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

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 speed is 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.

1. 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., 2012), 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).

1.1. 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, 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. (1999) 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; Figure 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).

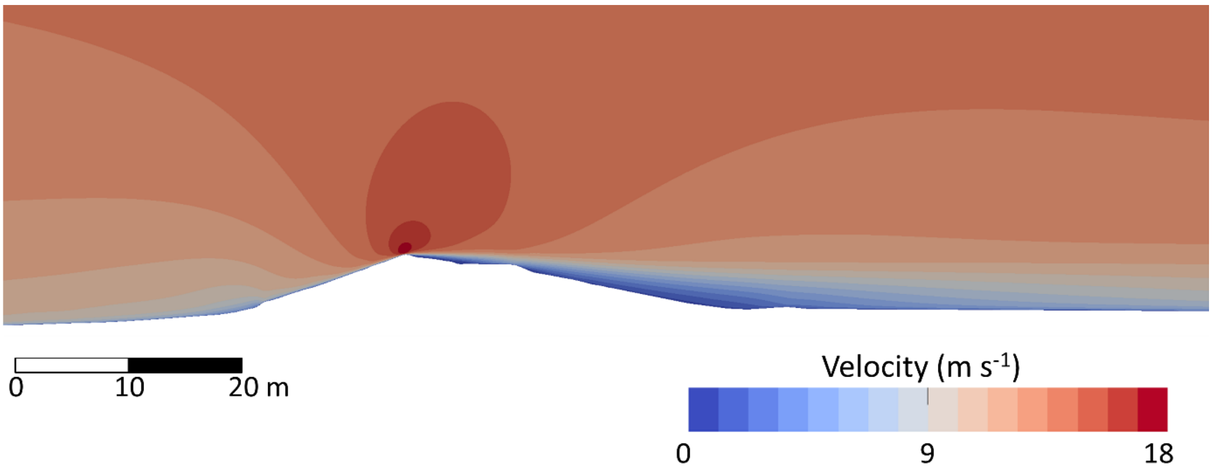

Figure 1. Isovels of wind speed above a two-dimensional surface of the foredune at Greenwich Dunes, Prince Edward Island, Canada. Incident wind direction is from left to right. Figure adapted from Hesp and Smyth (2016).

As well as the relatively simple dune topography depicted in Figure 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., 2018; 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 blowouts, 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 blowouts, 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., 2020). 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.

2. Methods

2.1 Study Sites

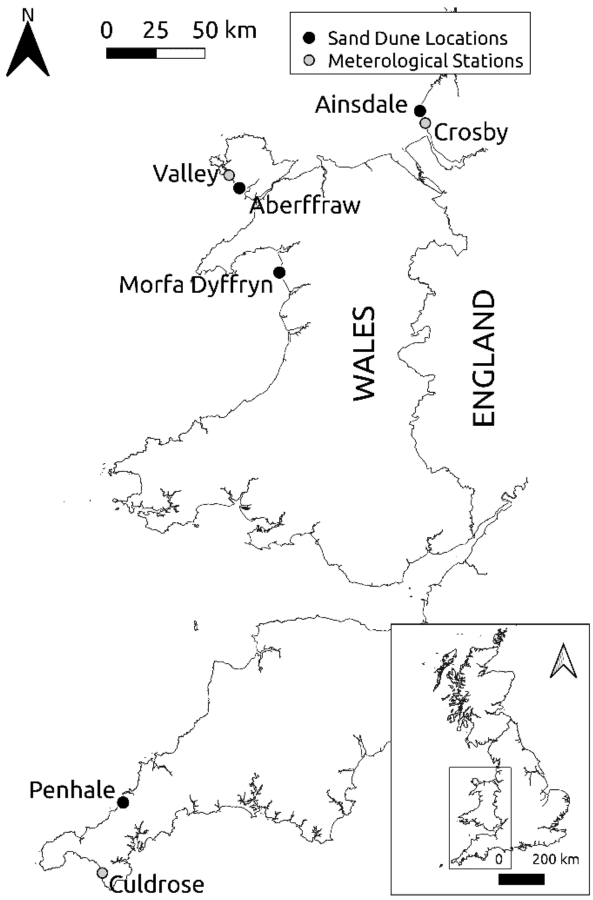

Figure 2. Map of western England and Wales denoting the four study sites investigated.

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 (Figure 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) (Figure 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.

2.1.1 Ainsdale

Ainsdale sand dunes are located on the Sefton Coast in northwest England (Figure 2). The area investigated forms part of the Ainsdale Sand Dunes National Nature Reserve and Ainsdale and Birkdale Sandhills Local Nature Reserve (Figure 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 (Figure 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 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 Figure 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 (Section 2.2. and Figure 3a).

2.1.2. Aberffraw

Aberffraw is a coastal dune field on the southwest coast of Anglesey (Figure 2). The dunes cover approximately 2.9 km^2 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 (Figure 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).

2.1.3. Morfa Dyffryn

Morfa Dyffryn is located in Gwynedd on the northeast coast of Cardigan Bay, in north Wales (Figure 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 km^2 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 (Figure 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 Dyffryn the 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).

2.1.4. Penhale Dunes

Penhale Dunes are located in Cornwall, England (Figure 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 (Figure 3d). The majority of the dunes are fixed and colonised by marram and red fescue grasses.

2.2. Bare sand delineation

For each location, areas of bare sand were mapped in QGIS using the semi-automatic classification 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 (Figure 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).

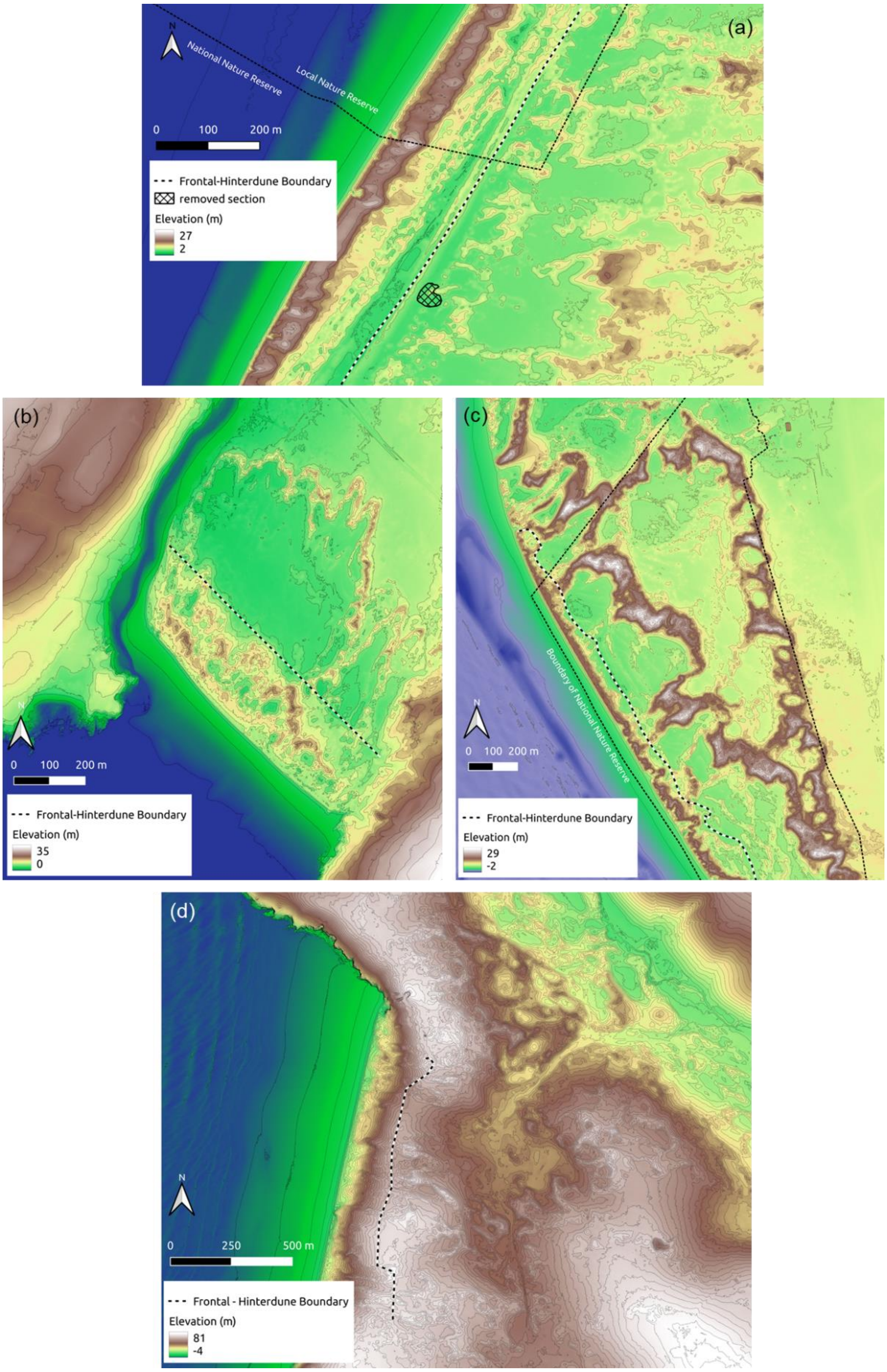

Figure 3. Digital Terrain Models (DTMs) and 2 m contours of the dunes investigated at (a) Ainsdale (b) Aberffraw (c) Morfa Dyffryn (d) Penhale Dunes. Information regarding the LiDAR tiles used is available in Table 1 and Appendix 1.

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

2.3. 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 OpenFOAM. 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 Pressure Linked Equations (SIMPLE) algorithm. This method produces a time-averaged 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 equations 1, 2 and 3 (Richards and Hoxey, 1993; Blocken et al., 2007):

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

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

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

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).

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}

The topographic surface of each study site was generated from 2 m resolution airborne LiDAR surveys performed between 2014 and 2016 (Table 1). Cell size in each computational domain decreased gradually size from 40 x 40 x 40 m at the highest vertical extent to 2.5 x 2.5 x 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}.

2.4. 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 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 ±90° to the prevailing wind direction. These cells are referred to as the 'stoss slope' in section 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. (2021), 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 (Figure 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), 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 Figures 3 - 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.

3. 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.

3.1. 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 (Figure 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 (Figure 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 are 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 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 (Figure 5) and Morfa Dyffryn (Figure 6). The steepest slopes at Ainsdale were on the frontal dune, particularly on the erosional walls of trough shaped blowouts. The hinterdune 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.

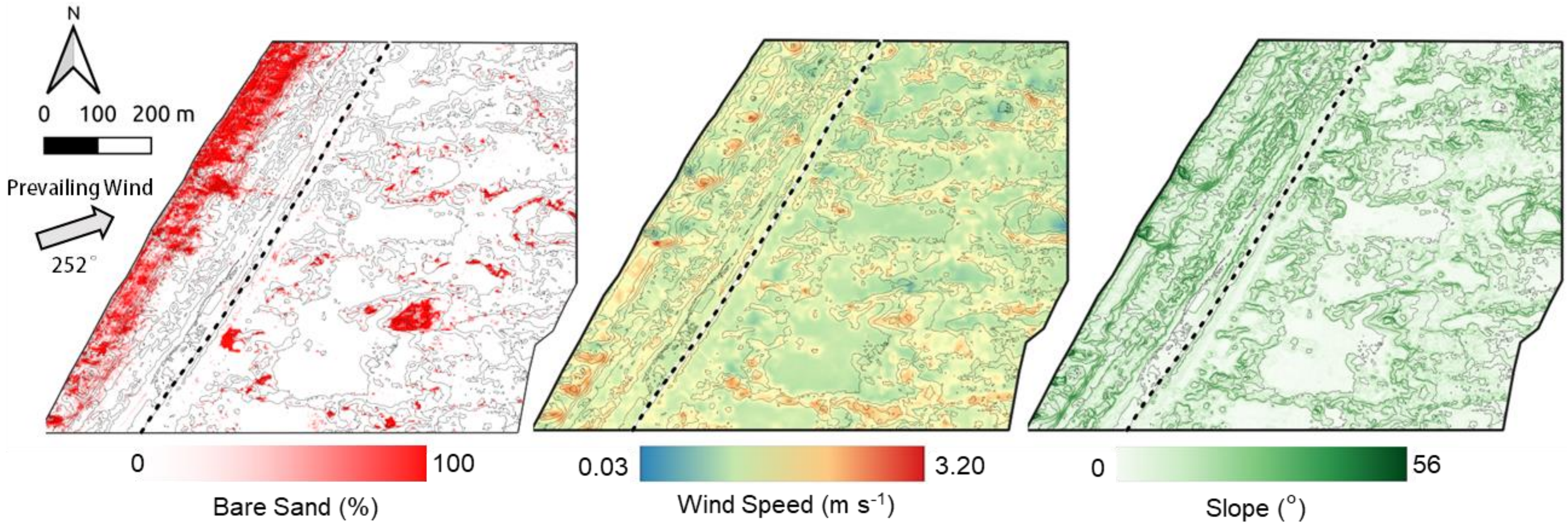

Figure 4. 2.5 m^2 resolution grid squares of bare sand percentage, wind speed (0.4 m above the surface) and dune slope at Ainsdale. The dashed line represents the frontal-hinterdune boundary. Contours are at 2 m intervals. The bare sand classification map represents the first replicate of three.

3.2 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. Topographical 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 (Figure 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 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°.

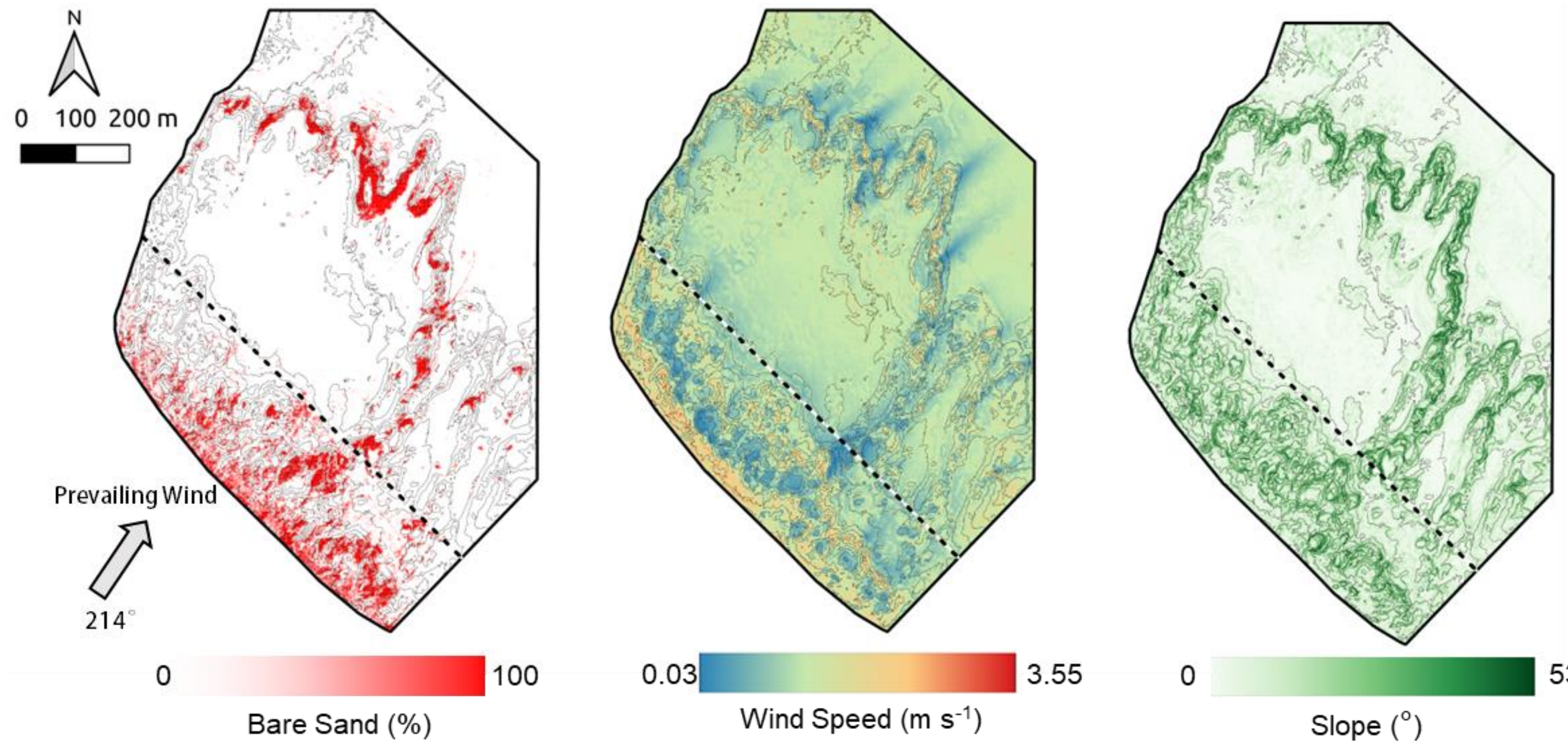

Figure 5. 2.5 m^2 resolution grid squares of bare sand percentage, wind speed (0.4 m above the surface) and dune slope at Aberffraw. The dashed line represents the frontal-hinterdune boundary. Contours are at 2 m intervals. The bare sand classification map represents the first replicate of three.

3.3. Morfa Dyffryn

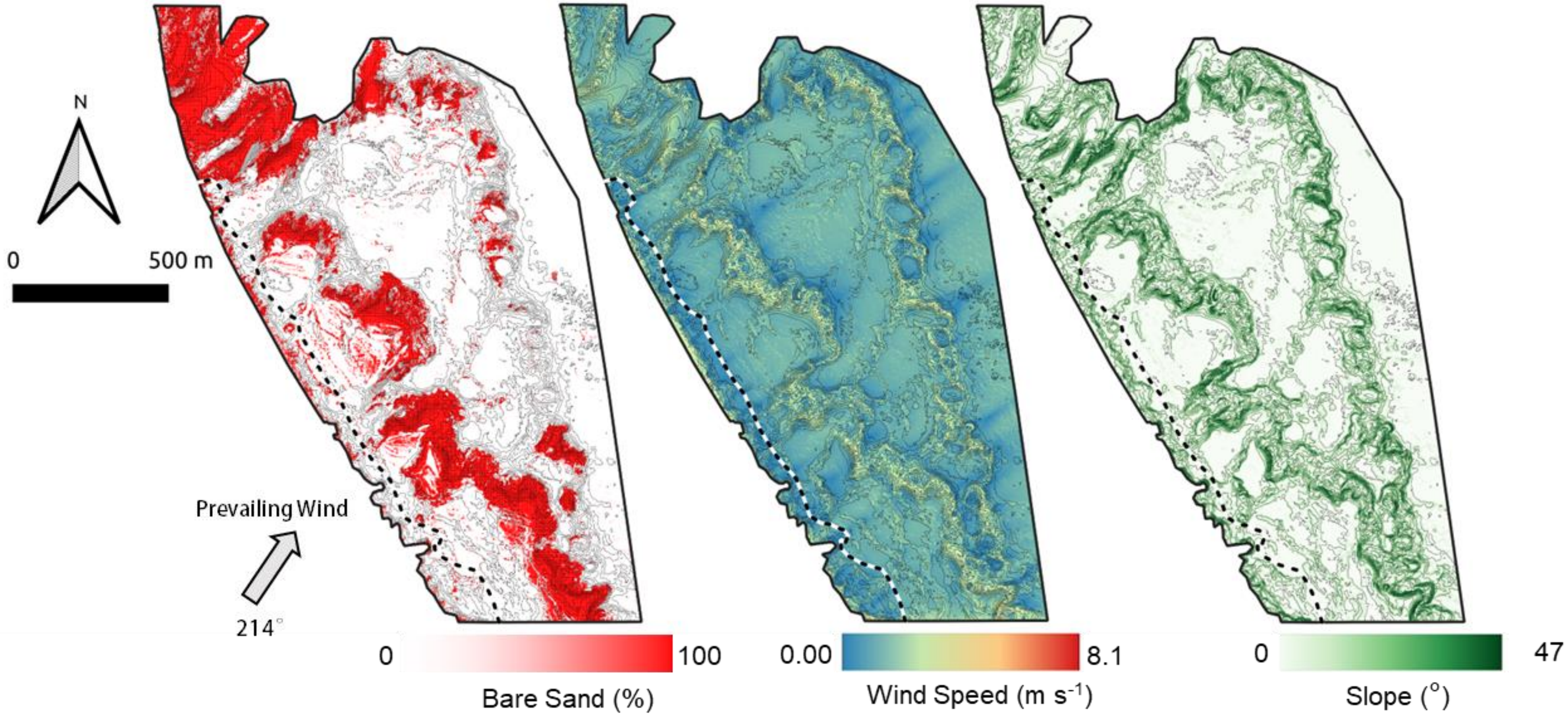

Figure 6. 2. Figure 6. 2.5 m^2 resolution grid squares of bare sand percentage, wind speed (0.4 m above the surface) and dune slope at Morfa Dyffryn. The dashed line represents the frontal-hinterdune boundary. Contours are at 2 m intervals. The bare sand classification map represents the first replicate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

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 (Figure 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 (Figure 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 (Figure 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.

3.4. 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 (Figure 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.

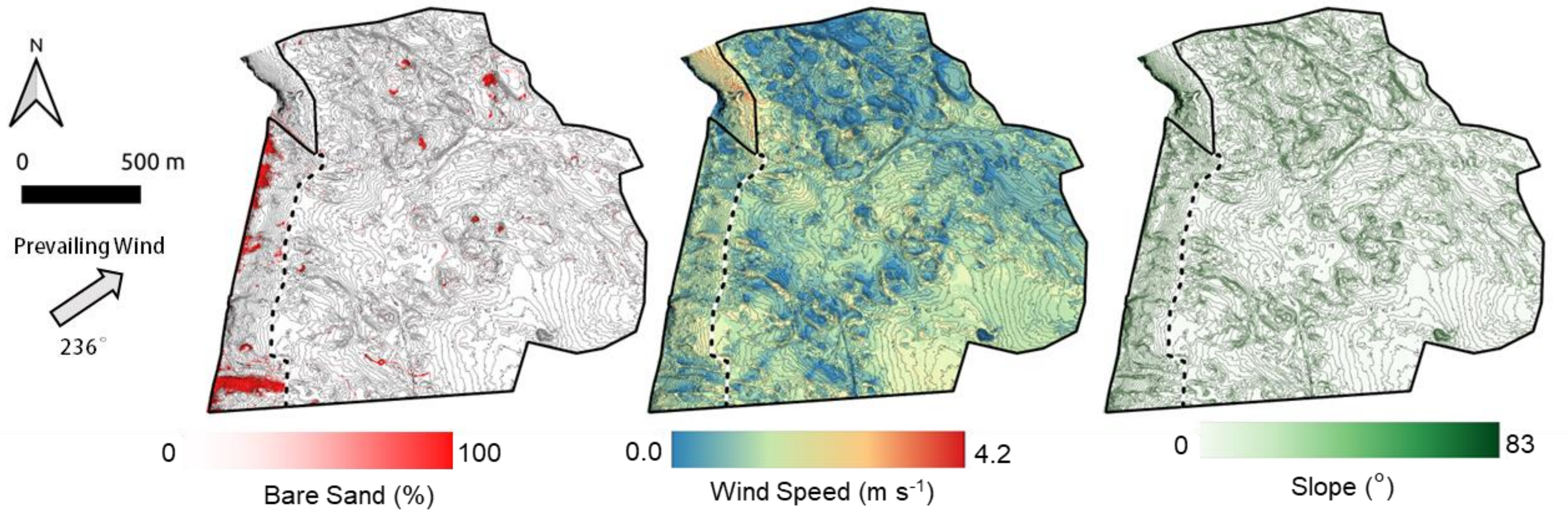

Figure 7. 2.5 m^2 resolution grid squares of bare sand percentage, wind speed (0.4 m above the surface) and dune slope at Penhale Dunes. The dashed line represents the frontal-hinterdune boundary. Contours are at 2 m intervals. The bare sand classification map represents the first replicate of three.

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 (Figure 7).

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

3.4. Statistics

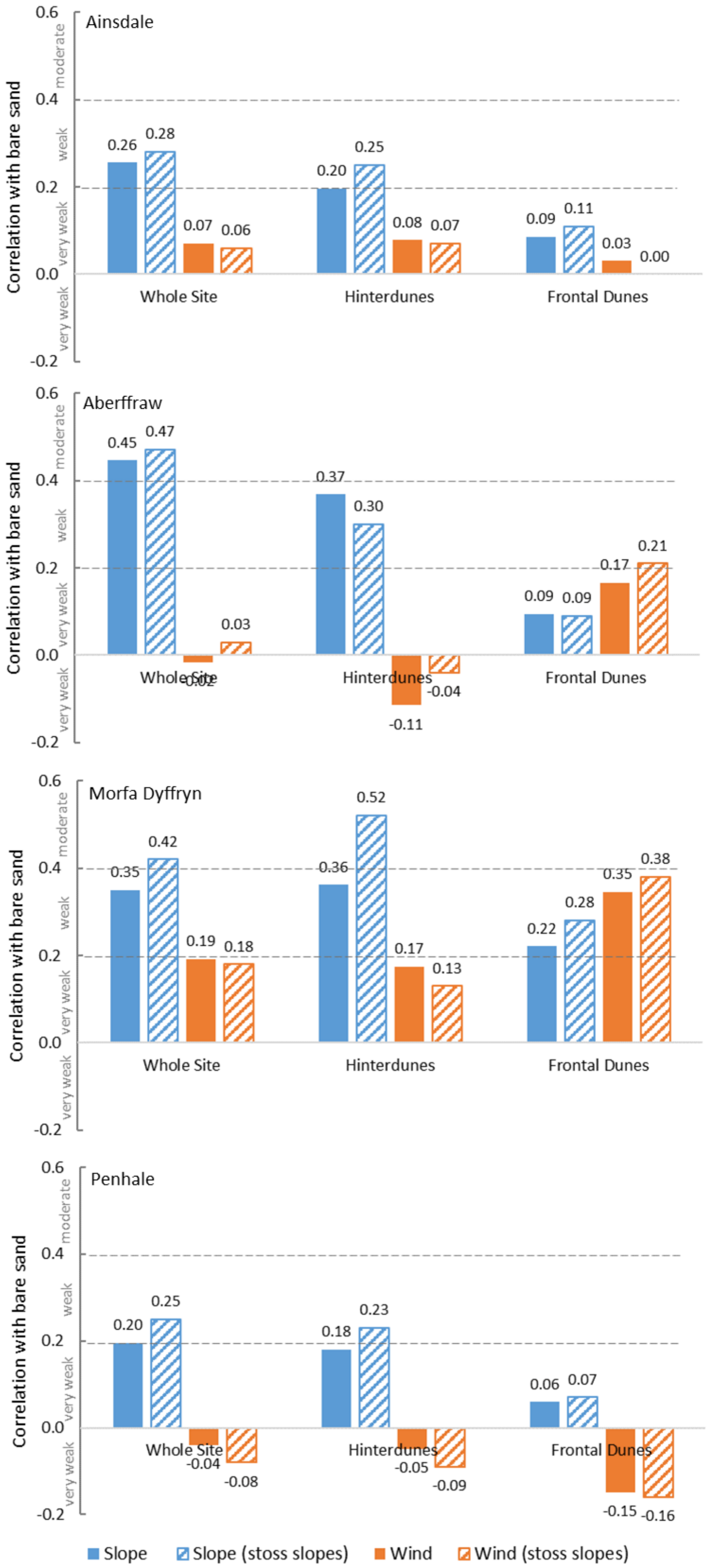

Figure 8. Spearman's rank correlation coefficients of bare sand, slope angle and wind speed calculated at 0.4 m above the surface at each coastal site dune investigated. Hatched bars indicate where only slopes ±90° to the prevailing incident wind were considered (stoss slopes).

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 (Figure 8). At a whole site scale, the strength of correlation between slope and bare sand increased when only slopes that were orientated ±90° to the prevailing wind were considered. Examining only those windward slopes also increased the strength of correlation in the hinterdunes for each site apart from at Aberffraw (Figure 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 (Figure 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.

4. 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 (Figure 9).

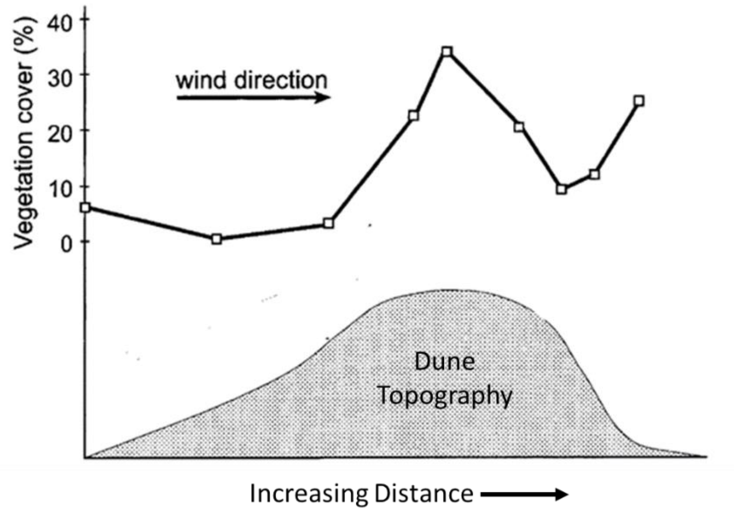

Figure 9. Conceptualised vegetation cover over parabolic dunes in a rapidly vegetating coastal dune field in Israel (Image adapted from Tsoar and Blumberg (2002))

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 Figure 4c in Barchyn and Hugenholtz (2013))

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

4.1 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 (Figure 4). In these erosional hollows bare sand is likely present not due to high wind speed, 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 (Figure 4), between the foredune and line of parabolic dunes at Aberffraw (Figure 5), both areas of wet dune slack and grassland, and the central and southern areas of hinterdune at Penhale (Figure 6). (3) There are also multiple examples where substantial patches of bare sand have been created by human footfall (frontal dunes at Ainsdale, Figure 4) and vehicular access to the dunes (hinterdunes at Aberffraw, Figure 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 (Figure 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 19^{th} to late 20^{th} century (Provoost et al., 2011) and were explicitly recorded by Ranwell and Boar (1986) at Aberffraw.

4.2 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 (Figure 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 in classified areas of bare sand, differences in the statistical correlations of bare sand, wind speed and slope between classification iterations were negligible.

4.3 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°.

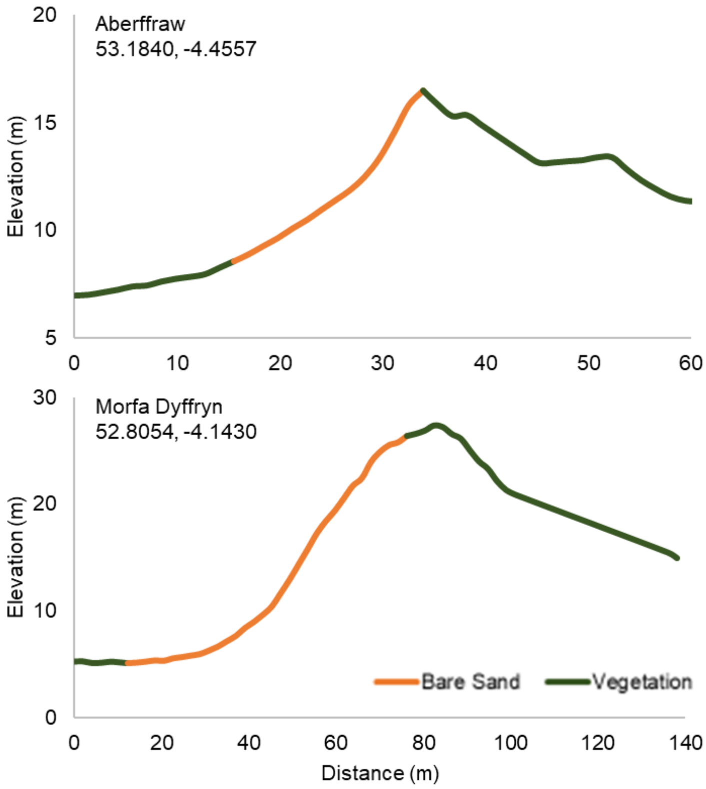

Figure 10. Representative topographic profiles of persistent bare sand patches at Aberffraw and Morfa Dyffryn. Topographic and vegetation profiles were generated from data in 2015 at Aberffraw and 2014 at Morfa Dyffryn (Table 1). Coordinates refer to the exact location of each dune and are in WGS 84 Web Mercator (Google Earth).

5. 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 in coastal dunes 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.

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Data availability

The processed bare sand, wind speed, aspect and slope data at each dune site used in this study is available from: https://doi.org/10.5285/972599af-0cc3-4e0e-a4dc-2fab7a6dfc85

References

Arens, S. M., & Geelen, L. H. W. T. (2006). Dune landscape rejuvenation by intended destabilisation in the Amsterdam water supply dunes. *Journal of coastal research*, *22*(5), 1094-1107.

Arens, S. M., Slings, Q., & De Vries, C. N. (2004). Mobility of a remobilised parabolic dune in Kennemerland, The Netherlands. Geomorphology, 59(1-4), 175-188.

Bagnold, R. A. (1941). *The physics of blown sand and desert dunes*. Courier Corporation.

Bailey, S. D., & Bristow, C. S. (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. *Restoration of coastal dunes*, 173-185.

Barchyn, T. E., & Hugenholtz, C. H. (2013). Reactivation of supply-limited dune fields from blowouts: A conceptual framework for state characterization. Geomorphology, 201, 172-182.

Bauer, B. O., & Wakes, S. J. (2022). CFD simulations of wind flow across scarped foredunes: Implications for sediment pathways and beach–dune recovery after storms. *Earth Surface Processes and Landforms*, *47*(12), 2989-3015.

Blocken, B., Stathopoulos, T., & Carmeliet, J. (2007). CFD simulation of the atmospheric boundary layer: wall function problems. Atmospheric environment, 41(2), 238-252.

Bradley, A. V., Haughan, A. E., Al-Dughairi, A., & McLaren, S. J. (2019). Spatial variability in shrub vegetation across dune forms in central Saudi Arabia. Journal of Arid Environments, 161, 72-84.

Carter, R. W. G., & Stone, G. W. (1989). Mechanisms associated with the erosion of sand dune cliffs, Magilligan, Northern Ireland. Earth Surface Processes and Landforms, 14(1), 1-10.

Carter, R. W. G., Hesp, P. A., & Nordstrom, K. F. (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(0.1), 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”. *Journal of Coastal Conservation*, *24*(3), 1-4.

Darke, I. B., Eamer, J. B., Beaugrand, H. E., & Walker, I. J. (2013). Monitoring considerations for a dynamic dune restoration project: Pacific Rim National Park Reserve, British Columbia, Canada. Earth surface processes and landforms, 38(9), 983-993.

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

Delgado-Fernandez, I., Smyth, T. A., Jackson, D. W., Smith, A. B., & Davidson-Arnott, R. G. (2018). Event-scale dynamics of a parabolic dune and its relevance for mesoscale evolution. Journal of Geophysical Research: Earth Surface, 123(11), 3084-3100.

Durán, O., Silva, M. V. N., Bezerra, L. J. C., Herrmann, H. J., & Maia, L. P. (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, G. S., Bennett, S. W., Olyphant, G. A., Bauch, N. J., Ferguson, V., Gellasch, C. A., ... & Woodfield, M. C. (1998). Windflow circulation patterns in a coastal dune blowout, south coast of Lake Michigan. Journal of Coastal Research, 451-460.

Gao, J., Kennedy, D. M., & Konlechner, T. M. (2020). Coastal dune mobility over the past century: A global review. Progress in Physical Geography: Earth and Environment, 44(6), 814-836.

Gares, P. A. (1992). Topographic changes associated with coastal dune blowouts at Island Beach State Park, New Jersey. Earth Surface Processes and Landforms, 17(6), 589-604.

Gares, P. A., & Nordstrom, K. F. (1995). A cyclic model of foredune blowout evolution for a leeward coast: Island Beach, New Jersey. Annals of the Association of American Geographers, 85(1), 1-20.

Garzon, J. L., Costas, S., & Ferreira, O. (2022). Biotic and abiotic factors governing dune response to storm events. Earth Surface Processes and Landforms, 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, P. A., & Pringle, A. (2001). Wind flow and topographic steering within a trough blowout. *Journal of Coastal Research*, 597-601.

Hesp, P. A., & Smyth, T. A. (2016). Jet flow over foredunes. Earth Surface Processes and Landforms, 41(12), 1727-1735.

Hesp, P.A. and Smyth, T.A.G. (2019). Anchored Dunes. In Aeolian Geomorphology (eds I. Livingstone and A. Warren). Wiley.

Hesp, P. A., & Smyth, T. A. (2021). CFD flow dynamics over model scarps and slopes. *Physical Geography*, *42*(1), 1-24.

Hesp, P. A., Smyth, T. A., Nielsen, P., Walker, I. J., Bauer, B. O., & Davidson-Arnott, R. (2015). Flow deflection over a foredune. Geomorphology, 230, 64-74.

Hugenholtz, C. H., & Wolfe, S. A. (2006). Morphodynamics and climate controls of two aeolian blowouts on the northern Great Plains, Canada. Earth Surface Processes and Landforms: The Journal of the British Geomorphological Research Group, 31(12), 1540-1557.

Jackson, D. W., Costas, S., González-Villanueva, R., & Cooper, A. (2019). A global 'greening'of coastal dunes: An integrated consequence of climate change?. Global and Planetary Change, 182, 103026.

Kolesar, K. R., Mavko, M., Burgess, E., Nguyen, N., & Schaaf, M. D. (2022). A modified resultant drift potential for more accurate prediction of sand transportation in the vicinity of the Keeler Dunes, California. *Aeolian Research*, *58*, 100819.

Konlechner, T. M., Ryu, W., Hilton, M. J., & Sherman, D. J. (2015). Evolution of foredune texture following dynamic restoration, Doughboy Bay, Stewart Island, New Zealand. Aeolian Research, 19, 203-214.

Lancaster, N. (1988). Development of linear dunes in the southwestern Kalahari, southern Africa. Journal of arid environments, 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. Science of the Total Environment, 782, 146813.

Levin, N., & Ben-Dor, E. (2004). Monitoring sand dune stabilization along the coastal dunes of Ashdod-Nizanim, Israel, 1945–1999. Journal of arid Environments, 58(3), 335-355.

Levin, N., Ben-Dor, E., Kidron, G. J., & Yaakov, Y. (2008b). Estimation of surface roughness (z0) over a stabilizing coastal dune field based on vegetation and topography. Earth Surface Processes and Landforms: The Journal of the British Geomorphological Research Group, 33(10), 1520-1541.

Levin, N., Kidron, G. J., & BEN-DOR, E. Y. A. L. (2008a). A field quantification of coastal dune perennial plants as indicators of surface stability, erosion or deposition. Sedimentology, 55(4), 751-772.

Li, F. R., Zhao, L. Y., Zhang, H., Liu, J. L., Lu, H. Y., & Kang, L. F. (2009). Habitat degradation, topography and rainfall variability interact to determine seed distribution and recruitment in a sand dune grassland. Journal of Vegetation Science, 20(5), 847-859.

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

McAtee, J. W., & Drawe, D. L. (1980). Human impact on beach and foredune vegetation of North Padre Island, Texas. Environmental Management, 4(6), 527-538.

Millington, J. A., Fullen, M. A., Moore, G. M., Booth, C. A., Trueman, I. C., Worsley, A. T., & Richardson, N. (2008). Morphodynamics of the Morfa Dyffryn coastal dunes, mid-Wales: photographic survey 1988–2007. Environmental problems in coastal Regions VII. WIT Transactions on the Built Environment, 99, 211-220.

Parsons, D. R., Walker, I. J., & Wiggs, G. F. (2004). Numerical modelling of flow structures over idealized transverse aeolian dunes of varying geometry. Geomorphology, 59(1-4), 149-164.

Piscioneri, N., Smyth, T. A., & Hesp, P. A. (2019). Flow dynamics over a foredune scarp. *Earth Surface Processes and Landforms*, *44*(5), 1064-1076.

Provoost, S., Jones, M.L.M. and Edmondson, S.E., 2011. Changes in landscape and vegetation of coastal dunes in northwest Europe: a review. Journal of Coastal Conservation, 15(1), pp.207-226.

Pye, K. (1982). Morphological development of coastal dunes in a humid tropical environment, Cape Bedford and Cape Flattery, North Queensland. Geografiska Annaler: Series A, Physical Geography, 64(3-4), 213-227.

Pye, K., & Blott, S. J. (2020). Is 're-mobilisation'nature restoration or nature destruction? A commentary. Discussion. Journal of Coastal Conservation, 24(1), 1-4.

Pye, K., Blott, S. J., Forbes, N., & Maskell, L. C. (2020). Geomorphological and ecological change in a coastal foreland dune system, Sandscale haws, Cumbria, UK: the management challenges posed by climate change. Journal of Coastal Conservation, 24(6), 1-34.

Pye, K., Blott, S.J. and Howe, M.A., 2014. Coastal dune stabilization in Wales and requirements for rejuvenation. Journal of coastal conservation, 18(1), pp.27-54.

Ranwell, D. (1958). Movement of vegetated sand dunes at Newborough Warren, Anglesey. The Journal of Ecology, 83-100.

Ranwell, D. S., & 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. Journal of Coastal Conservation, 13(1), 15-23.

Richards, P. J., & Hoxey, R. P. (1993). Appropriate boundary conditions for computational wind engineering models using the k-ϵ turbulence model. Journal of wind engineering and industrial aerodynamics, 46, 145-153.

Rust, I. C. (1990). Coastal dunes as indicators of environmental change. South African Journal of Science, 86(7), 299.

Smith, A., Gares, P. A., Wasklewicz, T., Hesp, P. A., & Walker, I. J. (2017). Three years of morphologic changes at a bowl blowout, Cape Cod, USA. Geomorphology, 295, 452-466.

Smyth, T. A., & Hesp, P. A. (2016). Numerical modelling of turbulent flow structures in a trough blowout. *Journal of Coastal Research*, (75 (10075)), 328-332.

Smyth, T. A., Delgado-Fernandez, I., Jackson, D. W., 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, T. A., Hesp, P. A., Walker, I. J., Wasklewicz, T., Gares, P. A., & Smith, A. B. (2019). Topographic change and numerically modelled near surface wind flow in a bowl blowout. Earth Surface Processes and Landforms, 44(10), 1988-1999.

Smyth, T. A., Jackson, D. W., & Cooper, J. A. G. (2013). Three dimensional airflow patterns within a coastal trough–bowl blowout during fresh breeze to hurricane force winds. Aeolian Research, 9, 111-123.

Smyth, T. A., Wilson, R., Rooney, P., & Yates, K. L. (2022). Extent, accuracy and repeatability of bare sand and vegetation cover in dunes mapped from aerial imagery is highly variable. Aeolian Research, 56, 100799.

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).

Tsoar, H. (2005). Sand dunes mobility and stability in relation to climate. Physica A: Statistical Mechanics and its Applications, 357(1), 50-56.

Tsoar, H., & Blumberg, D. G. (2002). Formation of parabolic dunes from barchan and transverse dunes along Israel's Mediterranean coast. Earth Surface Processes and Landforms, 27(11), 1147-1161.

Turner, A., Amos, S. L., & Williams, T. (2021). Coastal dunes as a sink and secondary source of marine plastics: A study at Perran Beach, southwest England. Marine Pollution Bulletin, 173, 113133.

van Dijk, D. (2021). Foredune dynamics at a Lake Michigan site during rising and high lake levels. Journal of Great Lakes Research, 47(6), 1581-1593.

Walker, I. J. (2020). Aeolian (windblown) sand transport over beaches. In D. W. T. Jackson, & A. D. Short (Eds.), Sandy Beach Morphodynamics (pp. 213-253). Elsevier.

Walker, I. J., Hesp, P. A., & Smyth, T. A. (2022). Airflow dynamics over unvegetated and vegetated dunes. Journal: Treatise on Geomorphology, 415-453.

Wiggs, G. F. S., Atherton, R. J., & Baird, A. J. (2004). Thresholds of aeolian sand transport: establishing suitable values. *Sedimentology*, *51*(1), 95-108.

Wiggs, G. F., Thomas, D. S., Bullard, J. E., & Livingstone, I. (1995). Dune mobility and vegetation cover in the southwest Kalahari Desert. Earth surface processes and landforms, 20(6), 515-529.

Williams, A. T., & Davies, P. (2001). Coastal dunes of Wales; vulnerability and protection. *Journal of Coastal Conservation*, *7*(2), 145-154.

Zeng, Z., Ziegler, A. D., Searchinger, T., Yang, L., Chen, A., Ju, K., ... & Liu, J. (2019). A reversal in global terrestrial stilling and its implications for wind energy production. Nature Climate Change, 9(12), 979-985.