highest pressure of use index of the 26 Welsh dune systems they surveyed. As part of the National Nature Reserve management of the site there is an ongoing programme of scrub clearance and winter grazing by cattle (Pye et al. 2014).

Penhale dunes

Penhale Dunes are located in Cornwall, England (Fig. 2) and span 4 km along the coast and 1.6 km inland. The dunes are amongst the tallest in Britain rising to over 81 m having climbed a bedrock escarpment from the beach (Pye and Blott 2020). Both Penhale dunes and Perran Beach, located to the west of the dunes, are popular tourist destinations (Turner et al. 2021). As a result, numerous paths span the site connecting Perran Sands Holiday Park located in the south of the dunes to the beach (Fig. 3d). The majority of the dunes are fixed and colonised by marram and red fescue grasses.

Bare sand delineation

For each location, areas of bare sand were mapped in QGIS using the semi-automaticclassification plugin (SCP) and the minimum distance algorithm on true-colour RGB images (Congedo 2016). At each site only areas of coastal dune devoid of clear changes in land use (e.g., the airfield at Morfa Dyffryn and pine plantations at Ainsdale) were mapped (Fig. 3). Classification using the minimum distance algorithm method in QGIS was selected due to its 'substantial'observed agreement (82%) between ground truth classification and remotely sensed classification in a coastal sand dune environment (Smyth et al. 2022). The minimum distance algorithm calculates the mean spectra of each predefined class and assigns a pixel to a class that has the least distance to the mean (Mather and Tso 2016).

Classification at each image resolution was trained using 60 regions of interest (ROIs) in total. 30 regions of interest represented vegetation and 30 regions of interest represented bare sand. Training polygons selected represented the range of colours present in each class and were approximately equal in size. The aerial imagery analysed at each site had a pixel resolution of 0.25 m, was cloud free and taken in spring or summer (Table 1). In all cases the aerial imagery was captured within one year of the LiDAR derived digital terrain model (Table 1). Due to the variation in the extent and repeatability of bare sand mapping in sand dunes using semi-automatic classification methods (Smyth et al. 2022), bare sand classification was repeated three times for each site. Statistical analysis was performed using each iteration of classified map and an average of the three iterations was used.

Table 1 Summary of aerial imagery (0.25 m resolution) and LiDAR derived Digital Terrain Model collection periods. All data was sourced using EDINA Aerial Digimap Service © Getmapping Plc (see appendix 1 for full details)

Site	Aerial Imagery Month and Year	Date of Digital Terrain Model Year	Digital Terrain Model Resolution
Abberfraw	April 2015	2015	1.0 m
Ainsdale	April 2015	2016	1.0 m
Morfa Dyffryn	July 2014	2014	2.0 m
Penhale	May 2016	2017	1.0 m

Abiotic factors

Wind speed at 0.4 m above the surface of the digital terrain model at each site was calculated using computational fluid dynamics (CFD). Computational fluid dynamics is a numerical method of solving fluid flow using the Navier-Stokes equations. Computational fluid dynamics modelling was performed using the open-source software Open-FOAM. The turbulence model, boundary conditions and cell sizes used were the same as those applied for calculating near surface wind flow over a 100 m long parabolic dune as part of a separate study by Delgado-Fernandez et al. (2018). Delgado-Fernandez et al. (2018) reported robust correlations between modelled and measured wind direction (R^2=0.89, n=21, p<0.01) and wind speed (R^2 = 0.81, n=21, p<0.01). Wind flow was calculated using the Semi-Implicit Method for PressureLinked Equations (SIMPLE) algorithm. This method produces a steady state solution, using the Reynolds-averaged Navier-Stokes equations (RANS). Turbulence was modelled using the Re-normalisation group (RNG) k-ε method as it has compared well with measured wind flow over complex dune topography (Smyth et al. 2013; Hesp et al. 2015; Delgado-Fernandez et al. 2018).

The incident wind speed at each site was described by the average wind speed and direction of above threshold velocity winds (>6 m s^{-1}) from the meteorological station closest or most appropriate to the study site for the 10 preceding years (Table 2). A sediment threshold velocity of 6 m s^{-1} at 10 m above the surface was selected for this study as it is a common threshold value used for resultant drift potential calculations (Kolesar et al. 2022). A threshold velocity value of 6 m s^{-1} is also close to the minimum threshold velocity of sediment transport at 10 m above a drying beach surface at Aberffraw as calculated by Wiggs et al. (2004), who measured a value of 7.15 m s^{-1} assuming the presence of a logarithmic boundary on the beach and a surface roughness height of sand (0.0005 m (Bagnold 1941)). It should be acknowledged that sediment threshold velocity is a highly variable value that is impacted by changes in instantaneous changes in sediment, local bedform morphology, crusting of the sediment (salt or biological) and moisture fluctuations (Walker 2020). As a result, sand transport events below threshold velocity and the non-occurrence of sand transport

events at wind speeds above threshold are frequent (Wiggs et al. 2004). From this meteorological data, shear velocity (u_*) was calculated and vertical profiles of wind velocity (U), turbulent kinetic energy (k) and turbulent dissipation rate (ε) produced using Eqs. 1, 2 and 3 (Richards and Hoxey 1993; Blocken et al. 2007):

$$U(z) = \frac{u_*}{\kappa} ln\left(\frac{z + z_0}{z_0}\right) \tag{1}$$

$$k(z) = \frac{u_*^2}{\sqrt{C_\mu}} \tag{2}$$

$$\epsilon(z) = \frac{u_*^3}{\kappa(z + z_0)} \tag{3}$$

Where z is the height above the surface, κ is the von Karman constant (0.42), z_0 is the surface roughness length and C a constant of 0.09 (Richards and Hoxey 1993). For all simulations and the calculation of shear velocity (u_*), z_0 was prescribed a uniform value of 0.175 m, the average z_0 value calculated for an Ammophila arenaria vegetated slack by Levin et al. (2008b). 0.175 m was also the surface roughness length used by Delgado et al. (2018).

The topographic surface of each study site was generated from airborne LiDAR surveys performed between 2014 and 2017 (Table 1). Cell size in each computational domain decreased gradually size from $40 \times 40 \times 40$ m at the highest vertical extent to $2.5 \times 2.5 \times 2.5$ m at the surface. In all cases the height of the computational domain was at least 5 times greater than the highest topographic point at that study site. Wind data was sampled 0.4 m above the surface of the dune and the numerical modelling of wind flow was considered complete when residuals for velocity (U) reached 10^{-4}.

Statistical analysis

Spearman's rank correlation was used to test the strength and direction of the relationship between bare sand and wind speed, bare sand and surface aspect and bare sand and surface slope at a spatial resolution of 2.5 m^2 at each site. Within each 2.5 m^2 area, wind speed, surface aspect and surface slope were averaged, and bare sand calculated

Table 2 Incident wind flow conditions at each study site based on average above threshold velocity winds (>6 m s^{-1}) from the meteorological station closest or most appropriate to the study site for the 10 preceding years before the aerial image of each site was collected

Study Site	Met. Station	Compass bearing and dist. from study site	Wind Dir.	Wind Speed
Ainsdale	Crosby	SSE 10 km	252°	9.22 m s^{-1}
Aberffraw	Valley	NW 9 km	214°	9.48 m s^{-1}
Morfa Dyffryn	Valley	NW 56 km	214°	9.48 m s^{-1}
Penhale	Culdrose	SSW 32 km	236°	8.76 m s^{-1}

as a percentage. At a site scale, this created approximately 375,000 points (0.94 km^2) to test at Penhale, 290,000 points (0.73 km^2) at Morfa Dyffryn, 100,000 points (0.25 km^2) at Aberffraw and 88,000 points (0.22 km^2) at Ainsdale. Additional statistical analysis was also performed at each site using only the cells with an aspect $\pm 90°$ to the prevailing wind direction. These cells are referred to as the 'stoss slope' in Sect. 3. 4. Spearman's rank was selected as it is a non-parametric test that does not require the data to be normally distributed, nor for the variables to be linearly related. Correlation (r_s) ranges from -1 to $+1$ and the strength of the relationship between variables is described as very weak to very strong in 0.2 increments (0.00–0.19 very weak, 0.20–0.39 weak, 0.40–0.59 moderate, 0.60–0.79 strong, 0.80–1.00 very strong). Multi-regression analysis, such as that applied to biotic and abiotic factors impacting dune response by Garzon et al. (2022), could not be applied to the data as the variables were not classified as independent (tested using the Durbin-Watson statistic) nor were the standardised residuals of the regression line normally distributed.

Statistical analysis was applied at three different geomorphic units for each location: (1) the entire site (2) the frontal dunes and (3) the hinterdunes (Fig. 3). In this study the 'frontal dune' refers to the area spanning from the crest of the foredune to the lowest point of the foredunes landward (lee) slope. The seaward slope of the foredune was not included in the analysis as bare sand is often caused by factors not accounted for in this study e.g., wave erosion (van Dijk 2021) and anthropogenic trampling and vehicle use (McAtee and Drawe 1980). Additionally, the seaward slope of the foredune at Morfa Dyffryn (Millington et al. 2008) and at Penhale (Turner et al. 2021) was substantially scarped creating a very steep slope which produced a dark shadow in much of the aerial imagery. Heavy shadowing of areas of bare sand makes accurate mapping extremely

difficult, increasing the likelihood of error (Smyth et al. 2022). In this study the 'hinterdune' refers to the dune area landward of the foredune. The boundary between frontal dune and hinterdune is highlighted for each site in Figs. 3, 4, 5, 6 and 7. All geomorphic units were defined manually in a geographical information system using a combination of both the aerial imagery and LiDAR generated digital terrain model.

Results

The results of bare sand, wind speed and surface slope are presented for each of the four sites. The average aspect at a 2.5 m^2 resolution for each site is not presented as it demonstrated a very weak relationship with bare sand at every site and at all geomorphic scales.

Ainsdale

The percentage area classified as bare sand ranged from 9.2 to 9.9% at Ainsdale, a percentage relative range (percentage relative range $= \frac{range}{average} * 100$) of 7.3% (Table 3). The majority of the bare sand classified at Ainsdale was located on the area classified as frontal dune (Fig. 4). In the hinterdune, bare sand was largely limited to an active blowout complex near the centre of the study area and on the slopes of fixed parabolic dunes, although some discrete small patches of bare sand can be observed throughout the site (Fig. 4). Topographical acceleration of wind flow 0.4 m above the surface was not uniform over the frontal dune. Wind speeds in the deflation basins of several relatively deep foredune blowouts in the centre of the study area were substantially lower than wind speeds above the surface of relatively high topography that surround them. Wind flow acceleration occurred on the stoss slope of dunes throughout the hinterland, however
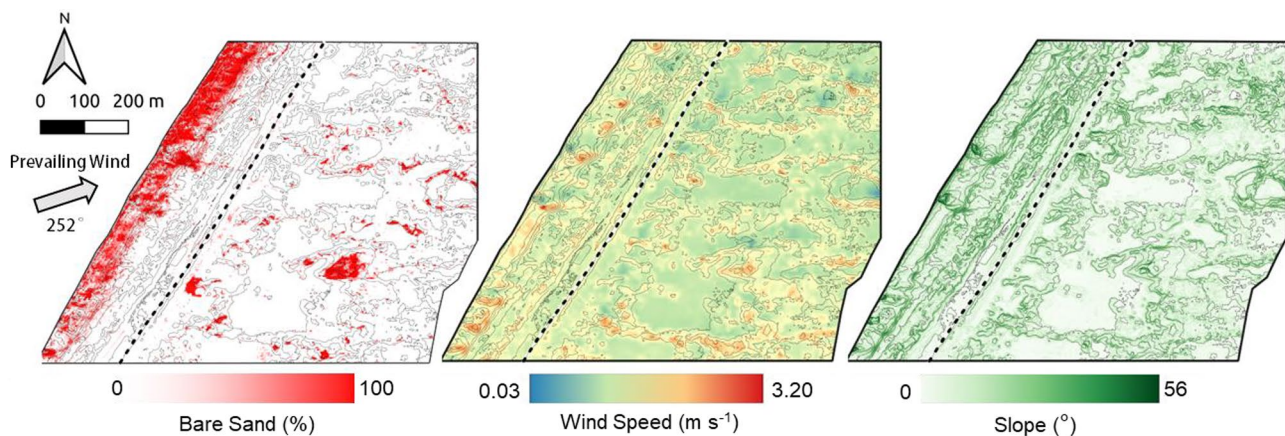

Fig. 4 2.5 m^2 resolution grid squares of bare sand percentage, wind speed (0.4 m above the surface) and dune slope at Ainsdale. The dashed line represents the frontal-hinterdune boundary. Contours are at 2 m intervals. The bare sand classification map represents the first replicate of three
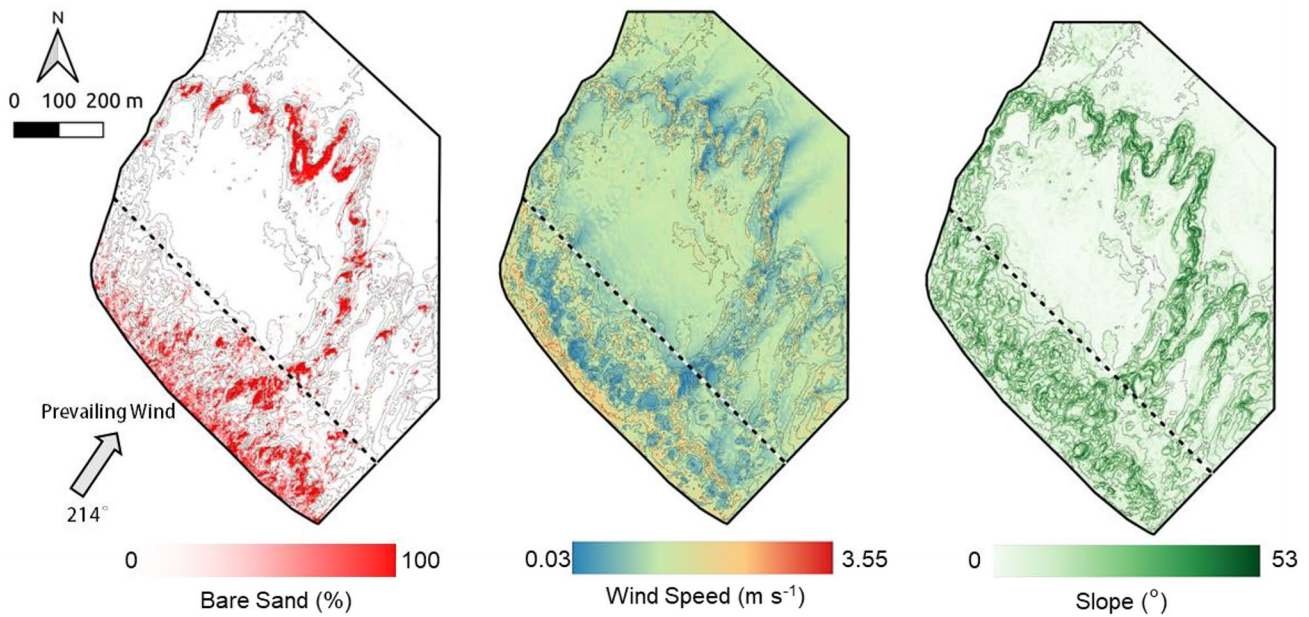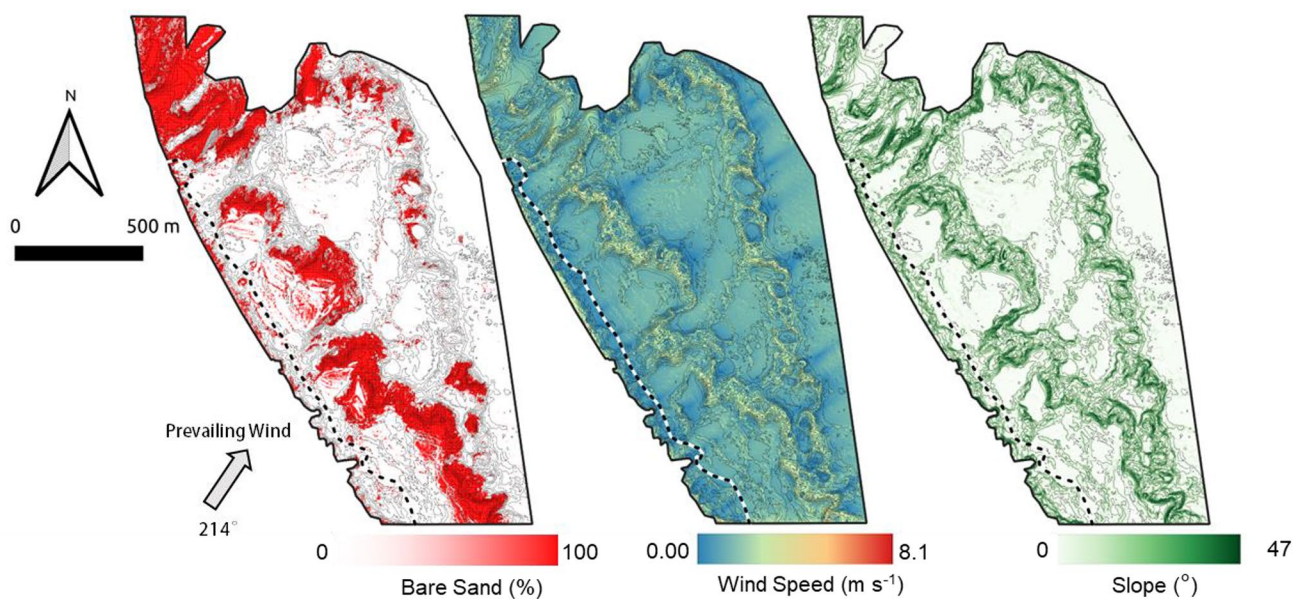
Fig. 5 2.5 m² resolution grid squares of bare sand percentage, wind speed (0.4 m above the surface) and dune slope at Aberffraw. The dashed line represents the frontal-hinterdune boundary. Contours are at 2 m intervals. The bare sand classification map represents the first replicate of three

Fig. 6 2.5 m² resolution grid squares of bare sand percentage, wind speed (0.4 m above the surface) and dune slope at Morfa Dyffryn. The dashed line represents the frontal-hinterdune boundary. Contours are at 2 m intervals. The bare sand classification map represents the first replicate of three

clearly defined regions of low wind speed in the lee of tall dunes in the hinterland was not observed as was the case at Aberffraw (Fig. 5) and Morfa Dyffryn (Fig. 6). The steepest slopes at Ainsdale were on the frontal dune, particularly on the erosional walls of trough shaped blowouts. The hinter-dune at Ainsdale was relatively flat with approximately a third of the area having a slope of less than five degrees. The steepest slopes were recorded on fixed parabolic dunes and isolated fixed dune ridges.

Aberffraw

Aberffraw had the highest variability of classified bare sand area ranging from 7.6 to 9.3%, with an average of 8.5%

(Table 3). Figure 5 shows that bare sand was predominantly located on the crest and immediately in the lee of the foredune, as well as on the stoss slopes of the semi-fixed parabolic dunes in the hinterdune. Topographic acceleration of flow at the crest of the foredune was relatively uniform. A consistent decrease in wind speed occurred in the lee of the foredune. In land from the foredune, where the surface of the dune was relatively flat and homogenous, wind speed recovered to approximately 1.75 m s^{-1} until reaching the digitate morphology of the semi-fixed parabolic dunes, where wind was accelerated to a similar velocity as that measured on the foredune crest (Fig. 5). Unlike in the lee of the foredune, where the zone of retarded wind speed formed a spatially homogenous zone perpendicular to the dune crest, distinct narrow regions of low wind speed extended beyond the crest of the depositional lobe of the parabolic dune. The stoss slopes of the parabolic dunes were also the steepest slopes measured in the study area, reaching a maximum angle of 53°.

Morfa dyffryn

On average, 26.6% of the study area at Morfa Dyffryn was classified as bare sand, with a range of 0.9% (Table 3), the lowest percentage relative range for all four sites investigated (3.4%). The northern section of the study area, where no foredune was present, was classified almost predominantly as bare sand (Fig. 5).In the rest of the study area, bare sand was concentrated on the erosional walls of the semi fixed parabolic dunes that were predominantly orientated opposite to the prevailing wind, 100 to 200 m inland of the foredune (Fig. 6).

The maximum wind speed measured at Morfa Dyffryn was substantially higher than the other three sites, despite similar incident wind speeds at each location (Table 2). However, cells with wind speeds greater than or equal to 6 m s^{-1} were limited to only 0.14% of the total study site. Furthermore, the mean wind speed across the entire site was not significantly different to the other locations investigated in this study. Like at Aberffraw (Fig. 5), wind speeds were greatest on the erosional walls of the parabolic dunes

and blowouts that faced the prevailing wind direction. The steepest dune slopes at Morfa Dyffryn also coincided with the erosional walls of the semi-fixed parabolic dunes at the site and the large regions of bare sand in the north of the study site.

Penhale dunes

Penhale dunes recorded the lowest average area of bare sand (2.9%) but the largest percentage relative range (31%) between the three bare sand classifications (Table 3). Within the study area at Penhale, bare sand was predominantly restricted to the frontal dunes (Fig. 7). Substantial areas of bare sand were located on the steep slopes of the northern section of the frontal dune, immediately inland of the scarped foredune edge. A large blowout/parabolic dune at the southern end of the site, which extended approximately 300 m inland from the beach, was also classified as bare sand. Visual inspection of the aerial imagery shows that the deflation basin of the landform appears to act as an access point to the beach from the caravan park and nearby car park. Isolated, discrete patches of bare sand were classified in the north of the hinterdune but these were much smaller than areas of bare sand classified on the frontal dune.

Wind speed at Penhale was generally greatest above the surface of the frontal dune, however due to the lack of a distinct lee slope downwind of the dune crest, flow deceleration in the lee of the frontal dune did not occur as at other sites in this study.

The north of the hinterdune was dominated by large areas of low wind speed within topographic depressions which appear to be the deflation basins of relict blowouts. Areas of high-speed wind flow in the hinterdune were restricted to the upper slopes of the erosional walls of now vegetated parabolic dunes and blowouts. The steepest slopes of any of the dunes investigated were measured at Penhale (62°), however these steep dune slopes only found in small areas on the frontal dune (only 10 m^2 had an average slope equal or greater than 60°). Large areas in the southeast of the hinterdune were mostly flat and lacked any substantial changes in slope (Fig. 7).

Statistics

Taking into consideration the whole study area and hinterdune for each location investigated, slope was always a stronger predictor of bare sand than wind speed, demonstrating a significant positive relationship at each site, ranging in strength from very weak at Penhale to moderate at Aberffraw (Fig. 8). At a whole site scale, the strength of correlation between slope and bare sand increased when only slopes that were orientated $\pm 90°$ to the prevailing wind

Table 3 The summary statistics and relative range ($\frac{range}{average} * 100$) from three iterations of bare sand classification performed at each site.

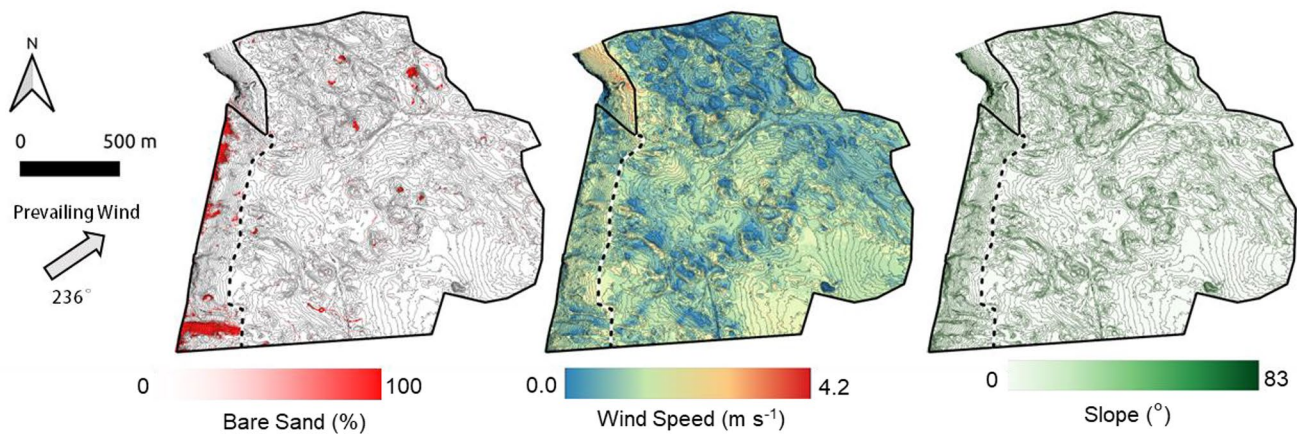| Site	Classified Bare Sand				
	Average	Max	Min	Range	% Relative Range
Ainsdale	9.6	9.9	9.2	0.7	7.3
Aberffraw	8.5	9.3	7.6	1.7	20.0
Morfa Dyffryn	26.6	27.0	26.1	0.9	3.4
Penhale	2.9	3.4	2.5	0.9	31.0

Fig. 7 2.5 m² resolution grid squares of bare sand percentage, wind speed (0.4 m above the surface) and dune slope at Penhale Dunes. The dashed line represents the frontal-hinterdune boundary. Contours are at 2 m intervals. The bare sand classification map represents the first replicate of three

were considered. Examining only those windward slopes also increased the strength of correlation in the hinterdunes at each site except Aberffraw (Fig. 8).

Overall, the strength of the relationships between bare sand and the factors examined was weaker for the frontal dune than the hinterdune or whole site, with all of the correlations for the frontal dune classed as weak or very weak. At both Aberffraw and Morfa Dyffryn wind speed was a stronger predictor of bare sand than slope on the frontal dunes. These positive correlations increased in strength when only the windward slopes were considered (Fig. 8). At Penhale a very weak negative correlation existed between wind speed and bare sand on the frontal dunes which also increased in strength when only the windward slopes of the dune were included.

Discussion

The statistical results of this study suggest that at a landscape scale, regions of bare sand on coastal dunes are best predicted by dune slope and that the strength of prediction is greatest when only windward slopes (± 90° to the prevailing wind) are considered. We hypothesise this positive correlation between bare sand and slope angle in coastal dunes is driven by the frequency of geomorphic activity on steep windward slopes, which is sufficient to inhibit the colonisation of vegetation. At a landscape scale geomorphic activity is likely triggered by three key agents: (1) Wind erosion of sediment, due to topographic acceleration as streamlines are compressed and wind speed accelerates over a slope (Walker et al. 2022); (2) Granular avalanches of unconsolidated sediment, particularly during periods of dry weather (Carter et al. 1990; Fraser et al. 1998) and (3) rotational slumping of unstable slopes (Carter and Stone 1989). Steep dune slopes

may also be the focus of anthropogenic recreation activities such as 'sand dune sledding', however the impact of these activities, whilst intense, are likely to be limited spatially to a relatively small areas of highest topographic relief, rather than act ubiquitously at a landscape scale.

The hypothesis that regions of bare sand on coastal dunes are best predicted by dune slope is supported by several other studies that have attributed bare sand presence to prevalent geomorphic activity on the steep windward slope of dunes. For example, when investigating a rapidly vegetating coastal dunefield in Israel, Tsoar and Blumberg (2002) concluded that very low percentages of vegetation cover on the windward slopes of parabolic dunes were in response to regular sediment erosion, a point also noted by Ranwell (1958) in Wales. Tsoar and Blumberg (2002) also attributed the relatively high vegetation cover at the crest of parabolic dunes to a lack of geomorphic change (Fig. 9).

In their descriptions of coastal dune stabilisation in Wales, Pye et al. (2014) reported that for at least five sites (Aberffraw, Newborough Warren, Morfa Harlech, Morfa Dyffryn, Methyr Mawr) the remaining patches of bare sand were present on the windward faces of parabolic dunes. Studies in the Netherlands (Arens et al. 2004), Australia (Pye 1982) and Brazil (Durán et al. 2008) have also recorded that the steep stoss slopes of parabolic dunes are amongst the final slopes to become vegetated in otherwise fixed coastal dune landscapes. Barchyn and Hugenholtz (2013) in their observations of blowout activity in Bigstick Sand Hills, Canada, also document that the steeper slopes of blowouts are amongst the last areas of bare sand to be vegetated by coloniser species (see Fig. 4c in Barchyn and Hugenholtz (2013).

In this study, the two sites which had the greatest positive correlation between bare sand and slope, Aberffraw (Fig. 5) and Morfa Dyffryn (Fig. 6), both contained clearly defined

Fig. 8 Spearman's rank correlation coefficients of bare sand, slope angle and wind speed calculated at 0.4 m above the surface at each coastal site dune investigated. Hatched bars indicate where only slopes ± 90° to the prevailing incident wind were considered (stoss slopes)
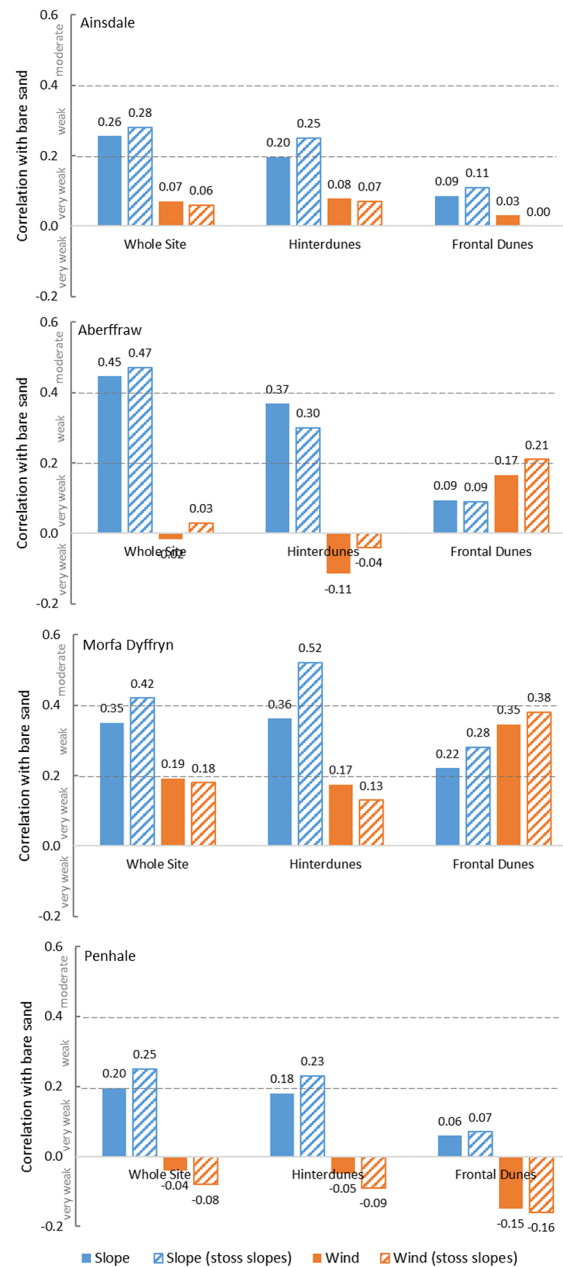
digitate ridges of semi-fixed parabolic dunes, landforms which encompassed steep windward slopes of bare sand. A lack of steep windward slopes associated with parabolic dunes within the 'frontal dunes' at any of the sites (Figs. 4–7) may also partially explain the weaker correlation between bare sand dune slope in the frontal dune. The frontal dune is also typically subject to a range of geomorphic agents that were not accounted for in this study, for example anthropogenic trampling and wave action. Additionally, the steeply scarped and often unvegetated seaward slopes of the foredunes (e.g., Ainsdale and Morfa Dyffryn) were not included in any analysis as accurate classification of these slopes was impossible due to heavy shadowing.

In comparison to the areas where parabolic dunes were present, relatively little bare sand was classified within blowouts at any of the sites (Fig. 3). This phenomenon is particularly evident in the northern area of the hinterdune at Penhale which contains a relatively high density of vegetated bowl blowouts (Fig. 7). Unlike parabolic dunes, which have an opening generally in the direction of the prevailing wind and can funnel wind from a broad range of incident wind directions toward the erosional walls (Smyth et al. 2020b), bowl, and many trough shaped blowouts, have deflation basins that are encompassed by an erosional wall creating a substantial area of low wind speed. A morphology referred to as a 'closed' blowout by Carter et al. (1990). The 'closed'
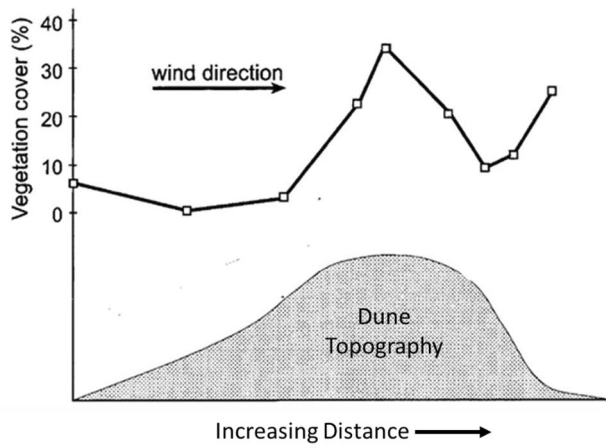
Fig. 9 Conceptualised vegetation cover over parabolic dunes in a rapidly vegetating coastal dune field in Israel (Image adapted from Tsoar and Blumberg (2002)

morphology of these landforms often limits substantial erosion to the upper slopes of the windward facing erosional wall, where wind speeds peak in magnitude (Smith et al. 2017; Smyth et al. 2019). These patterns of erosion contrast with the windward slopes of parabolic dunes, where most erosion occurs in the middle of the windward slope (Tsoar and Blumberg 2002). An additional factor that may limit blowout activity is the availability of an upwind sediment source. Where a sediment source upwind of a blowout is present, separation and/or retardation of windflow within the deflation basin may result in the deposition and filling of a deflation basin with sediment during an active transport event rather than net erosion (Smith et al. 2017; Smyth et al. 2019). This evolution of geomorphic activity over time in blowouts, from an incipient phase, erosion phase and finally accretion phase has been documented by Gares (1992) and Gares and Nordstrom (1995). A decade after blowouts had formed on Island Beach State Park, USA, Gares (1992) noted that the steep erosional walls of the blowouts became increasingly gentle and subsequently colonised by vegetation. This description from Gares (1992) of the vegetating of coastal dune blowouts provides further evidence to support the statistical analysis in this study that steepness of slope is an important factor in maintaining bare sand. Where active blowouts did exist within the coastal dunes investigated, they were predominantly in areas of relatively high elevation in comparison to the surrounding dunes, such as on the crests or trailing arms of relict parabolic dunes e.g., Ainsdale (Smyth et al. 2020b) and Morfa Harlech (Pye et al. 2014).

## Why didn't wind speed predict better?

Given that much of the previous literature on geomorphic activity and bare sand had attributed the presence of bare sand on coastal dunes to increased wind speeds (Pye et al.

2014; Arens et al. 2004; Pye 1982; Rust 1990), the limited capacity of wind speed as a predictor of bare sand in this study is somewhat surprising. We offer 3 reasons why this may be the case: (1) Bare sand may be associated with the deflation basin in a parabolic dune or blowout. Although bare sand is often present within the deflation basins of active blowouts and parabolic dunes, the wind speed at these locations relative to the surrounding landscape is substantially lower (Fig. 4). In these erosional hollows bare sand is likely to be present not due to erosion and high wind speeds, but due to the transportation and reworking of sediment from the surrounding erosional walls. (2) At a landscape scale, wind speed is not a good predictor of bare sand as large sections of dune have relatively high wind speeds but very low percentages of bare sand. This is notably evident landward of the foredune at Ainsdale (Fig. 4), between the foredune and line of parabolic dunes at Aberffraw (Fig. 5) (both areas of wet dune slack and grassland), and the central and southern areas of hinterdune at Penhale (Fig. 6). (3) There are also multiple examples where substantial patches of bare sand have been created by human footfall (frontal dunes at Ainsdale, Fig. 4) and vehicular access to the dunes (hinterdunes at Aberffraw, Fig. 5). This vegetation removal and resulting exposure of the underlying sand is most evident at Penhale, where several linear 'paths' of bare sand are visible in the classified bare sand map (Fig. 7). Conversely, it is also likely that many areas of high wind speed have become vegetated due to historical dune stabilisation works. Activities such as fencing and marram planting were common throughout northwest Europe in the 19th to late 20th century (Provoost et al. 2011) and were explicitly recorded by Ranwell and Boar (1986) at Aberffraw.

Bare sand classification

The results of the bare sand classification in this study show that the percentage relative range of bare sand generally increases as the overall percentage of bare sand decreases. Where bare sand is predominantly composed of large, clearly defined mobile landforms e.g., Morfa Dyffryn (Fig. 6), the relative range of bare sand classified is relatively low (3.4%, Table 3). However, at sites where the patches of bare sand are smaller and less clearly defined, the relative range in the classified area of bare sand is comparatively large (e.g., 31% at Penhale, Table 3). This variability in the repeatability of bare sand classification between sites was also observed by Smyth et al. (2022) who reported that the extent, accuracy, and repeatability of bare sand classification on an embryonic coastal dune was highly variable. Smyth et al. (2022) also hypothesised that land cover classification may be more accurate in areas with sharp boundaries between dense vegetation and active dune sand e.g., Morfa Dyffryn. However, it should be noted that even at sites with the greatest range

Fig. 10 Representative topographic profiles of persistent bare sand patches at Aberffraw and Morfa Dyffryn. Topographic and vegetation profiles were generated from data in 2015 at Aberffraw and 2014 at Morfa Dyffryn (Table 1). Coordinates refer to the exact location of each dune and are in WGS 84 Web Mercator (Google Earth)
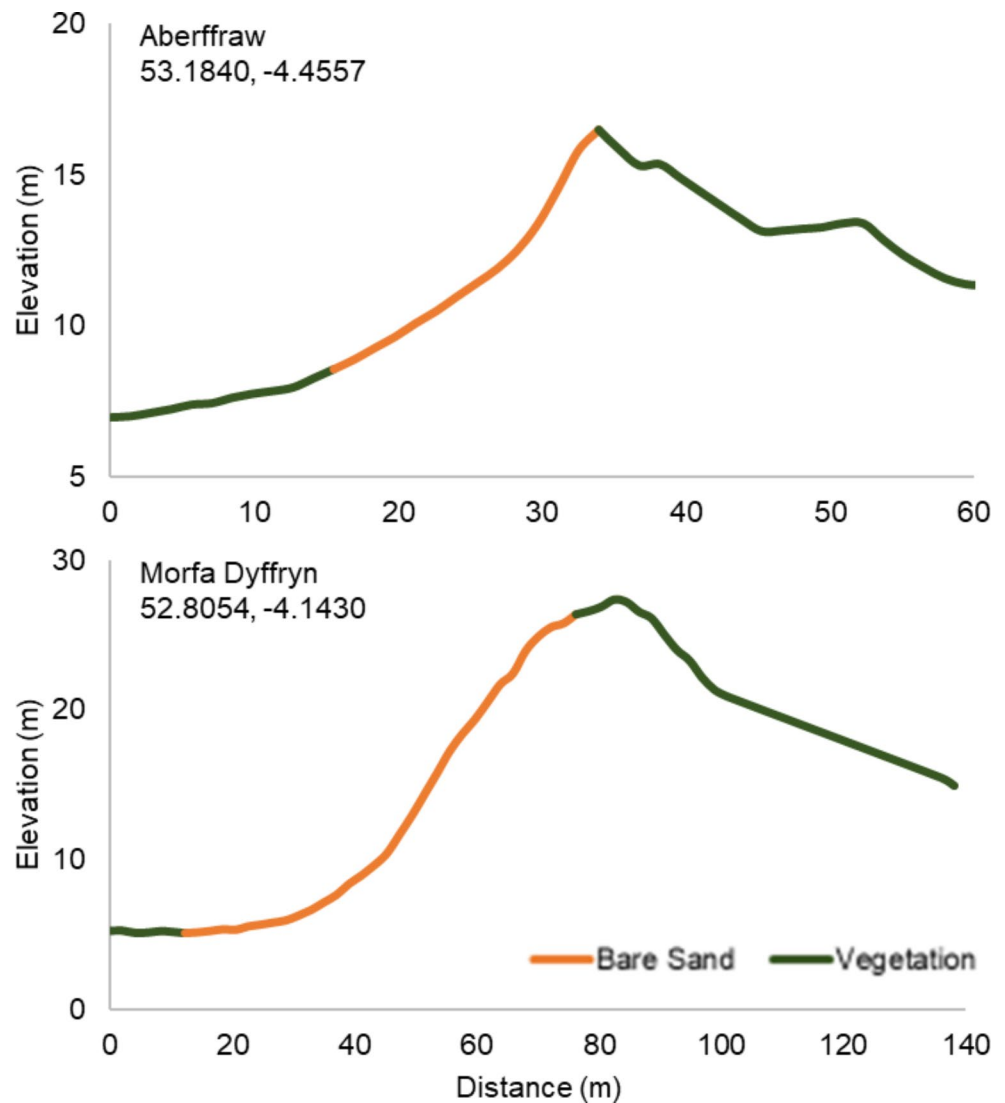
in classified areas of bare sand, differences in the statistical correlations of bare sand, wind speed and slope between classification iterations were negligible.

Implications for dune management

In coastal dunes in which land managers wish to 'rejuvenate' areas of bare sand (Pye et al. 2014), the findings of this study suggest that to maximise the sustainability of bare sand patches, particularly within the hinterdune, rejuvenation efforts should target steep windward dune slopes. We also suggest that any reprofiling of the dune face should mimic the concave profiles of active slope faces on parabolic dunes (Arens et al. 2004; Tsoar and Blumberg 2002). Figure 10 depicts the topographic profile of two patches of bare sand that have persisted at Aberffraw and Morfa Dyffryn since at least 1985 (Google Earth). The depicted dune

at Aberffraw has an average windward slope of 23° and the dune at Morfa Dyffryn and average slope of 19°.

Conclusion

By investigating existing patches of bare sand on four predominantly vegetated coastal dunefields our research found that dune slope, not near surface wind speed or aspect, best predicted the presence of bare sand. We suggest that bare sand on steep dune slopes is maintained by three factors: (1) Wind erosion of sediment, due to topographic acceleration as streamlines are compressed and wind speed accelerates over a slope (Walker et al. 2022); (2) Granular avalanches of unconsolidated sediment, particularly during periods of dry weather (Carter et al. 1990; Fraser et al. 1998) and (3) rotational slumping of unstable slopes (Carter and Stone 1989). We therefore recommend that to create sustainable

areas of bare sand in coastal dunes management efforts to 'rejuvenate' persistent geomorphic activity should focus on the steep windward slopes of dunes. Where appropriate, we recommend that any reprofiling of the dune slope should mimic the concave profiles of existing active slope faces on natural parabolic dunes and have an overall slope angle of between 18° and 23° from the dune toe to the crest.

References

Arens SM, Geelen LHWT (2006) Dune landscape rejuvenation by intended destabilisation in the Amsterdam water supply dunes. J Coastal Res 22(5):1094–1107

Arens SM, Slings Q, De Vries CN (2004) Mobility of a remobilised parabolic dune in Kennemerland, the Netherlands. Geomorphology 59(1–4):175–188

Bagnold RA (1941) The physics of blown sand and desert dunes. Courier Corporation

Bailey SD, Bristow CS (2004) Migration of parabolic dunes at Aberffraw, Anglesey, north Wales. Geomorphology 59(1–4):165–174

Bar P (2013) Restoration of coastal sand dunes for conservation of biodiversity: the israeli experience. Restor Coastal dunes, 173–185

Barchyn TE, Hugenholtz CH (2013) Reactivation of supply-limited dune fields from blowouts: a conceptual framework for state characterization. Geomorphology 201:172–182

Bauer BO, Wakes SJ (2022) CFD simulations of wind flow across scarped foredunes: implications for sediment pathways and beach–dune recovery after storms. Earth Surf Proc Land 47(12):2989–3015

Blocken B, Stathopoulos T, Carmeliet J (2007) CFD simulation of the atmospheric boundary layer: wall function problems. Atmos Environ 41(2):238–252

Bradley AV, Haughan AE, Al-Dughairi A, McLaren SJ (2019) Spatial variability in shrub vegetation across dune forms in central Saudi Arabia. J Arid Environ 161:72–84

Carter RWG, Stone GW (1989) Mechanisms associated with the erosion of sand dune cliffs, Magilligan, Northern Ireland. Earth Surf Proc Land 14(1):1–10

Carter RWG, Hesp PA, Nordstrom KF (1990) Erosional landforms in coastal dunes. In Coastal dunes. Form and process (pp. 217–250)

Congedo L (2016) Semi-automatic classification plugin documentation. Release 4(01):29

Creer J, Litt E, Ratcliffe J, Rees S, Thomas N, Smith P (2020) A comment on some of the conclusions made by Delgado-Fernandez et al.(2019). "Is 're-mobilisation' nature conservation or nature destruction? A commentary. J Coastal Conserv 24(3):1–4

Darke IB, Eamer JB, Beaugrand HE, Walker IJ (2013) Monitoring considerations for a dynamic dune restoration project: Pacific Rim National Park Reserve, British Columbia, Canada. Earth Surf Proc Land 38(9):983–993

Delgado-Fernandez I, Smyth TA, Jackson DW, Smith AB, Davidson-Arnott RG (2018) Event-scale dynamics of a parabolic dune and its relevance for mesoscale evolution. J Geophys Research: Earth Surf 123(11):3084–3100

Delgado-Fernandez I, O'Keeffe N, Davidson-Arnott R (2019) Natural and human controls on dune vegetation cover and disturbance. Science of The Total Environment

Durán O, Silva MVN, Bezerra LJC, Herrmann HJ, Maia LP (2008) Measurements and numerical simulations of the degree of activity and vegetation cover on parabolic dunes in north-eastern Brazil. Geomorphology 102(3–4):460–471

Fraser GS, Bennett SW, Olyphant GA, Bauch NJ, Ferguson V, Gellasch CA, …, Woodfield MC (1998) Windflow circulation patterns in a coastal dune blowout, south coast of Lake Michigan. J Coastal Res, 451–460

Gao J, Kennedy DM, Konlechner TM (2020) Coastal dune mobility over the past century: a global review. Progress in Physical Geography: Earth and Environment 44(6):814–836

Gares PA (1992) Topographic changes associated with coastal dune blowouts at Island Beach State Park, New Jersey. Earth Surf Proc Land 17(6):589–604

Gares PA, Nordstrom KF (1995) A cyclic model of foredune blowout evolution for a leeward coast: Island Beach, New Jersey. Ann Assoc Am Geogr 85(1):1–20

Garzon JL, Costas S, Ferreira O (2022) Biotic and abiotic factors governing dune response to storm events. Earth Surf Proc Land 47(4):1013–1031

Hansen E, DeVries-Zimmerman S, van Dijk D, Yurk B (2009) Patterns of wind flow and aeolian deposition on a parabolic dune on the southeastern shore of Lake Michigan. Geomorphology 105(1–2):147–157

Hesp PA, Pringle A (2001) Wind flow and topographic steering within a trough blowout. Journal of Coastal Research, 597–601

Hesp PA, Smyth TA (2016) Jet flow over foredunes. Earth Surf Proc Land 41(12):1727–1735

Hesp PA, Smyth TAG (2019) Anchored Dunes. In: Livingstone I, Warren A (eds) Aeolian geomorphology. Wiley

Hesp PA, Smyth TA (2021) CFD flow dynamics over model scarps and slopes. Phys Geogr 42(1):1–24

Hesp PA, Smyth TA, Nielsen P, Walker IJ, Bauer BO, Davidson-Arnott R (2015) Flow deflection over a foredune. Geomorphology 230:64–74

Hugenholtz CH, Wolfe SA (2006) Morphodynamics and climate controls of two aeolian blowouts on the northern Great Plains, Canada. Earth Surf Processes Landforms: J Br Geomorphological Res Group 31(12):1540–1557

Jackson DW, Costas S, González-Villanueva R, Cooper A (2019) A global 'greening'of coastal dunes: an integrated consequence of climate change? Glob Planet Change 182:103026

Kolesar KR, Mavko M, Burgess E, Nguyen N, Schaaf MD (2022) A modified resultant drift potential for more accurate prediction of

sand transportation in the vicinity of the Keeler Dunes, California. Aeolian Res 58:100819

Konlechner TM, Ryu W, Hilton MJ, Sherman DJ (2015) Evolution of foredune texture following dynamic restoration, Doughboy Bay, Stewart Island, New Zealand. Aeolian Res 19:203–214

Lancaster N (1988) Development of linear dunes in the southwestern Kalahari, southern Africa. J Arid Environ 14(3):233–244

Laporte-Fauret Q, Castelle B, Michalet R, Marieu V, Bujan S, Rosebery D (2021) Morphological and ecological responses of a managed coastal sand dune to experimental notches. Sci Total Environ 782:146813

Levin N, Ben-Dor E (2004) Monitoring sand dune stabilization along the coastal dunes of Ashdod-Nizanim, Israel, 1945–1999. J Arid Environ 58(3):335–355

Levin N, Kidron GJ, BEN-DOR EYAL (2008a) A field quantification of coastal dune perennial plants as indicators of surface stability, erosion or deposition. Sedimentology 55(4):751–772

Levin N, Ben-Dor E, Kidron GJ, Yaakov Y (2008b) Estimation of surface roughness (z0) over a stabilizing coastal dune field based on vegetation and topography. Earth Surf Processes Landforms: J Br Geomorphological Res Group 33(10):1520–1541

Li FR, Zhao LY, Zhang H, Liu JL, Lu HY, Kang LF (2009) Habitat degradation, topography and rainfall variability interact to determine seed distribution and recruitment in a sand dune grassland. J Veg Sci 20(5):847–859

Mather P, Tso B (2016) Classification methods for remotely sensed data. CRC press

McAtee JW, Drawe DL (1980) Human impact on beach and foredune vegetation of North Padre Island, Texas. Environ Manage 4(6):527–538

Millington JA, Fullen MA, Moore GM, Booth CA, Trueman IC, Worsley AT, Richardson N (2008) Morphodynamics of the Morfa Dyffryn coastal dunes, mid-Wales: photographic survey 1988–2007. Environmental problems in coastal regions VII. WIT Trans Built Environ 99:211–220

Parsons DR, Walker IJ, Wiggs GF (2004) Numerical modelling of flow structures over idealized transverse aeolian dunes of varying geometry. Geomorphology 59(1–4):149–164

Piscioneri N, Smyth TA, Hesp PA (2019) Flow dynamics over a foredune scarp. Earth Surf Proc Land 44(5):1064–1076

Provoost S, Jones MLM, Edmondson SE (2011) Changes in landscape and vegetation of coastal dunes in northwest Europe: a review. J Coastal Conserv 15(1):207–226

Pye K (1982) Morphological development of coastal dunes in a humid tropical environment, Cape Bedford and Cape Flattery, North Queensland. Geogr Annaler: Ser Phys Geogr 64(3–4):213–227

Pye K, Blott SJ (2020) Is 're-mobilisation'nature restoration or nature destruction? A commentary. Discussion J Coastal Conserv 24(1):1–4

Pye K, Blott SJ, Howe MA (2014) Coastal dune stabilization in Wales and requirements for rejuvenation. J Coastal Conserv 18(1):27–54

Pye K, Blott SJ, Forbes N, Maskell LC (2020) Geomorphological and ecological change in a coastal foreland dune system, Sandscale haws, Cumbria, UK: the management challenges posed by climate change. J Coastal Conserv 24(6):1–34

Ranwell D (1958) Movement of vegetated sand dunes at Newborough Warren, Anglesey. J Ecol, 83–100

Ranwell DS, Boar R (1986) Coastal dune management guide. Institute of terrestrial Ecology

Rhind P, Jones R (2009) A framework for the management of sand dune systems in Wales. J Coastal Conserv 13(1):15–23

Richards PJ, Hoxey RP (1993) Appropriate boundary conditions for computational wind engineering models using the k-ϵ turbulence model. J Wind Eng Ind Aerodyn 46:145–153

Rust IC (1990) Coastal dunes as indicators of environmental change. South Afr J Sci 86(7):299

Smith A, Gares PA, Wasklewicz T, Hesp PA, Walker IJ (2017) Three years of morphologic changes at a bowl blowout. Cape Cod USA Geomorphology 295:452–466

Smyth TA, Hesp PA (2016) Numerical modelling of turbulent flow structures in a trough blowout. J Coastal Res 75(10075):328–332

Smyth TA, Jackson DW, Cooper JAG (2013) Three dimensional airflow patterns within a coastal trough–bowl blowout during fresh breeze to hurricane force winds. Aeolian Res 9:111–123

Smyth TA, Hesp PA, Walker IJ, Wasklewicz T, Gares PA, Smith AB (2019) Topographic change and numerically modelled near surface wind flow in a bowl blowout. Earth Surf Proc Land 44(10):1988–1999

Smyth T, Thorpe E, Rooney P (2020a) Blowout evolution between 1999 and 2015 in Ainsdale Sand Dunes National Nature Reserve, England. North West Geography, 20(1)

Smyth TA, Delgado-Fernandez I, Jackson DW, Yurk B, Rooney P (2020b) Greedy parabolics: wind flow direction within the deflation basin of parabolic dunes is governed by deflation basin width and depth. Progress in Physical Geography: Earth and Environment 44(5):643–660

Smyth TA, Wilson R, Rooney P, Yates KL (2022) Extent, accuracy and repeatability of bare sand and vegetation cover in dunes mapped from aerial imagery is highly variable. Aeolian Res 56:100799

Tsoar H (2005) Sand dunes mobility and stability in relation to climate. Physica A 357(1):50–56

Tsoar H, Blumberg DG (2002) Formation of parabolic dunes from barchan and transverse dunes along Israel's Mediterranean coast. Earth Surf Proc Land 27(11):1147–1161

Turner A, Amos SL, Williams T (2021) Coastal dunes as a sink and secondary source of marine plastics: a study at Perran Beach, southwest England. Mar Pollut Bull 173:113133

van Dijk D (2021) Foredune dynamics at a Lake Michigan site during rising and high lake levels. J Great Lakes Res 47(6):1581–1593

Walker IJ (2020) Aeolian (windblown) sand transport over beaches. In: Jackson DWT, Short AD (eds) Sandy Beach Morphodynamics. Elsevier, pp 213–253

Walker IJ, Hesp PA, Smyth TA (2022) Airflow dynamics over unvegetated and vegetated dunes. Treatise on Geomorphology, Journal, pp 415–453

Wiggs GF, Thomas DS, Bullard JE, Livingstone I (1995) Dune mobility and vegetation cover in the southwest Kalahari Desert. Earth Surf Proc Land 20(6):515–529

Wiggs GFS, Atherton RJ, Baird AJ (2004) Thresholds of aeolian sand transport: establishing suitable values. Sedimentology 51(1):95–108

Williams AT, Davies P (2001) Coastal dunes of Wales; vulnerability and protection. J Coastal Conserv 7(2):145–154

Zeng Z, Ziegler AD, Searchinger T, Yang L, Chen A, Ju K, …, Liu J (2019) A reversal in global terrestrial stilling and its implications for wind energy production. Nat Clim Change 9(12):979–985