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Archean to Recent aeolian sand systems and their sedimentary record: Current understanding and future prospects

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ABSTRACT

The sedimentary record of aeolian sand systems extends from the Archean to the Quaternary, yet current understanding of aeolian sedimentary processes and product remains limited. Most preserved aeolian successions represent inland sand-sea or dunefield (erg) deposits, whereas coastal systems are primarily known from the Cenozoic. The complexity of aeolian sedimentary processes and facies

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variability are under-represented and excessively simplified in current facies models, which are not sufficiently refined to reliably account for the complexity inherent in bedform morphology and migratory behaviour, and therefore cannot be used to consistently account for and predict the nature of the preserved sedimentary record in terms of formative processes. Archean and Neoproterozoic aeolian successions remain poorly constrained. Palaeozoic ergs developed and accumulated in relation to the palaeogeographic location of land-masses and desert belts. During the Triassic, widespread desert conditions prevailed across much of Europe. During the Jurassic, extensive ergs developed in North America and gave rise to anomalously thick aeolian sucessions. Cretaceous aeolian successions are widespread in South America, Africa, Asia, and locally in Europe (Spain) and the USA. Several Eocene to Pliocene successions represent the direct precursors to present-day systems. Quaternary systems include major sand seas (ergs) in low-lattitude and mid-latitude arid regions, Pleistocene carbonate and Holocene–Modern siliciclastic coastal systems. The sedimentary record of most modern aeolian systems remains largely unknown. The majority of palaeoenvironmental reconstructions of aeolian systems envisage transverse dunes, whereas successions representing linear and star dunes remain under-recognized. Research questions that remain to be answered include: (i) what factors control the preservation potential of different types of aeolian bedforms and what are the characteristics of the deposits of different bedform types that can be used for effective reconstruction of original bedform morphology; (ii) what specific set of controlling conditions allow for sustained bedform climb versus episodic sequence accumulation and preservation; (iii) can sophisticated four-dimensional models be developed for complex patterns of spatial and temporal transition between different mechanisms of accumulation and preservation; and (iv) is it reasonable to assume that the deposits of preserved aeolian successions necessarily represent an unbiased record of the conditions that prevailed during episodes of Earth history when large-scale aeolian systems were active, or has the evidence to support the existence of other major desert basins been lost for many periods throughout Earth history?

Keywords: Aeolian, Archean, dunes, ergs, Mesozoic, Neogene, Palaeogene, Palaeozoic, preservation, Proterozoic, Quaternary.

INTRODUCTION

How can geologists best account for the preserved expression of aeolian sedimentary successions and relate such deposits to the varied set of processes responsible for their generation? "*The answer my friend is blowin' in the wind*" (Dylan, 1963), and has been for at least 3.2 billion years. The aim of this study is to present an overview of the current state of the science relating to the sedimentology of aeolian sand systems and their preserved successions. Specific objectives are as follows: (i) to demonstrate the variability and complexity of the sedimentology of recent and ancient aeolian sand systems; (ii) to show how the spatial and temporal distribution of aeolian systems and preserved successions has varied throughout Earth history; (iii) to discuss the main mechanisms for the construction, accumulation and preservation of aeolian systems; and (iv) to present some future perspectives relating to issues that currently remain unresolved in aeolian sedimentology, thereby highlighting research targets and opportunities for the future. This study is supported by a suite of complementary material arranged in a series of tables that detail many of the best-known and most representative examples of siliciclastic as well as some carbonate aeolian sand seas and coastal dunefields from the Archean and Proterozoic, Palaeozoic, Mesozoic and Cenozoic eras (see also Blakey *et al*., 1988; Tedford *et al.*, 2005; Veiga *et al.*, 2011a; Simpson *et al*., 2012). Although this work represents an attempt to compile an authoritative database of case-study examples for all periods in Earth history, many smaller and lesser-known aeolian systems have been omitted due to space limitations. The references contained in the supplementary tables of case studies (together with those references cited in the main manuscript) are contained in the supplementary file entitled 'References text and tables'.

AEOLIAN SAND SYSTEMS AND THEIR SEDIMENTARY RECORD: CURRENT UNDERSTANDING

Aeolian sand systems can be divided into inland sand sea and coastal dune systems. Inland aeolian sand seas (also known as ergs) and the aeolian dunefields present within these large-scale sediment systems comprise bedforms of different morphological types and sizes (ranging from ripples to megadunes or draas), areas of sand sheets, interdunes (including non-aeolian sediments), as well as related extradune environments of alluvial, fluvial, lacustrine, and marine affinity. Coastal dunefields likewise comprise various aeolian bedforms; many of these dune types – such as parabolic dunes – are also seen in inland systems, whereas others – such as coast-parallel dune ridges – are unique to coastal systems. Associated sediments include beach, wash-over fan, and lagonal facies.

Following Kocurek (1999) the creation of an aeolian stratigraphic record can be considered in three phases (Fig. 1): (i) sand sea (i.e. dunefield) construction; (ii) aeolian accumulation; and (iii) preservation of that accumulation. Construction of the world's largest modern, active sand seas occurs in arid regions that typically experience less than 150 mm annual precipitation, although sites of significant aeolian construction also occur in non-desert settings, especially along sandy coastlines. Although aeolian sediment transport takes place under a wide range in wind energy regimes (Fryberger, 1979), it is the directional variability of such regimes that plays a major role in determining dune type and therefore dictating the range of sedimentary structures that develop on bedforms, and the style and rate of accumulation of deposits of those bedforms (Wasson & Hyde, 1983). Many present-day actively constructing and accumulating sand seas are located at sites of relatively lower wind energy compared to upwind areas, such that sediment transport rates tend to decrease in the direction of transport, thereby encouraging sand deposition and accumulation. A down-wind reduction in sediment transport rate that leads to aeolian construction and accumulation may result from regional changes in atmospheric circulation patterns and wind regimes whereby wind speed decreases and/or directional variability increases (Wilson, 1973; Lancaster, 1999, 2013). Coastal dunefields typically develop along lowland coasts where plentiful sediment supply (often beach sand) is available for inland transport by persistent onshore winds (e.g. Klijn, 1990). The size and morphology of coastal dunes is dependent on vegetation cover, sand supply, beach-dune interaction, wind regime and coastal orientation with regards to persistent winds.

Aeolian dunefield construction (the initiation and growth of systems of bedforms) is a function of sediment supply, the availability of that supply for aeolian transport and the transport capacity of the wind (Kocurek & Lancaster, 1999). Sediment supply is the volume of sediment suitable for aeolian transport generated per unit time; supply may be contemporaneous or time-lagged (Kocurek, 1999) and can be derived from multiple sources. The proximity of a dunefield to its sediment source area is reflected in the response of the system to changes in boundary conditions. Dunefields that lie close to their sediment source (including most coastal dunefields) tend to be sensitive to variations in sediment supply, whereas systems that develop far from their ultimate source tend to be more sensitive to changes in sediment mobility or availability. Many sand seas are the depositional sinks of local to regional-scale sediment transport systems. Mineralogical and geochemical studies, aided in some instances by remote sensing data, can establish clear relations between source areas and sediment sinks (e.g. Scheidt *et al.*, 2011; Garzanti *et al.*, 2012). In many areas, however, these relations are not clear, and the source(s) of sand for major sand seas in the Sahara and elsewhere are poorly constrained (Garzanti *et al.*, 2003; Muhs, 2004). Regional wind patterns appear to show long-distance transport paths in the Sahara and Australia, but recent work also points to the importance of local sources in Australia (Pell *et al.*, 2000) and elsewhere (Muhs *et al.*, 2003). The sand in coastal dunefields is derived primarily from the beach; textural and geochemical studies of foredune deposits can give information on sediment provinces and transport pathways in the nearshore environment (Saye and Pye, 2006). Sediment availability is the susceptibility of surface grains to entrainment by the wind (Kocurek & Lancaster, 1999); stabilizing factors such as early intergranular cements (for example, gypsum), vegetation cover, coarse-grained lags, and elevated water tables all limit availability. Transport capacity is a measure of the potential sediment carrying capacity of the wind. Together these factors define the sediment system state (Kocurek & Lancaster, 1999), which can be used as a predictor of when and where episodes of aeolian construction will occur.

Following Kocurek & Havholm (1993), three princioal types of aeolian systems (Fig. 2) are recognized: (i) dry aeolian systems in which the water table and its capillary fringe are sufficiently far below the depositional surface that they have no effect on dune migration, sediment transport, and deposition; (ii) wet aeolian systems in which the water table and its capillary fringe are at or near the depositional surface, so that changes in moisture play an important role in the style and pattern of sediment accumulation (Kocurek & Havholm, 1993; Mountney, 2012), and in which interdune areas are damp or wet (flooded) and characterized by clastic, biogenic, and/or chemical sediments that are indicative of a near-surface water table; and (iii) stabilized aeolian systems in which factors such as vegetation, pedogenesis, permafrost or surface or near-surface cementation either episodically or continually act to stabilize the substrate while the system remains active overall, thereby encouraging aeolian construction and accumulation.

Aeolian accumulation to generate a body of strata requires a positive net sediment budget for which upstream sediment influx exceeds downstream outflux (Fig. 3). Special cases include aeolian accumulation in front of steep cliffs (e.g. Clemmensen *et al.*, 1997; Andreucci *et al.*, 2010a). By contrast, neutral budgets and negative budgets result in bypass and deflation (erosion), respectively. The positive net sediment budget required for aeolian accumulation needs either a downstream spatial decrease in the transport rate in response to airflow deceleration, or a temporal decrease in flow concentration in response to a reduction in dune size over time (Rubin & Hunter, 1982; Kocurek & Havholm, 1993). One commonly recognised mechanism for the accumulation of migrating dunes and draas (mega-bedforms) is via bedform climbing, whereby the angle of climb (which for large bedforms might typically be only a few tenths of a degree) is determined by the ratio between the rate of downwind bedform migration and the rate of rise of the accumulation surface (Fig. 3). Climb at low angles means that only the basal parts of large bedforms typically accumulate to generate cross-stratified sets (Fig. 3). Nevertheless, accumulated, vertically-stacked, cross-stratified sets recording the passage of multiple large bedforms are commonly each in excess of 10 m in thickness and some can attain thicknesses of >30 m (e.g. Mountney & Howell, 2000). The accumulation of sets via climbing and their composition of only the basal-most parts of the original bedforms from which they were constructed means that ancient aeolian accumulations are biased representations of original aeolian systems because they are composed of assemblages of lithofacies arranged into architectural elements that typically record only those processes that operated on the lowermost flanks of the original bedforms; such processes typically differ from those that operated on the higher parts of bedforms (e.g. Eastwood *et al.*, 2012).

Interdune migration bounding surfaces separate packages of strata that represent the accumulated deposits of successive migrating aeolian dunes and adjoining interdunes; superimposition bounding surfaces record the style of juxtaposition of smaller dunes on larger draas, and the style of migration of the smaller forms over the larger forms; reactivation surfaces record episodic changes in dune or draa lee-slope configuration, including temporal changes in steepness or orientation (Rubin, 1987;

Rubin & Carter, 2006). These bounding-surface types define and delineate architectural elements comprising packages of aeolian dune and interdune strata that are themselves composed internally of various arrangements of lithofacies (Brookfield, 1977; Chrintz & Clemmensen, 1993; Fryberger, 1993; Kocurek, 1991). The geometry and arrangement of these architectural elements are determined by: (i) the scale and morphology of the original dunes and interdunes; (ii) the style of migration of the dunes and interdunes over both time and space; and (iii) the style of accumulation, which in many systems is controlled by the angle of climb (e.g. Mountney & Thompson, 2002), although in other systems is known to be controlled by other mechanisms, including the infilling of local accommodation space between older remnant dunes (e.g. Langford *et al.*, 2008).

When and where the net sediment budget switches from positive to neutral or negative, aeolian accumulation ceases and bypass and deflation commence, respectively. Both bypass and deflation result in the generation of supersurfaces (Kocurek, 1988) that cap underlying accumulations. Such accumulations define aeolian sequences and their bounding supersurfaces can be considered sequence boundaries (Fig. 4). Deflation operates either until the net sediment flux becomes neutral or positive again, or until it progresses down to the water table (Stokes, 1968), which limits further deflation. Supersurfaces of allogenic origin tend to be regional in extent and truncate other bounding surface-types of autogenic origin, which themselves arise as a consequence of interdune migration and climb, bedform superimpositioning or bedform reactivation (Brookfield, 1977, Rubin, 1987; Mountney, 2006a). Some supersurfaces have been correlated laterally into adjoining non-aeolian environments where they merge into, for example, transgressive marine units (Havholm *et al.*, 1993; Blakey, 1996; Blakey *et al.*, 1996; Rodríguez-López *et al.*, 2013). Many supersurfaces that bound episodes of aeolian accumulation are paraconformities (diastems) considered to represent long-lived hiatuses in accumulation: supersurfaces with associated sedimentary features such as large and closely-spaced tree-size rhizoliths may take 10^4 to 10^5 years to form (Loope, 1985). Several authors have proposed that aeolian supersurface generation may occur as a result of Milankovitch-style orbital forcing operating with periodicities of 18 to 400 Kyr (Loope, 1985; Clemmensen *et al.*, 1994; Jordan & Mountney, 2010, 2012; Mountney, 2006b; Rodríguez-López *et al.*, 2012a). For many systems, the amount of time represented by aeolian accumulations probably is significantly less than that represented by intervening supersurfaces (e.g. Loope, 1985); thus, many preserved aeolian successions probably represent only a fraction of the geological time over which the aeolian systems were active, and the preserved record is therefore highly fragmentary and potentially biased toward a specific set of formative processes.

Long-term preservation of aeolian accumulations in the ancient record requires that the body of strata is placed below some regional baseline, beneath which erosion does not occur (Kocurek & Havholm, 1993). Thus, the rate of generation of accommodation space and the rate at which aeolian accumulations fill that space is a fundamental control on preserved architectural style (e.g. Howell & Mountney, 1997).

Approaches to the theoretical modelling of aeolian dune and interdune successions commenced with the development of purely qualitative depositional models for aeolian systems, many such examples of which were devised in the 1970s; commonly recognized packages of aeolian dune and interdune lithofacies were shown to occur as elements delineated by bounding surfaces (e.g. Brookfield, 1977). These models, which typically accounted for stratigraphic complexity in two spatial dimensions are so-called *static* aeolian depositional (or facies) models (Mountney, 2006a). One forward stratigraphic modelling approach to account for both spatial and temporal changes in aeolian architecture has led to the establishment of a conceptual framework for the classification of aeolian systems and their accumulated successions (Mountney, 2012). This framework identifies simple, *static* system architectures, that are generated by spatially and temporally invariable controls but additionally identifies and models *dynamic* system architectures in which spatial and temporal changes in dune morphology, scale and style of migration and accumulation (for example, angle of climb) give rise to more complex preserved architectures.

EVOLUTION OF AEOLIAN SAND SYSTEMS THROUGH EARTH HISTORY

Archean and Proterozoic aeolian sand systems (Table S1)

The oldest known aeolian system is the 3.2 to 3.0 Ga Lower Moodies Group of the Swaziland Supergroup in South Africa (Table S1), which accumulated in a series of intramontane extensional basins through which simple barchan dunes migrated (Simpson *et al.*, 2012). The palaeogeographical distribution of Archaean and Proterozoic aeolian systems was dictated by the worldwide geographical distribution of Archaean and Proterozoic cratons and, due to their extreme age, the preserved global record of aeolian systems from these Eons is highly fragmentary. Palaeoproterozoic aeolian systems accumulated in intracratonic sag basins, intracratonic and intercontinental extensional rift basins (both during rift and thermal phases of basin evolution), and transtensional basins (e.g. Rainbird *et al*., 2003; Master *et al*. 2010). Mesoproterozoic aeolian systems accumulated in intracratonic and intramontane basins, rift basins and transpressive strike-slip basins (e.g. Clemmensen, 1988; Martin & Thornea, 2002). Neoproterozoic aeolian systems accumulated in intracratonic rift basins (e.g. Grey *et al*. 2005; Sarkar *et al*., 2011). From this, it is clear that the majority of preserved Precambrian aeolian systems are synrift depositional systems in which preservation of aeolian deposits occurred during the rifting phases of supercontinents because of the associated increase in accommodation space (Eriksson & Simpson, 1998).

Pre-vegetation Archean and Proterozoic periods were not subject to palaeoenvironmental conditions that were especially well-suited to aeolian sediment accumulation and preservation (Eriksson & Simpson, 1998). The absence of vegetation as stabilizing agent favoured aqueous-reworking of pre-existing aeolian deposits, leading to their partial or total destruction, reworking and incorporation into a variety of coeval sedimentary environments (e.g. Tirsgaard & Øxnevad, 1998).

Proterozoic aeolian systems developed in association with a variety of coeval depositional systems and lead to fluvial–aeolian interactions, such as those from the Palaeoproterozoic Makgaben Formation (Waterberg Supergroup, South Africa, Eriksson *et al*., 2000 and Simpson *et al*., 2002), Amarook Formation (Wharton Group, Canada, Rainbird & Hadlari, 2000 and Rainbird *et al*., 2003) and Thelon Formation (Barrensland Group, Canada, Rainbird *et al*., 2003). Marine reworking of Precambrian aeolian deposits was apparently a widely-occurring process along Proterozoic coastlines; documented examples include the Palaeoproterozoic Quartzite Member, Mcheka Formation, Zimbabwe (Master *et al*., 2010), the Whitworth Formation, Haslingden Group, Australia (Simpson & Eriksson, 1993) and the Neoproterozoic Venkatpur Sandstone, India (Chakraborty, 1991).

Comparison of Proterozoic and Phanerozoic erg systems reveals a general trend toward the preservation of more complex aeolian systems during the Proterozoic. The general atmospheric circulation pattern influenced by palaeogeographic changes, palaeoland mass distributions and associated orogenic buildups, and the particular properties and characteristics of Archaean and Proterozoic atmospheres could have had a different effect on aeolian transport compared to equivalent processes that operated during the Phanerozoic. Studies of Han *et al.* (2014) have demonstrated that, for a particular wind speed, the ability of the air flow to transport sand decreases with lower air density; however, under the same conditions the saltation height increases. Taking into account that recent works like that of Som *et al.* (2012) have concluded that the density of the 2.7 Ga atmosphere was less than twice modern levels, it is possible that changes in air density over geological timescales could have influenced aeolian transport mechanisms, and this might be recorded by the predominance of different aeolian bedforms at different times. For example, documented examples of cosets of strata interpreted to represent the preserved accumulations of draa-scale bedforms are numerous for Precambrian successions. Furthermore, the preservation of very-coarse-grained (siliciclastic) Precambrian aeolian successions (for example, the Egalapemta Member, Mesoproterozoic, India; Biswas, 2005) is noteworthy. The specific dynamic configuration of the Precambrian atmosphere and its interaction with sediment grains could explain the occurrence of simple but giant transverse dunes with maximum preserved set thicknesses (more than 50 m thick) as the single aeolian dune cross-bedded set recorded from the Late Neoproterozoic McFadden Formation (Western Australia; Grey *et al*., 2005).

The absence of Phanerozoic cold (periglacial) aeolian dunefields compared to their occurrence in Precambrian times is noteable. The Neoproterozoic Bakoye 3 Formation, Bakoye Group from Mali

(Deynoux *et al.*, 1989) and the Neoproterozoic Whyalla Sandstone from Australia (Williams, 1998) constitute the only two examples on Pre-Cenozoic periglacial dunefields. This scarcity of periglacial dunefields in the fossil record could be a result of misinterpretation of the particular palaeoclimate setting in which some Precambrian and Phanerozoic aeolian systems formed. A re-evaluation of this topic is needed. It is known that glacial latitudes have changed through time (see Evans, 2003). It should be considered the possibility that cold deserts could have formed associated with glacier fronts even during the Phanerozoic. Particular attention should be paid to the latitudinal variation of the equilibrium-line altitude (ELA) as a control of glaciation through time (e.g. Isbell *et al.*, 2012).

Paleozoic aeolian sand systems (Table S2)

During the Cambrian, ergs were located in the main land masses in the Southern Hemisphere (Fig. 5). The Backbone Ranges Formation (Mackenzie Mountains, Canada; MacNaughton *et al.*, 1997) and the Wonewoc Formation (Wisconsin and Minnesota, USA; Dott *et al.*, 1986; Runkel *et al.*, 1998) accumulated in southern Laurentia where interaction between ergs and coastal systems adjacent to the Iapetus Ocean occurred (Fig. 5). The Amin Formation and the Lower Haima Group in Oman (Milson *et al.*, 1996) and the Lower Roan Formation in Zambia (Annels, 1989) accumulated as inland sandy deserts in Gondwana (Fig. 5). Many Cambrian ergs accumulated under the influence of easterly trade winds, between the Equator and 30º palaeolatitude (Fig. 5; see Dott *et al.*, 1986).

Cambro-Ordovician and Ordovician ergs were less widespread than those of Cambrian age (Fig. 5). Such aeolian systems again developed under the influence of active trade winds of subtropical high-pressure systems in land masses of the southern palaeo-hemisphere in Laurentia and Gondwana (Fig. 5) (for example, the Pedra Pintada Formation/Alloformation; Paim & Scherer, 2007; Almeida *et al.*, 2009). In particular, aeolian dunefields developed in southern Laurentia record marine–aeolian interactions characterized by complex associations of facies of both aeolian and aqueous origin, as is the case of the Cambro-Ordovician Nepean Formation (Postdam Group, Canada and USA; Malhame, 2007) and the Ordovician St. Peter Sandstone (Minnesota and Wisconsin, USA; Dott *et al.* 1986). In particular, the St. Peter erg succession records palaeowinds that are in agreement with more general reconstructions of the southern palaeo-trade wind belt (Dott *et al*. 1986).

Silurian and Siluro-Devonian aeolian successions are few in number. The main preserved systems accumulated in Western (Perth-Carnarvon Basin) (Trewin & Fallick, 2000) and Central (Amadeus Basins) Australia (Fig. 5) (Shaw *et al.,* 1991). The Swanshaw Sandstone Formation of Scotland constitutes a mixed aeolian–fluvial succession developed in the transtensional Lanark Basin (Smith *et al*., 2006). These Silurian and Siluro-Devonian systems developed in the southern palaeo-hemisphere desert belt.

During the Devonian, the assembly of Laurussia in response to the final stages of the plate collisions of the Caledonian Orogeny, and the northward migration of Gondwana, led to an increase of land masses present at subtropical latitudes that were subject to the influence of the southern palaeo-hemisphere desert belt (Fig. 5). This palaeogeographical configuration enabled the construction, accumulation and preservation of several major aeolian dunefield systems (Fig. 5).

The most representative Devonian aeolian systems are those forming part of the Old Red Sandstone of North West Europe (e.g. Browne *et al*., 2002; Morrisey *et al*., 2012) which accumulated in extensional basins formed as a result of the collapse of the over-thickened crustal belt resulting from Caledonian compressional tectonics (McClay *et al*., 1986). The Old Red Sandstone exhibits a variety of aeolian facies, many of which record wind–water interaction processes (for example, the Middle Devonian Yesnaby Sandstone Group, Lower Old Red Sandstone Supergroup, Scotland, Trewin & Thirlwall, 2002).

Devonian ergs are characterized by a variety of aeolian facies including aeolian sandsheet successions (for example, the Lower Clair Group, Clair Basin, UK; Nichols, 2005), aeolian dunefield successions composed of transverse dune deposits (for example, the Slieve Mish Group, Ireland; Horne, 1971), barchanoid dune deposits (for example, the Devonian of Scotland; Allen & Marshall, 1981) and draa deposits (for example, Kilmurry Sandstone Formation, Ireland, Dodd, 1986; Eday Sandstone, Eday Group, Scotland, Marshall *et al*., 1996). Devonian aeolian dunefield successions with draa and barchanoid dune deposits are preserved in north-east Greenland and these demonstrate aeolian interaction with ephemeral streams and terminal fans (see Olsen & Larsen, 1993). Other Devonian aeolian systems have been recorded from Antarctica (New Mountain Sandstone; Gilmer, 2008) and Australia (for example, the Langra Formation, Jones, 1972; the Tandalgoo Sandstone, Thornton, 1990).

The majority of Carboniferous aeolian systems are Pennsylvanian in age, with several spanning the Pennsylvanian–Permian boundary. Some examples of early Carboniferous aeolian systems include the Devonian to Mid-Carboniferous Khusayyayn Formation in Saudi Arabia (Stump & Van der Eem, 1995), the recently recognized aeolian systems of the Late Mississippian Loyalhanna Member, Pennsylvania, USA (the Mauch Chunk Formation and Appalachian Formation; Swezey *et al*., 2012) and the Devonian–Mississippian Harder Bjerg Formation in Greenland (Olsen & Larsen, 1993; Fig. 6).

During the Variscan Orogeny, Gondwana and Laurussia collided creating the Supercontinent Pangaea; Pennsylvanian to Permian aeolian systems were constructed in both the Northern and Southern Hemispheres (Fig. 6). The main systems crop out in North and South America, with well-developed examples including the Early Pennsylvanian Juruá Sandstone Formation (the Solimões Basin, Brazil; Elias *et al*. 2007), the Pennsylvanian Tyrwhitt and Tobermory Sandstone Formations (the Rocky Mountain Supergroup, Canada; Stewart & Walker, 1980) and the Pennsylvanian–Middle Permian Cangapi Formation (the Cuevo Group, Tarija Basin, Bolivia and Argentina; Hernández & Echevarría 2009), the Late Carboniferous–Early Permian Patquía Formation (the Paganzo Group, Paganzo Basin, Argentina; Caselli & Limarino, 2002; Geuna *et al*., 2010). In Saudi Arabia, the Carboniferous–Permian Unayzah Formation constitutes an economically important gas reservoir succession (Melvin & Haine, 2004; Melvin *et al*., 2010).

The best-known Pennsylvanian–Permian ergs are reported from the USA. The ‘Tensleep Complex’ (the Tensleep Sandstone, Casper Formation, Quadrant Sandstone; e.g. Peterson, 1988), the Honaker Trial Formation (e.g. Wiliams, 2009) and the Cutler Group, including the lower Cutler beds (e.g.

Jordan & Mountney, 2010, 2012; Wakefield & Mounteny 2013), the Rico Formation (e.g. Loope, 1985; Chan & Kocurek, 1988) and the Weber Sandstone (e.g. Doe & Dott, 1980; Driese, 1985) are all successions that exhibit well-exposed examples of central-erg and erg-margin systems (see Blakey *et al*., 1988 for further examples).

During the Permian, the construction of extensive erg systems across large parts of Pangaea was favoured by the location of land masses of this supercontinent in subtropical latitudes under the influence of the southern and northern palaeo-hemisphere desert zones (Fig. 6). The best-known Permian aeolian systems are the Rotliegend Group (Rotliegendes) of the North Sea and North-west Europe, and the Permian aeolian systems from the USA (Fig. 6). Major reserves of gas (and some oil) exist in Permian Rotliegend desert sandstone hydrocarbon reservoirs of North-west Europe, in particular in the Southern Permian Basin of the North Sea and some localities of the Northern Permian Basin (Glennie, 1970; 1972; 1998; Glennie & Buller, 1983).

In the Southern Permian Basin of the North Sea, aeolian dune deposits accumulated between wadi channels originating from the Variscan Highlands and the extensive sabkha and desert lake located southwards of the Ringkøbing-Fyn High (Glennie, 1972). Reconstructed aeolian dune types of the Rotliegend Group include transverse-crescentic dunes, barchans, longitudinal/linear and star dunes (for example, the Brodick Beds, Clemmensen & Abrahamsen, 1983; the Leman Sandstone Formation, Sweet, 1999; the Penrith Sandstone, Turner *et al*., 1995; Lovell *et al*., 2006), as well as interdraa, draa-plinth and draa-centre deposits (for example, Yellow Sands, Clemmensen, 1989; Chrintz & Clemmensen, 1993). Complex wind patterns resulted in the construction of barchanoid draa with superimposed oblique crescentic and linear dunes (for example, the Bridgnorth Sandstone Formation, UK, Steele, 1981; Benton *et al*., 2002).

Permian sedimentary basins of the USA contain extensive and complex aeolian depositional systems and record a variety of facies and processes (see Blakey *et al*., 1988 for compilation). Examples of these Permian aeolian units include the Schnebly Sandstone Formation, which comprises deposits of an aeolian dunefield associated with evaporite and carbonate deposits (Blakey & Middleton, 1983;

Blakey, 1990), and the Lyons Sandstone Formation with deposits of parabolic dunes and blowout-type interdunes (McKee, 1979).

Several Permian units in the USA preserve complete examples of central-erg sequences that demonstrate evidence for a complex merging relation with marine erg-margin systems. Examples include the White Rim Sandstone (e.g. Chan, 1989; Tewes & Loope, 1992, Kamola & Huntoon, 1994), the De Chelly Sandstone (e.g. Blakey, 1990; Stanesco, 1991), the Yeso Formation (Mack & Dinterman, 2002) and the Upper Minnelusa Formation (e.g. Fryberger, 1984, 1993).

The Permian Coconino Sandstone constitutes the accumulation of an inland dry erg system formed by climbing barchans or barchanoid-ridge and transverse dunes (e.g. Blakey & Middleton, 1983; Blakey, 1990; 1996). Some Permian ergs display examples of fluvial systems reworking the aeolian sands; this is the case for the Rush Springs Sandstone (the Whitehorse Group, Kocurek & Kirkland, 1998; Poland & Simms, 2012), the Cedar Mesa Sandstone of the Paradox foreland basin (Langford & Chan, 1988, 1989; Mountney & Jagger, 2004; Mountney, 2006a; Langford *et al*., 2008) and the overlying Organ Rock Formation (Cain & Mountney, 2009, 2011).

Permian aeolian systems developed in the Southern Hemisphere in Pangaea include the Pirambóia Formation in Brazil (Paraná Basin, Dias & Scherer, 2008), the Buena Vista Formation in Uruguay (Northern Uruguayan Basin, Goso *et al*., 2001) and the Permian aeolian systems of Argentina from the retroarc Paganzo Basin (for example, the Andapaico Formation and the De la Cuesta Formation, Spalleti *et al*., 2010; Correa *et al*., 2012).

Mesozoic aeolian sand systems (Table S3)

Throughout much of the Triassic, widespread aeolian desert and semi-desert conditions prevailed across much of northern Pangaea. The majority of Triassic ergs were located in equatorial to mid-latitudes in the Northern Hemisphere and most of these appear to be aligned following a north–south trend close to the eastern margin of Northern Gondwana (Fig. 7). Triassic erg systems of north-eastern

Pangaea include the Buntsandstein of Europe, which is characterized by a thick accumulation of red beds that record a variety of aeolian and mixed aeolian–fluvial–lacustrine successions (e.g. Clemmensen, 1985; Mader, 1985a, Mader & Laming, 1985; Tietze *et al*., 1997). The Triassic Buntsandstein facies in the north-eastern Iberian Chain (central eastern Spain), previously considered to be fluvial in origin, is now known to contain an evolving erg system (Soria *et al.*, 2011), which comprises a succession that records the transition from a wadi belt, via an inner erg-margin, to a central-erg system.

The equivalent lithostratigraphic unit to the Buntsandstein in the UK and Ireland is the Sherwood Sandstone Group (‘New Red Sandstone’) which is present in a series of rift basins in both onshore and offshore settings (Fig. 7; Brookfield, 2004, 2008; Tyrrell *et al*., 2009). Aeolian dunefields were mostly characterized by bedforms of modest size, many with damp or wet interdunes controlled by water table, as recorded, for example, by the Wilmslow Formation (Øxnevad, 1991; Bloomfield *et al*., 2006) and the Helsby Formation (Bloomfield *et al*., 2006; Mountney & Thompson, 2002) of the Cheshire Basin. In Scotland, the Hopeman Sandstone probably straddles the Permian–Triassic boundary and is characterized by deposits of the preserved remnants of a series of star dune and draa bedforms representing a small fragment of what is inferred to have been a very extensive dry aeolian system (Clemmensen, 1987; Glennie & Hurst, 2007; Hurst & Glennie, 2008). In the subsurface of the East Irish Sea Basin, Triassic aeolian deposits form important reservoirs for gas (Cowan & Boycott-Brown, 2003; Meadows, 2006).

Triassic ergs contructed close to the palaeo-equator (for example, the Oukaimeden Sandstone Formation, Morocco) record small aeolian dunes developed on floodplains of ephemeral fluvial systems (Fabuel-Perez *et al*., 2009; Mader & Redfern, 2011). Other Triassic ergs are located close to the western margin of Northern Pangaea and examples include the Nugget Sandstone of Utah and Wyoming (Fig. 7) (Sprinkel *et al*., 2011) which may be at least in part of Lower Jurassic age.

Thick and geographically widespread Jurassic aeolian desert erg successions of the Colorado Plateau region are extensively documented and are arguably the most intensely studied of all aeolian successions. Many authors have considered these successions collectively in terms of the regional palaeogeographic, palaeoclimatic and palaeotectonic setting (Fig. 7; e.g. Kocurek & Dott, 1983; Blakey *et al*., 1988; Dickinson & Gehrels, 2003; Loope *et al*., 2004). The best-known Jurassic aeolian successions of the south-western United States include the Wingate and Navajo sandstones (and stratigraphic equivalents) of the Lower Jurassic Glen Canyon Group, and the Page and Entrada sandstones (and equivalents) of the Middle Jurassic San Rafael Group. The Wingate Sandstone represents a largely dry aeolian system representative of an erg-centre setting (with compound draa development) but also demonstrates styles of interaction with fluvial deposits of the Moenave Formation in its erg-margin setting (Clemmensen & Blakey, 1989; Clemmensen *et al*., 1989; Tanner & Lucas, 2007). The Navajo Sandstone of the Glen Canyon Group is one of the most intensely studied sedimentary formations of any type and is well-exposed across much of the Colorado Plateau region, where it attains a thickness of nearly 700 m in south-western Utah. The succession represents the preserved remnant of a giant erg that was present across much of the western part of Pangaea (Fig. 7; Hunter & Rubin, 1983; Chan & Archer, 2000); this system was subject to seasonal wind reversals associated with annual monsoons that occurred each summer when more humid and cooler conditions prevailed and wind reversal occurred (Loope & Rowe, 2003; Loope *et al.*, 2008). The Page Sandstone of Arizona and southern Utah represents accumulation in an erg system close to the margin of an interior seaway; the system is composed of vertically stacked, progradational erg sequences that overlie marine deposits of the Carmel Formation such that the two units intertongue (Havholm & Kocurek, 1994; Jones & Blakey, 1997; Dickinson *et al*., 2010). The Entrada Sandstone of the San Rafael Group is exposed extensively across much of the Colorado Plateau region and represents the accumulated deposits of a coastal to inland aeolian system that was characterized by a complex arrangement of aeolian dune, damp and wet interdune, and sabkha elements (Kocurek, 1980, 1981a,b; Crabaugh & Kocurek, 1993; Crabaugh & Kocurek, 1998). Relic dune topography is preserved in places at the top of the succession as a result of later marine transgression (Benan & Kocurek, 2000).

The Upper Jurassic Norphlet Sandstone represents the accumulated deposits of a major aeolian erg succession that is known principally from the subsurface of Alabama, the shallow-water Gulf of

Mexico around Mobile Bay, and further offshore in the deep-water part of the Gulf of Mexico, where it forms a major oil reservoir (Taylor *et al*., 2004; Mankiewicz *et al*., 2009; Ajdukiewicz *et al*., 2010). Numerous aeolian successions of Jurassic age are documented from South America and especially from Brazil. Examples include the Pedreira Sandstone of Paraná Basin in Brazil, which is characterized by climbing aeolian dune sets with intervening damp and wet interdune units (Nowatzki & Kern, 2000) and the Guará Formation of southern Brazil, which records composite crescentic aeolian dune sets and cosets, and aeolian sand-sheet elements interbedded with distal flood deposits and fluvial channel-fill elements (Scherer & Lavina, 2006).

The backarc Neuquén basin of Argentina records a series of Jurassic aeolian successions. The Lotena Formation preserves a record of aeolian–fluvial interactions (Veiga *et al.*, 2011a). A tectonic inversion during the Late Jurassic led to the desiccation of the entire basin, giving rise to a complex array of continental facies for which aeolian deposits form a major part (Spalletti & Veiga, 2007, Spalletti *et al.*, 2011). In the southern part of the basin, an upward vertical transition from fluvial-dominated to aeolian-dominated deposition, probably arising from a climatic shift to drier conditions, is recorded as part of the Kimmeridgian Quebrada del Sapo and Tordillo formations (Zavala *et al.*, 2005a; Veiga & Spalletti, 2007). The Tordillo Formation represents migration and accumulation of transverse and barchan dunes in a style that generated a complex hierarchy of internal bounding surfaces within a largely dry aeolian system in which only thin dry interdune elements accumulated (Zavala *et al.*, 2005a). To the east, aeolian accumulation was more significant and led to the preservation of *ca* 300 m thick sequence of mainly aeolian deposits of the Sierras Blancas and Catriel formations, which include deposits of dune, wet and dry interdune and aeolian sandsheet elements (Maretto *et al.*, 2002; Spellati *et al.*, 2011).The Piramboia Formation of Entre Rios Province, Argentina, is an aeolian unit considered to be primarily of Lower Jurassic age (Silva & Fernandez, 2004).

The Lower Jurassic Clarens Formation, which forms a unit of the Karoo Supergroup in South Africa records a progressive upward transition from the deposits of a wet aeolian system that developed alongside coeval ephemeral fluvial systems to a dry aeolian system dominated by stacked cross-bedded aeolian dune sets (Bordy & Catuneanu, 2002; Holzförster, 2007).

Cretaceous aeolian successions, together with those that probably span the Jurassic–Cretaceous boundary, are numerous in South America and many have been the focus of detailed study over several decades. The Botucatu Formation of the Paraná Basin (São Paulo and Paraná states, Brazil) – which was originally thought to be Triassic in age (Bigarella, 1979a) – has lateral equivalents in Parnaiba Basin of northern Brazil and is a near-equivalent of the Bauru, Guará, Sambaiba, Sanga do Cabral and Piramboia formations, as well as of the Etjo Sandstone in Namibia (Mountney *et al.*, 1999 a,b), and possibly the Kudu Formation, offshore Namibia (Mello *et al.*, 2011). Relic aeolian dune forms and degraded topography are preserved at the top of the succession where it is overlain by flood basalts of the Serra Geral Formation and other flood basalts related to the Etendeka-Paraná Large Igneous Province (Scherer, 2002; Waichel *et al.*, 2008). The Serra Geral Formation records the exceptional preservation of relic aeolian dune topography of a dry aeolian system by flood basalts including various types of completely preserved dunes and sand-deformation features, including sand diapirs and peperite-like breccia.

In the backarc Neuquén Basin of west-central Argentina, Lower Cretaceous sandy aeolian accumulations are numerous and mainly related to lowstand periods and to the possible disconnection of the basin from the proto-Pacific Ocean (Howell *et al.*, 2005). These successions constitute important conventional oil and gas reservoirs. Aeolian deposits have been described from part of the proximal system of the Valanginian Mulichinco Formation in the subsurface of the basin (Zavala *et al.*, 2005b) and, more marginally, as part of environments of fluvial–aeolian interaction (Schwarz *et al.*, 2011). One of the best-described aeolian systems in the basin is the Hauterivian Avilé Member of the Agrio Formation (Rossi, 2001; Veiga *et al.*, 2011b). Within this non-marine unit, aeolian deposits are locally important and record a complex vertical evolution related to high-frequency climatic changes and to the development of multiple supersurfaces associated with aeolian deflation and fluvial flooding (Veiga *et al.*, 2002). Finally, the Baremian Lower Troncoso Member of the Huitrín Formation is characterized by the transition from fluvial to aeolian deposits (Veiga *et al.*, 2005). For both the Avilé and Troncoso members, marine inundation of the dunefields following transgression led to the preservation of relic dune topography, as well as to the development of a complex set of facies related to the deformation and reworking of the aeolian sands during the transgression (Strömbäck *et al.*, 2005; Veiga *et al.*, 2011b). Aeolian deposits have also been described in the Upper Cretaceous record of the Neuquén Basin as part of the Neuquén Group (Sánchez *et al.*, 2008).

In Africa, the Lower Cretaceous Etjo Sandstone Formation of north-west Namibia is a predominantly dry aeolian system in which relic aeolian dune bedforms with up to 100 m of topographic relief have been preserved following inundation by flood basalts of the Etendeka igneous province (Mountney *et al*., 1999a, b; Howell & Mountney, 2001; Mountney & Howell, 2000). The lower part of the succession records exceptionally thick examples of simple cross-bedded sets of aeolian dune origin (individual simple sets up to 52 m thick), with preservation probably having been enabled by the migration of a large dune into a pre-existing topographic depression. Aeolian sandstone occurs interleaved with flood basalts at multiple levels within the upper part of the succession, which forms the lower part of the overlying Etendeka Group Large Igneous Province (Jerram *et al*., 1999a, b; Jerram *et al.*, 2000a, b).

Various formations composed of Cretaceous strata of aeolian dune origin are present in several basins of China, including many in the Gobi Desert region of Inner Mongolia. Cretaceous aeolian dune accumulations are recorded from the Sichuan, Ordos, Kuche, Tarim Basin and the Kuqa basins, and especially in Inner Mongolia and surrounding regions. The preserved aeolian dunefield deposits preserve evidence for the development of both dry and wet (water-table controlled) aeolian systems (Xie *et al*., 2005; Jiang *et al*., 2008). The Upper Cretaceous (Campanian) Djadokhta Formation of the Ulan Nur Basin and the area around Tugrikiin Shiree and Ukhaa Tolgod (Nemegt Basin, Mongolia) represent dunefields that experienced heavy rainfall events resulting in the development of perched water tables, early calcite cementation, and dune collapse due to sediment gravity sliding (Jerzykiewicz *et al*., 1993; Loope *et al*., 1999; Seike *et al*., 2010).

In Europe (Spain), the mid-Cretaceous Iberian Desert System, represented by the Utrillas Group (that includes the previously known middle and upper parts of the Escucha Formation and the whole Utrillas Formation) developed from the early Albian to the early Cenomanian along the western Tethyan margin (Iberian Basin, eastern margin of Iberia) between the Tethys Ocean (to the east) and the highland Variscan Iberian Massif (to the west) over an area of more than 20,000 km^2 (Fig. 7; Rodríguez-López, 2008; Rodríguez-López *et al.*, 2008). The mid-Cretaceous Iberian Desert System displays a tripartite spatial configuration: a back-erg characterized by aeolian–fluvial (wadi) interactions, a central-erg characterized by thick accumulation of linear draa, other compound-draa sandstones and desert roses, and a fore-erg in which the interaction between compound aeolian dunes (draas) and coastal sedimentary environments (lagoons, tidal creeks, tidal deltas and marshes) occurred (Rodríguez-López *et al.* 2006; 2008; 2010; 2012a). The sedimentary record of this desert basin displays different erg sequences bounded by supersurfaces (Rodríguez-López *et al.*, 2013).

Palaeogene aeolian sand systems (Table S4)

Palaeogene erg systems have been the subject of only relatively modest investigation, mainly as part of regional studies; it is therefore difficult to draw conclusions regarding their distribution and development. Only one example of aeolian accumulation has been described for the Palaeogene of Europe (Fig. 8) and this corresponds to the Sables de Fontainebleu Formation (or Fontainebleau Formation) of early Oligocene age (Alimen, 1936). This unit is part of the fill of the Paris Basin and is composed of a 50 to 70 m thick succession of clean, fine-grained, well-sorted sand arranged in accumulations expressed as elongated ridges and is thought to represent the preserved topography of an ancient coastal barrier system (Thiry *et al*., 1988; Cojan & Thiry, 1992).

Palaeogene aeolian systems of North America are restricted to those of the Oligocene of western USA. The most important example corresponds to the 'Chuska Erg', an extensive sand sea (*ca* 140,000 km^2) developed in the uplifted Colorado Plateau between 33.5 Ma and 27 Ma (Lucas & Cather, 2003; Cather *et al*., 2008). The accumulated record of this sand sea (known as the Nabora Pass Member of the Chuska Sandstone) attains a maximum thickness of 535 m and records the northerly migration of transverse dunes (Cathers *et al.*, 2008). Also in the western USA, in the Great

Plains of South Dakota, Nebraska and Wyoming, volcaniclastic aeolian deposits have been described as part of the Brule Formation of the White River Group (Tedford *et al.*, 2005).

Few aeolian successions have been described for the Palaeogene of Africa and these are mainly of Middle to Upper Eocene age. The Hadida Formation, which developed in the Tindoouf-Ouarzazate Basin of Morocco, includes medium-grained, cross-bedded sandstone that occurs intercalated in a >300 m thick sequence mainly composed of gypsiferous mudstones (Swezey, 2009; Tesón *et al.*, 2010); this succession is regarded as the earliest record of the Saharan system (Swezey, 2006). Apart from these proto-Saharan deposits, sandy aeolian successions have been described as part of the Palaeocene fill of the Congo Basin in the West African margin of Gabon in the form of deposits originally described as the '*Gres Polymorphes*' (De Ploey *et al.*, 1968) that comprise a 180 m thick succession of cross-stratified sandstones with individual sets several metres thick interpreted as aeolian deposits (Batéké Sands, Séranne *et al.*, 2008). Aeolian deposits have also been described in the Fayum District in Egypt as part of the Qasr El-Sagha Formation (El-Fawal *et al.*, 2011). These include a 45 m thick succession of Middle to Upper Eocene age, previously described as channelized delta plain deposits (Bown & Kraus, 1988) but more recently reinterpreted as part of a desertification phase that caps a prograding deltaic system (El-Fawal *et al.*, 2011).

The only example of Cenozoic aeolian accumulation from Oceania comes from southern Australia, where the upper portion of the Middle to Upper Eocene Ooldea Sand (Barton Sand) represents a barrier dune complex developed during the transgression of the Eucla Basin (Hou *et al.*, 2006).

Neogene aeolian sand systems (Table S4)

Neogene sandy aeolian systems are relatively common and have been described worldwide. Most of these systems owe their origin to local climatic and tectonic factors. However, as the position of most major continental landmasses has not changed significantly since the Miocene, many Neogene aeolian systems have apparently been controlled by climatic conditions similar to those experienced by present-day desert systems, and such successions therefore constitute the precursors of some of the most important Quaternary aeolian systems, as in the Sahara, Kalahari and Namib sand seas.

Only one example of Miocene aeolian accumulation in Europe has been identified. It corresponds to the Vale de Chelas Sands in the Lower Tejo Basin of Western Iberia, a 10 m thick succession of cross-bedded sandstones of aeolian origin, interpreted as a coastal system (Telles Antunes *et al.*, 1999; Pais *et al.*, 2012). Pliocene aeolian deposits in Europe have been described in more detail and are mainly related to the onset of the Northern Hemisphere glaciation that led to stronger westerly winds. In central Spain, the Middle to Upper Pliocene Escorihuela Formation records the accumulation and preservation of an aeolian dunefield related to syn-sedimentary activity of normal faults in a syn-rift environment. This system is dominated by the interaction between constructive and destructive episodes related to high-frequency climatic changes (Rodríguez-López *et al.*, 2012b). Accumulation during the Middle to Upper Pliocene is also recorded in the northern Apennines of Italy, and is associated with extensional tectonics and the alternation between relatively more humid and more arid episodes in the Valdarno Basin (Ghinassi *et al.*, 2004). Here, deposits of the Rena Blanca Sand Unit are dominated by the superimposition of wetting–drying–wetting cycles that record high-frequency climatic oscillations, each apparently of *ca* 40 ka duration (Ghinassi *et al.*, 2004).

In the USA, several Miocene aeolian systems have been described and are mainly controlled by local tectonic conditions associated with warm and dry climatic conditions. In western and central USA, some local systems have developed related to extensional basins such as the Zia Formation of the Santa Fe Group in the Albuquerque Basin (Galusha & Blick, 1971) and the Ojo Caliente Sandstone of the Tesuque Formation in the La Española Basin (Koning *et al.*, 2004), both related to the large structure of the Río Grande Rift. These systems give rise to locally thick successions (up to 160 m) related to the development of dunefields that were strongly influenced by local conditions. Aeolian deposits have also been described in the High Plains of the USA, including the Early Miocene Arikaree Formation (Bart, 1977) in south-east Wyoming, where large-scale, cross-bedded sandstones have been related to the accumulation of barchan and transverse dunes. Finally, an aeolian origin has also been reported for a *ca* 100 m thick succession of the Comondú Group in Baja California, Mexico, related to the infill of the forearc basin developed between the Late Oligocene and Early Miocene (Umhoefer *at al.*, 2001).

Neogene aeolian accumulations in South America are related to the complex evolution of the Andes. The compressional regime in the western margin of South America led to the development of a complex foreland with multiple basins that formed important sites of accommodation that were themselves subject to an arid local climate regime. This resulted in the accumulation of several aeolian units, some with local names that record this stage of evolution, especially during the Miocene. For instance, the Petaca Formation in southern Bolivia (Uba *et al.*, 2005) and the Aguada Member of the Chacras Formation (Voss, 2002) in the Salar de Antofalla in north-west Argentina probably commenced accumulation in the latest Oligocene, but underwent their most important phase of accumulation during the Lower to Middle Miocene. These units are between 100 m and 150 m thick and record the interaction between aeolian and fluvial systems. One of the best described examples of an aeolian system developed in the Andean foreland is the Lower Miocene Vallecito Formation (Tripaldi & Limarino, 2005). This unit attains a maximum thickness of 1,200 m and comprises a complex facies arrangement that records the interaction of dunes, aeolian sandsheets and wet interdunes that interact with fluvial and lacustrine systems. The succession records a large aeolian system developed as the first syn-orogenic fill of the Andean foreland in this part of north-western Argentina (Tripaldi & Limarino, 2005). Another Lower Miocene unit with similar characteristics is the Pachaco Formation in the Precordillera of San Juan in western Argentina. The middle member of this formation is 700 m thick and records the accumulation of a large dunefield dominated by barchan and seif dunes and draas (Milana *et al*., 1993). The Angastaco Formation is related to an aeolian dunefield associated with fluvial systems that developed in the Lower Miocene of north-western Argentina (Do Campo *et al.*, 2010). Both the Mariño Formation in the Precordillera of Mendoza (Irigoyen *et al.*, 2000) and the Santo Domingo Member of the El Durazno Formation in the Sierra de Famatina (Dávila & Astini, 2003) also record synorogenic aeolian systems associated with the development of the Andean foreland during the Middle Miocene. In southern Patagonia, aeolian deposits have been also described as part of the Lower to Middle Miocene Santa Cruz Formation (Pinturas Formation of Bown & Larriestra, 1990) and in distal portions of the Andean Foreland and in the passive South American margin, sandy aeolian facies have been described as part of the Río Negro Formation associated with a marine transgression from the Atlantic (Zavala & Freije, 2001).

Most of the Neogene aeolian systems of Africa are closely related to Quaternary systems and they record early accumulation conditions, with some differences due to changes in climate over the past 10 Myr (Fig. 8). The oldest record is the Middle Miocene Tsondab Sandstone Formation in Namibia which comprises a succession up to 220 m thick of cross stratified and massive sandstones with the local development of pedogenic carbonates and palaeosols (Ward, 1988; Kocurek *et al.*, 1999; Ségalen *et al.*, 2004). This unit is interpreted as a proto-Namib sand sea that was influenced by winds that blew from the south/south-west, as today, but which developed under more humid conditions than those experienced today (Kocurek *et al.*, 1999). The succession comprises two sequences, each separated by a stabilization surface; deposits record the preservation of north-trending linear dunes that gradually undertook a lateral component of migration to the east. These linear forms supported superimposed dunes, similar to the large linear bedforms of the present-day Namib Desert, mainly as a consequence of a sustained wind regime that has been established since the Miocene (Ségalen *et al.*, 2004). Elsewhere in the Namib, up to four aeolian sequences are recognized (Senut *et al.*, 1998), comprising the deposits of star, linear and transverse dunes (Ségalen *et al.*, 2004).

Miocene aeolian deposits have been described in the Chad Basin and these are associated with the early hominid specimens in the Toros-Menalla site 266, in northern Chad, central Africa (Schuster *et al.*, 2002; Vignaud *et al.*, 2002). Accumulation of these aeolian successions was related to the early development of the Sahara (Schuster *et al.*, 2002; Vignaud *et al.*, 2002) although the relevance of this finding and its implication for pre-Quaternary desert development has been disputed (Swezey, 2006).

Upper Miocene to Lower Pliocene aeolian deposits are also present in the Western Cape of South Africa. In this area the Prospect Hill Formation is composed of calcarenites with shell fragments that record the development of a coastal dune system overlying sandy beach deposits (Franceschini & Compton, 2004).

The Garet Uedda Formation (or Members U and V of the Sahabi Formation) in Libya has been described as Upper Pliocene age and an aeolian origin has been proposed for this 25 m thick sequence of quartzitic sands interbedded with sandy shales (Tawadros, 2012). Some relic forms of the Kalahari Desert might be as old as Upper Pliocene and they record the aeolian reworking of fluvial sands (Lancaster, 2000; Haddon & McCarthy, 2005). These units have different formal names (Gordonia

Formation, Kalahari Sand, Batéké Sands and Zambezi Formation) and they can be correlated from Zaire and south-western Angola in the north to Botswana, Namibia, and northern South Africa in the south. Deposits of these successions are mainly unconsolidated sand and their distribution suggest a larger sand sea than the Quaternary Kalahari Desert, giving rise to the largest sand body on Earth (the 'Mega Kalahari') that covered over 2.5 million km^2 (Thomas and Shaw, 1990).

Neogene aeolian activity in Asia was dominated by the accumulation of thick sequences of fine-grained deposits (loess) in the Loess and Tibetan Plateaus and aeolian deposits interbedded with alluvial deposits (e.g. Zheng *et al.*, 2003). Sandstone aeolian accumulations are not common and only one example has been identified in the Middle East: the Shuwaihat Formation, a Middle Miocene unit exposed on a 16 m high cliff near Abu Dhabi in the United Arab Emirates (UAE; Whybrow *et al.*, 1999) interpreted as the record of the interaction between transverse and barchanoid dunes and a continental sabkha (Bristow, 1999).

Quaternary aeolian sand systems

Sand seas and dunefields occurring today developed during the Quaternary Era, during which significant changes in climate and sea-level related to glacial–interglacial cyclicity affected the supply, availability and mobility of sediment. Their accumulation and present configuration therefore reflects the legacy of these changes, in addition to contemporary processes. Quaternary aeolian sand systems occur on all continents and at all latitudes, from the Antarctic to the Arctic (Fig. 9). Quaternary aeolian systems are here sub-divided into those located inland and those occurring along the coast.

Inland sand seas and dunefields (Table S5)

Inland dune systems occur widely, with a concentration in low to mid-latitude arid regions of the Northern Hemisphere (35–50°N), especially in the arid regions of central Asia, on the semiarid Great Plains of North America, and in low latitude desert areas of Africa, Arabia and Australia (15–30°N and 15–30°S). Their geological setting varies, with many sand seas in Africa, Australia and Arabia occurring in cratonic basins; central Asian and South American sand seas, by contrast, are located mostly in foreland basins (Fig. 10). Dune types also vary, with linear dunes comprising about 50% of all dunes and dominating in many areas of the Sahara, south-east Arabia, Australia and southern Africa (e.g. Lancaster, 1999; Pye & Tsoar, 1990). Crescentic dunes comprise 40% of dunes and dominate sand seas in the northern Sahara, many parts of Arabia, and parts of central Asia and China. Star dunes comprise around 8% of dunes in low latitude inland sand seas, mainly in areas where topography creates complex wind regimes.

Wet systems form inland where the depositional surface intersects local perched or regional groundwater tables. Good examples of wet systems are the White Sands dunefield in New Mexico (Fig. 11; Kocurek *et al.*, 2007) and the Liwa area of the United Arab Emirates (Glennie, 2005; Stokes & Bray, 2005). Changes in sea-level, climate and/or vertical crustal movements that affect the groundwater table may result in a system changing over time from wet to dry or *vice versa*, as in the Wahiba Sands of Oman (Radies *et al*., 2004). Likewise, spatial changes in groundwater levels result in some parts of the system being dry and others wet, as in the Rub' al Khali sand sea of Saudi Arabia (Glennie, 1970; Al-Masrahy & Mountney, 2013).

It appears that the majority of modern and Quaternary aeolian sand systems operate as dry systems in which the water table is significantly below the depositional surface, such that it has no effect on the dynamics of the dune system. The major controls are therefore sand supply, its availability for transport and mobility (magnitude and frequency of winds capable of transporting sand). The Namib Sand Sea (Fig. 12) is a good and comprehensively-studied example of a dry system, sourced principally by sand from the interior of southern Africa via the Orange River (Garzanti et al., 2012). Its accumulation is interpreted to be the product of regional changes in wind regime, which result in a reduction of transport rates in the direction of transport, leading to deposition of sand by bedform climbing and dune growth. Estimates of angles of bedform climb made by Lancaster (1989) range from 0.003° for the linear dunes to 0.03 to 0.16° for crescentic dunes in the southern part of the sand sea. There is a clear spatial pattern of sand accumulation in central areas of the sand sea, represented by the equivalent or spread out sand thickness of complex linear dunes that reach 150 to 200 m in height (Bullard *et al*., 2011; Lancaster, 1989) (Fig. 12). Isotopic and sediment budget estimates for the age of the sand sea converge at around one million years (Vermeesch *et al*., 2010), but many of the dunes are relatively young, with OSL ages of 17 to 24 ka for compound linear dunes in the southern sand sea (Bubenzer *et al*., 2007) and less that 6 ka for linear dunes on the northern margin of the sand sea (Bristow *et al*., 2007). Although there is evidence for episodic dune accumulation in the Namib, it appears that the persistence of arid to hyper-arid conditions throughout the Quaternary has promoted development of a very well-organized dune system.

Stabilized dune systems are those in which vegetation or some other factor (for example, permafrost and pedogenesis) periodically or continually stabilizes the substrate while the system remains active overall. Large-scale stabilized systems include the Thar Desert sand sea in India, much of the South-western Kalahari and the Simpson-Strzlecki in Australia (Fig. 13) and parts of the Negev-Sinai sand sea (Egypt and Israel). Characteristic dune forms of these systems are vegetated linear dunes and parabolic dunes. Quaternary climatic changes also resulted in some dune systems becoming stabilized at times of increased precipitation, as in the early Holocene, when well-developed soils developed on the dunes of the southern Sahara (Kocurek *et al*., 1991; Lancaster *et al*., 2002). Such episodes of soil formation are important to preservation of older episodes of dune accumulation.

Many studies have indicated that dunes in arid Australia have accumulated episodically and some are of great age (up to one million years) (Fujioka *et al*., 2009). In many locations, the cores of linear dunes may exceed 380 ka in age, with multiple late Pleistocene and Holocene accumulation episodes (Fitzsimmons *et al*., 2007; Lomax *et al*., 2011). These episodes of dune growth have taken place without complete reworking of the dunes, in large part because of pedogenic alteration and stabilization of the deposits of older dune accumulation episodes by aeolian addition of clay and carbonate derived from nearby alluvial and lacustrine environments (Cohen *et al*., 2010; Hesse, 2011). Successive episodes of stabilization, reworking and dune growth result in an accretionary structure for the dunes, often associated with lateral migration in addition to dune extension (Rubin, 1990). Similar structures have been observed in stabilized (vegetated) linear dunes in other regions (e.g. Roskin *et al*., 2011; Telfer, 2011).

Coastal systems (Table S5)

Coastal dunefields may be divided into those composed of siliciclastic material and those composed of carbonate (or mixed carbonate and siliciclastic) material. The former type typically develops along humid-type, mid-high latitude coasts, whereas the carbonate-rich dunes/dunefields form along arid to semi-arid, mid-low latitude, coasts bordering productive carbonate platforms.

Dune types vary and their size and morphology is dependent on a number of factors including vegetation cover, sand supply, wind regime and coastal setting. Sand blown off the beach typically forms partly vegetated and fixed foredunes. Aeolian erosion of the foredunes can lead to the formation of blowouts and parabolic dunes and/or transgressive dunefields (Hesp, 1999). Along cliffed coasts special dune types including echo dunes and climbing dunes can develop (e.g. Clemmensen *et al*., 1997).

Most coastal dunefields in North-west Europe can be classified as wet as the groundwater table typically is close to the surface. Depth to the groundwater table is in many cases linked to sea-level, especially in subsiding coastal basins (Kocurek *et al*., 2001; Mountney & Russell, 2009). In other examples, especially in systems developed on uplifting coastal areas like the northern part of Denmark, dune dynamics are influenced by high precipitation rates and the formation of perched groundwater tables (Clemmensen *et al*., 2009; Pedersen & Clemmensen, 2005). Dunefields develop both on retreating and prograding coasts. Both types may share many sedimentary characteristics, but prograding systems tend to develop successive lines of stabilized foredune ridges (e.g. Bristow & Pucillo, 2006; Madsen *et al*., 2007; Reimann *et al*., 2011), whereas retreating systems more commonly experience phases of transgressive dune formation in the form of inland migrating parabolic dunes (Clemmensen *et al*., 2001a; Pedersen & Clemmensen 2005).

The Lodbjerg and Hvidbjerg coastal dunefields provide examples of wet-stabilized siliciclastic systems developed on a retreating coast (Figs 14 and 15). These two systems form part of an almost unbroken belt of coastal dunefields that flank the North Sea coast of Jutland, Denmark (Clemmensen *et al*., 2009; Pedersen & Clemmensen, 2005). Luminescence dating of the sand units and radiocarbon dating of the peaty palaeosols has made it possible to establish a detailed chronology of dunefield evolution (Fig. 16). Episodes of transgressive dune formation that occurred around 2200 BC, 800 BC, 100 AD, 1050 to 1200 AD, and between 1550 and 1650 AD were linked to periods of increased storminess (cool, wet summers), whereas stabilization took place during periods of decreased storminess (Clemmensen *et al*., 2009). The series of ages obtained by Clemmensen *et al*. (2001a) indicate accumulation of around 10 m of aeolian sand (below present groundwater table) since 2200 BC, at an average rate of 2.4 mm/yr.

Swina barrier coastal dune system is another example of a wet-stabilized siliciclastic system developed on a prograding shoreline. The Swina barrier is situated in north-west Poland along the southern part of the Baltic Sea; the dunefield is developed on top of two sandy spits that have formed between Pleistocene headlands (Reimann *et al*., 2011). Spit formation and shoreline progradation have taken place during the past 6600 years. The coast now forms a smooth and curved shoreline segment and is still prograding. The Swina barrier system is sourced by sand eroded from nearby headlands. Luminescence dating of the dunes indicates six hiatuses in foredune building, at 2100 BC, 900 BC, 200 BC, 200 AD, 600 AD, 1000 AD and 1600 AD. It is concluded that most of these phases of foredune erosion and instability were caused by climatic shifts to a cooler and windier climate. The transgressive dune formation *ca* 1600 AD was linked to increased storminess during the 'Little Ice Age' and this episode of dune formation seems to be contemporaneous with phases of increased aeolian activity in other dune systems in North-west Europe (e.g. Clarke & Rendell, 2009; Clemmensen & Murray, 2006; Clemmensen *et al*., 2009).

Carbonate-rich aeolian systems are commonly developed in mid-latitude and low-latitude, arid and semi-arid climate belts; these aeolian systems occur in a variety of settings including lowland and cliffed coasts (e.g. Brooke, 2001; Frébourg et al., 2008). Due to the climatic setting of these systems they are most logically classified as dry. The carbonate sand is lithified soon after deposition (Guern & Davaud, 2005) thereby forming one mechanism of preservation that is poorly known from siliciclastic systems. These lithified carbonate-rich aeolian deposits are termed aeolianites (Brooke, 2001).

Particularly well-developed carbonate aeolian systems occur in the Western Mediterranean region (Clemmensen *et al*., 1997; Fornós *et al*., 2009; Andreucci *et al*., 2010a). Quaternary successions with carbonate-rich aeolian sand units crop out quasi-continuously along the north-west coast of Sardinia near the town of Alghero (Andreucci *et al*., 2010a,b; 2014). The aeolian units, which are lithified, occur along a cliffed coast and can be subdivided into cliff-front dune accumulations and valley-head sand ramps (Andreucci *et al.*, 2010a; Fig. 17). Note also other major Quaternary aeolianites in South Africa (e.g. Roberts, 2008)

AEOLIAN RESEARCH: THE WAY FORWARD AND FUTURE RESEARCH PROSPECTS

Aeolian facies and sequence stratigraphic models: a useful approach to capturing complexity in aeolian successions?

Relating preserved aeolian stratigraphy to original bedform morphology and behaviour

Although it is now possible to effectively describe in detail both: (i) the morphological characteristics of modern bedforms and larger dune fields; and (ii) the geometry of architectural elements of preserved aeolian successions, notably by using the forward stratigraphic modelling techniques developed by Rubin (1987) and Rubin & Carter (2006), several problems remain regarding how to relate ancient preserved sets of aeolian strata to the morphology and migratory behaviour of the original bedforms. In particular, in cases where large compound and complex morphological dune types have accumulated in desert basins in which the rate of accommodation generation has been highly variable over time or space, for example in response to spatially and temporally variable synsedimentary tectonic activity (e.g. Rodríguez-López *et al.*, 2013).

The accumulated sedimentary record of most modern inland dunefields remains largely unknown, with only fragmentary glimpses of aeolian sedimentary architectures having so far been revealed from modern aeolian systems via techniques such as trenching (e.g. McKee, 1966) and ground-penetrating radar (GPR) studies (e.g. Bristow et al., 2000).

Conversely, relatively few ancient aeolian successions are known that preserve, intact, the original morphologies of the bedforms that gave rise to the architecturally complex set and cosets of aeolian cross-bedding that dominate the ancient sedimentary record (e.g. Mountney et al., 1999). Thus, the development of aeolian facies models that are used to relate modern dunes and dunefields to accumulated deposits remains problematic in terms of how best to interpret the preserved sedimentary architectures of ancient aeolian deposits.

Despite great progress on understanding of the geomorphology of Quaternary inland sand seas, there are few data relating to their stratigraphic and sedimentological record. In many areas, Quaternary sand seas and dunefields have not left a significant accumulation and the bedforms present in many modern dunefields are known to be partially or completely legacy landforms inherited from LGM times (e.g. Lancaster et al., 2002); thus, such forms do not necessarily reflect the currently prevailing climatic and sediment supply conditions. Elsewhere, information from the subsurface (for example, GPR data, cores and well logs) does not exist or is proprietary. The costs and logistics of acquiring such datasets are often prohibitive, but where they have been developed, the understanding of the sedimentary record of inland sand seas has been transformed and relations between contemporary processes and the sedimentary record established (Bristow *et al.*, 2000; 2005; 2007; Derickson *et al.*, 2008; Kocurek *et al*., 2007; Radies *et al*., 2004; Stokes & Bray, 2005). There is a clear need to advance knowledge in this area by a coordinated program of research that seeks to link dune processes to the sedimentary record. One valuable step in this direction is the development of a methodology for reconstructing wind direction, wind speed and duration of wind events from aeolian cross-strata (e.g. Eastwood *et al.*, 2012). The stratigraphic record of Quaternary (Holocene) coastal dunefields is better documented because such systems have been investigated by GPR mapping, outcrop and core studies, and placed in a chronological framework by luminescence and radiocarbon dating (e.g. Clemmensen *et al.,* 2001a; Pedersen & Clemmensen 2005); but also in these systems there is a need to link aeolian processes more closely to the sedimentary record. Sedimentary units with flat-bedded strata form a large portion of the stratigraphic record of these Holocene coastal systems (Clemmensen *et al.,* 2001a), but it is not yet clear whether these facies represent the trailing edges of inland migrating parabolic dunes or featureless sand plains formed in between the parabolic dunes.

The sedimentary structures of modern dunes need to be described more systematically in the literature. This includes the internal structures of fixed linear dune ridges as well as those of many transgressive (actively migrating) dunes including the parabolic dunes of coastal areas. Studies of the internal sedimentary structures of dunes can be conducted from some natural exposures, including blow-outs, but the most systematic observation is provided from ground-penetrating radar mapping of the dune bodies (e.g. Bristow *et al*., 1996; 2000a,b; Clemmensen *et al*., 2007; Tamura *et al*., 2011). Such studies of internal structures should be accompanied by geomorphological studies, map studies of dune evolution and data on wind patterns. Bristow and Hill (1998) attempted to reconstruct changes in dune morphodynamics and ancient wind conditions from preserved sets of cross-stratification in the Miocene Shuwaithat Formation, by using a reconstruction of localized wind patterns in relation to position on dune bedform slopes. This detailed reconstruction differs in scale to the more commonly attempted reconstructions of regional palaeowinds.

To date, relatively few pre-Quaternary coastal dune deposits (particularly in siliciclastic environments) have been described in the literature. Many ancient coastal dune deposits may have been overlooked and more studies of the internal structures of such modern dunes are needed to create a database of varied examples of sufficient size that will enable facies models for coastal aeolian dune systems to be 'distilled', and from which criteria for the recognition of characteristic stratigraphic architectures indicative of different coastal-dune types can be established and applied to the stratigraphic record.

Recognition of the deposits of different aeolian bedform types in the preserved record

One fundamental problem that remains in relation to the interpretation of original dune-type from the stratigraphic architecture preserved in the rock record is that the majority of reconstructions envisage transverse (and related barchanoid) dune types, whereas successions representing linear and star (and other) bedform types remain apparently under-recognized, despite such bedforms forming the majority of bedforms in modern dunefields. Given that there is no reason to suspect that linear and star bedforms were less abundant in the geological past, this suggests that current models and methods for the recognition and reconstruction of primary bedform type from preserved sedimentary architectures are not sufficiently refined to reliably distinguish bedform type. The reasons for these limitations in current facies models probably involve one or more of the complex factors considered below.

Linear bedforms, which make up *ca* 50% of the dunes in the world's modern dune fields, principally transport sediment in an orientation close to parallel to the trend of their crestlines, but Bristow *et al.* (2000a, 2007) used GPR techniques to convincingly demonstrate that a minor component of transverse motion in such bedforms favours the preferential preservation of cross-strata that dip in the direction of that transverse component of migration. Thus, such linear forms tend to produce cross strata that dip toward an azimuth that is at a high orientation to the migration direction of the primary bedform. The accumulation of sets of such cross-strata is therefore very difficult to distinguish from the deposits of transverse bedforms. This has significant implications for the reconstruction of palaeowind directions based on analysis of dip-azimuth data determined from preserved foresets, which, for linear bedforms, could be in orientations oblique to the dominant wind direction (see Scherer, 2000). Although recognition of linear draas is not an easy task, sedimentological and architectural features have been proposed as useful tools for their recognition in the sedimentary record (see Scherer, 2000 and Rodríguez-López, *et al.* 2008).

Many linear and star bedforms in present-day dunefields are isolated from neighbouring bedforms by extensive interdune flats and this suggests that such dunefields are not necessarily currently undergoing active accumulation, even in cases where the bedforms might be undergoing active construction (see Kocurek, 1999). Indeed, many large-scale examples of such bedforms are currently partially or fully stabilized by vegetation, by surface crusts or because the present-day wind regime is not capable of inducing the migration of such large forms (Bristow & Mountney, 2013). As such, linear and star forms (and their deposits) might have a reduced accumulation and, by implication, preservation potential such that successions representing the preserved deposits of such bedforms are indeed rare in the ancient stratigraphic record.

Draa architecture and apparent scarcity in the sedimentary record

Numerous facies models have been proposed with which to predict the likely arrangement of complex cosets of cross-strata arising from the accumulation of compound or complex draas representative of landforms comprising two or more scales of bedform that occur superimposed upon one another and which might undergo migration in different directions and styles, and at different rates (e.g. Brookfield, 1977; Rubin and Hunter, 1985; Rubin, 1987). Large draas are ubiquitous in the central parts of most of the world's major desert dune fields where interdunes have been reduced to isolated depressions and neighbouring bedforms can be shown to be migrating over one another to leave an accumulation. From a conceptual standpoint, the preservation potential of such draa deposits ought to be relatively good where such bedforms accumulate in gradually subsiding basins or under the influence of a sustained episode of water-table rise. However, documented examples of draa deposits preserved in the ancient geological record are relatively scarce (e.g. Herries, 1993; Clemmensen, 1989; Chrintz & Clemmensen, 1993; Mountney, 2006a; Veiga & Spalletti, 2007; Langford et al, 2008; Rodríguez-López *et al.*, 2008; 2012a). Obvious research questions that remain to be addressed in relation to this issue are as follows: (i) are the preserved deposits of large draas actually relatively common in the ancient aeolian record but remain under-recognised because detailed studies focussed specifically on their reconstruction have not yet been undertaken, either because outcropping examples have not been recognised or because the style of outcrop is not appropriate for their convincing 3D reconstruction; and (ii) is the preservation potential of large draas low such that well-preserved examples are actually rather rare in the ancient aeolian record? In relation to this second point, how might large draa bedforms become preserved? In situations where such bedforms migrate over one another, angles of climb would probably be very low and, as such, only the very lowermost flanks of these large bedforms would accumulate. Such deposits would tend to be dominated by sets of low-angle inclined cross-strata dominated by wind-ripple deposits that represent only the lower plinths of large bedforms. Reconstruction of the original bedform morphology from this fragmentary record would be problematic. The potential for the preservation of such large bedforms essentially *intact* is low and would probably require exceptional circumstances, such as inundation by flood basalts (e.g. Mountney *et al.*, 1999a,b; Jerram *et al.*, 1999a, b) or important transgressive events (Strömbäck *et al.*, 2005; Rodríguez-López *et al.*, 2012a).

Mechanisms of accumulation and preservation: alternatives to bedform climbing

Many past and current facies models that attempt to account for the preserved sedimentary architecture of aeolian successions invoke bedform climbing as a mechanism to explain the accumulation of thick preserved successions of aeolian dune and interdune strata. In its most basic form, the bedform-climb model envisages a train of contemporaneously active bedforms separated by interdunes that undertake gradual and progressive accumulation as they migrate over time, thereby accumulating a series of stacked sets of cross-bedded dune and interdune strata. Crucially, accumulation via such a mechanism requires a sustained episode of accumulation that is concurrent with ongoing bedform migration. Although bedform climbing can be convincingly shown to have been responsible for some preserved successions (e.g. Mountney & Thompson, 2002), others are more likely to have accumulated in response to punctuated accumulation whereby the preserved record represents numerous sequences, each bounded by supersurfaces that represent potentially protracted episodes of non-deposition (bypass) or erosion (deflation) (e.g. Loope, 1985). Although accumulation via bedform climbing and accumulation to form individual sequences separated by supersurfaces have traditionally been considered competing end-member models (e.g. Kocurek, 1991), several studies now convincingly demonstrate that both mechanisms for accumulation can occur and that transitions between accumulation via bedform climb and via sequence and supersurface generation can occur both spatially and temporally for some preserved aeolian successions (e.g. Mountney, 2012). Indeed, other mechanisms for the accumulation of aeolian successions are also possible (and arguably likely): the model of Paola and Borgman (1991) for the generation of sets of cross-strata in response to the migration of bedforms of pseudo-random height and scour depth and in the absence of net deposition (i.e. a zero angle of net climb) describes one such mechanism, although the long-term preservation potential of sets generated by such a mechanism remains uncertain because the passage of later bedforms would render the deposits prone to reworking in the absence of a component of accommodation generation.

Some obvious research questions that remain to be answered include: (i) what proportion of the known aeolian preserved record arose via sustained episodes of bedform climb versus episodic sequence and supersurface generation, respectively; (ii) what specific set of controlling conditions allow for sustained bedform climb versus what set of conditions allow for episodic sequence accumulation and preservation; and (iii) can sophisticated 4D models be developed with which to account for complex patterns of spatial and temporal transition between different mechanisms of accumulation and preservation and, if so, will such models be dependent on the development of high-resolution dating techniques, meaning that they will likely be best suited to the analysis of Quaternary aeolian successions?

Encapsulation of time in the aeolian stratigraphic record

It has long been recognised that the preserved sedimentary record is highly fragmentary (e.g. Barrel, 1917) such that the majority of geological time is encapsulated within unconformities and this notion is summarized succinctly by Ager (1993) in the often-recounted statement that: "the stratigraphical record is of one long gap with only very occasional sedimentation". Indeed, the aeolian record is likely to be especially fragmentary because both accumulation of aeolian deposits and also their long-term preservation demand a rather specific set of circumstances to conspire fortuitously (Kocurek, 1999) such that accumulation can take place at a time when accommodation space is being generated to enable preservation. Indeed, wind-blown sands deposited on thick continental crust are especially likely to be riddled with unconformities (e.g. Loope, 1985).

Several significant research questions remain to be addressed regarding the duration required to construct, accumulate and ultimately preserve aeolian system deposits. Specifically, how much time is represented by thick accumulations of aeolian strata as a fraction of the total time episode over which the succession accumulated? How much time is encapsulated within the unconformities (i.e. supersurfaces) present in successions of multiple stacked aeolian erg sequences and what are the fundamental allogenic controls that act at the basin scale to govern the timing of episodes erg construction and accumulation versus episodes of supersurface generation (see Rodríguez-López *et al.*, 2013)? Are episodes of aeolian construction and accumulation and episoides of supersurface formation gradual or abrupt, or does the time required for their development vary according to the underlying nature of the climatic and tectonic controls? The role of local synsedimentary tectonics (basin setting) and the differences between ergs accumulated in different tectonic settings (for example, foreland-basins verus rift basins) remains a largely open research question. How do rates of accommodation generation (via basin subsidence), uplift and denudation vary between desert basins and how might such rates be reconstructed from the preserved aeolian record? Given that the preserved stratigraphical record is generally accepted to be highly fragmentary (e.g. Loope, 1985), are specific episodes during which accumulation and preservation occurred necessarily representative of conditions that prevailed throughout the entire episode over which aeolian systems were active more generally but for which no record was preserved? Alternatively (and arguably, more likely) is the preserved aeolian satratigraphic record inherently biased such that certain types of aeolian dunefields have a greater chance of preservation because the climatic and tectonic settings in which they are constructed and accumulated favour preservation? What specific sets of circumstances are required to enable accumulation and preservation? Over what duration can an aeolian sequence accumulate given favourable conditions?

Attempts to address these research issues face a fundamental problem: the availability of appropriate dating techniques with which to establish the time periods over which accumulation and preservation occurred. Attempts to date aeolian sequences are best suited to studies of Quaternary successions for which techniques such as high-resolution OSL and radiocarbon dating can be applied; yet establishing the presence, lateral extent and significance of major supersurfaces of regional extent is best suited to ancient outcropping successions for which key stratal surfaces can be traced over distances of many hundreds to thousands of square kilometres (e.g. Havholm et al, 1993; Havholm and Kocurek, 1994; Blakey et al. 1996).

An additional problem is that supersurfaces may not necessarily form isochonously across an entire basin; rather, their development may proceed in a progressive manner across the basin as the conditions responsible for their development gradually change over time and space (cf. Porter, 1986). Such a complex evolutionary history could mean that a single supersurface may potentially be time transgressive and may exhibit different characterisitcs across its area in response to spatial variation in the allocyclic controls responsible for its development at a basin scale (Rodríguez-López *et al.*, 2013).

The role of the water table in enabling accumulation and preservation

Although many aeolian systems and preserved successions are considered to represent dry aeolian systems, a significant number of other examples reveal evidence to suggest that the water table (or its near-surface capillary fringe) played a significant role in assisting accumulation and enabling long-term preservation (Mountney, 2012; cf. Kocurek & Havholm, 1993). Although several well-known aeolian successions, including parts of the Permian Cedar Mesa Sandstone (Mountney & Jagger, 2004) and Jurassic Entrada Sandstone (Crabaugh & Kocurek, 1993), can be shown to have accumulated via bedform migration and climb that was coincident with a gradual and progressive rise in the relative water table, it is also clear that these succession are also divided into a series of separate sequences bounded by supersurfaces. Several questions remain unresolved in relation to the role of the water table in enabling aeolian accumulation and preservation; for example, over what period can a relative rise in water-table level be sustained to enable wet system accumulation and how thick might the resulting preserved wet aeolian sequences be?

Can alternatives to the 'climbing wet system' model be developed to account for the origin of individual aeolian dune sets preserved between sets of interdune strata indicative of a wet or damp substrate? Might such single aeolian dune sets record the sporadic accumulation and preservation of thin aeolian sequences during episodes when controlling conditions conspired to fortuitously enable preservation? If so, might the intervening ‘interdune’ accumulations actually record protracted episodes during which accumulation of damp sand flats occurred at very slow rates (cf. Kocurek, 1981b; Crabaugh and Kocurek, 1993)?

Why are climbing wet aeolian systems only relatively rarely recognised in modern settings (e.g. Mountney & Russell, 2009), yet are increasingly interpreted from ancient successions? Might this reflect a bias that is inherent in the preserved stratigraphical record, whereby wet aeolian systems have an increased preservation potential because they progressively subside beneath a relative water table as they accumulate and this protects these deposits from later deflation (cf. Stokes, 1968; Loope, 1985; Kocurek and Havholm, 1993)?

The role of groundwater level in preserving aeolian sand accumulations in coastal settings has been emphasized by many authors (e.g. Kocurek *et al.,* 2001; Clemmensen *et al*., 2001b; Mountney & Russell, 2009), but few studies have discussed in detail what actually controls the groundwater level in coastal systems (e.g. Bristow *et al.*, 2000b). Sea-level is frequently cited as a dominant control on groundwater level (Havholm and Kocurek 1994; Kocurek *et al.*, 2001), but in many coastal dune systems in northern Denmark, for example, groundwater level is up to 15 m above sea-level (e.g. Clemmensen *et al.,* 2001a). Clearly factors other than sea-level are controlling the groundwater level in these coastal systems.

The problem of dating aeolian sediments and rocks

Age control for episodes of dune construction, stability, and reworking, as well as rates of dune accumulation for many Middle and Late Pleistocene and Holocene aeolian systems is being provided by luminescence dating techniques (e.g. Murray & Clemmensen, 2001; Lancaster, 2008; Singhvi & Porat, 2008; Fornós *et al*., 2009; Andreucci *et al*., 2010b; Reimann *et al*., 2011; Thomas, 2013). In most studies, luminescence dating is based on quartz grains having typical grain sizes between 0.18 mm and 0.25 mm. A luminescence age is derived by dividing the dose absorbed from ionising radiation during burial by the dose rate; the latter originates from the natural radioactivity (U, Th, ^{40}K and decay products) in the sediments and, to some degree, with cosmic radiation.

Supplementary age control can be obtained using K-rich feldspar grains (Buylaert *et al.*, 2012), and although quartz optically stimulated luminescence usually is limited to the last *ca* 100.000 years, accurate ages up to 600.000 years can be obtained using feldspar infrared stimulated luminescence (Buylaert *et al*., 2012). Additional age control is provided by radiocarbon dating of organic-rich soils, which are common components of many Holocene coastal aeolian successions (e.g. Clemmensen *et al.*, 2001a; 2009; Arbogast *et al.,* 2002) but are rare in low and mid-latitude desert dune systems. Holocene aeolian sand systems may contain artefacts, and archaeological studies of these items can add to the chronological understanding of the evolution of the system (e.g. Kocurek *et al.*, 1991; Liversage & Robinson, 1993; Roberts, 2008). Some Quaternary aeolian deposits interbedded with shell-bearing beach or shallow marine deposits have been dated by means of U-series analysis or by use of aminostratigraphy (e.g. Hearty *et al.*, 1986; Hillaire-Marcel *et al*., 1996), and a few Quaternary aeolian systems have been dated by magnetostratigraphy and susceptibility stratigraphic analysis (e.g. Nielsen *et al.*, 2004).

Pre-Quaternary aeolian successions are hard to date precisely due to the lack of material that can be dated numerically. Aeolian successions that are associated with lava flows, however, can be placed in a chronological framework (for example, the Cretaceous Etjo Formation of Namibia; Mountney & Howell, 2000). Most siliciclastic aeolian deposits contain no fossils that can be used to place them in a precise biostratigraphical framework. However, some rare exceptions include the Tertiary Tsondab Sandstone of Namibia (Pickford & Senut, 1999). In contrast, carbonate aeolianites contain abundant microfossils that potentially could be useful in biostratigraphical analysis (see Babić, *et al.*, 2013). However, the application of such dating techniques should be undertaken with care to eliminate the risk of possible error associated with aeolian reworking of older bioclasts which could be incorporated in more recent dune systems (see Teller *et al.*, 2000).

Aeolian systems interbedded with fossiliferous marine strata can be placed in a biostratigraphical framework and commonly also interpreted in a sequence or climate stratigraphical context (for example, the Late Carboniferous Hermosa Formation, Utah, USA, Atchley and Loope, 1993; the Pennsylvanian to Permian lower Cutler beds, Utah, USA, Jordan and Mountney, 2010, 2012). Additional methods that have been used in chronological studies of pre-Quaternary aeolian successions include combined palaeomagnetic and cyclostratigraphical analysis (for example, aeolian sandstones associated with lacustrine sediments in the Triassic of eastern USA and East Greenland; Kent & Tauxe, 2005).

Quantitative modelling of aeolian systems

Quantitative forward stratigraphic modelling provides a tool with which to demonstrate the link between: (i) aeolian bedform morphology; (ii) the processes operating on and between such bedforms; (iii) the style of migratory bedform behaviour; and (iv) the sedimentary architecture of preserved aeolian successions, which is characterized by packages of lithofacies arranged into architectural elements with predictable geometries that are delineated by bounding surfaces of various types.

Building on the innovative geometrical modelling approach of Rubin (1987), there is a clear need to develop an effective approach to the modelling and characterization of the distribution of aeolian lithofacies within larger-scale architectural elements and establishment of a method to unequivocally demonstrate the primary processes responsible for the generation of such lithofacies distributions. Forward stratigraphic models for aeolian systems that account for both spatial and temporal variations in original bedform morphology, scale and style of migratory behaviour are needed. Such *dynamic* models (cf. Mountney, 2006b, 2012) will be useful to predict the 4D evolution of aeolian systems and the 3D distribution of resultant aeolian architectural elements, which will vary both spatially and vertically (as a function of temporal system evolution).

Another outstanding research issue is the establishment of a modelling technique with which to demonstrate the relation between the development of bounding surfaces due to autogenic processes, such as bedform migration and climb, and those that might be influenced by allogenic controls. For example, bedform climbing to generate an accumulation (which is generally considered to be an autogenic process) requires a sustained supply of sediment that is available for transport, and a wind that is capable of carrying that sediment to the site of accumulation (Kocurek & Lancaster, 1999); each of these controls will probably be influenced by allogenic factors, such as climate change, tectonism or relative sea-level change (in the case of near-coast aeolian systems). It is now evident that many processes once considered to be either allogenic or autogenic in origin are, in fact, more complex and reflect combined extrinsic and intrinsic controls (see Veiga *et al.*, 2005; Mountney, 2012; Rodríguez-López *et al.*, 2012a).

Forward stratigraphic modelling should aim to provide a quantitative approach to the 4D sequence stratigraphic modelling of aeolian successions, whereby the generation of aeolian sequences and their bounding supersurfaces (for example, deflation surfaces) can be accounted for in terms of formative processes. Coupled with advances in the application of dating techniques to Quaternary aeolian successions, such modelling approaches could be used to demonstrate the longevity (or otherwise) of episodes of aeolian dune-field construction and accumulation versus episodes of destruction and supersurface development (cf. Loope, 1985; Kocurek & Havholm, 1993; Mountney, 2006a; Jordan & Mountney, 2012; Rodríguez-López *et al.*, 2012a, 2013).

Crucially, any adopted modelling approach must be undertaken in combination with detailed studies of outcropping successions, whereby meticulous recording of key stratal surfaces (such as supersurfaces of regional extent) using recently available digital technologies such as laser scanning (LiDAR) and photogrammetry will provide the hard data with which to constrain and justify modelled synthetic stratigraphies. One particularly important application of this combined and integrated outcrop analysis and conceptual modelling technique will be to test scenarios for which accumulation and preservation might take place in response to bedform climb, versus scenarios for which episodic and punctuated accumulation of aeolian sequences bounded by supersurfaces might occur.

Potential applications of new approaches to forward stratigraphic modelling of aeolian systems and their preserved successions have been applied to better understand aeolian hydrocarbon reservoirs such as the Permian Rotliegend Group of North Sea (Howell & Mountney, 1997), the Permian Unayzah Formation of Saudi Arabia (Melvin *et al.*, 2010a) and the Jurassic Norphlet Sandstone of the Gulf of Mexico (Story, 1998; Ajdukiewicz *et al*., 2010). Future aeolian modelling applications include: (i) the development of more realistic object-based reservoir models for the improved characterization of net and non-net aeolian reservoir intervals; (ii) the prediction of the likely distribution and lateral extent of low permeability barrier and baffles to flow, for example due to the presence of mudstone units in interdune elements; (iii) the prediction of fluid-flow pathways (Lindquist, 1988; Garden *et al*., 2005); and (iv) a technique for reconstructing 3D aeolian architecture from attributes present in 1D core samples (e.g. Romain & Mountney, 2014). Similar approaches could also be used for the characterization of water aquifers (e.g. Bloomfield *et al*., 2006) and for assessment of aeolian successions as potential sites for the underground sequestration of CO_2 (Chiaramonte *et al*., 2008).

How can studies of Quaternary dunefields and sand seas inform understanding of ancient preserved aeolian successions in the rock record?

Preceding sections have highlighted the issues and challenges involved in understanding the preserved record of aeolian strata. In many cases, the record of Quaternary aeolian accumulation can provide data to inform these questions.

Relating preserved strata to original bedform morphology requires better documentation of the sedimentary structures and architecture of modern dunes of all types. Such documentation will also aid the recognition of different dune types in the sedimentary record and will address the question posed by Rubin & Hunter (1985): why are deposits of longitudinal (linear) dunes rarely recognized in the sedimentary record? Ground penetrating radar (GPR) can provide such data in an efficient manner, but attempts at 3D reconstructions of the sedimentary architecture of modern dunes are rare (see Bristow *et al.*, 2007, for an example). There exists a clear need for a comprehensive program of 3D GPR imaging of a representative selection of modern dunes to provide a 'library' of architectural styles for comparison with the rock record. Such studies will need to be guided by selection of appropriate spatial scales for acquisition of GPR data and make use of innovative software applications for visualization of these complex datasets.

The nature of the intervals of time represented by episodes of accumulation separated by bounding surfaces of different orders can now be addressed, at least for Quaternary successions, by high resolution OSL dating of dunes and their accumulations and radiocarbon dating of associated organic-rich soils, coupled with stratigraphic information from exposures or GPR (e.g. Clemmensen *et al.*, 2001ab; Pedersen & Clemmensen, 2005). A promising development is the modelling of dune accumulation using a probabilistic approach (Telfer *et al.*, 2010; Bailey & Thomas, 2013). The discontinuous nature of many dune accumulations is becoming very clear. Optically stimulated luminescence dating studies in several areas suggest that intervals of rapid accumulation of sand separated by long hiatuses is a common feature of dune accumulation (e.g. Atkinson *et al.*, 2011; Bray and Stokes, 2003; Bristow *et al.*, 2007; Telfer, 2011). However, Leighton *et al.* (2013) have conducted a detailed sampling of aeolian sedimentary units in large linear dunes in the UAE, and conclude that stratigraphy alone is not sufficient at the studied sites to guide OSL sampling. For example, significant unconformities that represent a hiatus in dune accumulation are not distinguishable in the field from bounding surfaces representing more rapid changes in dune dynamics, whereas unconformities separating stratigraphic units of similar sedimentary properties do not always conform to the duration of associated hiatuses determined from OSL ages of the units.

On a broader spatial and temporal scale, a synthesis of multiple data sets to determine the boundary conditions of tectonics, topography and climate in which Quaternary aeolian sand systems have developed would provide a valuable baseline against which to evaluate ancient sand seas and dunefields. Such an approach requires a data intensive investigation that transcends the efforts made by Breed *et al*. (1979) and will require collection of data that describe aspects of spatially changing dune and interdune morphology and migratory behaviour across dunefields and their marginal areas (e.g. Al-Masrahy & Mountney, 2013). Coastal dunefields should be treated separately and dunefield evolution in particular related to coastal setting, sea-level and climate change. Similar approaches, most notably in relation to Distributive Fluvial Systems (DFS), have transformed our ability to understand these systems and have significantly altered perspectives on how and where these systems form and are preserved in the rock record (Hartley *et al.*, 2010; Weissmann *et al.*, 2010), leading to a new interest in the sedimentology of these features and their relations to groundwater and hydrocarbon reservoirs. The work of Livingstone *et al*. (2010) in Namibia is a promising start to this type of study.

What will be the ultimate preserved record of the great Last Glacial Maximum sand seas?

The period during the last Glacial Maximum (19 to 26 ka) and immediately following it (11.5 to 19 ka) is characterized by widespread evidence of inland dune and draa construction. Preservation of the deposits of these landforms and their eventual incorporation into the rock record is, however, much less certain and requires that preservation space (sensu Kocurek, 1998) exists as a result of subsidence and burial, sea-level rise and/or a relative or absolute rise in the water table. There are few documented examples of any of these situations. Subsidence alone is usually insufficient to preserve aeolian deposits unless accompanied by a rise in water table that limits bedform migration and erosion. Such conditions may have existed locally during early Holocene wet phases in the Sahara and Arabia, but subsequent aridification has resulted in falling water tables and reworking of accumulations, as in the Liwa area of the UAE where Marine Isotope Stage 5 (MIS-5) and Holocene aeolian deposits dominate (Stokes & Bray, 2005). Similarly, during the early Holocene, dunes in the southern and western Sahara and parts of Arabia were stabilized by soil formation (Talbot, 1985) and this stabilized surface has survived in many places to the present day.

A promising environment for preservation of Last Glacial Maximum dune deposits may occur where dunes from inland migrated or extended on to the continental shelf during glacial periods of low sea-level and were subsequently preserved by shallow marine and/or coastal sabkha deposits during the early Holocene sea-level rise. Such scenarios exist in Mauritania (Hanebuth *et al.*, 2013; Lancaster *et al.*, 2002), in the Arabian Gulf (Al-Hinai *et al.*, 1987; Shinn, 1973), where crescentic dunes have built a clastic wedge offshore of Qatar, overlain by coastal sabkha deposits, and in north-western Australia, where linear dunes are preserved by estuarine muds deposited during the Holocene transgression (Jennings, 1975). Hanebuth *et al.* (2013) present a detailed analysis of the architecture of aeolian and shallow marine deposits off northern Mauritania, which indicates preservation of deposits of an inland dune complex of MIS-4 age, but not of the MIS-2 (LGM) dunes, despite the intensity of this aeolian constructional phase. These authors hypothesize that LGM dune deposits are not preserved because they were not cemented prior to the sea-level rise. Alternatively, the position of major dune complexes may have shifted to the south between MIS-4 and MIS-2. In either case, the present study demonstrates the likely sporadic nature of preservation of inland dune deposits in the Quaternary sedimentary record.

During the Last Glacial Period global sea-level was low and during the Last Glacial Maximum sea-level had dropped to a level *ca* 120 m below that of the present day. Luminescence dating indicates that these conditions of low sea-level frequently led to inland wind transport of bioclastic sand from exposed shelf areas (e.g. Zhou *et al*., 1994; Preusser *et al*., 2002; Radies *et al*., 2004; Fornós *et al*., 2009; Andreucci *et al.*, 2010a). Sand – commonly in the form of transgressive dunes – migrated far inland building large (coastal) dunefields such as the Wahiba Sands, Sultanate of Oman (Radies *et al*., 2004), or more isolated aeolian accumulations such as cliff-front dune or ramp deposits on Mallorca and Sadinia, Western Mediterranean (e.g. Clemmensen *et al*., 1997, Fornòs *et al*., 2009; Andreucci *et al*., 2010). Preservation of dune sand in the Wahiba Sands took place during a subsequent episode of increased wetness and elevated groundwater table (Radies *at al.*, 2004), whereas preservation of the cliff-front aeolian accumulations in the Western Mediterranean was linked to early cementation (Fornòs *et al*., 2009).

CONCLUSIONS

Wind has been an agent responsible for eroding, transporting and depositing sediments on Earth since at least 3.2 Ga. The study of the sedimentary record of aeolian deposits has allowed resolution of some fundamental questions regarding sedimentology and stratigraphic architecture. However, important questions regarding the complexity of the relative role of different allocyclic controls, preservation mechanisms and facies variability remain to be resolved. Multidisciplinary approaches are required to address a series of research problems of fundamental importance: has the terrestrial atmosphere had the same properties throughout Earth history? Facies models of aeolian systems are currently too simplistic and do not account for the complexity inherent in facies patterns. The sedimentary record is likely to be fundamentally biased because preserved aeolian deposits represent only a fraction of the dunefields that have persisted throughout Earth history. Furthermore, common mechanisms for aeolian accumulation tend to preserve only the lowermost parts of dune aprons, such that attempts to reconstruct aeolian palaeoenvironments are based on data that reflects only a small part of an original dunefield system. The development of detailed facies and sequence stratigraphic models will require high-resolution dating techniques, meaning that they will probably be best devised via the analysis of Quaternary aeolian successions. At the basin scale, the significance of syn-sedimentary tectonics has probably been understudied with respect to other allocyclic controls (for example, climate) known to influence erg architecture and supersurface generation. This study serves to outline the current limitation in current understanding of aeolian systems and their preserved successions: it is hoped that this discussion will stimulate thought and renewed activity in addressing shortcomings in existing knowledge.

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FIGURE CAPTIONS

Figure . 1. Schematic diagram showing the three-phase creation of the aeolian rock record and main controlling factors as proposed by Kocurek (1999).

Figure 2. Characteristics of dry, wet and stabilizing aeolian systems illustrating the role of aerodynamic configuration, water table level and stabilizing agent as controls on accumulation space. After Kocurek (1998).

Figure 3. Spectrum of preserved dune and interdune architectures resulting from temporally and spatially invariable (i.e. static) aeolian system behaviour. The angle of bedform climb defines fields of accumulation and deflation, with bypass occurring when the angle climb is zero. Within the field of accumulation, preserved sedimentary architecture is partly determined by the proportion of the accumulation surface covered by dunes. Accumulating dry aeolian systems typically require 100% dune cover whereby dunes have been constructed to a size where interdunes are reduced to isolated depressions between bedforms. Accumulating wet or stabilizing systems have less than 100% dune cover. The angle of climb is determined by the ratio of the vertical accumulation rate and bedform migration rate. The stratal configurations are scale-independent and can potentially occur in systems of any size; after Mountney (2012). Bedform spacing is the crest to crest (or toe to toe) distance between adjacent bedforms in an orientation perpendicular to the trend of elongate bedform crestlines; dune wavelength records the extent of a bedform in an orientation perpendicular to the trend of the bedform crestline and this may vary from a maximum dune wavelength to a minimum dune wavelength within one dune segment as a function of bedform sinuosity (Al-Masrahy & Mountney, 2013). Bedform spacing and dune wavelength will be the same for straight-crested dunes that lack intervening interdune flats.

Figure 4. Spectrum of interdune geometries generated by variations in the frequency and magnitude of water table change, the rate of dune migration and the net aeolian sediment budget. (A) Entrada

Sandstone, Kocurek (1981a). (B) Navajo Sandstone (Herries, 1993). (C) and (D) Helsby Sandstone Formation (Mountney & Thompson, 2002). (E) Cedar Mesa Sandstone (Langford & Chan, 1988, 1989; Mountney & Jagger, 2004). (F) White Sands (Simpson & Loope, 1985; Loope & Simpson, 1992). Modified after Mountney & Thompson (2002).

Figure 5. Palaeogeographic distribution of siliciclastic aeolian sand systems during: (A) late Cambrian; (B) middle Ordovician; (C) middle Silurian; and (D) early Devonian. Palaeogeographic maps from Scotese (2001) PAEOMAP Project.

Figure 6. Palaeogeographic distribution of siliciclastic aeolian sand systems during: (A) early Carboniferous; (B) late Carboniferous; and (C) late Permian. Palaeogeographic maps from Scotese (2001) PAEOMAP Project.

Figure 7. Palaeogeographic distribution of siliciclastic aeolian sand systems during: (A) early Triassic; (B) early Jurassic; (C) late Jurassic; and (D) late Cretaceous. Palaeogeographic maps from Scotese (2001) PAEOMAP Project. For each map, the aeolian successions shown are not all necessarily of the same age and therefore were not necessarily all active at the same time. Furthermore, the global palaeogeographies depicted in each map might not necessarily be entirely accurate for aeolian successions that are slightly older or younger than the age shown in the maps. For example, the early Cretaceous Botucatu Sandstone of the Paraná Basin of Brazil and the Etjo Sandstone of north-western Namibia are considered to represent preserved portions of the same erg system that developed prior to the onset of opening of the South Atlantic, yet the palaeogeography map shown depicts an interval in the Late Cretaceous, shortly after the onset of the opening of the south Atlantic.

Figure 8. Palaeogeographic distribution of siliciclastic aeolian sand systems during: (A) Eocene; (B) Miocene; and (C) last glacial maximum (18 ka). Palaeogeographic maps from Scotese (2001) PAEOMAP Project.

Figure 9. Location of major low-latitude and mid-latitude inland sand seas and dunefields, as well as coastal carbonate aeolianite deposits. After Sun & Muhs (2007).

Figure 10. Tectonic setting of major inland sand seas and dune fields Tectonic data from Fugro Tellus Sedimentary Basins of the World Map http://www.datapages.com/AssociatedWebsites/GISOpenFiles/FugroTellusSedimentaryBasinsoftheWorldMap.aspx. See also Table S5.

Figure 11. Example of a wet aeolian system: White Sands, New Mexico, USA: (A) cross-section of dunefield showing relations between aeolian and lacustrine sediments (after Kocurek *et al.*, 2007); (B) crescentic ridges migrating across surface of older crescentic dune deposits shown by exposed cross-bedding; (C) trench in interdune area showing cross-bedded dune structures.

Figure 12. Example of a dry aeolian system: Namib Sand Sea, Namibia. Cross-section of central sand sea at 24.2°S; elevation data from ASTER GDEM; extent of Tsondab Sandstone Formation after Ward (1988). Landsat image of area for comparison.

Figure 13. Example of a stabilizing aeolian system: Strzelecki Desert, Australia. Along-dune profile showing multiple episodes of dune accumulation spanning the past 120 ka. From Cohen et al. (2010). 'II' to 'VII' – Marine isotope stages.

Figure 14. Ground-penetrating radar mapping of sedimentary architecture; coastal dunefield at Lodbjerg, Denmark (Clemmensen *et al*., 2001a). The surface of the modern dunefield is situated *ca* 15 m above sea-level, and the dunefield is truncated by a coastal cliff toward the North Sea. Partly active cliff-top dunes are developed along the cliff.

Figure 15. Coastal cliff section at Lodbjerg, Denmark. A Wechselian till at the base of the section is overlain by 15 to 20 m of aeolian sand with peaty soils (dark horizons). Terminology of aeolian events after Clemmensen *et al.* (2009).

Figure 16. Simplified aeolian stratigraphy, Lodbjerg coastal dunefield. Note that most previous events of aeolian sand movement and transgressive dune formation led to the accumulation of aeolian sand-sheet deposits. Cross-bedded strata of inland migrating parabolic dunes are only preserved locally. The modern cliff-top dunes have eroded part of the underlying aeolian succession. Radiocarbon dating of the peaty soils and optically stimulated luminescence dating of the aeolian sand indicate that major periods of aeolian sand movement and transgressive dune formation were initiated at *ca* 2200 BC, 700 to 800 BC and AD 1100 to 1200. The cliff-top dunes have formed since about AD 1900. Terminology of aeolian events after Clemmensen *et al*. (2009).

Figure 17. Valley-head sand ramp accumulation, north-west Sardinia (Andreucci et al., 2010a). The sand ramp is composed of marine bioclastic-rich sand that was blown inland by dominant north-westerly winds *ca* 75,000 years ago. Several episodes of ascending dune accumulation are revealed in the cliff section.

COMPLEMENTARY MATERIAL. FIGURE CAPTIONS.

Table S1. Compilation of main Archean, Palaeoproterozoic, Mesoproterozoic and Neoproterozoic aeolian sand systems.

Table S2. Compilation of main Cambrian, Ordovician, Silurian, Devonian, Carboniferous and Permian aeolian sand systems.

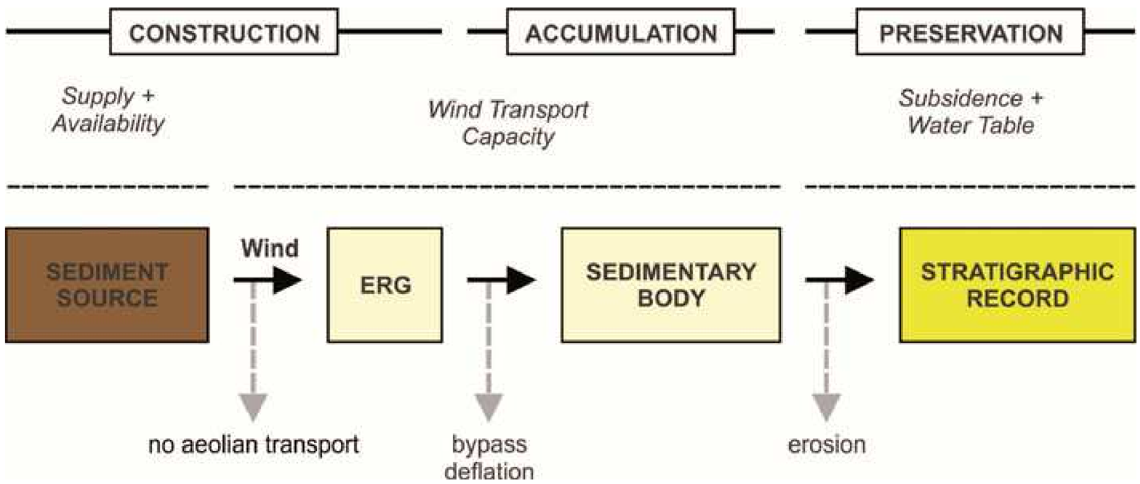
Table S3. Compilation of main Triassic, Jurassic and Cretaceous aeolian sand systems.

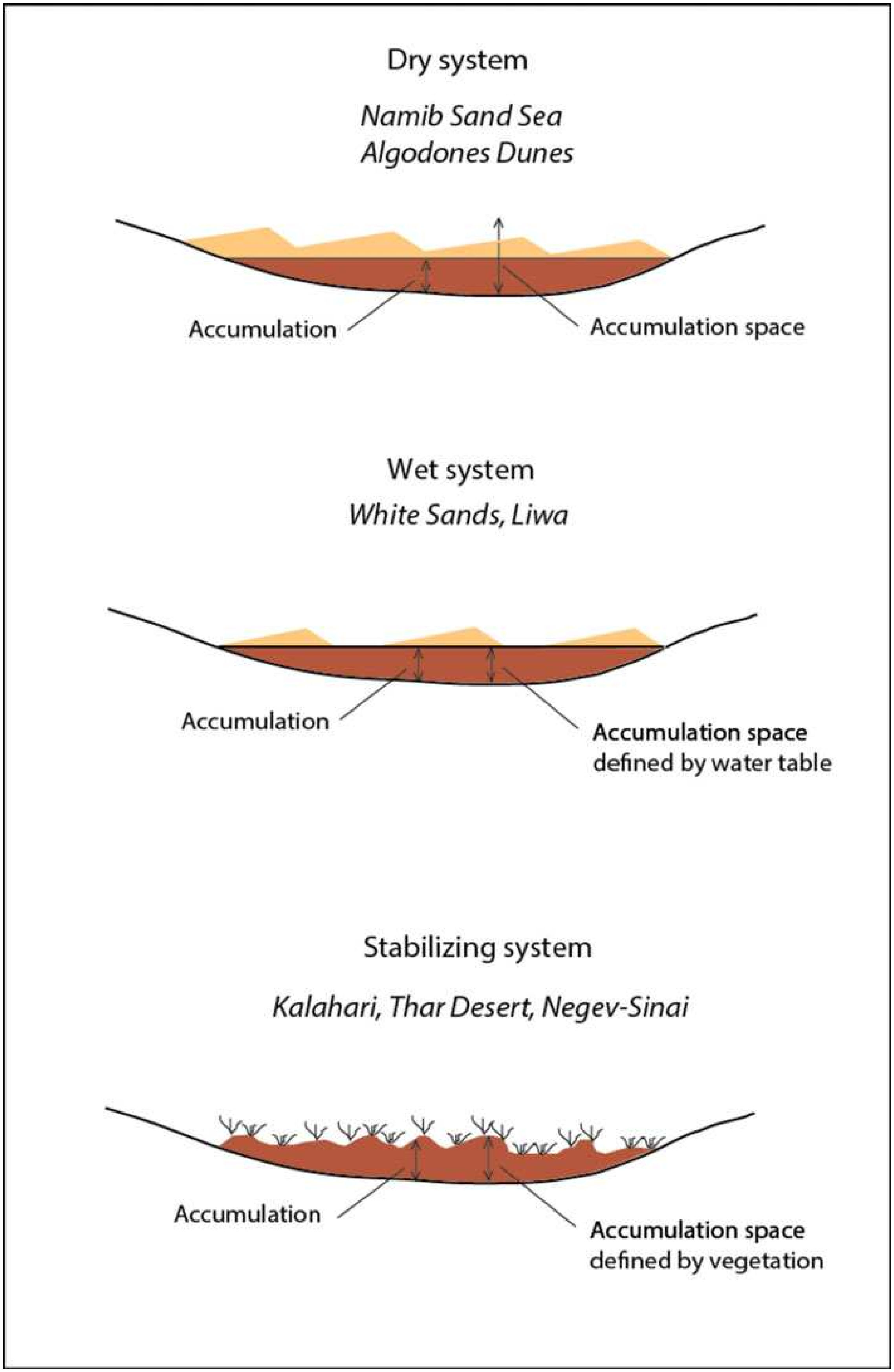
Table S4. Compilation of main Palaeogene and Neogene aeolian sand systems.

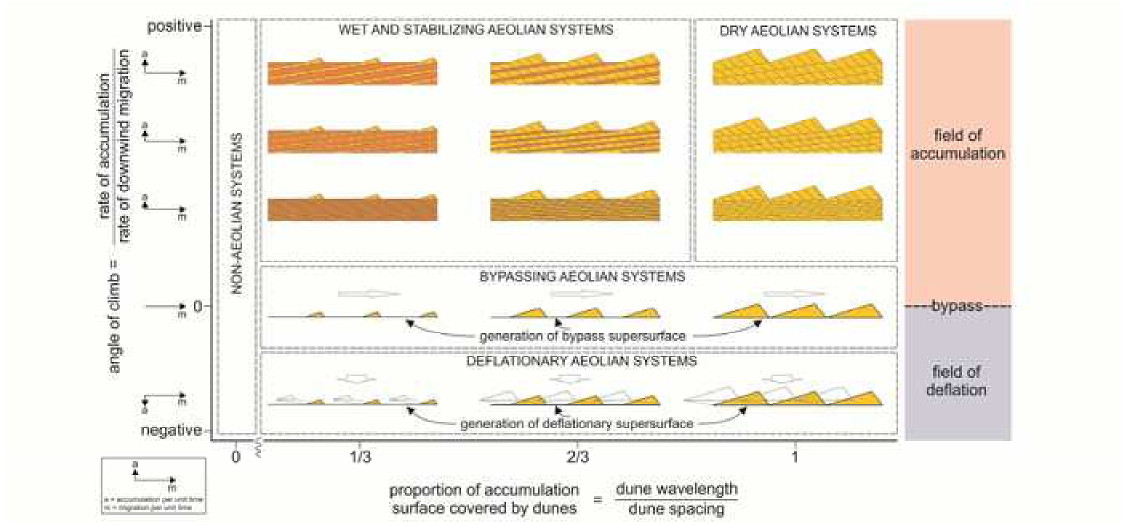
Table S5. Compilation of main Quaternary inland dunefiled (ergs).

Table S6. Compilation of main Quaternary coastal dunefields.

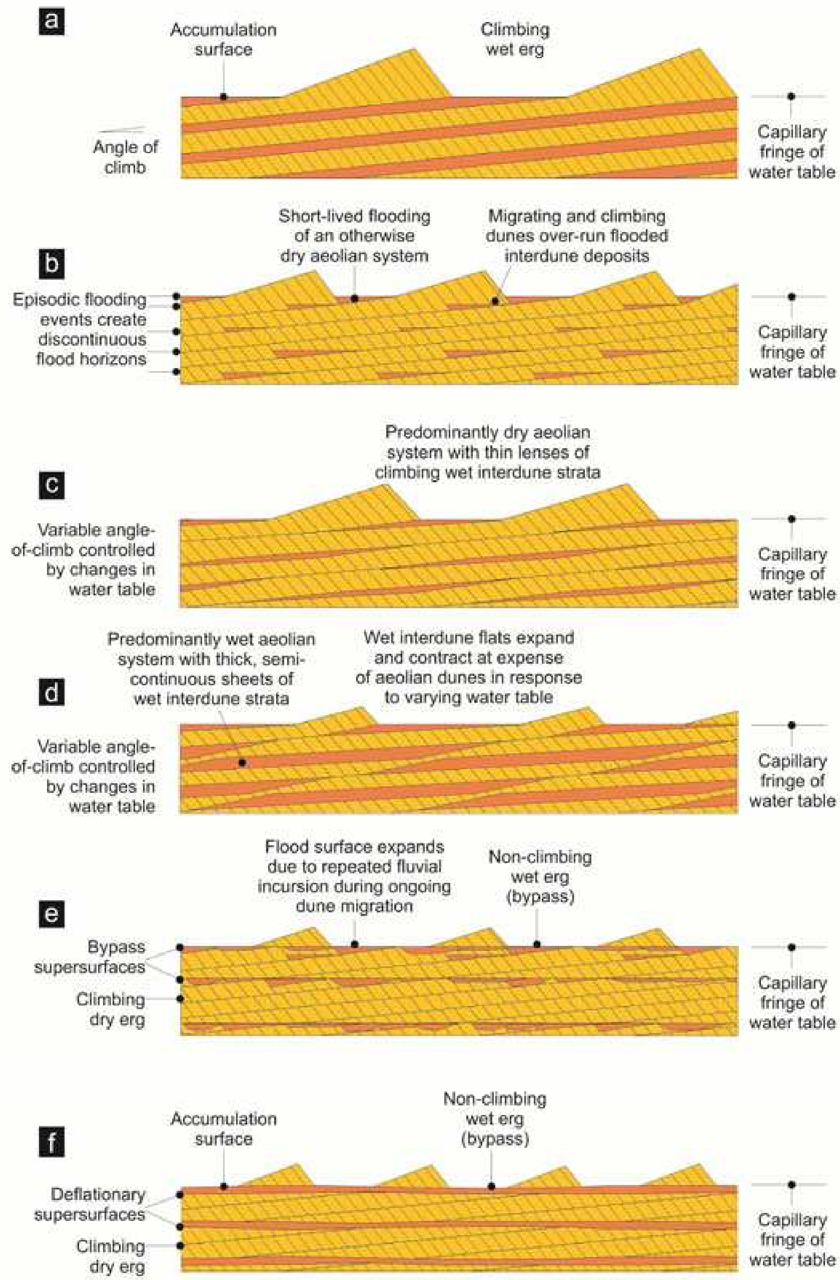
S7. References text and tables.


CONSTRUCTION
ACCUMULATION
PRESERVATION
Supply + Availability
Wind Transport Capacity
Subsidence + Water Table
SEDIMENT SOURCE
Wind
ERG
SEDIMENTARY BODY
STRATIGRAPHIC RECORD
no aeolian transport
bypass deflation
erosion

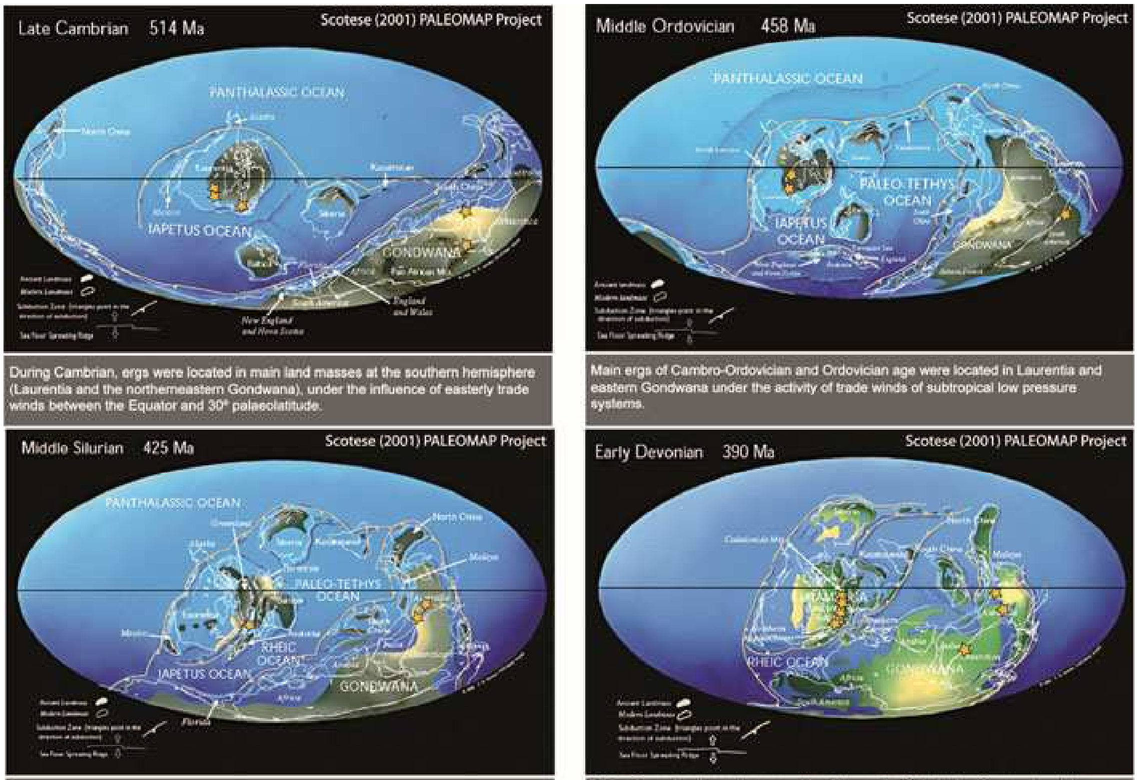

Dry system
Namib Sand Sea
Algodones Dunes
Accumulation
Accumulation space
Wet system
White Sands, Liwa
Accumulation
Accumulation space
defined by water table
Stabilizing system
Kalahari, Thar Desert, Negev-Sinai
Accumulation
Accumulation space
defined by vegetation



positive
negative
angle of climb = rate of accumulation / rate of downwind migration
NON-AEOLIAN SYSTEMS
WET AND STABILIZING AEOLIAN SYSTEMS
DRY AEOLIAN SYSTEMS
BYPASSING AEOLIAN SYSTEMS
generation of bypass supersurface
DEFLATIONARY AEOLIAN SYSTEMS
generation of deflationary supersurface
field of accumulation
bypass
field of deflation
0
1/3
2/3
1
proportion of accumulation surface covered by dunes = dune wavelength / dune spacing



a
Accumulation surface
Climbing wet erg
Angle of climb
Capillary fringe of water table
b
Short-lived flooding of an otherwise dry aeolian system
Migrating and climbing dunes over-run flooded interdune deposits
Episodic flooding events create discontinuous flood horizons
Capillary fringe of water table
c
Predominantly dry aeolian system with thin lenses of climbing wet interdune strata
Variable angle-of-climb controlled by changes in water table
Capillary fringe of water table
d
Predominantly wet aeolian system with thick, semi-continuous sheets of wet interdune strata
Wet interdune flats expand and contract at expense of aeolian dunes in response to varying water table
Variable angle-of-climb controlled by changes in water table
Capillary fringe of water table
e
Flood surface expands due to repeated fluvial incursion during ongoing dune migration
Non-climbing wet erg (bypass)
Bypass supersurfaces
Climbing dry erg
Capillary fringe of water table
f
Accumulation surface
Non-climbing wet erg (bypass)
Deflationary supersurfaces
Climbing dry erg
Capillary fringe of water table

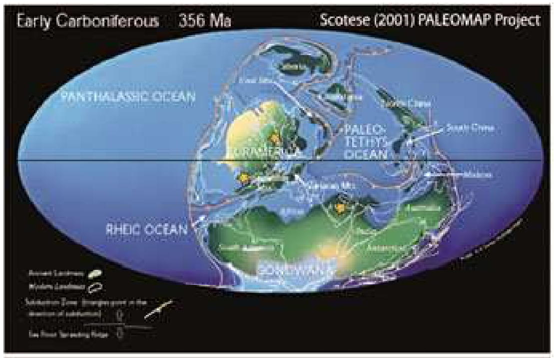


During Cambrian, ergs were located in main land masses at the southern hemisphere (Laurentia and the northerneastern Gondwana), under the influence of easterly trade winds between the Equator and 30° palaeolatitude.

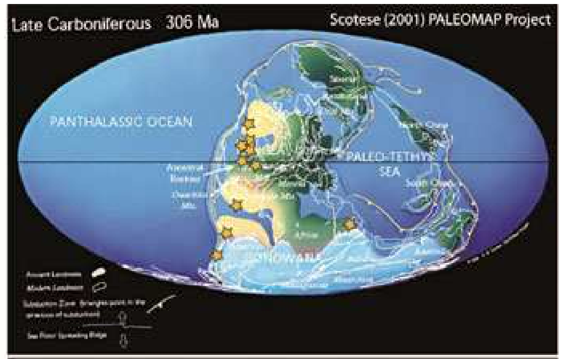
Main ergs of Cambro-Ordovician and Ordovician age were located in Laurentia and eastern Gondwana under the activity of trade winds of subtropical low pressure systems.

Main Silurian and Siluro-Devonian aeolian deposits are located in western/central Australia, located at the southern palaeo-hemisphere desert zone.

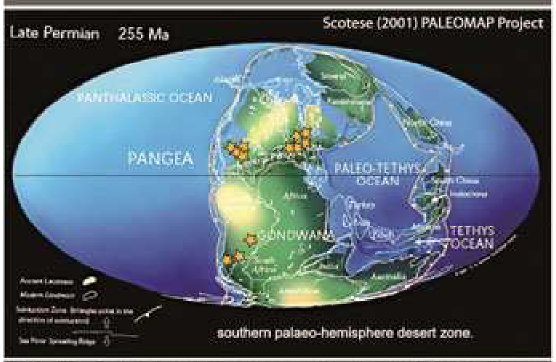
During Devonian times, the assembly of Laurrusia (Caledonian Orogeny) and the northern migration of Gondwana led to an increase of land masses at southern palaeohemisphere-desert zone. Old red Sandstone constitutes main Devonian aeolian record.



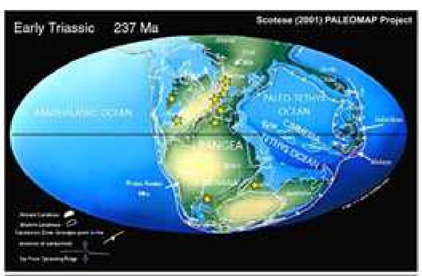
Lower Carboniferous aeolian systems were located in southern Laurrusia and at the northern margin of Gondwana.



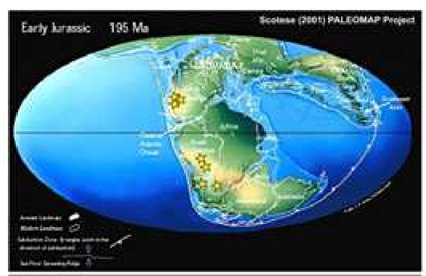
Late Carboniferous-Permian aeolian systems were located in Pangea super-continent both at northern and southern hemisphere. Other aeolian systems developed in northern Gondwana (Bolivia and Argentina).



During Permian times extensive erg systems developed on Pangaea, favored by the location of this supercontinent in subtropical latitudes under the influence of the southern and northern palaeo-hemisphere desert zones.

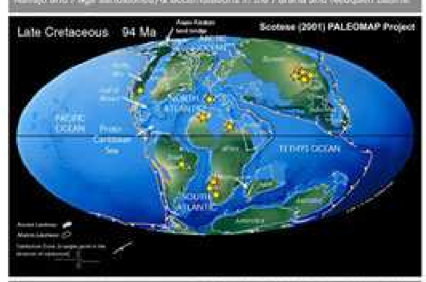


Triassic erg systems occupied a broad belt across the tropical latitudes of northern Pangea in areas now occupied by Europe and North America. Other southern hemisphere erg systems are recorded from Gondwana (S. Africa and India).

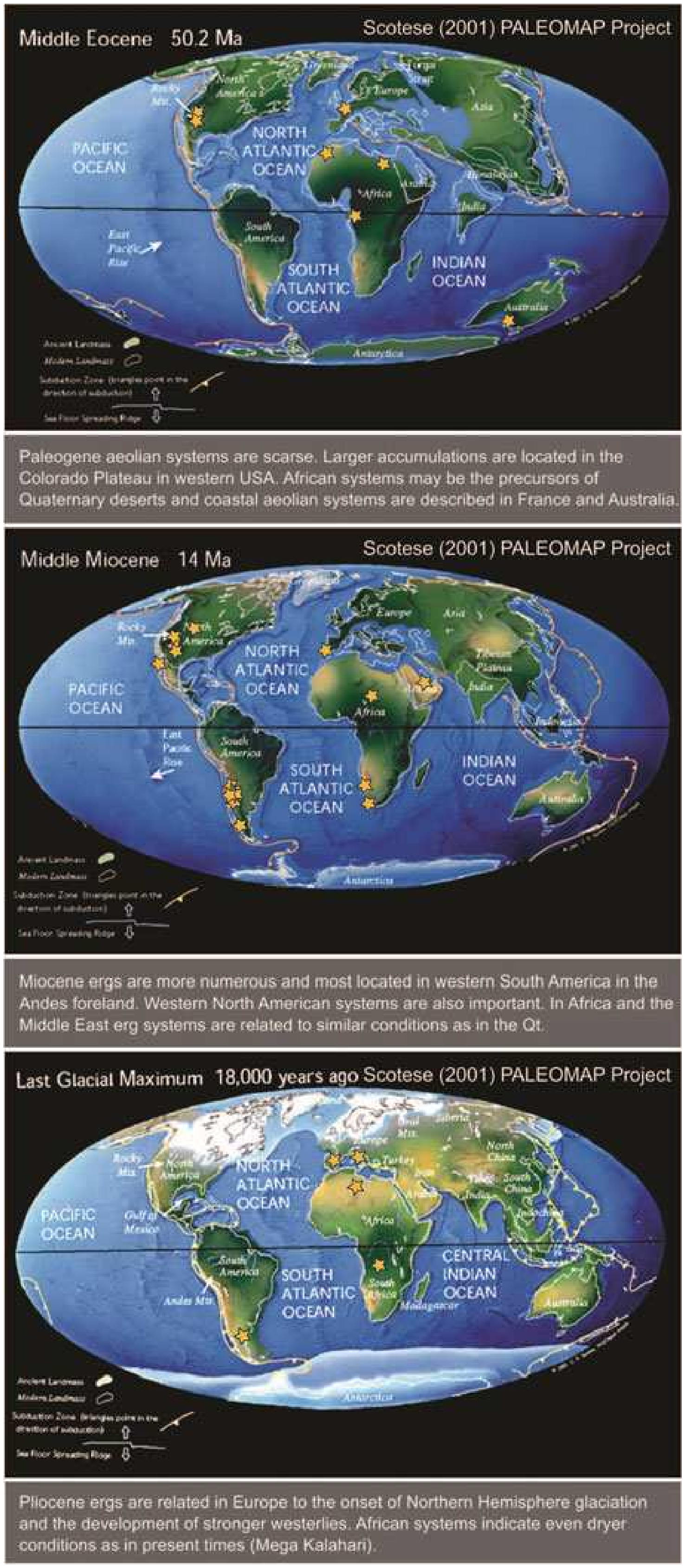


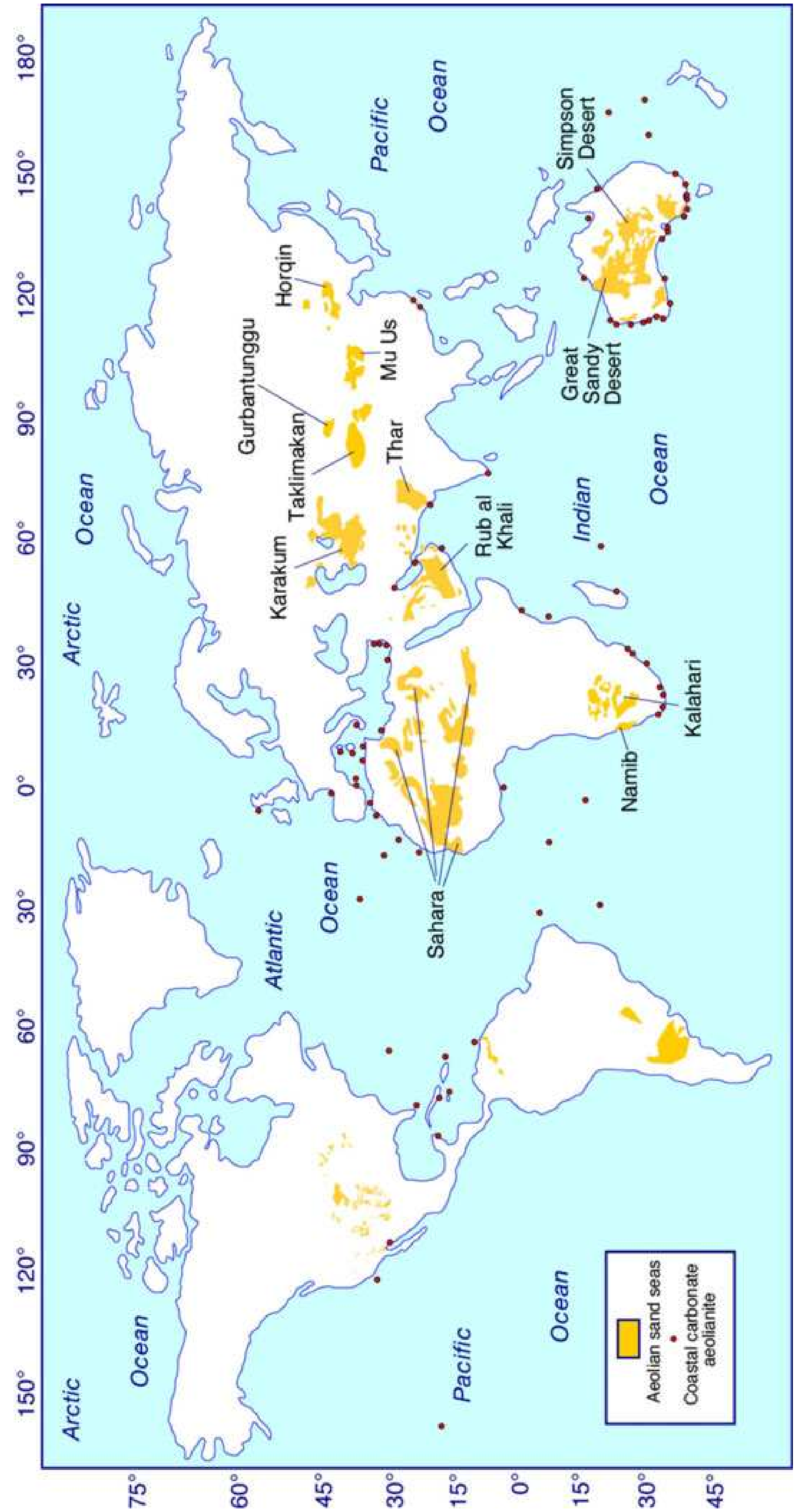
The main erg systems of the Early Jurassic include the geographically widespread accumulations of the Glen Canyon Group in the western US (including the Wingate, Navajo and Page sandstones) & accumulations in the Parana and Neuquen basins.

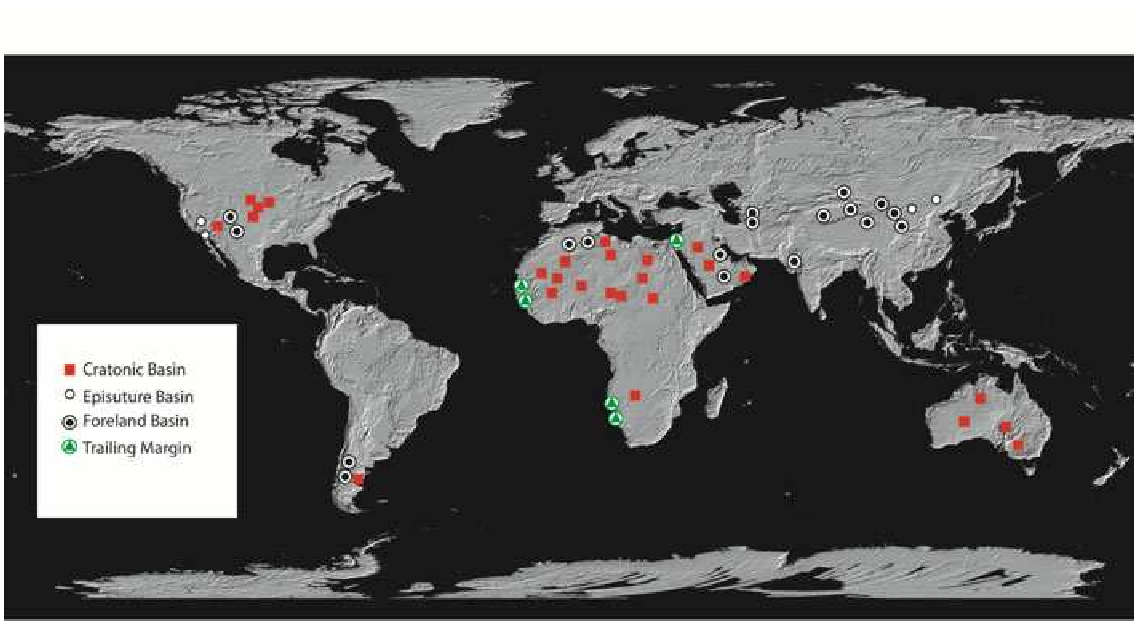
The main erg systems of the Late Jurassic include the geographically widespread accumulations of the San Rafael Group in the western US (including the Entrada Sandstone) & accumulations in the Parana Basin (Brazil) and southern Africa.



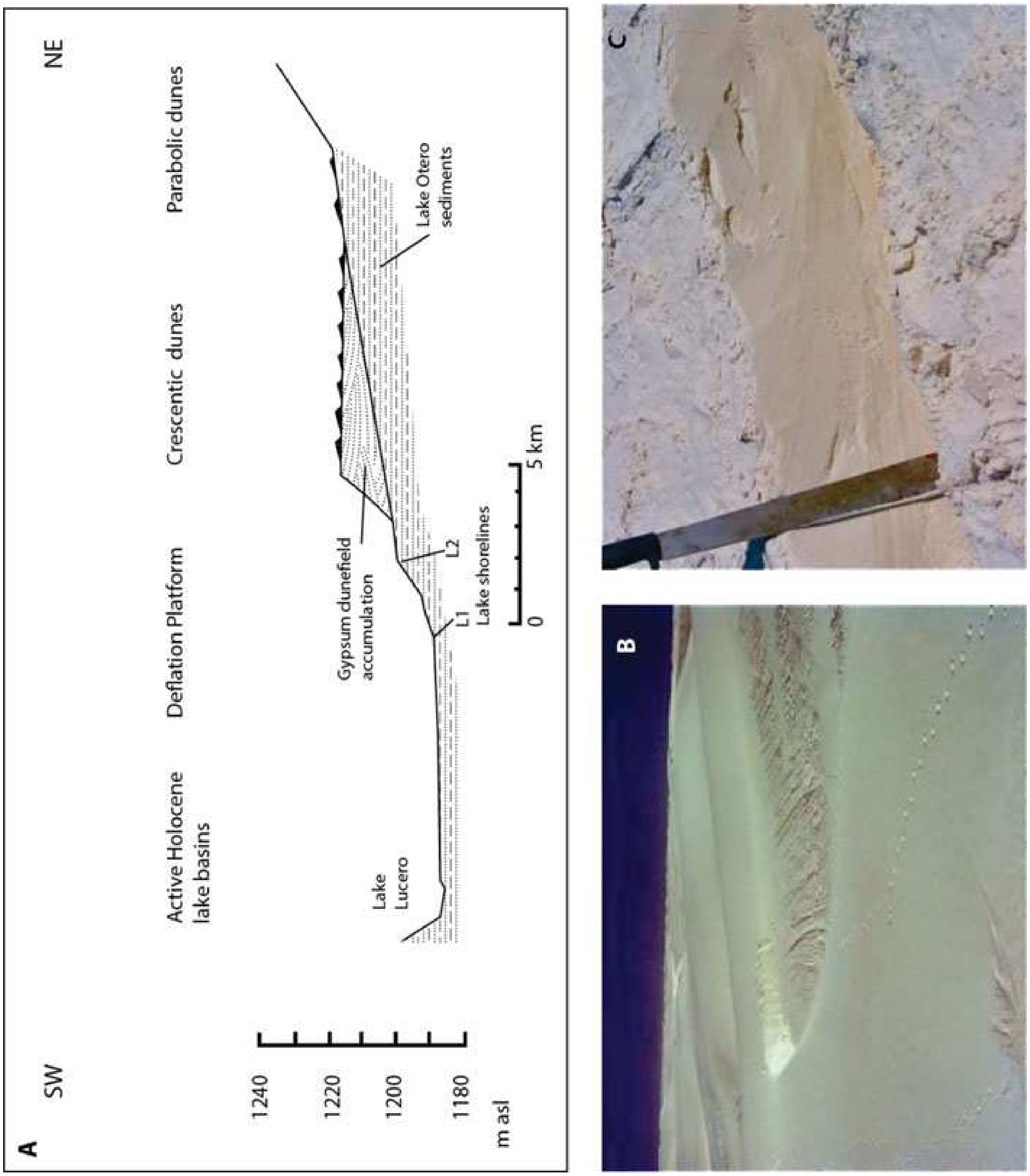
Cretaceous erg systems were geographically widely distributed in tropical and mid-latitudes in both the northern and southern hemispheres, with major dune fields occupying SW Africa & central Brazil prior to and immediately following the onset of break-up of West Gondwana. Additional major ergs were located in northern China


Middle Eocene 50.2 Ma
Scotese (2001) PALEOMAP Project
PACIFIC OCEAN
NORTH ATLANTIC OCEAN
SOUTH ATLANTIC OCEAN
INDIAN OCEAN
Paleogene aeolian systems are scarse. Larger accumulations are located in the Colorado Plateau in western USA. African systems may be the precursors of Quaternary deserts and coastal aeolian systems are described in France and Australia.
Middle Miocene 14 Ma
Scotese (2001) PALEOMAP Project
PACIFIC OCEAN
NORTH ATLANTIC OCEAN
SOUTH ATLANTIC OCEAN
INDIAN OCEAN
Miocene ergs are more numerous and most located in western South America in the Andes foreland. Western North American systems are also important. In Africa and the Middle East erg systems are related to similar conditions as in the Qt.
Last Glacial Maximum 18,000 years ago
Scotese (2001) PALEOMAP Project
PACIFIC OCEAN
NORTH ATLANTIC OCEAN
SOUTH ATLANTIC OCEAN
CENTRAL INDIAN OCEAN
Pliocene ergs are related in Europe to the onset of Northern Hemisphere glaciation and the development of stronger westerlies. African systems indicate even dryer conditions as in present times (Mega Kalahari).

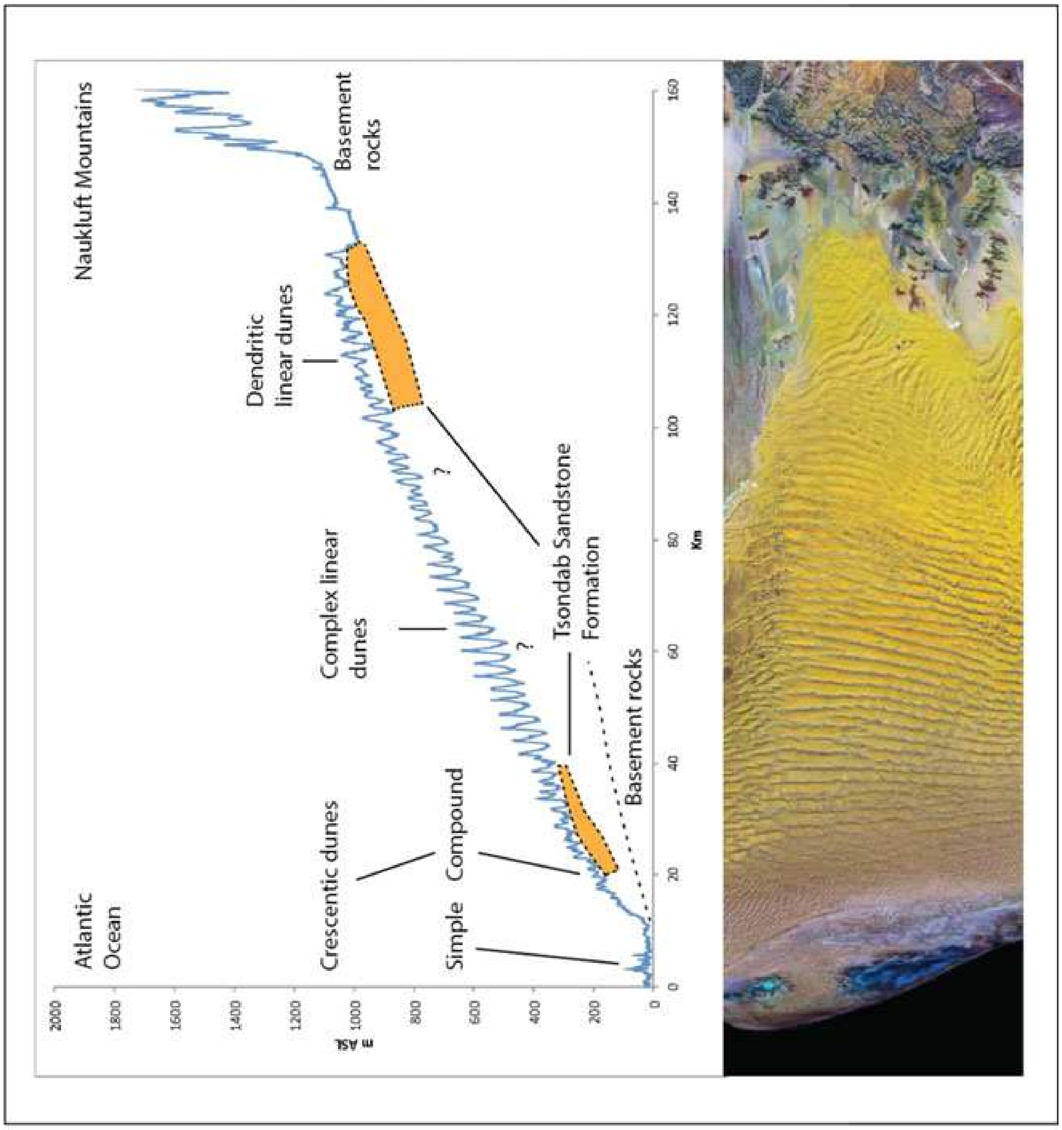

Arctic
Ocean
Pacific
Ocean
Atlantic
Ocean
Arctic
Ocean
Pacific
Ocean
Indian
Ocean
Sahara
Karakum
Taklimakan
Gurbantunggu
Thar
Mu Us
Horqin
Rub al
Khali
Namib
Kalahari
Great
Sandy
Desert
Simpson
Desert
Aeolian sand seas
Coastal carbonate
aeolianite



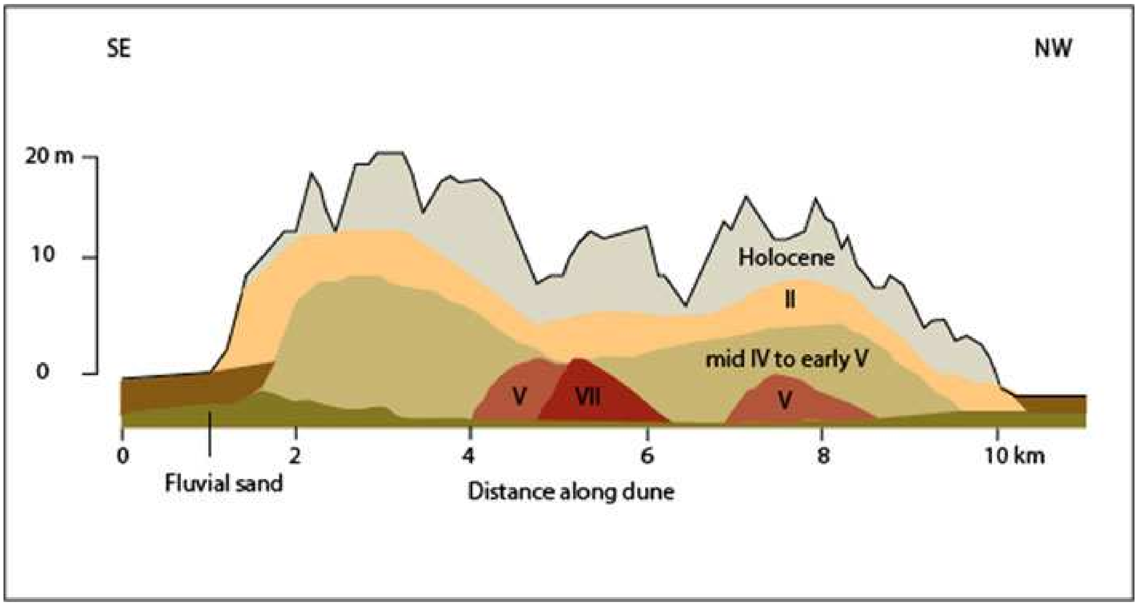
Cratonic Basin
Episuture Basin
Foreland Basin
Trailing Margin



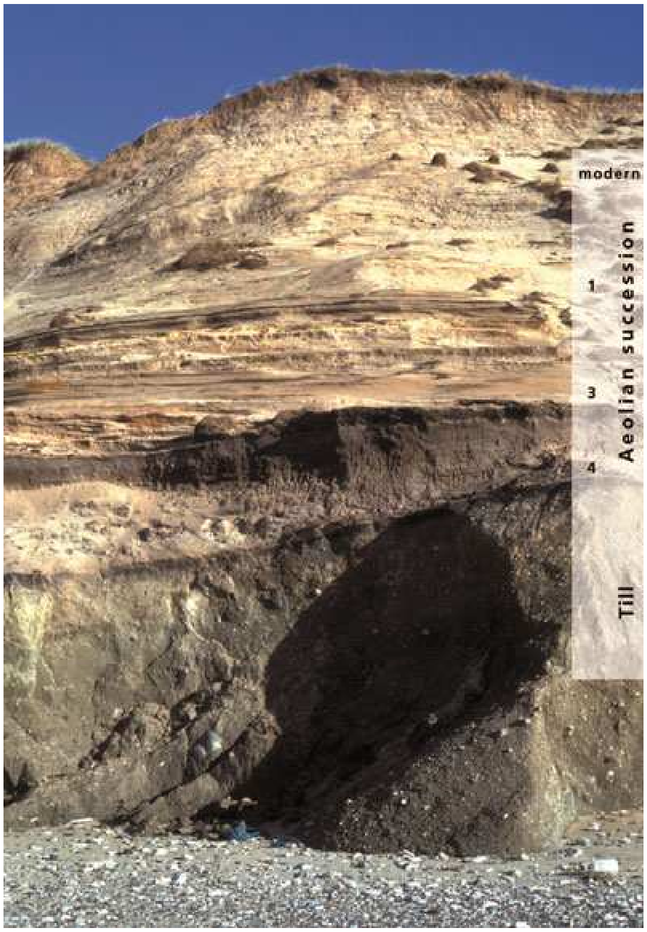
A
SW
NE
Active Holocene lake basins
Deflation Platform
Crescentic dunes
Parabolic dunes
1240
1220
1200
1180
m asl
Lake Lucero
Gypsum dunefield accumulation
L1
L2
Lake shorelines
Lake Otero sediments
0
5 km
B
C

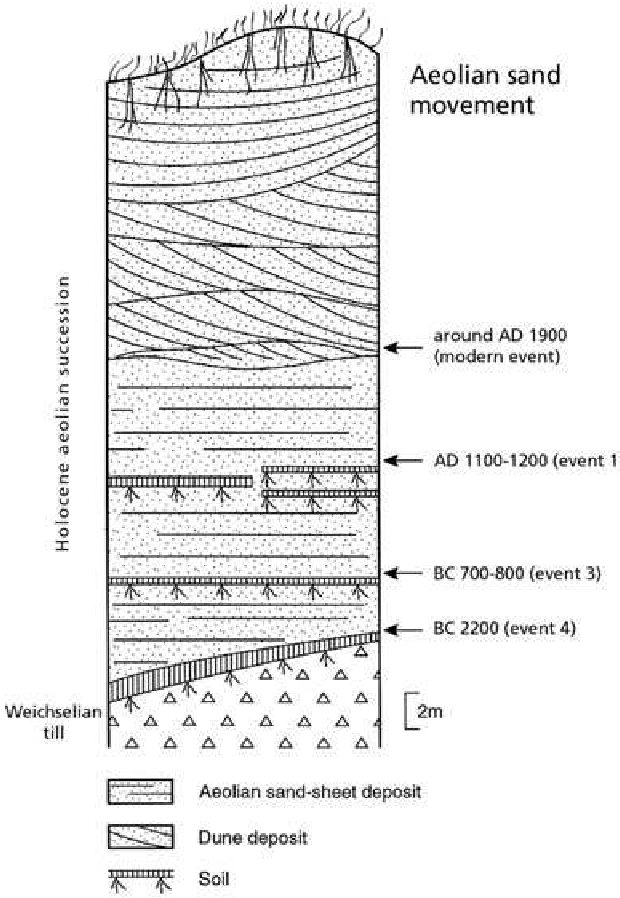


Naukluft Mountains
Basement rocks
Dendritic linear dunes
Complex linear dunes
Tsondab Sandstone Formation
Crescentic dunes
Simple
Compound
Basement rocks
Atlantic Ocean
m ASL
Km


SE
NW
20 m
10
0
Holocene
II
mid IV to early V
V
VII
V
0
2
4
6
8
10 km
Fluvial sand
Distance along dune


N
S


modern
Aeolian succession
1
3
4
Till



Aeolian sand movement
Holocene aeolian succession
around AD 1900 (modern event)
AD 1100-1200 (event 1)
BC 700-800 (event 3)
BC 2200 (event 4)
Weichselian till
2m
Aeolian sand-sheet deposit
Dune deposit
Soil


NW
SE