Title:
Parabolic dunes and their transformations under environmental and climatic changes: toward
a conceptual framework for understanding and prediction

Authors:
Na Yan¹ and Andreas C.W. Baas²
¹ Department of Geography, King’s College London, UK – na.yan@kcl.ac.uk
² Department of Geography, King’s College London, UK – andreas.baas@kcl.ac.uk

Contact Information:
Na Yan
Tel: +44(0)79-7724-9292
Email: na.yan@kcl.ac.uk; yanna813@gmail.com

Department of Geography
King’s College London
Strand, London
WC2R 2LS
United Kingdom
Abstract:
The formation and evolution of parabolic aeolian dunes depends on vegetation, and as such is particularly sensitive to changes in environmental controls (e.g., temperature, precipitation, and wind regime) as well as to human disturbances (e.g., grazing, agriculture, and recreation). Parabolic dunes can develop from the stabilisation of highly mobile barchans and transverse dunes as well as from blowouts, as a consequence of colonisation and establishment of vegetation when aeolian sand transport is reduced and/or when water stress is relieved (by increasing precipitation, for instance). Conversely, existing parabolic dunes can be activated and may be transformed into barchans and/or transverse dunes when vegetation suffers environmental or anthropogenic stresses. Predicted increases in temperature and drought severity in various regions raise concerns that dune activation and transformation may intensify, and this intensification would have far-reaching implications for environmental, social, and economic sustainability. To date, a broad examination of the development of parabolic dunes and their related transformations across a variety of climate gradients has been absent. This paper reviews existing literature, compares data on the morphology and development of parabolic dunes in a comprehensive global inventory, and scrutinises mechanisms of different dune transformations and the eco-geomorphic interactions involved. This knowledge is then integrated into a conceptual framework to facilitate understanding and prediction of potential aeolian dune transformations induced by changes in environmental controls and human activities. This conceptual framework can aid judicious land management policies for better adaptations to climatic changes.

Keywords:
Dune transformation; parabolic dune; aeolian; climatic change; eco-geomorphic response; vegetation change
1. Introduction

Desertification and associated land degradation in dry regions is responsible for increased emission and reduced sink of atmospheric carbon, currently accounting for about 4% of global emissions (Lal, 2001; Millennium Ecosystem Assessment, 2005). Land degradation and vegetation loss also result in severe reduction of global food production (Scherr and Yadav, 1996). Projections of future climatic change, in particular increases in temperature and drought severity and decreases in freshwater availability expected in various regions around the world (IPCC, 2013; Maestre et al., 2012), raise concerns that aeolian activity and desertification may be exacerbated by more active dune transformations, particularly the activation of dunes that are currently stabilised by vegetation and/or biogenic crusts (Ashkenazy et al., 2012; Forman et al., 1992; Lancaster, 1997; Le Houérou, 1996; Muhs and Maat, 1993; Muhs et al., 1996; Thomas et al., 2005; Thomas and Leason, 2005). Relatively small changes in climatological parameters may contribute to an abrupt change in vegetation cover and catastrophic shifts between states of eco-geomorphic systems (Bhirey et al., 2011; Lavee et al., 1998; Muckersie and Shepherd, 1995; Rietkerk et al., 2004; Sole, 2007; Yizhaq et al., 2007; Yizhaq et al., 2009).

Despite a growing awareness of great sensitivity of aeolian landforms to vegetation change as well as the diverse feedbacks between vegetation and sand erosion and burial, the complex eco-geomorphic interrelations between vegetation and dune landforms are not completely understood. Parabolic dunes, in particular, are common aeolian landforms that are strongly controlled by eco-geomorphic interactions. Such dunes often form where there is an adequate sand supply, unidirectional wind regime, and moderate vegetation cover (Hugenholtz et al., 2008; Hugenholtz, 2010; Lancaster, 1995; McKee and Bigarella, 1979). Under ameliorating vegetation conditions, parabolic dunes can develop from highly mobile non-vegetated dunes such as barchan dunes and transverse dunes (Hart et al., 2012; Hesp and
Walker, 2013; Reitz et al., 2010; Tsoar and Blumberg, 2002). When the vegetation cover decreases, however, parabolic dunes can be transformed back to highly mobile, non-parabolic dunes (Anton and Vincent, 1986; Hack, 1941). The development and transformations of parabolic dunes are also highly sensitive to changes in many environmental factors such as precipitation (Landsberg, 1956; Stetler and Gaylord, 1996), temperature (Wolfe and Hugenholtz, 2009), wind strength and variability (Hesp, 2002; Tsoar et al., 2009), as well as to changes in land management and other anthropogenic factors (Hesp, 2001; Tsoar and Blumberg, 2002). A brief discussion on the distribution, morphology and change of parabolic dunes was recently provided by Goudie (2011), but there has been no detailed examination of the differences in development of parabolic dunes and their related transformations on a global scale across a wide climatic gradient.

This paper reviews past research on parabolic dunes and their related transformations on a global scale, exploring mechanisms of different dune transformations and their indications in the context of climatic change, mediated by vegetation, which is then integrated into a conceptual framework of understanding and predicting dune transformations influenced by changes in environmental variables and human activities. Analysis of dune landform transformations within a conceptual framework allows for the examination of geomorphic responses to environmental fluctuations and climatic change on different temporal and spatial scales. This analysis also provides a better understanding of different dune transformation mechanisms and possible dunefield evolutions, and provides a framework for planning judicious land management practices.

2. Morphology, Development, and Migration of Parabolic Dunes

Simple parabolic dunes are U- or V-shaped dunes in plan with two trailing arms pointing upwind, a deflation basin contained within arms, and a depositional lobe at the downwind end
Vegetation, usually shrubs or trees, surrounding the parabolic dunes can resist widening of the deflation basin, whilst plants on the trailing arms can bind sand and maintain the parabolic shape of dunes. Many parabolic dunes have a slip face, and some large ones may have multiple crests and slip faces. As airflow approaches towards the dune crest, flow is compressed by the stoss slope, resulting in the increases in shear stress and sediment transport. Beyond the crest, flow expands, and may create a separation zone within which positive pressure causes reversal of flow back up the lee slope, forming ‘back-eddies’ (Delgado-Fernandez et al., 2013; Walker and Nickling, 2002). In the zones of flow separation, flow deceleration causes grainfall deposition which forms grainfall lamination on slip faces (Hunter, 1977). As deposition continues, avalanching occurs where the slope angle reaches the critical angle of repose. The resulting grainflow and sand flowage changes pre-existing stratification and forms cross-strata (Hugenholtz et al., 2007; Hunter, 1977).

Parabolic dunes can exhibit variable morphologies (Cooke et al., 1993; Kilibarda and Blockland, 2011) (Figure 1), governed by wind regime, sediment supply and local vegetation characteristics (Baas, 2007; Hack, 1941; Hugenholtz, 2010; Pye, 1990; Rubin and Hunter, 1987; Wasson and Hyde, 1983).

Figure 1. Diagram from Pye and Tsoar (1990) showing the following seven morphologies of parabolic dunes: (a) hairpin; (b) lunate; (c) hemicyclic; (d) digitate; (e) nested; (f) long-walled transgressive ridge with secondary transverse dunes; and (g) rake-like en-echelon dunes.
Elongated parabolic dunes with long-walled arms, also referred to as hairpin- or U-shaped dunes, develop in a strong unidirectional wind regime, whereas a greater directional variability results in much shorter trailing arms and imbricate dune forms (Gaylord and Dawson, 1987; Hesp and Walker, 2013; Pye, 1982; Pye, 1983c; Tinley, 1985). Cross-winds that blow oblique to the prevailing wind may lead to a left- or right-handed asymmetry in dune morphology. Where multiple discrete wind directions dominate at different times, hemicyclic- or digitate-shaped parabolic dunes may form (Filion and Morisset, 1983; Pye and Tsoar, 1990). The seasonal variations of winds also significantly influence airflow patterns and sediment transport over dunes (Byrne, 1997; Hansen et al., 2009).

Relatively abundant sediment supply is crucial for dunes to maintain their mobility and grow in height. The availability of external sediment supply from sandy beaches and foredunes largely controls the size of coastal dunefields (Aren et al. 2004). As dunes move forward, they can also grow in height by incorporating sand from their substrata underneath (Livingstone and Warren, 1996). If dunes move onto a non-sandy substratum in absence of an external sediment supply, the depositional lobes may flatten gradually due to continuous sand loss.

The ecological conditions and characteristics of the regional vegetation are the other essential factors determining the morphology of parabolic dunes. Digitate parabolic dunes are usually associated with the presence of a forest cover, as trees force dunes to move in divergent directions and facilitate the formation of high depositional lobes with steep windward and lee slopes (Buynevich et al., 2010; Filion and Morisset, 1983; Levin, 2011). If regeneration of the tree population is interrupted (by wildfires, for example), digitate parabolic dunes can further transform to hemicyclic dunes (Filion and Morisset, 1983). Long-lived perennial shrubs also may play an important role in trapping sediments, developing into nebkhas, and maintaining the shape of parabolic dunes (Hesp, 2008; Tsoar and Blumberg,
2002). Ephemeral annual plants, however, can only anchor dune surfaces temporarily and suffer abrupt changes under external pressures (e.g., precipitation, temperature and grazing intensities), and therefore exert minimal impacts.

Migration of parabolic dunes is principally controlled by the interplay between sand drift potential imparted by wind regime, moisture content and vegetation cover (Ash and Wasson, 1983; Bagnold, 1941; Fryberger, 1979; Lancaster, 1997; Lancaster and Baas, 1998).

The orientation of coastal parabolic dunes is largely determined by wind regime (Jennings, 1957; Landsberg, 1956), which is usually defined in terms of sand drift potential, reflecting the capacity of winds to transport sediments, as an index of regional wind energy (Arens et al., 2004; Fryberger, 1979; Levin, 2011; Levin et al., 2006; Tsoar, 2005; Wasson and Hyde, 1983).

Moisture content, which is largely controlled by precipitation and evapotranspiration, is another crucial factor for modifying sand transport and the associated dune migration (Lancaster, 1997; Tsoar, 2005). The spatial heterogeneity of soil moisture resulting from differences of local micro-environments (e.g., slope) can lead to spatial variations in sedimentation balance (Ritsema and Dekker, 1994; Stout, 2004). A greater moisture content of sand increases the critical shear velocity needed for the initiation of particle movement and inhibits sediment transport (Belly, 1964; Cornelis and Gabriels, 2003; Davidson-Arnott et al., 2008; Hugenholtz et al., 2009; Jackson and Nordstrom, 1998; Namikas and Sherman, 1995; Wiggs et al., 2004). Precipitation can, therefore, reduce the migration rate of parabolic dunes significantly (Arens et al., 2004). For example, at the Great Sand Dunes National Park and Preserve (Colorado, USA), some parabolic dunes were mobilised six times faster during drought periods than (preceding or subsequent) wet periods (Marín et al., 2005).
Unpredictable rainfall events in semi-arid regions, moreover, encourage the growth of vegetation characterised by a ‘pulse-activity’ response (Noy-Meir, 1973; Wand et al., 1999), and further increase surface roughness (Wolfe and Nickling, 1996). In the more humid coastal dune areas, however, precipitation variations can also significantly influence the local water table and subsequently change the sand availability, vegetation cover and dunefield mobility (Luna et al., 2012; Miot da Silva and Hesp, 2013). The spatial distribution and temporal variations of vegetation then alter the airflow dynamics over the surface, and influence the spatial heterogeneity in migration speeds of a single dune and of a dunefield as a whole (Kuriyama et al., 2005; Lancaster and Baas, 1998; Wasson and Nanninga, 1986; Wiggs et al., 1995).

Migration rates of parabolic dunes reported in literature (Table 1) naturally vary because of different local environmental settings but also depend on the measuring methods used. Typical measurement approaches, from short to long time scales, include ground surveys by setting transects and/or pins (Arens et al., 2004; Cooper, 1958; Ranwell, 1958; Wolfe and Lemmen, 1999), interpretation of multi-time aerial photographs and topographic maps (Anthonsen et al., 1996; Arens et al., 2004; Bailey and Bristow, 2004; Hesp, 2001; Hugenholtz et al., 2008; Marín et al., 2005; Pye, 1982; Siljeström and Clemente, 1990; Stetler and Gaylord, 1996; Tsoar and Blumberg, 2002), and chronological dating such as tree-ring dating and optical dating (Cooper, 1958; David et al., 1999; Wiles et al., 2003).

<table>
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<tr>
<th>Reference</th>
<th>Study Region</th>
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<td></td>
<td>Mt. Mitchell dune, Cape Flattery, Queensland</td>
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<td>Story, 1982</td>
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<td>Source</td>
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<td>Northern Cape York Peninsula</td>
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<td>Pye, 1983b</td>
<td>Temple Bay, Northern Cape York Peninsula</td>
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<td>Wolfe and Lemmen, 1999</td>
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<td>David et al., 1999</td>
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<td>Anthonsen et al., 1996</td>
<td>Råbjerg Mile, Skagen Odde</td>
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<td>Arens et al., 2004</td>
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<td>1.5-6.7</td>
<td>aerial photographic interpretation and field surveys by a total station</td>
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<td>remote sensing images (Landsat ETM)</td>
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<td>stereo aerial photographic interpretation</td>
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<td>Yurk et al., 2002</td>
<td>Holland, eastern shore of Lake Michigan</td>
<td>1.45</td>
<td>aerial photographic interpretation</td>
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In contrast to barchan dunes, which move forward as coherent entities, parabolic
dunes continuously change in form as they elongate downwind. Whilst the arms of parabolic
dunes are largely fixed in place by vegetation, the dune depositional lobes migrate at various
rates. A wide morphological variety reinforces a great spatial heterogeneity in dune mobility.
Precisely determining a migration rate of parabolic dunes is thus a challenge (Bailey and
Bristow, 2004; Girardi and Davis, 2010). Some studies have measured the advance of slip
faces (Cooper, 1958; Forman et al., 2008; Wolfe and Lemmen, 1999), whereas others have
used a linear-fit method or a crest-to-crest method (Bailey and Bristow, 2004). One study,
Furthermore, proposed a calculus method aided with GIS technology to determine the average
migration rates of lobes (Levin and Ben-Dor, 2004; Levin, 2011). Meanwhile, as a migrating
rate measured is an average over a certain time period and usually also an average of a
number of dunes in an area, measuring frequency and duration (in addition to the spatial
scope of study sites) are also key factors for determining the migration rate.

Because of climatic instability, a dune migration rate should be evaluated on a
sensible spatiotemporal scale (Lockwood, 2001). A migration rate based on a short period of
field experiments (usually a few events or years) can hardly be scaled up to provide an
adequate understanding of long-term dune behaviour (on a temporal scale of decades or
centuries). Likewise, chronological dating elucitates long-term historical trajectories (e.g.,
centuries) of dunefield development, but is insufficient to assist in a detailed understanding of
shorter-term variations (e.g., seasons or decades) (Aagaard et al., 2004; Sherman, 1995).
Changes in migration rates of parabolic dunes may be caused by external forces such
as environmental controls and human activities (Forman et al., 2008), but also by internal or
autocyclic adjustments of a geomorphological system (Brunsden, 2001). Some large parabolic dunes may migrate faster than smaller ones because of ample sand supply for wind entrainment and less vegetation impeding saltation (Marín et al., 2005). Relatively small surface roughness creates less turbulence, thereby enhancing the sand transport efficiency and dune migration rate. When a dune moves onto a thicker sandy substratum, the dune migrates slower because the substratum provides more abundant sand supply to the dune. Similarly, in a sand-starving environment, as the lobe of a parabolic dune shrinks over time the dune migrates at an increasing rate. Collectively, a dune migration rate is a poor indicator of mobility of a larger dune system.

In order to estimate impacts of physical and anthropogenic variables on the development of aeolian dunes and to anticipate potential changes in dunefield mobility in the context of environmental fluctuations and climatic change, it is necessary to choose an appropriate time scale. Because dune dynamics involve time-lags and hysteresis effects between climate and dune mobility, an appropriate time scale should ensure that geomorphological components of an aeolian system have had sufficient response time to adjust themselves to external conditions such as temperature and precipitation (Hugenholtz and Wolfe, 2005; Knight et al., 2004; Levin, 2011; Yizhaq et al., 2009). The response time of dunes, however, differs depending on the characteristics of different geomorphological components (e.g., the flora) (Overpeck et al., 1992), as well as on the spatial scale (e.g., local, regional or global) of climatic change to which the dunes are responding (Huggett, 1991). Moreover, a sensible spatial scale is needed to differentiate the spatial variability of individual dune mobility from entire dunefield mobility, a variability that arises from the specific history of single dunes, for instance related to localised anthropogenic impacts.
3. **Distribution of parabolic dunes**

A global distribution of parabolic dunefields was collected from a comprehensive literature review of approximately 250 publications, and all sites mentioned in literature sources were examined from Google Earth imagery. Presently discernible parabolic dunefields were compiled and mapped as shown in Figure 2. Some parabolic dunefields reported in literature have been reshaped by human activities (e.g., agriculture, recreation and urbanisation), and in some regions have been largely destroyed (e.g., in Portugal, Hungary, Poland and Brazil). In China, in particular, few coastal parabolic dunefields have survived the large-scale urbanisation and industrialisation. Some other dunefields could not be identified because of insufficient image resolution and/or because they had become covered with dense vegetation.

Research on parabolic dunes started in the first half of the 20th century in the United States (Cooper, 1958; Hack, 1941; Melton, 1940), Australia (Jennings, 1957), New Zealand (Brothers, 1954) and Europe (Lefevre, 1931; Landsberg, 1956; Paul, 1944). Early research was mainly limited to qualitative description of dune morphology and local distribution, an attempt at morphology-based classification, and associated conjectures regarding dune origins and formative processes. The crucial role of wind regime on the development and morphology of parabolic dunes has been well-recognised (Bagnold, 1941; Fryberger, 1979; Jennings, 1957), but only a number of studies quantitatively measured dune morphology and migration rates (Brothers, 1954; Cooper, 1958; Ranwell, 1958).

The importance of parabolic dunes did not gain much attention until the 1980s, when research gradually expanded to a number of different regions including India (Wasson et al., 1983), Australia (Pye, 1982; Story, 1982), Fiji (Kirkpatrick and Hassall, 1981), South Africa (Eriksson et al., 1989) and Saudi Arabia (Anton and Vincent, 1986). During this period, research focused on field measurements and understandings of physical processes (flow...
dynamics and sand transport) and controlling factors (wind regime, sediment supply and vegetation cover). Different dune transformations have been noted in various regions, yet detailed investigation has been absent. Increasingly wide use of advanced technology such as aerial photographs, nevertheless, expanded research on a much larger spatial and longer temporal scale, which facilitated the systematic exploration and comparison of parabolic dunes in various environments and also facilitated the development of schematic models regarding dune formation and classification (David et al., 1999; Pye, 1982; Wolfe and David, 1997).

From 1995 onwards, more research on parabolic dunes has been conducted across different climatic regions both on the coast and inland. In particular, the use of computer modelling and simulation, expanded from the Werner Model (Werner, 1995), has enabled exploration of fundamental principles underlying the dynamics of dune patterns and testing of possible assumptions based on real-world observations and investigations. Parabolic dunes with trailing arms developing from blowouts have been successfully simulated by the Discrete Eco-geomorphic Aeolian Landscape (DECAL) model (Baas, 2002; Nield and Baas, 2008), whilst a dune transformation from barchan to parabolic form has been simulated with a continuous model (Duran et al., 2008). GIS technology, advances in remote sensing (e.g., LiDAR), and advances in luminescence dating techniques have further accelerated the capability and scope of investigations (Anthonsen et al., 1996; Levin and Ben-Dor, 2004; Tsoar and Blumberg, 2002; Swezey et al., 2013; Wolfe and Hugenholtz, 2009).

Presently, with progressive concerns about the potential impacts of climatic change on aeolian dune environments and associated impacts of human behaviour, parabolic dunes are receiving increased attention because of their sensitivity to changes in environmental conditions. Research on potential activation of currently stabilised parabolic dunefields imparted by climatic variations has been conducted in a few regions such as the Canadian
Prairies (Hugenholtz and Wolfe, 2005; Wolfe, 1997) and Israel (Tsoar, 2005), and will likely continue to be an important topic.

As indicated in Figure 2 and Table 2, parabolic dunes are widely spread across a large range of climatic gradients from hot equatorial savannah (Fernandez et al., 2009; Hesp et al., 2010; Hesp, 2008; Porat and Botha, 2008; Pye, 1982; Shulmeister and Lees, 1992) to warm climates (Anthonsen et al., 1996; Arens et al., 2004; Clemmensen et al., 2007; Hart et al., 2012; Morkunaite et al., 2011; Tsoar and Blumberg, 2002; Zular et al., 2013) to cold climates (Bélanger and Filion, 1991; Bhiry et al., 2011; Eyles and Meulendyk, 2012; McKee, 1966), and from humid climates (Bailey and Bristow, 2004; Bigarella et al., 2006; Levin, 2011; Ranwell, 1958; Wakes et al., 2010) to arid climates (Hack, 1941; Hugenholtz et al., 2010; Hugenholtz et al., 2008; Reitz et al., 2010; Wolfe and Lemmen, 1999; Yan et al., 2010) to hyper-arid desert climates (Anton and Vincent, 1986; Carter et al., 1990; Eriksson et al., 1989; Kar et al., 1998; Nichol and Brooke, 2011). In contrast to highly mobile barchan or transverse dunes, which may be distributed extensively across a large region, parabolic dunes are usually restricted to relatively small areas, mostly arranged in belts in coastal settings, or along inland river valleys and lake shores.

Parabolic dunes in coastal settings are strongly influenced by the geometrical alignment of the coastline relative to onshore winds (Jennings, 1957). Unidirectional onshore winds are preferable for the development of parabolic dunes, and such dunes are often associated with the initiation of blowouts on previously vegetated foredunes. Blowouts develop when vegetation cover is breached by either natural process such as increased wind erosion during periods of drought or storminess or human activity such as excessive grazing (Hesp, 2002). Sand exposed in a blowout is transported and deposited in the leeward margin, developing a
bare lobe. A parabolic dune forms as the bare lobe migrates inland (cf. Section 4.3). These coastal parabolic dunes are widely seen in humid, sub-humid, and semi-arid regions, usually in conjunction with coastal foredunes in such areas as the west coast of Manawatu, New Zealand (Hesp, 2001), the Oregon coasts of the United States (Cooper, 1958), and the west and northeast coasts of Australia (Carter et al., 1990; Nichol and Brooke, 2011; Pye, 1982; Shepherd and Eliot, 1995).

Elongated parabolic dunes occur in coastal settings where relatively abundant sand supply continuously supplements sand loss from mobile lobes, without which dunes would otherwise be stabilised by vegetation. Elongated parabolic dunes are usually present in equatorial or warm climates with an ample annual precipitation yet interspersed with periodic dry seasons. These areas are generally dissipative systems and well-covered by dense forests or scrubs. Seasonal dry periods accompanied by strong onshore winds expose abundant sediments previously inundated by inter-dune lakes, which enables dune lobes to maintain mobility whereas vegetation on arms remains intact, forming long-walled arms. An alternation between wet and dry periods and strong onshore winds occurring in dry seasons are crucial in the elongation of parabolic dunes such as those on the east coasts of Australia, South America and Africa (Barbosa and Dominguez, 2004; Porat and Botha, 2008; Pye, 1982).

Digitate and hemicyclic parabolic dunes may develop on coasts that are exposed to a wind regime with multidirectional onshore winds. Presence of a woodland cover is of particular importance in forcing bare lobes moving inland in divergent directions, as can be seen on the northern coast of Brazil (Buynievich et al., 2010), and the east coasts of Fraser Island (Levin, 2011) and Queensland (Pye, 1982) in Australia.
Other important controls on the development of coastal parabolic dunes include wave power, beach morphology, storm surge, and sea level change. High wave energy dissipative and intermediate beaches have a wide and flat/gently sloping backshore. Without considerable flow disturbance, onshore winds can maintain high velocities and have a great potential for continuous landward sand transport, thereby providing abundant sediment supply for dune development (Short and Hesp, 1982). Strong storms may cause powerful wave action that removes vegetation and scars foredunes, and initiate blowouts (Hesp, 2002). The frequency and magnitude of storms, therefore, contributes significantly to shoreline destruction and coastal dune development. Sea level rise induces the near-shore profile to keep adjusting itself to a new level, which is likely to increase dynamics in sediment exchange and potentially provide greater sediment supply for aeolian sand transport (Carter, 1991; Hesp and Thom, 1990; Psuty and Silveira, 2010).

Coastal parabolic dunes are also usually found adjacent to river mouths or estuaries, where sediment from rivers provides an abundant sand supply for wind transport, as can be seen in the areas at the mouth of the Sigatoka River in Fiji (Kirkpatrick and Hassall, 1981), on the south coast of Wilderness Dune Cordons in South Africa (Illenberger and Rust, 1988), on the Oregon coast in the United States (Cooper, 1958), and on the São Francisco River Strand Plain and northeast coast of Brazil (Barbosa and Domínguez, 2004; Duran et al., 2008).

In contrast to the relatively extensive research on coastal parabolic dunes - across eighteen countries - inland parabolic dunes have been investigated in only five countries. Inland parabolic dunes are usually found in arid and semi-arid regions adjacent to river valleys and along lake margins. Their formation and development is strongly governed by regional controls: the orographic conditions and the distance to rivers or lakes.
Many of the inland parabolic dunes in western North America are derived from river sediments. In this setting, mountain ridges alter regional climate regime and the local biogeomorphic interactions, as is the case for the widespread dunefields on the Great Plains along the eastern side of the Rocky Mountains (Forman et al., 1992; Halfen et al., 2010; Holliday, 2001; Hugenholtz, 2010; Madole, 1995; Muhs et al., 1996). The Rocky Mountains block moisture from the Pacific Ocean in the west, casting areas in the east in rain shadow (Hugenholtz et al., 2010). Most parabolic dunes in this region are stabilised under prevailing climate conditions, and LiDAR images reveal that they have been transformed from barchans under recent climate warming (Wolfe and Hugenholtz, 2009). The stabilised parabolic dunes derived from fluvial sediments are also found in the south-eastern United States such as in the Savannah River valley in Jasper County of South Carolina (Swezey et al., 2013) and on the Coastal Plain of Georgia (Ivester and Leigh, 2003).

Inland parabolic dunes can also be derived from lake sediments. The White Sands, for example, consist of gypsum sediments that precipitated as a saline lake evaporated (Kocurek et al., 2007; Langford, 2003; Scheidt et al., 2010). A small scale of parabolic dunes on the eastern shore of Lake Michigan is associated with the development of blowouts (Arbogast et al., 2002; Hansen et al., 2009; Hansen et al., 2010). On the eastern shore of Hudson Bay imbricate parabolic dunes have developed under multidirectional winds (Filion and Morisset, 1983).

In contrast to coastal parabolic dunes, which usually develop either from expansion and activation of blowouts or from stabilisation of transgressive dunefields (Hesp and Walker, 2013), inland parabolic dunes usually develop from barchan or transverse dunes. When barchans move into an environment with more abundant vegetation (e.g., closer to a river or a higher water table), their horns are invaded and anchored by grasses and shrubs first. The remaining bare lobes then move forward, leaving behind the stabilised horns (cf. Section 4.2).
Ample sand availability from mobile dunefields upwind may enable such parabolic dunes to maintain a high mobility. Examples of such parabolic dunes are widely distributed on the eastern margins of White Sands in New Mexico (McKee, 1966; Reitz et al., 2010), in the west of Fremont County and on the eastern Snake River Plain in Idaho (Chadwick and Dalke, 1965; Forman et al., 2003), and in the east of the Horqin Desert in Inner Mongolia (Yan, 2010).

Arms of inland parabolic dunes usually have relatively low relief compared with the arms of coastal parabolic dunes because grasses and shrubs rather than trees dominate these inland regions. As dune arms are frequently overridden or cut through by following dunes, elongated parabolic dunes are not commonly seen inland (Halfen et al., 2010; Marín et al., 2005), with the exception of those in the Thar Desert of India and Pakistan (Wasson et al., 1983). Although trees hardly survive in an arid desert environment, patches of stunted trees have been shown to initiate the development of small parabolic dunes in the Kalahari Desert, South Africa (Eriksson et al., 1989).
Figure 2. Global distribution of parabolic dunes. Köppen-Geiger climate zone is adapted from Kottke et al., 2006.

Table 2. Global distribution of parabolic dunes

<table>
<thead>
<tr>
<th>Region</th>
<th>Representative Literature</th>
<th>Köppen-Geiger Climate Zone &amp; Main Climate</th>
<th>Mean Annual Precipitation (mm)</th>
<th>Mean Annual Temperature (˚C)</th>
<th>Location</th>
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</thead>
<tbody>
<tr>
<td>Valdes Peninsula, Northeastern Patagonia</td>
<td>del Valle et al., 2010</td>
<td>BSk: arid steppe climate</td>
<td>steppe 231</td>
<td>cold arid 13</td>
<td>coastal</td>
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<tr>
<td>Cape Bedford and Cape Flattery, Queensland</td>
<td>Pye, 1982; 1984</td>
<td>Aw: equatorial savannah with dry winter</td>
<td>winter dry 1784</td>
<td>hot 27</td>
<td>coastal</td>
</tr>
<tr>
<td>Cervantes-Dongara coast</td>
<td>Shepherd and Eliot, 1995</td>
<td>Csa: warm temperate climate with dry summer</td>
<td>summer dry 534</td>
<td>hot summer 20</td>
<td>coastal</td>
</tr>
<tr>
<td>Eyre Peninsula</td>
<td>Dutkiewicz and Prescott, 1997</td>
<td>Csb: warm temperate climate with dry summer</td>
<td>summer dry 383</td>
<td>warm summer 18</td>
<td>coastal</td>
</tr>
<tr>
<td>Fraser Island</td>
<td>Ward, 2006; Levin, 2011</td>
<td>Cfa: warm temperate climate</td>
<td>fully humid 1200</td>
<td>hot summer 22</td>
<td>coastal</td>
</tr>
<tr>
<td>Groote Eylandt</td>
<td>Shulmeister and Lees, 1992</td>
<td>Cfs: warm temperate climate with dry winter</td>
<td>winter dry 1350</td>
<td>hot 26</td>
<td>coastal</td>
</tr>
<tr>
<td>King Island, Tasmania</td>
<td>Jennings, 1957</td>
<td>Csb: warm temperate climate with dry summer</td>
<td>summer dry 811</td>
<td>warm summer 13</td>
<td>coastal</td>
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<tr>
<td>north of Carnarvon</td>
<td>Carter et al., 1990</td>
<td>BWh: desert climate</td>
<td>desert 241</td>
<td>hot arid 31</td>
<td>coastal</td>
</tr>
<tr>
<td>Northern Cape York Peninsula, Queensland</td>
<td>Pye, 1983b; 1984</td>
<td>Aw: equatorial savannah with dry winter</td>
<td>winter dry 1745</td>
<td>hot 26</td>
<td>coastal</td>
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<tr>
<td>Northern Territory coast</td>
<td>Story, 1982</td>
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<td>winter dry 1200</td>
<td>hot 27</td>
<td>coastal</td>
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<td>Point Cloates, Carnarvon shelf</td>
<td>Nichol and Brooke, 2011</td>
<td>BWh: desert climate</td>
<td>desert 226</td>
<td>hot arid 22</td>
<td>coastal</td>
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<td>Location</td>
<td>Reference(s)</td>
<td>Climate Type</td>
<td>Characteristics</td>
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<td>Ramsay Bay, Hinchinbrook Island, Queensland</td>
<td>Pye, 1983a; 1984; Pye and Mazzullo, 1994</td>
<td>Am: equatorial monsoon</td>
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<td>Ramsay Bay, Hinchinbrook Island, Queensland</td>
<td>Pye, 1983a; 1984; Pye and Mazzullo, 1994</td>
<td>Csb: warm temperate Mediterranean climate with dry summer</td>
<td>summer dry 400</td>
<td>warm summer 16</td>
<td>coastal</td>
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<td>River Murray mouth region</td>
<td>Murray-Wallace et al., 2010</td>
<td>Csb: warm temperate Mediterranean climate with dry summer</td>
<td>summer dry 400</td>
<td>warm summer 16</td>
<td>coastal</td>
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<td>Brazil</td>
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<td>Atalaia Beach, Pará State</td>
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<td>Am: equatorial monsoon</td>
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<td>Fortaleza, Ceará</td>
<td>Duran et al., 2008</td>
<td>As: equatorial savannah with dry summer</td>
<td>summer dry 1642</td>
<td>hot 27</td>
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<td>Lagow dune field, Santa Catarina Island</td>
<td>Bigarella et al., 2005; 2006</td>
<td>Cfa: warm temperate subtropical climate</td>
<td>fully humid 1521</td>
<td>hot summer 21</td>
<td>coastal</td>
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<tr>
<td>Rio de Janeiro coast</td>
<td>Fernandez et al., 2009</td>
<td>Aw: equatorial savannah with dry winter</td>
<td>winter dry 771</td>
<td>hot 24</td>
<td>coastal</td>
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<tr>
<td>São Francisco do Sul coastal barrier</td>
<td>Zalar et al., 2013</td>
<td>Cfa: warm temperate subtropical climate</td>
<td>fully humid 1250</td>
<td>hot summer 18</td>
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<tr>
<td>São Francisco River Strand Plain</td>
<td>Barbosa and Dominguez, 2004</td>
<td>As: equatorial savannah with dry summer</td>
<td>summer dry 1700</td>
<td>hot 24</td>
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<tr>
<td>Canada</td>
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<tr>
<td>Bigstick Sand Hills, Saskatchewan</td>
<td>Hugenholtz et al., 2007; Hugenholtz, 2010</td>
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<td>steppe 380</td>
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<td>eastern coast of Hudson Bay, Northern Québec</td>
<td>Filon and Morisset, 1983; Belanger and Filion, 1991; Bhury et al., 2011</td>
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<td>Îles de la Madeleine, Quebec</td>
<td>Giles and McCann, 1997</td>
<td>Dfb: snow climate</td>
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<td>Lake Huron coast, Ontario</td>
<td>Byrne, 1997; Eyles and Meulendyk, 2012</td>
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<td>northern Great Plains, Saskatchewan</td>
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<td>China</td>
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<td>Ebinur Lake district, Xinjiang</td>
<td>Jia et al., 2012</td>
<td>BWk: arid desert climate</td>
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<td>Hobq Desert, Ordos, Inner Mongolia</td>
<td>Yan, 2010; Zhang et al., 2011</td>
<td>BSk: arid steppe climate</td>
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<td>Horqin Desert, Inner Mongolia</td>
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<td>Hulunbuir Grassland, Inner Mongolia</td>
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<td>Anholt, Kattegat</td>
<td>Clemmensen et al., 2007</td>
<td>Cfb: warm temperate climate</td>
<td>fully humid 478</td>
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<tr>
<td>Lodbjerg, northwest coast of Jutland</td>
<td>Clemmensen et al., 2001</td>
<td>Cfb: warm temperate climate</td>
<td>fully humid 359</td>
<td>warm summer 9</td>
<td>coastal</td>
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<td>Råbjerg Mile, Skagen Odde</td>
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<td>fully humid 706</td>
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<tr>
<td>Vejers, west coast of Jutland</td>
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<td>fully humid 355</td>
<td>warm summer 9</td>
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</tr>
</tbody>
</table>

Fiji

| Sigatoka sand dunes, Viti Levu | Kirkpatrick and Hassall, 1981 | Af: equatorial rainforest | fully humid 1862 | hot 30 | coastal |

France

| northern shore | Meurisse et al., 2005 | Cfb: warm temperate climate | fully humid 592 | warm summer 11 | coastal |
| South-western coast | Bertran et al. 2011 | Cfb: warm temperate climate | fully humid 823 | warm summer 11 | coastal |

India & Pakistan

<p>| Thar Desert | Wasson et al., 1983; Goossens et al., 1993; Kar | BWh: tropical desert climate | desert 172 | hot arid 27 | inland, river bank |</p>
<table>
<thead>
<tr>
<th>Country</th>
<th>Location and Description</th>
<th>Climate Type</th>
<th>Temperature and Humidity</th>
<th>Location Type</th>
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<td>southeastern Mediterranean Coast</td>
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<td>Csa: warm temperate Mediterranean climate with dry summer</td>
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<td>Adjabiya coast</td>
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<td><strong>Lithuania</strong></td>
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<td>Curonian Spit, southeastern Baltic Sea Coast</td>
<td>Morkunaite et al., 2011</td>
<td>Cfb: warm temperate climate, intermediate between marine and continental</td>
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<td><strong>Mexico</strong></td>
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<td>El Farallon-La Mancha Dunefield</td>
<td>Hesp et al., 2010</td>
<td>Aw: equatorial savannah with dry winter</td>
<td>winter dry 1200; hot 24</td>
<td>coastal</td>
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<td><strong>Netherlands</strong></td>
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<td>fully humid 847; warm summer 10</td>
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<td>Mason Bay, Stewart Island</td>
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<td>Anton and Vincent, 1986</td>
<td>BWh: desert climate, Indian Ocean Monsoonal</td>
<td>desert 88; hot arid 27</td>
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<td><strong>South Africa</strong></td>
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<td>Maputaland coastal plain</td>
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<td>southern Kalahari Desert</td>
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<td>Wilderness Dune Cordons</td>
<td>Hellström, 1996; Illeberger, 1996</td>
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<td>Doñana National Park</td>
<td>Siljeström and Clemente, 1990</td>
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<td>coastal</td>
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<tr>
<td>Liencres dune system, Cantabria</td>
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<td>Cfb: warm temperate climate</td>
<td>fully humid 1150; warm summer 14</td>
<td>coastal</td>
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<td>Servera et al., 2009</td>
<td>Csa: warm temperate Mediterranean climate with dry summer</td>
<td>summer dry 427; hot summer 18</td>
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<td>Cfb: warm temperate maritime climate</td>
<td>fully humid 1434; warm summer 11</td>
<td>coastal</td>
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<td>Sands of Forvie, Scotland</td>
<td>Robertson-Rintoul, 1990; Ritchie, 2000</td>
<td>Cfb: warm temperate maritime climate</td>
<td>fully humid 750; warm summer 9</td>
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<td><strong>United States</strong></td>
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<td>Cape Cod National Seashore, Massachusetts</td>
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<td>Cfa: warm temperate climate</td>
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<td>Casper Dunefield, Casper, Wyoming</td>
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<td>Eastern Colorado</td>
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<td>eastern Upper Michigan</td>
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<td>fully humid 3500, warm summer 6</td>
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<td>Great Bend Sand Prairies, Kansas</td>
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<td>Cfa: warm temperate continental climate, semi-arid to sub-humid</td>
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<td>Great Sand Dunes National Park and Preserve, Colorado</td>
<td>Marín et al., 2005; Forman et al., 2006</td>
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<td>Hanford, Washington</td>
<td>Stetler and Gaylord, 1996</td>
<td>BSk: arid steppe climate</td>
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<tr>
<td>High Plains of Colorado</td>
<td>Forman et al., 1992; Muhs et al., 1996</td>
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<td>Holland, eastern shore of Lake Michigan</td>
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<td>Dfb: snow continental climate</td>
<td>fully humid 2738, warm summer 9</td>
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<td>Lanphere Dunes, Northern California</td>
<td>Craig, 2000</td>
<td>Csb: warm temperate climate with dry summer</td>
<td>summer dry 969, warm summer 12</td>
<td>coastal</td>
</tr>
<tr>
<td>Navajo County, Arizona</td>
<td>Hack, 1941</td>
<td>BSk: steppe climate, true desert to humid mountain climate</td>
<td>steppe 210, cold arid 12</td>
<td>inland, river bank</td>
</tr>
<tr>
<td>north-western Bahamas</td>
<td>Kindler and Strasser, 2000</td>
<td>Aw: equatorial savannah with dry winter</td>
<td>winter dry 1120, Hot 24</td>
<td>coastal</td>
</tr>
<tr>
<td>Oregon coast</td>
<td>Cooper, 1958</td>
<td>Csb: warm temperate climate with dry summer</td>
<td>summer dry 1794, warm summer 12</td>
<td>coastal</td>
</tr>
<tr>
<td>Petoskey State Park, Michigan</td>
<td>Lepczyk and Arborgast, 2005</td>
<td>Dfb: snow continental climate</td>
<td>fully humid 813, warm summer 7</td>
<td>inland, lake shore</td>
</tr>
<tr>
<td>Savannah River valley, South Carolina</td>
<td>Swezey et al., 2013</td>
<td>Cfa: warm temperate climate</td>
<td>fully humid 1298, hot summer 18</td>
<td>inland, river bank</td>
</tr>
<tr>
<td>St. Anthony, Idaho</td>
<td>Chadwick and Dalke, 1965</td>
<td>BSk: semi-arid steppe climate</td>
<td>steppe 340, cold arid 13</td>
<td>inland, river bank</td>
</tr>
<tr>
<td>south Texas</td>
<td>Forman et al., 2009</td>
<td>Cfa: warm temperate subtropical climate</td>
<td>fully humid 806, hot summer 22</td>
<td>coastal</td>
</tr>
<tr>
<td>Southern High Plains of Texas and New Mexico</td>
<td>Holliday, 2001</td>
<td>BSk: semi-arid steppe climate</td>
<td>steppe 409, cold arid 8</td>
<td>inland, river bank</td>
</tr>
<tr>
<td>Walking Dunefield, Napeague, New York</td>
<td>Girardi and Davis, 2010</td>
<td>Cfa: warm temperate climate</td>
<td>fully humid 1217, hot summer 11</td>
<td>coastal</td>
</tr>
<tr>
<td>White Sands Dunefield, New Mexico</td>
<td>McKee, 1966; Reitz et al., 2010</td>
<td>BSk: arid steppe climate</td>
<td>steppe 264, cold arid 16</td>
<td>inland, river bank</td>
</tr>
<tr>
<td>Wilderness State Park, northern lower Michigan</td>
<td>Lichter, 1995</td>
<td>Dfb: snow continental climate</td>
<td>fully humid 767, warm summer 5</td>
<td>inland, lake shore</td>
</tr>
</tbody>
</table>

### 3.1 Coastal parabolic dunes

Coastal parabolic dunes are extensively developed in Australia. Hairpin-, hemicyclic-, and digitate-shaped parabolic dunes are found on the northeast coast (Figure 3a), and some of these dunes overlap with or are nested within others, developing a compound dune form (Levin, 2011; Pye, 1982; Pye, 1983a; Pye, 1983c; Pye, 1984; Pye and Mazzullo, 1994; Shulmeister and Lees, 1992; Ward, 2006). These parabolic dunes are generally stabilised by...
vegetation except some in the Cape Bedford – Cape Flattery dunefields and on the east coast of Northern Cape York Peninsula. This region is controlled by an equatorial savannah climate or an equatorial monsoon climate with a strong seasonal variation of humidity in the north, to a fully humid climate down to the south. In comparison to the very elongated parabolic dunes on the east coast of Australia, parabolic dunes on the west coast (Figure 3b) are more mobile and somewhat less elongated (Carter et al., 1990; Nichol and Brooke, 2011; Shepherd and Eliot, 1995). The climate there is hot and arid, with a desert climate in the northwest, to a seasonal humid climate (dry summers) in the southeast. In the Carnarvon dunefield, for example, the mean annual precipitation is only ~200 mm. Parabolic dunes are also present on King Island (Jennings, 1957) and on the south coast of South Australia (Dutkiewicz and Prescott, 1997; Murray-Wallace et al., 2010), but on a relatively smaller scale, governed by a temperate climate with dry summers (Figure 3c).

Figure 3. Coastal parabolic dunes in Australia, at the same scale.

Hairpin-shaped parabolic dunes are present along the west coast of Auckland in New Zealand (Brothers, 1954), along the west coast of Maputaland Plain and south coast of
Wilderness Dune Cordons in South Africa (Hellström, 1996; Illenberger, 1996; Porat and Botha, 2008), and at the mouth of the Sigatoka River in Fiji (Kirkpatrick and Hassall, 1981).

In these coastal regions, most of the parabolic dunes are fully vegetated with minor aeolian sediment transport. In some of these locations, the movement of parabolic dunes is impeded by a forest canopy, and the dunes have developed digitate-shaped lobes.

On the coasts of Mason Bay and Manawatu Plain in New Zealand, however, parabolic dunes are highly mobile and they move inland continuously (Figure 4), even though the climate is fully humid with a mean annual precipitation of ~900 mm (Clement et al., 2010; Hesp, 2001; Hart et al., 2012; Wakes et al., 2010).

In Brazil, active parabolic dunefields are primarily concentrated on the equatorial coast (in an equatorial savannah climate with dry summers) in particular on the São Francisco River Strand Plain (Barbosa and Dominguez, 2004) and Fortaleza coast (Duran et al., 2008).

In these areas, the parabolic dunes display a long-walled transgressive ridge/lobe with relatively thin and short trailing arms (Figure 5). Parabolic dunes at the São Francisco do Sul coastal barrier, however, are influenced by a humid subtropical climate and are fully...
stabilised by a forest canopy only with minor development of blowouts (Zular et al., 2013). Due to the impacts of human activity, few active parabolic dunes survive on the Atalaia Beach in Pará State (Buynevich et al., 2010; Miot da Silva and Hesp, 2010) and the coast of Rio de Janeiro (Fernandez et al., 2009).

Figure 5. Coastal parabolic dunes on the São Francisco River Strand Plain in Brazil

Lunate parabolic dunes are present in Oregon on the west coast of the United States (Cooper, 1958), and on the Mediterranean coast of Israel (Ardon et al., 2009; Tsoar and Blumberg, 2002). The Israeli coastal dunefield described in this literature, however, may also be interpreted as vegetated transverse dunes, and only a few have developed into somewhat parabolic shape, although this dunefield forms the basis for the parabolic dune stabilisation mechanism proposed by Tsoar and Blumberg (2002), see section 4.2. These parabolic dunes have lunate-shaped lobes yet lack well-defined trailing arms (Figure 6). A temperate climate with dry summers dominates these regions, and aeolian activity alternates with periods of stabilisation by vegetation cover. Small active parabolic dunefields also occur on the east coast of the United States, such as Cape Cod National Seashore in Massachusetts (Forman et
al., 2008; Winker, 1992) and Walking Dunefield in New York (Girardi and Davis, 2010), both of which are in a temperate climate with an annual precipitation greater than 1000 mm. Presently partial-submerged parabolic dunes are also found in the north-western Bahamas (Kindler and Strasser, 2000).

In Canada, coastal parabolic dunes are characterised by niveo-aeolian deposits that formed in a continental setting where winds transported snow and sand coincidently during the winter (Bélanger and Filion, 1991; Bhiry et al., 2011; Filion and Morisset, 1983; Giles and McCann, 1997). The cold and dry climatic conditions during winter facilitate niveo-aeolian activity compared with cool humid conditions during summer.

In Europe, coastal parabolic dunes are usually stabilised by vegetation and are relic features that formed under different climatic conditions, and the dunes are hence not easily identified because of their complex forms, full coverage by vegetation, and/or destruction by human activities. Stabilised parabolic dunes can be identified in northern Denmark.
(Clemmensen et al., 2001), the Netherlands (Jungerius and Riksen, 2010), and Scotland in the United Kingdom (Ritchie, 2000; Robertson-Rintoul, 1990), whereas active parabolic dunes can be discernible on the coasts of France (Meurisse et al., 2005), Doñana National Park and Cantabrian Coast in Spain (Arteaga et al., 2008; Siljeström and Clemente, 1990), Curonian Spit in Lithuania (Morkunaite et al., 2011), Wales in the United Kingdom (Bailey and Bristow, 2004), and Vejers in Denmark (Clemmensen et al., 1996).

A large coastal parabolic dunefield is present along the edge of the Jafurah Desert in Saudi Arabia (Figure 7), and this dunefield extends from the inland area to the coast (Anton and Vincent, 1986). Although the Jafurah Desert has a mean annual precipitation less than 100 mm, the ubiquitous presence of shallow groundwater enables the survival of sparse desert scrubs, which lead to local erosion that forms parabolic dunes.

![Figure 7. Coastal parabolic dunes in the Jafurah Desert, Saudi Arabia](image)

### 3.2 Inland parabolic dunes

Inland parabolic dunes are located in many areas of western North America, both in the United States and Canada. On the Great Plains of western North America, large-scale aeolian sediment mobilisation is attributed to orographic impacts by the Rocky Mountains (Odynsky,
1958; Smith, 1952). Parabolic dunes are scattered widely across the Canadian Prairies, where the climate is arid continental steppe with a mean annual precipitation of ~400 mm (Hugenholtz et al., 2010; Wolfe and Hugenholtz, 2009). Most parabolic dunes here are fully stabilised by vegetation (Figure 8), although a highly active parabolic dune with development of blowouts has been studied in detail in the Bigstick Sand Hills of Saskatchewan (Hugenholtz, 2010; Hugenholtz et al., 2008; Hugenholtz et al., 2009).

Figure 8. Inland parabolic dunes in the Canadian Prairies

Compared with those on the Canadian Prairies, parabolic dunes on the United States Prairies exhibit greater mobility. This region is governed by a semi-arid steppe climate but with varying precipitation depending on altitude and orography. Hairpin-shaped parabolic dunes occur in the Casper Dune Field in Wyoming (Figure 9) (Gaylord, 1982; Halfen et al., 2010), in the Great Sand Dunes National Park and Preserve in Colorado (Forman et al., 2006; Marín et al., 2005), and in various parts of eastern Colorado (Madole, 1995). These parabolic dunes are fully- or semi-stabilised by vegetation with trailing arms anchored by grasses and shrubs. Small active compound parabolic dunes are widespread on the High Plains of north-
eastern Colorado (Figure 10), Texas, and New Mexico (Forman et al., 1992; Holliday, 2001; Muhs et al., 1996), and on the Great Bend Sand Prairies in Kansas (Arbogast, 1996).

In the United States west of the Great Plains, parabolic dunes are found on the eastern margin of White Sands in New Mexico (Figure 11) (McKee, 1966; Reitz et al., 2010), in the Navajo County in Arizona (Hack, 1941), and in the Hanford dunefield in eastern Washington (Stetler and Gaylord, 1996). These parabolic dunes, on average, have a greater mobility with evident bare lobes, compared with those on the High Plains (Figure 12).
Inland parabolic dunes are present along some of the shores of the Great Lakes of North America. For example, parabolic dunes are well-developed along the east shore of Lake Michigan and along the south shore of Lake Huron (Arbogast et al., 2002; Byrne, 1997; Eyles and Meulendy克, 2012; Hansen et al., 2009; Hansen et al., 2010; Lepczyk and Arbogast, 2005; Lichter, 1995; Timmons et al., 2007). Under the influence of a fully humid continental climate, most of these parabolic dunes are fully vegetated. Only a few that arise from blowouts are active at present.

In the Thar Desert of India and Pakistan, inland parabolic dunes link and override each other, presenting a clustered ‘rake-like’ appearance (Wasson et al., 1983). The irregular
and asymmetric noses of parabolic dunes have developed during the process of dune
overriding, in which one side of a parabolic dune is cut off when another parabolic dune
moving across the nose. This process also causes the dune clusters to be irregular. The
clustered parabolic dunes near Shergarh in Pakistan possess rounded noses and exhibit U-
shaped dune morphology (Figure 13), whereas those between Barmer and Jaisalmer in India
display V-shaped dune morphology and have exceedingly elongated arms (Figure 14).

Figure 13. U-shaped parabolic dunes near Shergarh in the Thar Desert. These parabolic dunes are clustered forming individual dune groups.

Figure 14. V-shaped parabolic dunes between Barmer and Jaisalmer in the Thar Desert. These parabolic dunes are imbricated to various
degrees, and have very elongated arms.
Inland parabolic dunes in the Kalahari Desert in South Africa (Figure 15) have developed in patches under a hot and arid climate. Eriksson et al. (1989) suggested that the formation of these parabolic dunes is associated with the presence of stunted trees. Winds may continuously erode the base of the stunted trees, eventually forming blowouts and small parabolic dunes behind these blowouts.

There are a variety of parabolic dunes in northern China (Yan, 2010; Yan et al., 2010). Parabolic dunes developed from blowouts are concentrated on the Hulunbuir grasslands (Zhuang and Hasi, 2005), where extensive human activity, in particular grazing and off-road driving, breaches the vegetated surface and exposes sand underneath, which initiates aeolian sand transport. The spatial arrangement of parabolic dunes in this area is hence highly controlled by local socio-economic activities, especially the related transportation networks and grazing behaviour. The Horqin Desert is located on the southeast of mountains governed by an arid steppe climate, and exhibits a transition from transverse dunes in the west to parabolic dunes in the east in similarity with White Sands in New Mexico. Highly active parabolic dunes occur extensively between the Xilamulun River and the Laoha River along
the north bank of the Laoha River (Figure 16). In contrast, parabolic dunes in the Hobq
Desert are fully- or partially-stabilised by shrubs (Figure 17), most of which are along the
east bank of the Xuhaitu River (Yan, 2010; Zhang et al., 2011). This region is characterised
by a strong seasonality of both precipitation and wind regime, resulting in dunes migrating
periodically. Parabolic dunes are also found in the Ebinur Lake District (Jia et al., 2012) and
the Take Ermu Ku’er Desert (Zeng, 2008) of Xinjiang.

Figure 16. Inland parabolic dunes in the Horqin Desert, north-eastern China

Figure 17. Inland parabolic dunes in the Hobq Desert, northern China
4. **Parabolic dune related transformations**

Parabolic dunes play a significant role in dune transformations. Parabolic dunes, can not only develop from highly mobile barchan dunes, transverse dunes, and coastal transgressive dunes (McKee, 1966; Reitz et al., 2010; Stetler and Gaylord, 1996; Tsoar and Blumberg, 2002; Wolfe and Hugenholtz, 2009; Hesp and Walker, 2013), but also from blowouts (Baas and Nield, 2007; Girardi and Davis, 2010; Hesp, 2001), and stabilised transverse dunes and coastal foredunes (Carter et al., 1990; Hesp, 2002; Klijn, 1990; Muckersie and Shepherd, 1995; Nield and Baas, 2008). Well-vegetated parabolic dunes, on the other hand, can also be activated and transformed into more mobile barchan dunes and transverse dunes mediated by external pressures of both environmental changes and anthropogenic disturbances (Anton and Vincent, 1986; Hack, 1941; Hesp, 2001). In order to understand the different dune transformation mechanisms and their indications in the context of global climate change, the following sections first examine how vegetation plays a significant role in shaping aeolian dune morphology, and then explore how eco-geomorphic interactions lead to different dune transformations.

4.1 **Eco-geomorphic interactions in aeolian dune environments**

The development of a vegetated dunefield depends on the ability of vegetation to limit sand movement on the one hand, and on the ability of aeolian sand transport to limit the growth of vegetation on the other (Ashkenazy et al., 2012; Baas and Nield, 2010; Hack, 1941). Vegetation shapes local aeolian dune landscapes through processes of physical and biochemical interactions (Figure 18).
Almost all aeolian sand transport happens within 50 cm above the surface (McEwan and Willetts, 1993; Sherman et al., 1998). Physically, vegetation may shelter an erodible surface and decrease the available surface area for wind erosion. Comprising the primary roughness elements in aeolian dune environments, vegetation partitions shear stresses and hence decreases particle saltation and erosion potential because only finer grains with relatively smaller threshold shear velocities can be dislodged and transported by winds (Ash and Wasson, 1983; Gillies et al., 2007; Gillies et al., 2010; Levin et al., 2008; Musick and Gillette, 1990; Nordstrom et al., 2007; Raupach, 1992; Raupach et al., 1993; Turpin et al., 2010; Wiggs et al., 1995; Wolfe and Nickling, 1993; Wolfe and Nickling, 1996). Sand transport is effectively stopped when vegetation ground cover reaches ~20% (Kuriyama et al., 2005; Lancaster and Baas, 1998; Wiggs et al., 1995), although it is recognised that the threshold very much depends on the plant species, plant geometry and structure, and spatial distribution (Buckley, 1987). The roughness density or lateral cover, defined as the ratio of the total frontal-silhouette areas (perpendicular face of a plant which is vertical to the wind direction) of roughness elements to the total surface area (Marshall, 1971), is commonly used...
to evaluate the degree to which vegetation protects surfaces against wind erosion (Burri et al., 2011; Musick and Gillette, 1990; Raupach, 1992; Raupach, 1994; Raupach et al., 1993).

Whilst vegetation limits the capacity of winds to transport sand, vegetation considerably promotes its potential for trapping blown sand. This process can have significant biochemical implications for the micro-environment. By continuously trapping finer grains carried by winds, plants alter the particle-size distribution of sediments in their vicinity, in addition to altering soil texture and structure (Jungerius et al., 1995; Shields and Drouet, 1962). A relatively stable surface, a better water-retaining soil structure, and greater nutrient content (from plant roots and litter) promote the formation and development of crusts underlying plants (Johansen, 1993). Biological soil crusts (usually arising from cyanobacteria, mosses, and lichens) further increase soil stability and resistance to wind erosion by binding surface particles, and contribute nutrients to plants by fixing atmospheric nitrogen (Abed et al., 2013; Belnap, 2002; Belnap and Gillette, 1997; Belnap and Gillette, 1998; Delgado-Baquerizo et al., 2013; Drahorad et al., 2013; Eldridge and Leys, 2003; Johnson et al., 2007; Pluis, 1994; Shields et al., 1957; Thiet et al., 2005). Biological soil crusts, moreover, can resist long periods of droughts and desiccation, and can recover biological activity quickly as long as sufficient water is available from dew or precipitation (Veste et al., 2001; West, 1990).

The formation of Calcretes and gypcretes can also play an important role in reducing aeolian sediment transport and stabilising mobile dunes (Amit, 1995; Chen, 1997; Dijkmans et al., 1986; Galloway et al., 1992; Pye, 1980; Swezey, 2003; Warren, 1983). Aeolian dunes in southern Tunisia, for example, are stabilised by 0.1 to 0.5 m thick gypcretes (Swezey, 2003).

Changes in the characteristics of topsoil significantly alter hydrological regimes of aeolian dune environments, in particular water distribution and budget. In aeolian dune environments, precipitation can easily be lost due to the low moisture tension and the high hydraulic conductivity pertaining to sand (Tsoar and Blumberg, 2002). Plant canopies and
plant litter increase interception, lower soil bulk density, and act as a buffer protecting water from rapid loss (West, 1990). Biological crusts also influence infiltration, percolation, moisture retention, overland flow, and water redistribution (Belnap, 2006; Chamizo et al., 2012; Johansen, 1993; Rodríguez-Caballero et al., 2013; Tsoar and Moller, 1986; Verrecchia et al., 1995; West, 1990; Yair, 1990).

The complex mosaic pattern of vegetation in aeolian dune environments is responsible for spatial differences in flow dynamics over the surface and the associated patterns of sediment erosion and deposition (Ranwell, 1958; Willis and Yemm, 1961). Plants act as obstacles that cause sand to accumulate in their vicinity, thereby modifying the topography of dunes that they inhabit (Raupach, 1992; Wolfe and Nickling, 1993). Resulting sand accretion in and behind plants can lead to the formation of nebkhas (Tengberg, 1995; Tengberg and Chen, 1998) and shadow dunes (Gunatilaka, 1989; Hesp, 1981), respectively. Changes in micro-topography further alter airflow patterns over surfaces and shape dune landforms at a larger spatial scale (Frank and Kocurek, 1996a; Frank and Kocurek, 1996b). Nevertheless, vegetation, sand transport, and dune development are interactive within the context of climatological background, which is subject to a variety of environmental fluctuations because of seasonal or year-to-year variations in wind regime, temperature, precipitation, water table, and salinity. During drought or windy periods, for example, intense aeolian sediment transport can reshape dune landscapes significantly (Anderson and Walker, 2006; Byrne, 1997).

In order to survive in aeolian dune environments, plants employ both avoidance and tolerance strategies to cope with environmental stresses such as high wind velocities, sand blasting, sand accretion, wind erosion, unstable substratum, high soil temperature, and nutrient deficiency (Hesp, 1991; Maun, 1994; Maun, 1998). Seedling recruitment, for instance, usually happens during periods of low wind energy and high moisture availability.
(Maun, 1994). Survival on an eroding surface is usually a challenge for most plants. They are likely die of desiccation when their roots are exposed to the air by constant erosion (Lee and Ignaciuk, 1985; Maun, 1981).

Many plants, however, have various capabilities of withstanding sand burial.

According to the tolerance to sand accretion, Maun (1998) classified plant species in aeolian dune communities into the following three categories: non-tolerant, sand-tolerant, and sand-dependent. In his model (Figure 19), an individual plant may show the following different responses as sand burial increases: I) a negative response that causes the plant to die soon; II) no response and the plant grows normally within a certain level of sand accretion; and III) a stimulation of plant growth within a certain level of sand accretion. Despite a broad spectrum of the maximum tolerance to sand burial, every plant species will show a negative response beyond a certain limit (Dech and Maun, 2005; Gilbert and Ripley, 2010; Levinsh, 2006; Kent et al., 2001; Maun, 1998; Maun and Lapierre, 1984; Wagner, 1964; Zhang and Maun, 1994).

![Figure 19. A model of interaction between vegetation response and increasing levels of sand accretion in aeolian dune environments (adapted from Maun, 1998). I. negative vegetation response; II. no vegetation response within a limited level of sand accretion; and III. a stimulation of vegetation growth within a certain level of sand accretion.](Image)

Burial acts as a strong selective force by eliminating sensitive plant species, decreasing the abundance of less tolerant species, and increasing the abundance of sand-tolerant and sand-dependent species (Dech and Maun, 2005; Eldred and Maun, 1982;...
Martinez et al., 2001; Maun, 1994; Maun, 1998; Maun and Perumal, 1999; Moreno-Casasola, 1986). Many studies have indicated that some plant species are well-adapted to dynamic and recurrent burial, and that some plant species even require a certain amount of regular burial in order to maintain their high vigour (Bendali et al., 1990; Harris and Davy, 1987; Maun, 1998; Olson, 1958).

4.2 Transformations from barchan/transverse dunes to parabolic dunes

In contrast to parabolic dunes with two trailing arms pointing upwind, typical barchan dunes are crescent-shaped with two horns extending downwind and a slip face in their interiors. Barchan dunes develop under conditions of a high-energy unidirectional wind, a low sediment supply and a sparse vegetation cover (Hack, 1941; Rubin and Hunter, 1987; Wasson and Hyde, 1983). As sand supply increases, barchan dunes merge laterally into crescentic dunes with continuously sinuous ridges (Lancaster, 1995; Reitz et al., 2010). If sand supply is ample, then transverse dunes emerge with taller and continuous slip faces.

An increase in vegetation cover may lead to a transformation from barchan dunes or transverse dunes to parabolic dunes. This phenomenon has been studied in many countries including Israel (Tsoar and Blumberg, 2002), Brazil (Duran et al., 2008), Denmark (Anthonsen et al., 1996; Landsberg, 1956), the United Kingdom (Landsberg, 1956), New Zealand (Hesp, 2001), Canada (Wolfe and Hugenholtz, 2009) and the United States (Hack, 1941; McKee, 1966; Stetler and Gaylord, 1996) (Table 3). Most of these studies attribute the transformation to a progressive increase in vegetation cover caused by either climatic change or human disturbances. In many of these studies, however, the exact transformation process and the underlying mechanism remain unclear.
### Table 3. Studies on transformations from barchan or transverse dunes to parabolic dunes

<table>
<thead>
<tr>
<th>Reference</th>
<th>Study Region</th>
<th>Method</th>
<th>Transformation Cause</th>
</tr>
</thead>
<tbody>
<tr>
<td>Hack, 1941</td>
<td>Navajo Country, Arizona, US</td>
<td>field surveys and photo-engravings</td>
<td>aggressive plants survive sand burial</td>
</tr>
<tr>
<td>Landsberg, 1956</td>
<td>UK and Denmark</td>
<td>field observation</td>
<td>wetter climate</td>
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<td>McKee, 1966</td>
<td>White Sands, New Mexico, US</td>
<td>analysis of cross-stratification</td>
<td>arms anchored by vegetation</td>
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<tr>
<td>Muckersie and Shepherd, 1995</td>
<td>New Zealand</td>
<td>radiocarbon dating</td>
<td>El Nino-Southern Oscillation</td>
</tr>
<tr>
<td>Anthonsen et al., 1996</td>
<td>Rahbjerg Mile, Skagen Odde, Denmark</td>
<td>topographical maps and aerial photographic interpretation</td>
<td>climate changes (wind regime and vegetation cover)</td>
</tr>
<tr>
<td>Stetler and Gaylord, 1996</td>
<td>Hanford, Washington, US</td>
<td>A regional climate model</td>
<td>increased precipitation</td>
</tr>
<tr>
<td>Hesp, 2001</td>
<td>Manawatu, New Zealand</td>
<td>aerial photographic interpretation</td>
<td>human activity (vegetated by farmers and less grazing pressure)</td>
</tr>
<tr>
<td>Tsoar and Blumberg, 2002</td>
<td>Southeastern Mediterranean Coast, Israel</td>
<td>aerial photographic interpretation</td>
<td>decrease of human pressure (stop of agricultural land-use and reduced grazing activity)</td>
</tr>
<tr>
<td>Duran and Herrmann, 2006</td>
<td>Ceara, Brazil</td>
<td>continuous modelling</td>
<td>the ratio between dune characteristic erosion rate and vegetation growth velocity</td>
</tr>
<tr>
<td>Duran et al., 2008</td>
<td>Ceara, Brazil</td>
<td>field surveys, QuickBird panchromatic satellite images interpretation and continuous modelling</td>
<td>vegetation growth rate defined by two parameters: the initial vegetation growth rate when no sand transport occurs and the maximum height a plant can reach</td>
</tr>
<tr>
<td>Ardon et al., 2009</td>
<td>southern coastal plain of Israel</td>
<td>field surveys and aerial photographic interpretation</td>
<td>land use change and emergence of shrubs on the crest</td>
</tr>
<tr>
<td>Wolfe and Hugenholtz, 2009</td>
<td>Northern Great Plains, Canada</td>
<td>LIDAR imagery interpretation and optical stimulation luminescence dating</td>
<td>climate warming</td>
</tr>
<tr>
<td>Reitz et al., 2010</td>
<td>White Sands, New Mexico, US</td>
<td>LIDAR imagery and aerial photographic interpretation</td>
<td>dune surface erosion/deposition rates decrease below a threshold of half the vegetation growth rate</td>
</tr>
<tr>
<td>Barchyn and Hugenholtz, 2012a</td>
<td>-</td>
<td>CA modelling</td>
<td>climate shift</td>
</tr>
<tr>
<td>Barchyn and Hugenholtz, 2012b</td>
<td>-</td>
<td>process-based hypothesis</td>
<td>the slipface deposition rate &lt; the peak deposition tolerance of vegetation</td>
</tr>
<tr>
<td>Hart et al., 2012</td>
<td>Mason Bay, New Zealand</td>
<td>field surveys and aerial photographic interpretation</td>
<td>marram grass invasion</td>
</tr>
</tbody>
</table>

Some studies have suggested that the transformation is rooted in the anchor-like function of vegetation on the horns of barchan dunes (Livingstone and Warren, 1996; Muckersie and Shepherd, 1995; Reitz et al., 2010; Robertson-Rintoul, 1990; Stetler and Gaylord, 1996; Wolfe and Hugenholtz, 2009). Vegetation retards movement of horns, acting as anchors, whilst the dune apex continues to migrate downwind. As vegetation cover extends from dune horns to the apex, the advancing apex leaves behind the protected trailing ridges, and the dune is gradually transformed from the barchan to the parabolic shape in plan. Duran and Hermann (2006) simulated this process using a continuum model starting with a single
barchan dune on a non-erodible bed. Barchyn and Hugenholtz (2012a) simulated a barchan-
to-parabolic dunefield transformation imposed by a climate shift using a cellular automaton
(CA) model (Baas and Nield, 2010; Nield and Baas, 2008). A few studies have suggested that
this transformation happens when the sand deposition rate decreases below a certain
threshold related to the vegetation growth rate (Barchyn and Hugenholtz, 2012b; Duran and
Herrmann, 2006; Reitz et al., 2010).

The ‘horns-anchoring’ mechanism described above is likely to happen on the
following three conditions: 1) barchan dunes are surrounded by well-vegetated land; 2)
vegetation species are sufficiently aggressive to withstand a certain amount of sand burial;
and 3) sediment availability is relatively limited. The first prerequisite can be fulfilled when a
barchan dune is moving onto an area with a greater vegetation cover (Reitz et al., 2010). As
an alternative, the interdune areas of barchan dunefields may be re-vegetated because of
changes in environmental factors, such as increased precipitation (Landsberg, 1956; Stetler
and Gaylord, 1996), reduced wind strength (Anthonsen et al., 1996), and climate warming
(Wolfe and Hugenholtz, 2009), or because of changes in anthropogenic pressure on the
environment, such as reduced grazing activity or artificial vegetation restoration (Hesp, 2001).
The second condition depends on the characteristics of local plant communities, in particular
the dominant species, which is also closely related to the regional climate parameters of wind
regime, temperature, and precipitation. The third condition is determined by sand sources,
specifically a limited external sand supply and a thin sandy substratum.

Tsoar and Blumberg (2002) proposed another potential mechanism driving the transformation
from barchan dunes or transverse dunes to parabolic dunes, specifying that the establishment
of vegetation on the crest of barchan dunes or transverse dunes initiates the transformation.
Their argument is that vegetation preferentially grows and recovers on the crest of dunes
where the erosion/deposition balance is neutral. The establishment of vegetation on the crest
then changes the airflow and sediment transport dynamics over the stoss slope. Sand eroded
from the stoss slope is partly trapped by clumps of plants forming isolated nebkhas, and the
associated abrupt reduction in sediment supply encourages plants to take root on the lee
slopes and slip faces. The accumulation of sand on the crest by nebkhas, meanwhile,
gradually changes the profile of the stoss slope from convex to concave. The subsequent
funnelling effect of the wind over the concave stoss slope can undercut the nebkhas and
expose plant roots on the central apex, but plants on dune sides remain intact and are left
behind by the mobile dune apex, developing into trailing arms eventually. Hugenholtz et al.
(2008) confirmed the important role of vegetation on the dune crest in trapping sand
transported from the stoss slope and in changing the dune profile.

In contrast to the previous mechanism in which vegetation is established close to the
groundwater table in interdune areas and on the horns of barchan dunes, in the mechanism
proposed by Tsoar and Blumberg (2002) vegetation starts to germinate and grow on the crest
of dunes where erosion and deposition balance each other. Some studies have shown that a
small amount of sand burial may prevent atmospheric desiccation, increase relative humidity
around seeds, and anchor seedlings into soil; therefore, a small amount of sand burial may be
vital to ensure successful seed germination and plant growth (Maun, 1998). Maun and
Lapierre (1986) suggested that a maximum germination rate occurs at a burial depth of 2-4
cm across all four studied dune species. A study on seven dune species by Zhang and Maun
(1994) has shown that the majority of seeds germinate at a depth of 5-10 cm, and that deeper
burial greater than 15 cm significantly inhibits seed germination. These results are consistent
with the findings from Lee and Ignaciuk (1985). A relatively stable surface with a slight
amount of net deposition is clearly crucial for seeds to take root.
This ‘nebkhas-initiation’ mechanism proposed by Tsoar and Blumberg (2002) involves the following steps/prerequisites: 1) seeds germinate successfully on dune crests during a less windy season; 2) sufficient precipitation and minor aeolian sediment mobilisation allow seedlings on the crests to thrive such that they can prepare themselves adequately for strong wind energy during the following windy season; and 3) plants develop tap roots that allow them to use relatively sustainable groundwater during a dry and windy season, ensuring that the plants are able to keep vitality, trap sand continually, and develop into nebkhas that can subsequently alter the profile of stoss slopes in shape. These three prerequisites are intimately connected with each other and fundamentally provide an opportunity for vegetation to survive on the crest of dunes, arising from some combination of increased precipitation, reduced windiness, and/or rise of the local groundwater table. These variables, in turn, respond to seasonal fluctuations, climate change, and/or human pressure.

To act as an initial trap and further develop into a nebkha that initiates this transformation process, the vegetation involved should be perennial and needs not only to grow up in height quickly, but also to display a branching growth pattern in order to achieve a high sand-trapping efficiency (Livingstone and Warren, 1996). Isolated plants that have lost their lower branches and leaves are much less efficient in decreasing wind velocity and sand transport, because almost all sand transport happens close to the ground. Extensive roots are necessary for vegetation survival on dune crests where soils have particularly low fertility and low capacity for retaining precipitation.

The interaction between vegetation and sand transport as well as the related mechanisms of the transformation from barchan/transverse dunes to parabolic dunes is likely to vary depending on vegetation species and the limiting factors that control vegetation germination and growth. Annual grasses such as *Agriophyllum squarrosum* are short-lived with shallow
roots, are unlikely to access groundwater, and hence only survive on precipitation. They usually germinate and grow rapidly after episodic rainfall events, but die of drought shortly afterwards. The impacts of annuals on dune morphology are, therefore, generally highly limited. Perennial vegetation, however, develops deep and/or widespread roots, and can exert different impacts depending on their growth forms. For perennial grasses that grow uniformly over the surface, the degree to which such grasses reduce sand transport is primarily determined by an overall coverage of grass assembly. Perennial vegetation in need of a large water supply is usually distributed relatively sparsely as discrete clumps and shrubs. Such perennial vegetation influences the local wind regime more as individual entities through their outstanding canopies.

The ‘horns-anchoring’ transformation mechanism tends to occur in an environment where water deficiency is the limiting factor for vegetation growth, whereas the ‘nebkhas-initiation’ mechanism tends to occur where wind erosion is the predominant limiting factor. In comparison with the ‘horns-anchoring’ mechanism in which any perennial species can play the anchoring role as long as it can withstand a small degree of wind erosion and sand burial (Reitz et al., 2010; Wolfe and Hugenholtz, 2009), perennial vegetation in the ‘nebkhas-initiation’ mechanism needs to develop tap roots in order to access groundwater from dune crests and at the same time has the capability of withstanding substantial amounts of sand burial as the nebkha form (Tsoar and Blumberg, 2002). The ‘nebkhas-initiation’ mechanism demands more specialised plant species and is hence less common. Parabolic dunes formed by the ‘nebkhas-initiation’ mechanism, furthermore, are much less elongated compared with that of the ‘horns-anchoring’ mechanism in which bare lobes can move forward unimpeded over a long period.
4.3 Transformations from blowouts to parabolic dunes

The evolution of blowouts can result in the formation of parabolic dunes, specifically on vegetated sandy surfaces (Gutierrezelorza et al., 2005; Landsberg, 1956; Pye, 1982). Coastal examples have been documented in Australia (Pye, 1982; Pye, 1983b), New Zealand (Brothers, 1954; Hesp, 2001), the United Kingdom (Ranwell, 1958), the Netherlands (Klijn, 1990), Spain (Gutierrezelorza et al., 2005), Brazil (Duran et al., 2008), and the United States (Girardi and Davis, 2010; Hansen et al., 2009), whereas inland examples have been documented in the southern Kalahari Desert in south Africa (Eriksson et al., 1989) and the Jafurah Desert in Saudi Arabia (Anton and Vincent, 1986) (Table 4). Some early studies refer to parabolic dunes developed from blowouts as “blowout dunes” (Brothers, 1954; Cooper, 1958; Eriksson et al., 1989; Melton, 1940).

Blowouts are saucer-, bowl-, cup-, or trough-shaped depressions that usually develop by wind erosion on a pre-existing sand deposit (Hesp et al., 2011; Hesp, 2002; Hesp and Hyde, 1996). The initiation of blowouts may be associated with both natural and anthropogenic disturbances to a vegetation cover such as wildfires, increased windiness, rises in sea level, increased frequency of drought, large volcanic eruptions, storms, overgrazing, trampling, and diseases (Gutierrezelorza et al., 2005; Muckersie and Shepherd, 1995; Pye and Tsoar, 1990; Watt, 1937). These agents locally exterminate vegetation, breach the surface crust, and create discrete bare patches that develop into hollows as erosion continues. These hollows may become enlarged, and turbulent eddies may form, which further accelerate expansion of the hollows (Hesp, 2002). The deepening of the hollows also encourages the funnelling of winds, which in turn can result in more intense erosion (Smyth et al., 2014).

Continued wind scour widens blowouts by steepening side walls and inducing side wall failure (Carter et al., 1990). Sand eroded from the deflation hollow is transported by winds.
and accumulates in the leeward margin, forming a depositional lobe, which is regarded as part of a blowout by some researchers (Hesp and Hyde, 1996; Robertson-Rintoul, 1990).

Cooper (1958) defined two primary types of blowouts: saucer blowouts and trough blowouts. Despite a wide range of variability in different aeolian environments, most blowouts can be classified as either of these two types (Hesp, 2002). Saucer blowouts are shallow, and semicircular- or saucer-shaped. Trough blowouts are deeper, and more elongated with longer lateral walls (Carter et al., 1990; Cooper, 1958; Hesp, 2002; Hesp and Hyde, 1996). Compared with a short, wide, radial depositional lobe found in the leeward margin of a saucer blowout, a trough blowout develops a tall, parabolic-shaped depositional lobe. In a trough blowout, corkscrew airflows may develop in the deflation basin. These airflows are then compressed and accelerated towards the crest of the depositional lobe, and subsequently decelerate rapidly due to flow expansion upon the exiting trough, leading to the development of a parabolic lobe (Hesp, 2002).

If a trough blowout and its lobe overcome the surrounding vegetation, they continue to migrate and develop downwind, enlarging overtime and evolving into a fully developed parabolic dune (Carter et al., 1990; Hesp and Hyde, 1996; Livingstone and Warren, 1996). The parabolic dune may continue to increase in size if the blowout provides a continuous sediment supply by down-wearing its deflation basin. The dune may cease to increase in size if the funnelling effect is lessened by progressive widening of the openings, if a groundwater table or resistant stratum are exposed (such as a caliche bed or a clay bed), or if revigorated growth of vegetation impedes further erosion (Cooper, 1958; Livingstone and Warren, 1996; Melton, 1940). Baas and Nield (2007) simulated the development of blowouts using a Discrete ECo-geomorphic Aeolian Landscape model (DECAL), and showed the significant role of the interplay between vegetation growth and sand transport in the transformation from blowouts to elongated parabolic dunes. Baas and Nield (2010) further showed that blowouts
can also expand laterally as they elongate, incorporate progressively increased sand, and
develop into more mobile transgressive and transverse ridges. The initiation of blowouts by
disturbances and their potential impacts on overall dunefield activation has been
conceptualised by Barchyn and Hugenholtz (2013) in the context of vegetation resilience and
sediment transport activity. They also showed that depth-limited blowouts can migrate and
elongate at a high rate, and are hence more difficult to be stabilised by vegetation. In addition
to cellular automaton models, Duran et al. (2008) applied a continuum model in a similar
context to simulate the development of a parabolic dune in north-eastern Brazil.

The formation of parabolic dunes from blowouts requires the following three
conditions: 1) a generally stabilised surface, which enables concentration of winds at isolated
points of weakness; 2) an underlying sandy substratum, which provides sufficient sediment
supply for forming a parabolic-shaped depositional lobe that must avoid the coalescence of
adjacent blowouts; and 3) predominantly unidirectional winds (Cooper, 1958; Eriksson et al.,
1989; Pye, 1983b). Blowouts developed on coastal dunes are in some instances regarded as a
symptom of a negative sand budget (Livingstone and Warren, 1996; Psuty, 1988). A decrease
in sand supply from a beach means that onshore winds have larger capacity to erode
foredunes. The higher parts of the dunes are usually more vulnerable to desiccation and
disturbance, and therefore more susceptible to initiation of a blowout. Blowouts are, however,
present widely on the coasts of Manawatu Plain in New Zealand where beaches are
progradational (Hesp, 2002). Sufficient sand supply and strong unidirectional winds are
essential for the further transformation of blowouts into parabolic dunes, which may be
secured by climatic changes such as a stronger wind regime, a more arid climate, and a
lowered groundwater table, or by land degradation induced by anthropogenic perturbation
such as grazing and human recreational activities.
Table 4. Studies on transformations from blowouts to parabolic dunes.

<table>
<thead>
<tr>
<th>Reference</th>
<th>Study Region</th>
<th>Method</th>
<th>Transformation Cause</th>
</tr>
</thead>
<tbody>
<tr>
<td>Melton, 1940</td>
<td>southern High Plains, US</td>
<td>aerial photographic interpretation</td>
<td>more arid climate, lowered ground-water surface</td>
</tr>
<tr>
<td>Hack, 1941</td>
<td>Navajo Country, Arizona, US</td>
<td>field surveys and photo-engravings</td>
<td>periglacial winds on sand covered by sparse vegetation</td>
</tr>
<tr>
<td>Brothers, 1954</td>
<td>Auckland, New Zealand</td>
<td>field observation</td>
<td>burning-off, animal tracks and topographic difference</td>
</tr>
<tr>
<td>Cooper, 1958</td>
<td>Oregon, US</td>
<td>description</td>
<td>windward slope stabilisation of saucer blowouts</td>
</tr>
<tr>
<td>Ranwell, 1958</td>
<td>Anglesey, Wales, UK</td>
<td>field surveys</td>
<td>natural forces</td>
</tr>
<tr>
<td>Pye, 1982</td>
<td>Cape Bedform and Cape Flattery, Queensland, Australia</td>
<td>aerial photographic interpretation</td>
<td>natural forces (fire, cyclones and lightning strikes) and local differences in vegetation cover or topography</td>
</tr>
<tr>
<td>Pye, 1983b</td>
<td>Northern Cape York Peninsula, Queensland, Australia</td>
<td>field surveys, aerial photographic interpretation and sediment analyses</td>
<td>increase in windiness or reduction in rainfall</td>
</tr>
<tr>
<td>Anton and Vincent, 1986</td>
<td>Jafurah Desert, Eastern Province, Saudi Arabia</td>
<td>field surveys and aerial photographic interpretation</td>
<td>preferable deflation of sand sheet areas</td>
</tr>
<tr>
<td>Eriksson et al., 1989</td>
<td>southern Kalahari Desert, South Africa</td>
<td>field surveys and aerial photographic interpretation</td>
<td>deflation of bare patches shaded by trees assisted by biological processes</td>
</tr>
<tr>
<td>Carter et al., 1990</td>
<td>west coast, Australia</td>
<td>description</td>
<td>trough blowouts evolve into parabolic dunes</td>
</tr>
<tr>
<td>Klijn, 1990</td>
<td>Younger Dunes, Netherlands</td>
<td>C14 and historical dating</td>
<td>climatic changes (sea level rise, and increased storm frequency and surges)</td>
</tr>
<tr>
<td>Hesp, 2001</td>
<td>Manawatu, New Zealand</td>
<td>aerial photographic interpretation</td>
<td>natural forces aided by human recreational activity</td>
</tr>
<tr>
<td>Hesp, 2002</td>
<td>-</td>
<td>description</td>
<td>high energy wind coasts</td>
</tr>
<tr>
<td>Gutierrezelozoa et al., 2005</td>
<td>Tierra de Pinares, Spain</td>
<td>field observation</td>
<td>climate changes</td>
</tr>
<tr>
<td>Baas and Nield, 2007</td>
<td>-</td>
<td>DECAL model</td>
<td>Dynamic interplay between sedimentation balance and vegetation effectiveness</td>
</tr>
<tr>
<td>Duran et al., 2008</td>
<td>Ceara, Brazil</td>
<td>field surveys, QuickBird panchromatic satellite images interpretation and continuous modelling</td>
<td>vegetation growth rate defined by two parameters: the initial vegetation growth rate when no sand transport occurs and the maximum height a plant can reach</td>
</tr>
<tr>
<td>Hansen et al., 2009</td>
<td>Green Mountain Beach dune, Holland, southeast shore of Lake Michigan, US</td>
<td>field surveys</td>
<td>steering of winds in the deflation area</td>
</tr>
<tr>
<td>Girardi and Davis, 2010</td>
<td>Walking Dunes, New York, US</td>
<td>aerial photographic interpretation and previous literatures</td>
<td>dune-vegetation interactions</td>
</tr>
</tbody>
</table>

4.4 Transformations from parabolic dunes to other dune morphologies

Under certain conditions, parabolic dunes may be transformed into other dune morphologies. Parabolic dunes are commonly developed in well-vegetated landscapes and under conditions of a restricted sediment supply (Hack, 1941; Lancaster, 1995). If a new sediment supply becomes available or if vegetation cover decreases below a certain level, parabolic dunes may be transformed into more mobile dunes (Livingstone and Warren, 1996; McKee, 1966). Some studies have indicated that parabolic dunes can lose vegetation, and are activated and...
transformed into transverse dunes (Anton and Vincent, 1986; Hack, 1941; Hesp, 2001). Some regions exhibit a downwind transition continuum from parabolic dunes to barchan dunes and transverse dunes (Pye and Tsoar, 1990). Studies on these transformations are, nevertheless, few and often limited to anecdotal descriptions (Table 5). The biomorphic interactions and physical processes underlying the activation of parabolic dunes and their transformations into highly mobile barchan dunes or transverse dunes have not been investigated in detail, whereas these transformations may have significant implications on local land management and social-economic development. Parabolic dunes may also develop into other dune forms such as dome dunes (Anton and Vincent, 1986).

### Table 5. Studies on transformations from parabolic dunes to other dune morphologies

<table>
<thead>
<tr>
<th>Transformation Type</th>
<th>Reference</th>
<th>Study Region</th>
<th>Method</th>
<th>Transformation Cause</th>
</tr>
</thead>
<tbody>
<tr>
<td>parabolic to transverse and barchan</td>
<td>Hack, 1941</td>
<td>Navajo Country, Arizona, US</td>
<td>field surveys and photo- engravings</td>
<td>vegetation destroyed by external sands from blowouts</td>
</tr>
<tr>
<td></td>
<td>Anton and Vincent, 1986</td>
<td>Jafurah Desert, Eastern Province, Saudi Arabia</td>
<td>field surveys and aerial photographic interpretation</td>
<td>decline of vegetation density possibly due to change in water tables, natural vegetation succession and over-grazing</td>
</tr>
<tr>
<td></td>
<td>Shulmeister and Lees, 1992</td>
<td>Groote Eylandt, Australia</td>
<td>thermo-luminescence dating</td>
<td>decline in vegetation cover because of decreased precipitation and/or increased aboriginal burning</td>
</tr>
<tr>
<td></td>
<td>Hesp, 2001</td>
<td>Manawatu, New Zealand</td>
<td>aerial photographic interpretation</td>
<td>human activity (burning, grazing, introduction of exotic species and wetland modification)</td>
</tr>
<tr>
<td></td>
<td>García-Hidalgo et al., 2002</td>
<td>Duero Basin, Spain</td>
<td>aerial photographic interpretation</td>
<td>noses of parabolic dunes are stopped by water and the arms continues to move forward</td>
</tr>
<tr>
<td>parabolic to dome</td>
<td>Anton and Vincent, 1986</td>
<td>Jafurah Desert, Eastern Province, Saudi Arabia</td>
<td>field surveys and aerial photographic interpretation</td>
<td>vegetation colonisation in the deflation hollow due to lower groundwater salinity</td>
</tr>
<tr>
<td>parabolic to longitudinal</td>
<td>Meurisse et al., 2005</td>
<td>northern shore, France</td>
<td>stratigraphy, 14C dating and sedimentology</td>
<td>climatic modifications, agricultural practices and sea level rise</td>
</tr>
</tbody>
</table>

5. **A conceptual framework for understanding parabolic dune transformations**

Parabolic dunes are distributed widely across a large range of climatic gradients associated with a variety of floristic regions responding to different environmental stresses. Parabolic dunes are found on coasts, river valleys and lake shores as well as the margins of deserts and steppes, where eco-geomorphic interactions are often highly sensitive to climatic variability.
and closely influenced by socio-economic activities (Livingstone and Warren, 1996). This paper outlines a conceptual framework to assist in understanding the development of parabolic dunes and related transformations into other dune morphologies. In turn, this framework may serve as a useful tool to help predict possible landscape development scenarios in different aeolian environments under climate changes. Dune landscapes are governed primarily by wind regime, sand availability, and vegetation cover (Hack, 1941). The following three determinant factors are therefore chosen in this framework: 1) sand availability; 2) wind strength (wind variability is disregarded because parabolic dunes are mostly developed under a unidirectional wind regime); and 3) drought stress, the major control for the growth of vegetation in an aeolian environment. The degree to which drought stress influences vegetation growth varies depending on the characteristics of vegetation species, such as the resilience to drought and the capability of withstanding wind erosion and sand burial. Drought stress is generally determined by the regional climate, and encompasses the combined impacts of temperature, precipitation, and groundwater dynamics.

Two parts are included in the framework. The first part illustrates how increases and decreases in wind strength and/or increases and decreases in existing drought stress lead to the development of parabolic dunes evolved from blowouts (mobilisation), or from barchan dunes and transverse dunes (stabilisation), as shown in Figure 20. Each pane represents an eco-geomorphic system with a different level of sand availability, in which horizontal and vertical axes denote changes in drought stress and wind strength respectively. Five representative panes are shown here for simplification, with sand availability increasing from left-most (Figure 20a) to right-most (Figure 20e).
Figure 20. Conceptual diagrams of dune transformations leading to development of parabolic dunes under changes in wind strength and drought stress. Vertical axis within each pane denotes a change in wind strength, either increasing (upward) or decreasing (downward) from the original climatic state at the origin. Horizontal axis within each pane denotes a change in drought stress relative to the original climatic state, either increasing (to the right) or decreasing (to the left). S, P, and T represent Stabilised dunefields, Parabolic dunefields, and Transverse dunefields respectively. The white zones represent conditions where the existing landform is not transformed. Purple dotted lines are examples of minimal boundaries of changes in wind strength and drought stress leading to development of parabolic dunes under a different dominant vegetation species with stronger erosion and sand burial tolerances. The sequence of panes represents a gradient of sand availability from low (left-most) to high (right-most).

The transformation from a blowout to a parabolic dune (Figure 20a-c) occurs when sufficient sand is available from the deflation basin of the blowout for aeolian transport, and is deposited downwind, forming a tall, parabolic-shaped lobe.

In an environment with a limited sand supply from a sandy substratum (Figure 20a), given an existing vegetation species (with its specific growth structure and behaviour), a blowout is likely to be transformed into a parabolic dune if more sand is acquired by increasing wind strength and/or drought stress (moving to the upper-right in Figure 20a).

Stronger winds have greater capability to erode soils and plants on the downwind edge of the blowout, and to expose more sand underneath as the blowout elongates by eliminating vegetation on its way. Greater drought stress provides more sand due to a loss of vegetation cover by desiccation. Extreme high wind strength may lead to the development of parabolic dunes from blowouts even if drought stress decreases slightly. Similarly, this transformation may happen during an extremely dry but less windy period.
We hypothesise that blowouts with a moderate sand supply (Figure 20b) can be transformed into parabolic dunes in an environment with smaller thresholds of increases in wind strength and/or drought stress compared with that of the blowouts in an environment with a limited sand supply. This difference may exist because more sand is available for transport when vegetation on the same surface area is removed.

In an environment with ample sand supply (Figure 20c), the available sand in a blowout is beyond the transport capacity of current winds, and wind strength becomes the limiting factor for sand transport. In this situation, an increase in drought stress alone (to any level) is insufficient to initiate a transformation, because sand transport is already at capacity. Drought stress impact is only effective if it is accompanied by an increase in wind strength capable of transporting the newly exposed sand. In a situation where drought stress does not change, an increase in wind strength leads more easily to a transformation to a parabolic dune than in the previous two cases (Figure 20a-b), because of a greater sand supply.

In all three cases discussed above, when wind strength and drought stress increase dramatically so that vegetation declines rapidly, blowouts elongate downwind, expand laterally, and interact/coalesce with nearby/adjacent blowouts, developing into a transverse dunefield. The thresholds of increased wind strength and/or drought stress needed for the transformation to transverse dunes are higher with greater sand availability (from Figure 20a to 20c): a thicker substratum below the blowouts offers a more abundant local sand supply, which causes the development of larger depositional lobes that migrate at a slower pace.

Conversely, a thinner substratum enables blowouts to extend downwind at a higher rate and encourages the interactions and lateral coalescence of nearby/adjacent blowouts. Large-scale coalescence activity and the disappearance of vegetated ridges between blowouts can then
mobilise the entire area, leading to the development of a transverse dunefield (as discussed in section 4.3).

Blowouts can also be stabilised (moving to the lower left in Figure 20a-c) when wind strength decreases so that sand transport diminishes and the associated impacts of erosion and sand burial on vegetation growth are relieved, and/or drought stress decreases so that vegetation cover is restored due to greater water availability. Blowouts with less sand availability are more easily stabilised because of limited initial sand transport activity.

In comparison to blowouts (Figure 20a-c), barchan dunes and transverse dunes develop in an environment with a greater sand supply (Figure 20d-e). These dunefields are very active with a limited, usually temporary, vegetation cover.

The transformations of barchan dunes and transverse dunes into parabolic dunes (moving to the lower left in Figure 20d-e) are associated with the processes of dune re-vegetation and stabilisation, achieved by 1) a decrease in wind strength so that the erosion and sand burial activity is alleviated to the capability of vegetation tolerances; and/or 2) a decrease in drought stress so that less vegetation dies of desiccation. As more abundant sand is available in a transverse dunefield compared with that of a barchan dunefield, the transformation from transverse dunes to parabolic dunes requires a more considerable decrease in wind strength and/or drought stress in order to provide a relatively stable surface for vegetation taking root. An extreme situation in which both the wind strength and the drought stress are reduced significantly may lead to the stabilisation of the whole dunefield (moving to the lower left corner in Figure 20d-e).

Vegetation species plays a key role in dune transformations by moving the boundaries between different transformations with regard to changes in wind strength and drought stress.
An example in Figure 20c presents the boundary of a transformation from blowouts to parabolic dunes under the control of a different vegetation species with greater erosion and burial tolerances (Veg Type b), for example related to different growth structure and behaviour, in comparison to the default species (Veg Type a). To transform into a parabolic dune, a blowout keeps elongating downwind and eroding vegetation. A vegetation species with greater erosion and deposition tolerances is able to resist more severe sand transport activity. Therefore, the transformation from blowouts to parabolic dunes necessitates greater minimal increases in wind strength and drought stress.

The boundary of a transformation from transverse dunes to parabolic dunes regarding changes in wind strength and drought stress is more easily met (Figure 20e) under the influence of a vegetation species with greater erosion and sand burial tolerances (Veg Type b). This relation is due to the fact that stronger erosion and deposition tolerances enable vegetation to survive greater sand transport. As a result, smaller minimal decreases in wind strength and/or drought stress are required.

Similar deductions can be made for other dune transformations (e.g., from blowouts to transverse dunes, and from barchan dunes to parabolic dunes) and the stabilisation of different dunefields. In general, vegetation species with greater erosion and deposition tolerances move boundaries of dune transformations towards the upper right of panes in Figure 21.

The other half of the framework shown in Figure 21 illustrates how changes in wind strength and drought stress result in the transformations from stabilised as well as active parabolic dunes, with varying sand availability, into highly mobile barchan dunes and transverse dunes.
Figure 2. Conceptual diagrams of dune transformations from parabolic dunes to barchan dunes and transverse dunes under changes in wind strength and drought stress. Upper sequence represents transformations from fully stabilised parabolic dunes along a gradient of sand availability; lower sequence shows transformations from active parabolic dunes along a gradient of sand availability. Vertical axis within each pane denotes a change in wind strength, either increasing (upward) or decreasing (downward) from the original climatic state at the origin. Horizontal axis within each pane denotes a change in drought stress relative to the original climatic state, either increasing (to the right) or decreasing (to the left). B, T, and S represent Barchan dunefields, Transverse dunefields, and Stabilised dunefields respectively. The white zones represent conditions where the existing landform is not transformed. The purple dotted line in pane (a) is an example representing boundaries of changes in wind strength and drought stress leading to the activation of parabolic dunes into barchan dunes under a different dominant vegetation species with stronger erosion and sand burial tolerances.

Activation of stabilised parabolic dunes cannot occur solely by increasing wind strength, because surfaces are fully covered with vegetation and there is no sand available for transport (Figure 21a-c). A sole increase in drought stress, however, can lead to widespread vegetation decline due to desiccation, exposing sand at the surface to wind energy. Under severe drought stress, parabolic dunes are likely to be activated and transformed into barchan dunes even if wind strength decreases slightly. An increase in wind strength encourages the activation of parabolic dunes only if an accompanying drought stress makes available
sufficient additional sand supply. The drought stress threshold for an increased wind strength being able to initiate a transformation from parabolic dunes to barchan dunes is independent of the sandy substratum thickness, but is governed by the characteristics of vegetation cover, in particular the vegetation resilience to drought and tolerances to erosion and sand burial.

Stabilised parabolic dunes with a thicker sandy substratum (from Figure 21a, to b, to c) require a smaller increase in drought stress to be transformed into barchan dunes, because more sand is exposed to winds under the same level of drought stress. When vegetation cover is largely destroyed by substantial increases in drought stress and/or wind strength, the activation of parabolic dunes can also lead to the development of transverse dunes, provided that sufficient sand is conserved in the dune environment (Figure 21c).

In comparison to stabilised parabolic dunes with the same level of sand availability (Figure 21a-c), the transformation of active parabolic dunes to barchan dunes requires relative smaller increases in wind strength and/or drought stress (Figure 21d-f).

For active parabolic dunes with a limited sand supply (Figure 21d), similar to their stabilised counterparts (Figure 21a), an increasing wind strength does not trigger a transformation until a greater drought stress creates an additional sand supply. The threshold of drought stress is lower for active parabolic dunes because more sand surface is already exposed to the wind. For the same reason, compared with that of their stabilised counterparts, the transformation from active parabolic dunes to barchan dunes requires a smaller increase in drought stress.

In a field of active parabolic dunes where sand availability is larger than sand transport capacity of winds (Figure 21e), parabolic dunes can be transformed into barchan dunes by increasing wind strength (and the associated sand transport capacity) alone, as well as by increasing drought stress. Although an increase in drought stress alone does not
enhance sand transport capacity, an increase in drought stress raises the vulnerability of vegetation on trailing arms of active parabolic dunes and on surrounding interdune areas to wind erosion and sand burial activity. As a result, bare lobes of active parabolic dunes move forward more easily, then break with arms, and transform into barchan dunes.

In comparison to active parabolic dunes with a less sand supply, active parabolic dunes with an abundant sand supply (Figure 21f) require a smaller increase in wind strength and/or drought stress to be transformed into barchan dunes. A significant increase in wind strength and/or drought stress, meanwhile, can result in severe activation of dunes and the transformation of parabolic dunes into transverse dunes.

Active parabolic dunes can also be fully-stabilised when wind strength is reduced and/or drought stress is attenuated. Aeolian environments with more sand availability require larger decreases in wind strength and/or drought stress in order to be stabilised by vegetation (Figure 21d-f).

Similar to the dune transformations in Figure 20, the boundaries of transformations from active parabolic dunes to barchan dunes and transverse dunes as well as stabilisation by vegetation are controlled by the characteristics of vegetation species. An example in Figure 21a shows the boundary of the transformation from stabilised parabolic dunes with a limited sand supply into barchan dunes under the influence of a different vegetation species with greater erosion and sand burial tolerances (Veg Type b). In Figure 21, the boundary denoting thresholds for increases in wind strength and drought stress for this different species moves towards the upper right of the pane because vegetation is able to survive in an environment with greater aeolian activity. Similarly, boundaries of other transformations in Figure 21 show the same behaviour. Specifically, vegetation species with greater resilience to aeolian sediment mobilisation move transformation boundaries towards the upper right of a system in Figure 21.
As discussed in section 2, a change in dune migration rate is not a good indicator for the state of activity of an aeolian environment. Vegetation responds to climatic fluctuations generally more quickly than dune morphology, thereby providing a buffer between climatic change and geomorphic responses (Phillips, 1995; Yizhaq et al., 2009). Dune transformation tendency, representing the evolutionary trend of dune landforms mediated by vegetation change and eco-geomorphic feedback, is closely related to the stability of a dunefield and can be potentially used as an indicator to anticipate the activity of aeolian dunes under the impacts of environmental and anthropogenic forces.

Dune transformation tendency can be defined as the likelihood of a dune transformation occurring under changes in wind strength and drought stress, arising from environmental fluctuations, climatic change, and human disturbances. For example, blowouts with a limited sand supply need relatively large increases in wind strength and/or drought stress to transform into parabolic dunes, as compared to blowouts with a moderate sand supply, and therefore the transformation tendency from blowouts to parabolic dunes is relatively large in settings with a moderate sand supply. In other words, blowouts with a moderate sand supply are more sensitive to changes in environmental controls on their transformations into parabolic dunes. Dune transformation tendency is influenced by regional environmental parameters and can be potentially identified through knowledge of local eco-geomorphic dynamics and processes on a time scale of decades (Viles and Goudie, 2003).

The framework presented in this paper provides a conceptual way to organise and compare different aeolian dune environments and potential dune transformations under the impacts of environmental fluctuations and climatic change. The representative aeolian environments shown in the diagrams, however, are not fixed, and any given dune in reality can be related to different diagrams when sand supply changes. For example, sand
availability can increase when more sand is blown from a beach into a field of parabolic
dunes due to a rise in sea level (Carter, 1991). A fall of the groundwater table can also
increase sand availability by increasing the thickness of a dry sandy substratum which can be
potentially mobilised and transported by winds (Swezey, 2003). Furthermore, changes of the
dominant vegetation species arising from either natural succession or human disturbances can
move the boundaries of transformations in an aeolian dune system.

6. Conclusions

The development of parabolic aeolian dunes and related transformations to and from other
dune morphologies are very sensitive to vegetation characteristics and to environmental
variations arising from both natural and anthropogenic disturbances. This paper reviews the
literature on parabolic dunes and their related transformations on a global scale to provide a
comprehensive inventory. Coastal and inland parabolic dunes differ on environmental
controls, transformation processes, and morphology. Coastal parabolic dunes are controlled
by geometrical alignment of the coastline, tidal range, wave power, and sea level change.
Coastal parabolic dunes are often associated with the initiation of blowouts on previously
vegetated foredunes by either natural forces such as storms or human disturbances such as
grazing, and are sometimes associated with the stabilisation of transgressive dunefields by
vegetation colonisation. In contrast, inland parabolic dunes are governed by orographic
conditions and groundwater availability related to nearby rivers or lakes. Inland parabolic
dunes are largely transformed from barchan dunes or transverse dunes, which are usually
found in arid and semi-arid regions. In relatively humid areas, coastal parabolic dunes can
migrate over trees and form arms of relatively high relief, whereas inland parabolic dunes
usually have arms of relatively low relief because grasses and shrubs dominate those areas.
Elongated parabolic dunes are often found on coasts where wet periods alternate with dry
periods accompanied by strong onshore winds (usually in equatorial or warm climates), but such elongated dunes are not commonly seen inland because the dune arms are frequently overridden or cut through by following dunes.

The transformation process from barchan dunes to parabolic dunes varies depending on vegetation species as well as the limiting factors of their germination and growth. The ‘horns-anchoring’ transformation mechanism is likely to happen when water deficiency is the limiting factor, whereas the ‘nebkhas-initiation’ transformation mechanism is likely to occur when vegetation growth is threatened primarily by wind erosion. Of the two mechanisms, the ‘nebkhas-initiation’ mechanism is less common because it requires more specialised vegetation species that can develop extensive roots utilising groundwater, while being able to withstand severe sand burial. A number of studies have explored the transformation from barchan dunes or blowouts to parabolic dunes. The transformations of parabolic dunes back into highly mobile barchan dunes or transverse dunes, however, have not drawn sufficient attention.

As parabolic dunes are widely distributed around the world across a broad climatic gradient, migration rates reported in literature vary because of local environmental settings, but also due to spatio-temporal scales and measuring methods employed. A change in the dune migration rate is, therefore, not a good indicator for the state of activity of an aeolian environment. The dune transformation tendency can potentially be used as a proxy for the stability of a dune system. The integrated framework presented in this paper provides a baseline for organising and comparing different aeolian dune environments, and for predicting possible transformation scenarios under changes of environmental controls. This framework shows how the impacts of changes in sand availability, wind strength, and drought stress vary in different aeolian settings. Further research is needed to quantify the exact thresholds of different transformations in various aeolian environments. Multiregional
comparisons across the globe remain a challenge and need a wide collaboration among different research communities. The integration of real-world surveying with computer modelling may serve as a useful approach to further testing of this framework.

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