

2024-04-06

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du Pont, SC

<https://pearl.plymouth.ac.uk/handle/10026.1/22473>

10.1016/j.earscirev.2024.104772

Earth-Science Reviews

Elsevier

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Complementary classifications of aeolian dunes based on morphology, dynamics, and fluid mechanics

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September 21, 2023

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Abstract

Dunes form where winds blow over a bed of mobile sediment grains – conditions that are common in our solar system. On Earth, dunes abound in arid continental interiors and along sandy coastlines. Dune fields have also been recognized on other planetary bodies, including Venus, Mars, Saturn’s moon Titan, and Pluto. Despite the relatively basic conditions required for their formation, dunes adopt a rich diversity of shapes, sizes, and behaviors in response to their boundary conditions and other environmental forcings. Thus, people around the globe and over centuries have developed a rich vocabulary to describe dunes and their complexity. In addition, many studies have been devoted to link dune shape to environmental forcings, usually by means of correlations. As a result, existing dune nomenclature often includes redundant terms with differing definitions across scientific communities. Although a worthy first step, correlation-based classifications can be misleading if not based on an underlying mechanics and if dune morphogenetic classes are not uniquely defined. Here, we synthesize existing dune terminology and put the last two decades of research on dune morphodynamics in perspective in proposing three simplified dune classification

schemes based on the state-of-the-art understanding of dune morphology, morphogenetic processes and coupling between sand bed, fluid flow and sediment transport. Together, these classifications provide a unified framework for geomorphologists, sedimentologists, geographers, physicists, and others to describe windblown sand dunes on Earth and beyond through their shape, dynamics, and size as a response to winds and boundary conditions.

1 Introduction – Previous Classifications and Aims

Dunes propagate and develop through the action of wind, constrained by other factors such as topography and vegetation. They are not only the result of present winds, but can integrate the history of winds including seasonal wind cycles and longer-term changes. This property helps to explain the richness of shapes and scales observed, and makes dunes witnesses of past winds and conditions. On Earth, dunes are used to study paleoclimates and test global circulation models [1, 2, 3]. The sedimentary structure, shape and orientation of dunes are also used to constrain climate models and the history of celestial bodies such as Venus, Mars and Titan, for which little climate data are available but where dunes are observed [4, 5, 6, 7, 8, 9, 10, 11, 12, 13, 14].

The discovery of planetary dunes and the use of comparative geomorphology make it increasingly necessary to define a dune in terms of physical processes, in particular to compare their sizes in different environments. Such physical definitions are especially critical when, for example, the scales and morphologies of different types of bedforms may overlap.

While sand transport mechanisms, sizes, and characteristic times may be different from one environment to another, dune shapes are similar (Fig. 1). General formation processes seem insensitive to the details of sediment transport, and such over-arching processes tend to prevail in the establishment of dune shape and the development of dune patterns and arrangement within fields. This area of research has advanced considerably in recent years through a combination of field studies, laboratory experiments, and models, and it is now possible to quantitatively relate wind regimes to dune shapes and dynamics.

The purpose of classification is to describe the relationship between different objects in such a way as to simplify these relationships and facilitate generalization [16]. Despite the overwhelming variety of terms used to denote dune types (table 1), there is an assumption that this variety collapses to a much smaller number of common groups, and often that dunes not only look different from each other, but are formed in different ways. Thus classification may help us to use a name to convey morphology and thereby comprehend dune formation processes and dynamics. Defining the categories into which dunes should be classified, however, is challenging. The debate is an old one, and very closely analogous to the long-standing biological taxonomic debates, using Charles Darwin’s terms, between ‘lumpers’ (those that seek to classify into as few categories as possible, allowing for more variability within each class) and ‘(hair) splitters’ (those that tend to permit more categories, but allow for less variability within a class). An ideal classification scheme should be comprehensive, mutually exclusive, internally consistent, unambiguous and easy to use, with morphology providing information about genesis. These ideals can be hard to reconcile when dealing with natural phenomena.

Often, different terms have developed in separate geographical locations for essentially the same phenomenon. In the early days of the study of dune geomorphology, this was perhaps more understandable, as the true global extent of some landforms would have been unknown, and thus locally-used terms were applied to landforms which were new to the eyes of (largely European) science; hence, for instance, the abundance of terms derived from the Arabic language (*e.g.* seif, draa). Yet, in some instances, regional terminology has persisted. For instance, the English-language term ‘pyramid dune’ (to describe a large dune with multiple crestlines, typically leading to a central highpoint) recently has been used almost exclusively in the Chinese-based literature, and the synonymous term ‘star dune’ is found more commonly elsewhere.

Early classifications of dunes were based primarily on field-based research, and thus were generally based on work in specific geographical locations, such as Sokolow’s work in Russia [22], Melton’s work in the High Plains, USA [23], Smith’s work in Nebraska [24] and Cooper’s work along the west coast of the USA [25, 26]. Some of these early classifications were reviewed by Mainguet [27], but in general remained regional in scope and therefore sometimes lacked more general application. That said, Aufrère did attempt some more global qualitative synthesis, based not only on his own studies in the Algerian Sahara, but also on the work of Vaughan Cornish and Richard Oldham (India), Sven Hedin (central Asia) and Cecil Madigan (Australia) [28]. From the outset, there were frequent endeavours to link formative factors to landform, and one of the first to attempt this semi-quantitatively was Hack, based on his work in northern Arizona, USA [29]. The resultant ternary diagram, based on wind strength, sand supply and vegetation

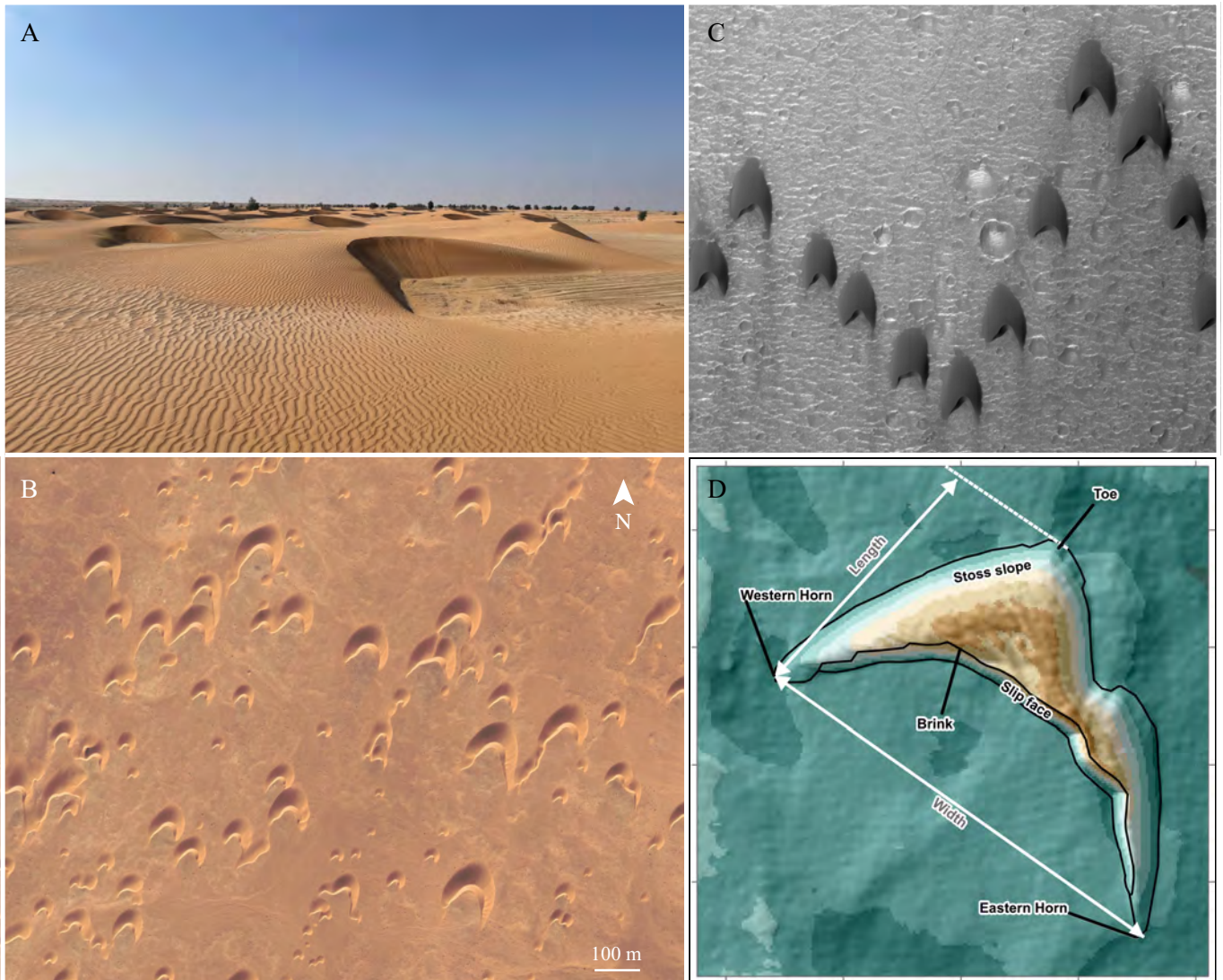


Figure 1: Barchan dunes in different environments. A: Barchan dunes in Oman (20.75°N, 57.611°W). B: Barchan dunes in Occidental Sahara (26.758°N, 13.378°E), date: 11/2021. C: Barchan dunes on Mars. D: Bathymetry image of a marine barchan dune off the east coast of the UK in the North Sea from [15]. The width scale of the image is 410 m. Credits: M. Lapôtre (A), Maxar Technologies (B), NASA/JPL-Caltech/University of Arizona (C), [15] (D).

66 cover worked well in that region, but lacked global applicability.

67

68 Developing a global (and possibly universal) classification required the global scope that satellite-borne remote
 69 sensing provided. The first attempt at a general characterisation of dune forms from global examples was that of
 70 McKee and co-workers [18] facilitated by the global coverage of Landsat imagery that became available in the 1970s.
 71 McKee also brought a body of work on the internal structure of dunes to this study, which was used to identify sand
 72 dune types based on their process of construction. Various versions of the McKee scheme have been subsequently
 73 adopted particularly in desert and aeolian geomorphology textbooks (e.g. [30]). McKee also introduced a terminology
 74 to allow for the co-existence of dunes at different scales [18]: *simple dunes* – single dunes or a single dune type;
 75 *compound dunes*, dunes of the same type at two or more scales; *complex dunes*, dunes of more than one type, usually
 76 at different scales.

77 While McKee's classification distinguished a number of fundamental forms on the basis of morphology and internal
 78 structure, other classifications, often developed as pedagogic tools for textbooks, have used a branching (dendritic)
 79 structure to show the interrelationships between dune types. The first of these was developed by Pye and Tsoar ([19],

abyssal	aklé	anchored	barchan	barchanoid
blowout	chaots	chequerboard	chevron	clay
cliff-top	climbing	complex	compound	coppice
crescentic	demkha	dome	draa	echo
ellipsoidal	elongating	embryo	falling	foredune
free	frontal	ghourd	hairpin	hummock
lee	linear	longitudinal	lunette	megadune
nail	nebkha	network	oblique	obstacle
oghroud	pancake	parabolic	parallel	pyramidal
phytogenetic hillock	polygonal	precipitation ridge	raked	relic/relict
retention ridge	rhourd	sand massif	sand mountain	sand ramp
shadow	seif	simple	snow	source-bordering
star	stellate	straight-crested-asymmetric	teardrop	terminal
transverse	true	unvegetated	vegetated	zalib
zibar				

Table 1: Dune type terminology. Terms are drawn from a number of English-language sources, although some terms used in English originate from other languages. Key sources for terms are [17, 18, 19, 20, 21].

p.162, Fig 6.7) and was subsequently modified by Cooke *et al.* [31] and then by Livingstone and Warren [17]. Each of these versions uses the distinction between *free* and *anchored* dunes as a starting point. *Free dunes* (called “true dunes” by Bagnold ([32] p.188) exist because of a lack of inhibition to the movement of particles. These are fundamental bedforms. Conversely, *anchored dunes* (which Bagnold termed “sand shadows and sand drifts”) owe their existence to the presence of an inhibition to particle movement, most usually topography or vegetation (see review by Hesp and Smyth p. 157-178 in [17]).

At the heart of geomorphological studies of dune form is the relationship between morphology and wind regime. The work of Wasson and Hyde [33], evaluated by Bullard and Livingstone [34], was a first attempt to use empirical data (largely from Australia) to distinguish dunes on the basis of available sand (measured by Wasson and Hyde as an estimate of the depth of available sand if spread as a sheet of uniform thickness) and the directional variability of the wind regime. Although this pioneering work provided a useful scheme for future studies, the limited data (largely from Australia), resulted in restricted ranges of dune type. Consequently, the results suggested that longitudinal dunes were restricted to areas with little sand. Using the same general approach, Livingstone and Warren [17] and Bishop *et al.* [35] revised this conclusion and showed that linear dunes can form where sand coverage is extensive. This scheme was subsequently adapted as the basis for a classification by Lancaster who added his own data from a wider range of sand seas to that of Wasson and Hyde [36, 37]. Lancaster added “other key parameters” to sand supply and wind regime.

Although many schemes have related dune form to the ‘resultant’ wind directions or sand-transport directions, Hunter *et al.* [38] and Rubin and Hunter [39, 40] were keen to point out that terminologies, and therefore classifications, implying those relationships were problematic. Their contention was that dunes can form with orientations that are neither parallel nor perpendicular to the resultant transport direction. Thus, it is misleading to use terms such as longitudinal and linear interchangeably: *linear* denotes a form in which one planform dimension greatly exceeds the other and is therefore a morphological description, whereas *longitudinal* implies an orientation that is approximately parallel to net sand transport. Many linear dunes are not aligned with the resultant transport directions but are oblique, *i.e.*, the planform long axis is at 15-75 degrees from the resultant sand transport direction [38]. Rubin and Hunter showed that dune orientation where the bed is fully covered with sand does not depend on the resultant transport direction, but on what they termed gross bedform normal transport [40]. The more recent experimental and theoretical work of Courrech du Pont *et al.* bridged the gap between the older studies that classified dunes at least partially on the basis of sand availability, and the new studies that showed that dune orientation is controlled by two mechanisms, one of which dominates where the bed is fully covered with sand and the other where the bed is partially starved [41].

Although many of the classification schemes discussed so far owe their origins to inland desert dunefields, throughout this period there was also development of an understanding of coastal dunes, as well as increasingly sophisticated work

115 from the physics community aimed at understanding fluid entrainment, transport and deposition of granular materials
116 and their coupling with the topography. Despite the fundamental process similarity between coastal and desert
117 environments highlighted by the physical modeling of dunes, some differences in nomenclature had begun to emerge.
118 Thus, whilst parabolic and transverse dunes were frequently discussed in both desert and coastal literature, the term
119 *foredune* is reported almost exclusively from the coastal dune domain (*e.g.* [42]). Conversely, the term *source-bordering*
120 *dune* [43], common in the desert literature to refer to a dune owing its existence to an immediately adjacent sediment
121 supply – which might be said to apply to any coastal dune if it evolves from the backshore like foredunes – is almost
122 absent from the coastal literature.

123 The observation in the early 1970s of dunes on Mars, tentatively from Mariner 6 [44] and more definitively from
124 Mariner 9 [45], opened new opportunities and challenges in dune classification. The subsequent discovery of aeolian
125 bedforms on at least six other solar system bodies (the process of their discovery is reviewed in [46]) emphasizes the
126 seeming near-ubiquity of dunes and other bedforms – indeed, they have so far been found on every solid planetary
127 body with an atmosphere, and some where the ‘atmosphere’ is at best extremely tenuous. There is something of an
128 irony in that in each case, landforms were identified as dunes by analogy with identified, and classified, terrestrial
129 dunes, and yet in some instances, the planetary dunes have thrown up additional complications for classification. For
130 instance, on Mars, barchans were one of the first types of dunes to be identified, and yet we now know there is greater
131 morphological diversity of Martian barchans than those on Earth, and some Martian dune morphologies do not readily
132 fit into McKee’s, or subsequent, classifications [47].

133
134 This study aims to provide a comprehensive classification of dunes that can form the basis for a nomenclature that
135 can be used both with terrestrial and extra-terrestrial systems, building on recent advances in the understanding of
136 those systems. We propose a new classification of dunes through three distinct and complementary classification-tree
137 diagrams that when combined should answer the fundamental questions of how to describe a shape, relate dune shape
138 and size to the external constraints (the wind regime, the environment and the boundary conditions), and infer the
139 dune dynamics. The first tree is based on dune morphology where a step-by-step recognition of shapes, possibly
140 interacting with their surroundings, discriminate the main dune types (Section 2). This tree is connected to the
141 second tree, which is based on mechanisms of dune growth and dynamics that links the shapes to external constraints
142 through morphogenetic processes (Section 3). The third tree is based on the fluid mechanics and sediment transport,
143 which set the characteristic scales and the range of possible sizes that can occur in different planetary environments
144 (Section 4). We illustrate the practical use and relevance of those three classifications through several case studies
145 and examples (Section 5). These different classifications involve many concepts and parameters. We present the
146 main ideas as necessary in the description of the trees and refer to appendix (Section 7) for detailed development
147 and explanation. It is our intention that together these classifications will lead to a convergence in understanding
148 dune-forming processes and dynamics and will provide a platform and nomenclature for future dune studies on Earth
149 and beyond.

150 2 Classification of aeolian dunes based on dune morphology

151 2.1 Purpose and approach

152 The morphology of a dune encompasses the three-dimensional geometry defined by the bedform surface. The readily
153 observable characteristics of dune morphology have been the bedrock of previous dune classification schemes (Section
154 1) and can be investigated either from the ground level, or using aerial photography or satellite imaging. Importantly,
155 classifications based on dune morphology do not require *a priori* knowledge of the local winds. We assembled here
156 a morphology-based dune identification tree (Fig. 2) that aims to assist in the categorization of a given dune based
157 on its observable shape, including crestline morphology in plan-view, and environment. The tree is structured with
158 the goal that someone with introductory training in geography, geomorphology, geology or other related disciplines
159 could identify dune type by answering the series of questions posed in the tree. We attempted to keep the number of
160 questions to the minimum necessary to discriminate between the different types of dunes.

161 The categorization tree in Figure 2 is appropriate to the majority of dunes found in nature. Rather than encom-
162 passing all possible cases, including niche situations or exotic environments, the classification tree aims to broadly
163 categorize and include the very large majority of the dune types found on Earth (Section 2.3.1 for discussion of
164 planetary applications and special cases).

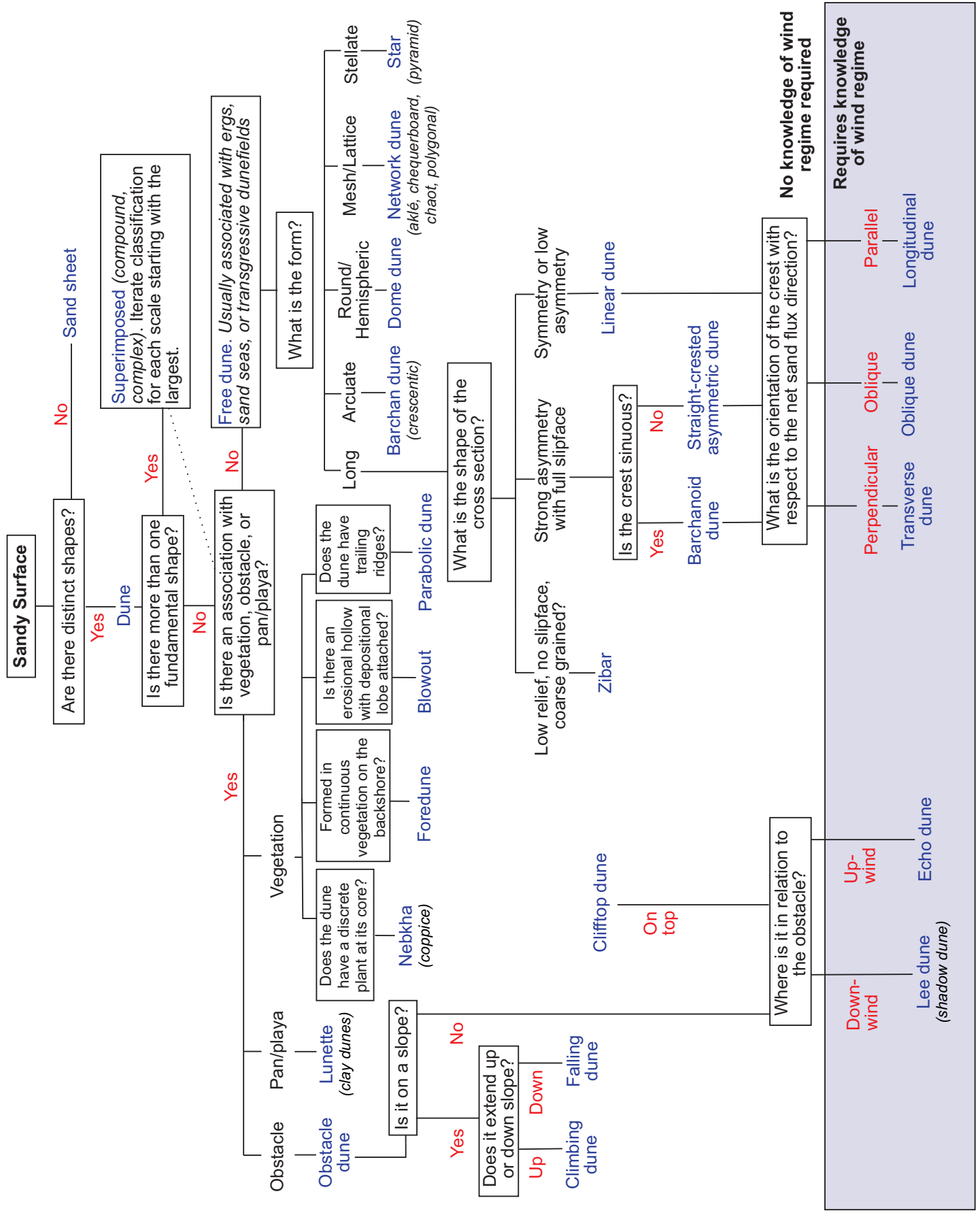


Figure 2: Identification tree for dune classification based on morphology.

2.2 Tree description – Selection of dune type from shape and interactions with surrounding landscape

Below is a walkthrough of the tree in Figure 2, highlighting some nuances in the classification where additional description is helpful. We start by requiring a sandy surface, and immediately differentiate between dunes and non-dunes (*e.g.*, sand sheets) using the presence of distinct shapes, and from there use observations about the morphology in plan-view to determine dune type. Although the vertical morphology of dunes holds important information, the plan-view morphology of dunes exhibits more significant global variability and is both more diagnostically useful and more convenient in part because planform morphology is readily visible in remotely sensed images. In addition, we refer to the crestline of a bedform, which loosely follows the planform morphology. Crestlines are defined by topographic high points along a dune’s upper surface and along with brinklines (*i.e.*, the often sharp transition from stoss to lee face), can be easily observed in aerial images. Once the dune type has been identified, it may imply information about the local wind regime. The inferred wind direction and/or variability is based on previous research and helps connect the identification in this tree to the growth mechanisms and fluid dynamics outlined in the other two classification trees in this work. Dunes may occur in desert sand seas and ergs [48, 49], on beaches of all kinds (*e.g.* on ocean shores, lake shores, playas, rivers and estuaries), and on coastal transgressive dunefields (coastal sand seas or ergs). For sandy surfaces with distinct shapes (dunes), the classification is intended to be applied to a single one of these shapes (morphologies). In many cases, these shapes are likely repeated in the immediate vicinity to form a dunefield. Dunes exhibiting similar shapes at different scales or multiple shapes at different scales are considered compound or complex dunes, respectively, and each morphology and scale can be interrogated separately.

Dune types are differentiated by the immediate non-dune boundary conditions that influence the dune morphology. Dunes whose morphology is integrally related to the presence of a local obstacle, vegetation, or playa are differentiated from dunes not directly related to such specific boundary conditions. Although any dune is influenced by boundary conditions such as local topography, water table, and vegetation [50, 51, 52, 21], we only consider dunes where these local conditions are clearly related to or control the morphology, and consider all other dunes *free dunes*. The former are more commonly found in local settings (*e.g.*, coastal systems, lakes, estuaries, rivers, playa margins), whereas the latter are more likely to populate extensive dune fields, ergs, and sand seas (*e.g.*, as seen in the Rub’ al Khali [18, 53], or Taklimakan deserts [54] and in many coastal transgressive dunefields [42]).

2.2.1 Free dunes

Free dunes (Figs. 1 & 3) are further subdivided based on their plan-view shape. Stellate morphologies with many arms in radial patterns are considered *star dunes*. Mesh or lattice patterns in dune crests indicate *network dunes*. Many other terms for this morphology also exist, but their defining characteristic is always a network of dune crests crossing one another, commonly at near perpendicular angles and at similar scales. Isolated dunes with arcuate shapes, sometimes called crescent- or half-moon shapes, are *barchan dunes*. Mound-shaped, round, and hemispheric sand bodies are called *dome dunes*. Round landform structures may also develop on sand sheets in early stages of development. These proto-dunes are differentiated from dome dunes by low heights and attachment to a sand sheet source. Proto-dunes are not included in the categorization tree because as transitional features between sand sheets and dunes they may not be identifiable as distinct shapes. However, we note them here for disambiguation and completeness [55, 56, 57, 58].

The remaining categories of free dunes are all morphologically ‘long’, meaning their length to width ratio is high and the crestline continuity in plan-view is large with respect to the bedform width or wavelength. The term ‘elongate’ was avoided given the use of the term in the dynamical process-based dune classification (Section 3). To differentiate between dune types with long morphologies in plan-view, additional information is needed on the shape of the cross section and in particular on its symmetry with respect to the crest line. Dunes with a symmetrical or slightly asymmetrical normal-to-crest profile are referred to as *linear dunes*. In dunes with some slight asymmetry in cross section, a slip face may be locally present. Linear dunes by definition exhibit long, straight-crested morphologies. Dunes whose cross section has a strong asymmetry with an extended slip face on one side are *barchanoid dunes* if the crest line is sinuous and simply *straight-crested asymmetric dunes* if not. Long dunes with low relief and coarse grains not developing a slipface are termed *zibars* [59, 60, 61].

Morphologically long dunes can be further classified according to their orientation with respect to the net transport direction. We note that the wind direction discussed when categorizing aeolian dunes should refer specifically to the direction of sand-transporting winds, as sub-threshold winds will not contribute to the mobilization of sand or, therefore, dune development (Section 7.2). Dunes whose crest is perpendicular to the net transport direction are

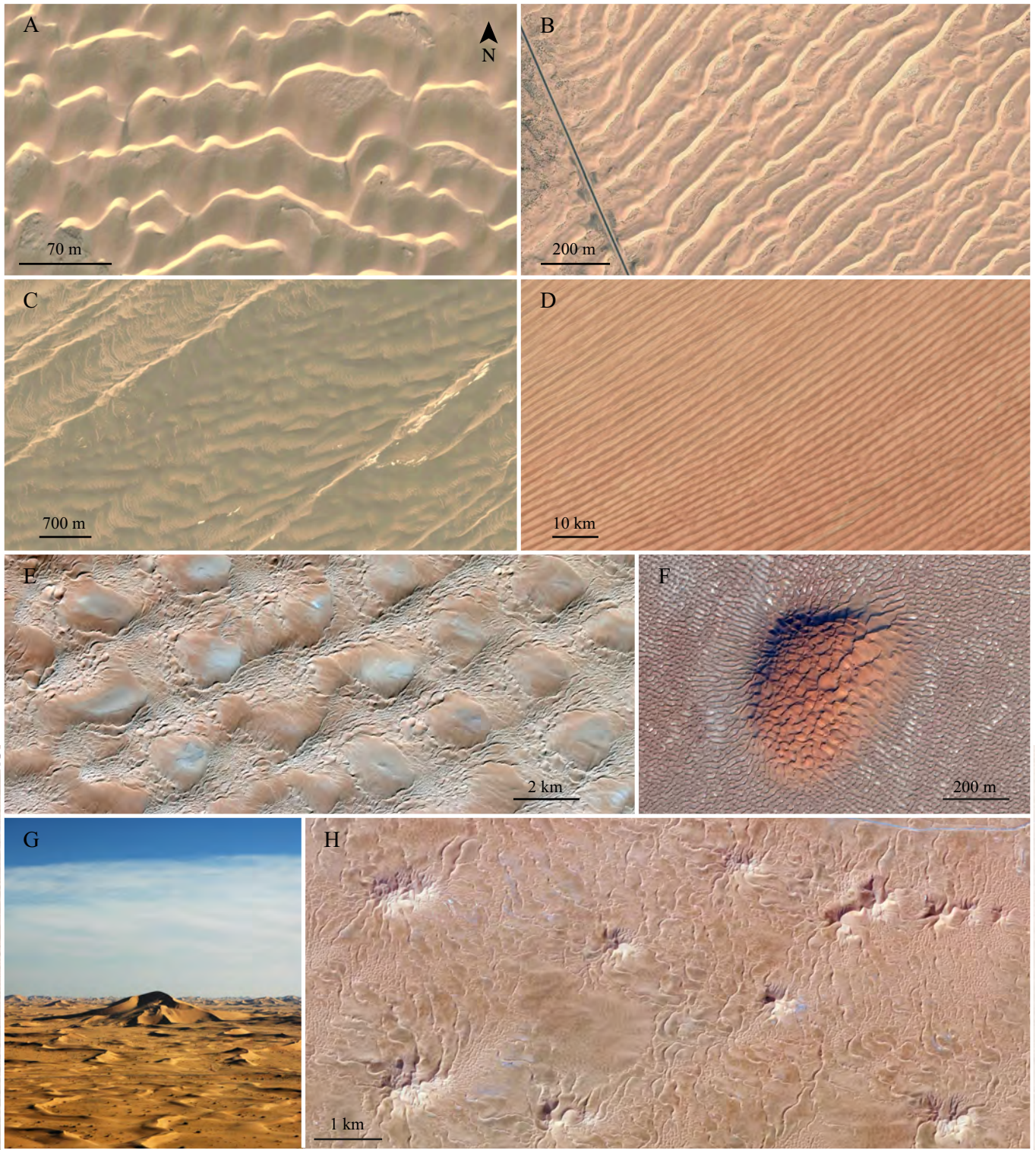


Figure 3: Free dunes. A: Barchanoid ridges in Occidental Sahara (27.166°N, 13.29°E), date: 11/2018. B: Straight-crested asymmetric dunes in the Mu-Uu desert in China (38.756°N, 107.936°E), date: 3/2011. C: Zibars (low relief dunes between linear dunes) in the Kumtagh desert in China (40.33°N, 92.655°E), date: 3/2021. D: Linear dunes in the Rub' al Khali desert (18.39°N, 48.058°E), date: 12/2016. E: Network dunes in Libya (25.136°N, 13.082°E), date: 9/2016. F: Dome dune in Oman (18.522°N, 53.549°E), date: 11/2014. G,H: Star dunes in Algeria (31.427°N, 7.302°E), date: 8/2012 (H). Credits: Maxar Technologies (A, B, E, F, H), CNES/Airbus (C), Landsat Copernicus (D), P. Claudin (G).

218 termed *transverse*. Those that are oriented parallel are referred to as *longitudinal*. *Oblique* dunes are in between.
219 Previous work proposed that oblique dunes be defined differing from parallel or perpendicular orientations by more
220 than 15 degrees [38, 40]. Such denominations are not morphological in nature. However, linear dunes typically
221 develop in wind regimes that are bi-modal to multi-directional and have parallel or oblique orientations (Section 5.2.2)
222 [40, 62, 63, 41, 64]. In some singular cases as in reversing flows, their orientation may be transverse analogous to
223 wave ripples or tidal sand waves [65]. Typical barchanoid or straight-crested asymmetric dunes are transverse and
224 develop slipfaces along some or all of the downwind side of their crests, allowing for identification of the dominant
225 sand-transporting wind direction from aerial photograph/satellite images or surface observation.

226 2.2.2 Dunes coupling to their surroundings

227 Dunes not considered as free dunes (*i.e.*, those with morphologies controlled by local material boundary conditions)
228 are subdivided into morphologies related to vegetation (Fig. 4), obstacles, or playas and pans (Fig. 5).

229 Although any dune type can become vegetated (*e.g.*, the Nebraska Sand Hills [66] and most coastal dune systems
230 [42]), here we specifically require that the vegetation be morphologically intrinsic to the dune [67, 68, 42, 21]. For
231 example, *nebkha dunes* form with a core of vegetation and develop as the vegetation traps sand by slowing winds
232 while also continuing to grow in place [69, 70]. Various other terms have been used to describe discrete dunes formed
233 in isolated plants such as coppice dunes, bush mounds, nabkha, phytogenetic hillocks, and others [70]. Dunes that
234 are formed by aeolian sand deposition in semi-continuous to continuous vegetation in coastal backshore systems are
235 considered *foredunes* [71, 72]. Note that nebkhas commonly form along the backshore of semi-arid to arid coasts and
236 may also be considered foredunes in these systems [73, 74]. Dunes that form principally by erosion within vegetation
237 forming erosional bowls, troughs, saucers (and other shapes) are called *blowout dunes* (or simply *blowouts*) [75, 72].
238 Blowouts normally display downwind attached depositional lobes thereby fulfilling the definition of a dune; defined
239 by Bagnold [32] as ‘a mound or hill of sand which rises to a single summit’ (p. 188). Note that erosional depressions
240 in the absence of a depositional lobe may also be referred to as blowouts, but then do not meet the definition of a
241 dune. *Parabolic dunes* develop when vegetation stabilizes the lateral margins of the depositional lobes, causing them
242 to develop U- or parabola-shaped (sometimes V-shaped) crestlines and plan view shapes. Parabolic dunes display
243 short to long trailing ridges whereas blowouts do not have trailing ridges [21]. The trailing ridges in the parabolic
244 dunes point in the opposite direction to the avalanche face, rather than in the same direction, as is the case with
245 barchan dunes.

246 Although dunes are almost exclusively formed from sand-sized grains, dunes may form with significant clay and
247 silt fractions along the margins of playa, salt lake, and evaporitic pan systems. Much of the clay and silt can occur as
248 sand-sized aggregates that behave dynamically as sand grains. These dunes are intrinsic to the pan/playa and have
249 been termed *clay dunes* and *lunette dunes* [76, 77, 78]. The term *lunette* is preferred here.

250 Finally, wind and, therefore, aeolian sediment transport respond to topographic obstacles, sometimes generating
251 dunes with morphologies tied to that topography. We refer to these collectively as *obstacle dunes* and differentiate
252 between them using the underlying slope, proximity to an obstacle, and direction of the wind [80]. As with some free
253 dunes discussed earlier, assessing the sub-type of an obstacle dune requires either knowledge of the sand-transporting
254 wind direction, or an inference based on the morphology of the dune. *Climbing-* and *falling dunes*, found on slopes
255 less steep than the angle of repose, may or may not exhibit slip faces, but develop in response to sediment transport
256 up- and down slope, respectively. Note that the use of ‘climbing’ in the geomorphic classification of dunes is distinct
257 from the stratigraphic use of the term (*e.g.*, climbing ripple structures, or bounding surfaces with an angle of climb)
258 which refers to bedform migration accompanied by net deposition. Depending on the local boundary conditions, dunes
259 can also form immediately upwind, downwind, or on top of a topographic obstacle that diverts the wind. Once the
260 position of the dune with respect to the wind direction and the obstacle is known, the dune can be identified as either
261 an *echo dune* (upwind of the obstacle, and approximately echoing the planform or morphology of the adjacent feature),
262 *lee dune* (downwind of the obstacle), or *cliff-top dune* (on top of the obstacle). We note that the obstacles controlling
263 dune development in this case are broadly defined. For example, echo dunes might form at the base of scarps, in
264 front of and along the margins of small to large boulders or obstacles, and lee dunes (including shadow dunes) form
265 downwind of obstacles (including vegetation and nebkhas). However, climbing-, falling-, and cliff-top-dunes all require
266 topographic slopes to form.



Figure 4: Dunes associated with vegetation. A: Fore dune at Fort Steven, Oregon (46.182°N, 123.982°W). B: Fore dune plain comprising multiple foredunes formed by coastal progradation. C-E: Nebkhas. F: Trough blowout 12 m deep and 600 m long on the Tibet plateau (35.992°N, 100.415°E). G: Bowl blowout 3 m deep at Cape Cod (42.075°N, 70.189°W). H & I: Parabolic dune in south Australia (34.912°S, 135.89°E) in 9/2016 (I). Credits: C. Bristow (A), P. Hesp (B, D, F, G, H), C. Narteau (C, E), CNES/Airbus (I).

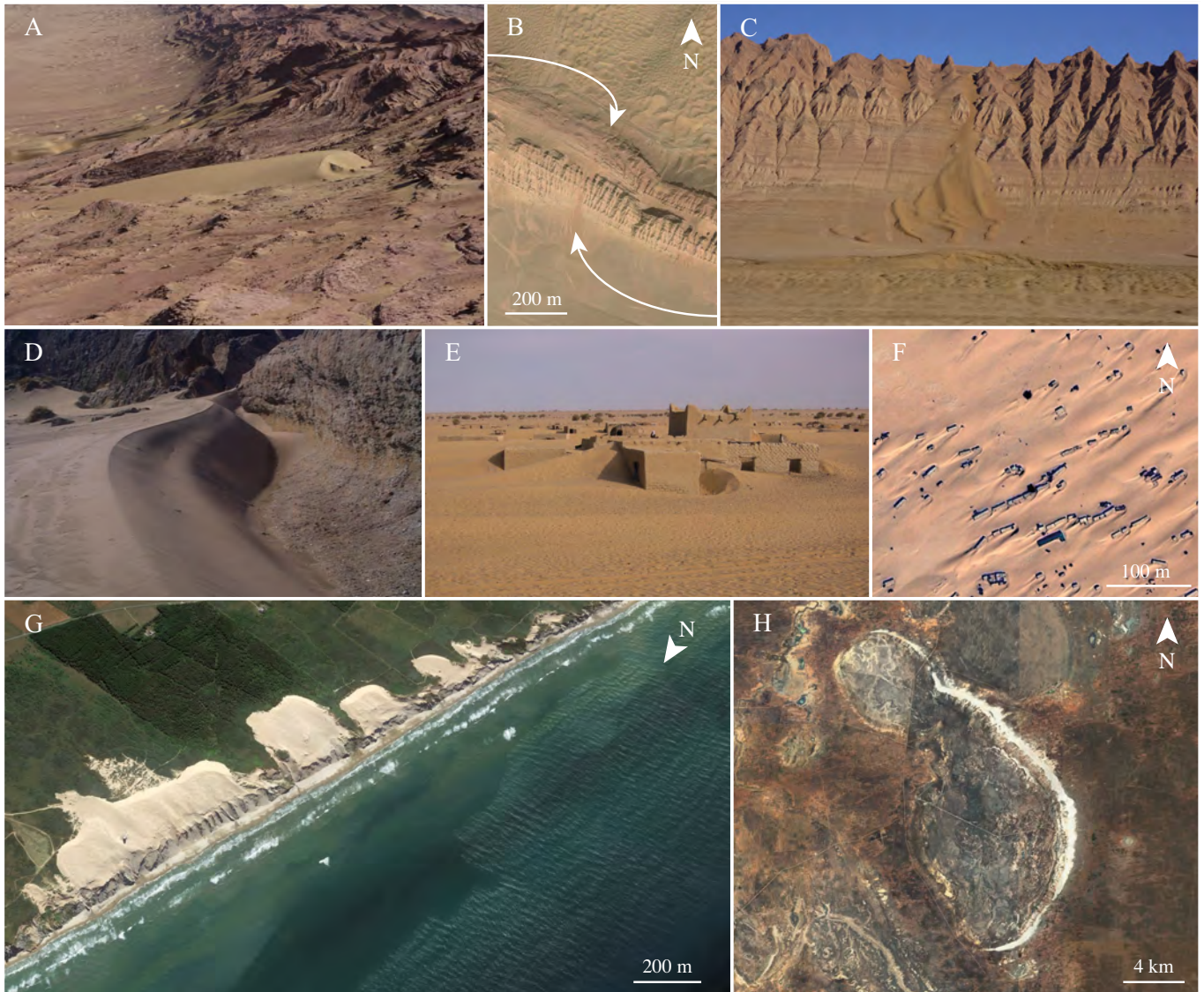


Figure 5: Dunes in association with an obstacle or pan. A-C: Climbing (A) and falling (C) dunes on either side of Mazartagh mountain range (38.676941°N, 80.378328°E) (B, date: 09/2019). D: Echo dune at Castlepoint, New Zealand (-40.9°, 176.23°), date: . E: Lee (downwind, left) and echo (upwind, right) dunes on either side of a building from Kouba Olanga in Chad, date: 2005. Panel F shows the context (15.75°N, 18.3°E), date: 01/2006. G: Cliff-top dune in Rubjerd Knude, Denmark (57.439336°N, 9.768589°E), date: 07/2021 [79]. H: Lunette dune bordering (now dry) Lake Mungo in Australia (33.74°S, 143.13°E), date: 02/2023. Credits: C. Narteau (A, C), CNES/Airbus (B, H), P. Hesp (D), C. Bristow (E), Maxar Technologies (F), Terra Metrics (G).

267 2.3 Discussion

268 2.3.1 Scope of the morphology-based classification

269 The morphology-based identification tree in Figure 2 is intended to encompass the majority of aeolian dune types
 270 observed on Earth. Other aeolian bedforms, including many varieties of ripples (*e.g.*, impact ripples, megaripples,
 271 granule ripples) are not included. Not every name for every type of dune is represented in the geomorphology tree. In
 272 a perusal of recent literature, we identified more than 50 different terms for types of dunes (table 1). Many of these
 273 terms were duplicative, or specific to niche environments or localities. To combat some of the confusion that arises
 274 from so many terms, in Figure 2 we note some equivalent terms in black below the dune identification in blue. In
 275 the future, we encourage the community to converge on the use of a single term for a given dune type, but here we

276 include the duplicative names in this work for reference. Because dune type recognition can be aimed to infer wind
277 regimes and environmental constraints, we avoided using dynamical properties or we included them at the end of the
278 classification. For example, we did not include the commonly used term *reversing dunes*. This term can be added
279 to further define a dune. A dune undergoing and integrating successive inversions generally does not have a fully
280 developed avalanche face on one side only.

281 Although the geomorphic dune identification tree was developed with terrestrial aeolian dunes and dune examples
282 in mind, the tree can be applied to dunes in any system or on any planet. Aeolian bedforms are known to develop
283 on many planetary bodies in our solar system, including Venus, Earth, Mars, Titan, Pluto, Io, and Comet 67P
284 [8, 81, 46, 82, 83, 84]. In most cases the morphologies of dunes observed on other planetary bodies are similar to those
285 observed on Earth [7, 85, 86, 82] and the tree in Figure 2 can be easily applied to these systems. Some planetary aeolian
286 bedforms have debatable analogs on Earth. For example, transverse aeolian ridges [87, 88] and large ripples [12, 89]
287 are both abundant on Mars, but bedforms of similar scales and morphologies are comparatively uncommon on Earth.
288 Large martian ‘ripples’ have no known direct analog among the many kinds of aeolian ripples on Earth, none of which
289 are included in this geomorphology-based classification tree. However, transverse aeolian ridges (TARs) on Mars can
290 be several meters high and would be considered by this classification to be transverse dunes, though dynamically the
291 processes forming each might be quite different. The application of the geomorphology tree to planetary systems must
292 be done with appropriate caution and context. Together with the other two trees, the tree based on dune morphology
293 will facilitate interplanetary comparison in future research and limit terminological confusion in the literature.

294 **2.3.2 Wind direction and variability inferred from morphology**

295 The classification based on dune morphology does not require prior knowledge of the local wind direction, except
296 in some cases when it is needed to disambiguate between sub-types. However, decades of research studying the
297 relationship between dune morphology and local winds has allowed us to make inferences about the local wind direction
298 when a given morphology is observed (*e.g.*, [32, 40, 62, 90, 41, 64]). As noted above, slipfaces develop on the net
299 downwind side of dunes, providing a morphological indicator of the leeward side of dunes with slip faces that are fully
300 developed on one side only.

301 Asymmetry in planform or cross-section in the morphology of a dune can also provide some information about
302 the local wind variability. For example, barchan dunes with dissimilar horn lengths form when two winds with an
303 obtuse divergence angle have differing strengths [91, 92, 64, 93, 94]. In barchanoid dunes, lee-side spurs that are
304 asymmetric in plan form (skewed rather than transverse to the main dune) indicate the main dune is asymmetrically
305 oriented relative to the winds (*i.e.*, not entirely transverse). Similarly, lee-side spurs that are perpendicular to the
306 main crests but are asymmetric in cross-section (having a steep lee side and gentle stoss side) indicate net transport
307 parallel to the lengths of the main dunes, thereby demonstrating that the main dunes are not perfectly transverse.
308 Net transport parallel to the crests of the main dunes can also be inferred from orientation of asymmetric peaks,
309 saddles, or superimposed dunes enabling significant complexity to be disentangled from observed dune morphologies
310 [95, 39, 40, 96, 41]. Superimposed dunes are common on linear dunes that can form with kilometer-scale wavelengths
311 and large enough widths to host subordinate dunes on their surfaces [97]. Specifically for long-crested dunes, the
312 transverse and longitudinal symmetry/asymmetry indicates whether the transverse and longitudinal net transport,
313 respectively, is zero or greater than zero.

314 This general approach has been applied to identification of deposits of transverse, oblique, and longitudinal dunes
315 in the stratigraphic record [95, 39, 98]. Stratigraphically, the approach is applied by determining whether the main
316 dunes migrated laterally or aggraded in place and whether or not the superimposed topographic features such as
317 dunes, spurs, peaks, and saddles migrated in a preferred direction along the lengths of the main dunes. Deposits of a
318 perfectly transverse dune preserve migration of the main dune without a preferred along-crest migration direction of
319 superimposed features; perfectly longitudinal dunes aggrade in place with superimposed features migrating consistently
320 in the same along-crest direction; and deposits of oblique dunes preserve both lateral migration of the main dune and
321 consistent along-crest migration of superimposed features.

322 Here we suggest that the same general approach can be used to distinguish transverse, oblique, and longitudinal
323 dunes using asymmetry rather than stratification to infer migration of the main dune and superimposed features
324 (Table 2.3.2). This approach comes with three limitations. First, it can only be applied where the main dune has
325 superimposed topography. Second, both the main dune and the superimposed bedform must integrate the wind
326 regime. Although some dunes are small enough to re-form seasonally, here we are restricted to dunes that coexist and
327 persist through complete cycles of flow [99]. Third, the determination of dune type may not necessarily match the
328 definition of Hunter, Richmond, and Alpha [38]. Specifically, dunes that meet the geomorphic criteria of oblique dunes
329 (Table 2.3.2) might have orientations outside the defined range of 15° to 75° relative to the net transport direction.

330 Nevertheless, this approach can identify dunes that are qualitatively neither transverse nor longitudinal, and it can
 331 do so using only morphologic criteria as shown in Section 5.3. The same information required to classify dunes as
 332 transverse, oblique, and longitudinal can be used to infer the net sand transport direction relative to the main dunes.
 333 Where dunes meet the geomorphic criteria of transverse or longitudinal dunes, the net sand transport direction is
 334 perpendicular or parallel to the dune crests, respectively. Where dunes are oblique, the net transport direction is
 335 constrained to the quadrant bounded by the migration direction of the main dunes and the direction of migration
 336 of superimposed features along the crests of the main dunes. These directions of migration are derived from the
 337 dip direction of the slipfaces. Note that these two migration directions are only used to constrain the net transport
 338 direction; the two migration directions do not necessarily correspond to two formative winds.

Across-crest (transverse) symmetry or asymmetry (identified by cross-section profile)	Along-crest (longitudinal) symmetry or asymmetry (identified by symmetry or asymmetry of superimposed topographic features such as peaks, saddles, crest sinuosities, spurs, or superimposed dunes)	
	Along-crest symmetrical	Along-crest asymmetrical
Across-crest symmetrical	Zero net transport in both the across-crest and along-crest directions suggests formation by opposed equal winds; singular conditions with zero net transport (analogous to some wave ripples)	Longitudinal dunes (net along-crest transport with no net across-crest transport)
Across-crest asymmetrical	Transverse dunes (net across-crest transport but no net along-crest transport)	Oblique dunes (net across-crest and along-crest transport)

Table 2: Identification of transverse, oblique, and longitudinal dunes based on symmetry/asymmetry of main dune and superimposed features.

339 When considering multiple related dune morphologies, perhaps occurring at different scales, it is important to
 340 consider the timescales of formation and how they compare to the time period of the wind regime. Smaller dunes
 341 migrate faster than larger dunes, and change morphology in response to a change in wind regime faster than larger
 342 dunes [32, 100]. The most recent winds are reflected by the smallest bedforms (all the way down to wind ripples),
 343 and larger and larger dunes reflect wind regimes as integrated over longer and longer timescales. Different scales
 344 also frequently differ in orientation or morphology for reasons other than changes through time in wind regime. The
 345 larger dunes may modify the apparent winds for smaller superimposed or adjacent dunes, or the different scales
 346 may interact [99, 101]. Often, the different scales experience different boundary conditions. For example, the main
 347 dune may extend on bedrock while the superimposed ones develop on the underlying main dune, *i.e.* a sand bed.
 348 Orientation and morphology of both the main dune and the superimposed ones then provide information about winds
 349 and potentially help differentiate between longitudinal and oblique dune types [41, 102]. Additional discussion of the
 350 relationship between morphology and process is provided in Section 3.

351 2.3.3 Dunefield patterns and changes in dune type

352 The geomorphological identification tree deals with individual dunes and their morphology, but dunes in nature are
 353 rarely found in isolation. More commonly, many dunes form a field, and that field may vary spatially in the morphology
 354 and size of individual dunes. Boundary conditions that change spatially (*e.g.*, sediment supply, antecedent topography,
 355 water table, vegetation) can cause variability in the morphology of the dunes [103, 104, 21]. Similarly, dunefield
 356 patterns mature spatially and in time, with dunes becoming larger and more widely spaced as they develop and
 357 migrate downwind [105]. With time, almost all dunefields and dune systems can be fully stabilized and many other
 358 dune forms can develop as a result of vegetative stabilization [106, 107, 42, 108]. Dunefield patterns are fundamentally
 359 constrained by these local boundary conditions, but changes in that pattern are also driven by time and interactions
 360 between the bedforms [109, 110, 111]. These interactions can act at different scales and be of different kinds, *e.g.*,
 361 aerodynamic in nature, a dune modifying the flow in its vicinity, or concern directly the mass exchange (sand loss and
 362 capture or collisions) [112, 113, 100, 114]. For example, in barchan dune fields, smaller dunes migrate faster than larger
 363 dunes, resulting in small upwind dunes colliding with slower downwind dunes. When colliding, dunes may split or merge
 364 leading to size regulation and spatial organization within the field as in dune corridors [113, 114, 115, 100, 116, 117, 118].

365 Barchan dunes may also link together to form barchanoid ridge dunes, which is an overall change in the morphology
366 of dunes in the field. It is rare to find a dune field composed of a single dune morphology in deserts, although it
367 does occur more frequently in coastal dunefields (*e.g.*, on Earth there are multiple examples of foredune plains, and
368 parabolic dunefields), and just as the morphology of a single dune can help determine information about the local
369 winds and sediment state, interpreting the dune morphologies of a field in aggregate, the spatial succession of forms
370 and types, can yield important information about the larger system [119, 120], as exemplified by the case study in
371 Section 5.

372 **3 Classification of aeolian dunes based on dune dynamics**

373 **3.1 Purpose and approach**

374 Aeolian dunes exhibit a wide variety of forms, and geomorphologists have wondered for more than a century what
375 processes control this variability. The goal of this classification is to link the various dune types (bottom row in Figure
376 6) with the formative processes in a dune field (upper rows in Figure 6). Specifically, we propose a tree structure to
377 identify the dynamical processes by which dunes are built and aligned according to wind regime, sand availability,
378 and other boundary conditions. This dynamics-based tree diagram thus links dune type to the sediment and wind
379 conditions, allowing for both forward modeling of dune morphologies in specified conditions (top-to-bottom) or inverse
380 interpretation of observed dune morphologies (bottom-to-top). The case studies in section 5.2 complete and illustrate
381 this classification based on dune dynamics and link it to the classification based on morphology (Figure 2).

382 The dynamics-based classification reflects the state of the art of the physical understanding of dune morphody-
383 namics derived from field observations, laboratory experiments, numerical simulations, and theoretical studies. Based
384 on these physical insights, environmental conditions are intentionally simplified or conceptualized, and cannot cover
385 the whole range of complexity observed in nature. In such a process-based approach to dune classification, we consider
386 that dunes types and general morphologies are in dynamic equilibrium with environmental conditions. Other factors
387 such as interactions between dunes, *e.g.* collisions, modify these general morphologies. Such morphological properties,
388 which are usually local and transient, are beyond the scope of this classification. As a first step, we focus on free
389 dunes composed of loose sand. Cohesion between grains and vegetation are not taken into account. We expect that
390 this diagram will evolve as the understanding of the formative conditions and processes progresses.

391 **3.2 Tree description – Selection of dune shape and crest orientation from formative** 392 **processes**

393 The tree in Figure 6 classifies dunes depending on the dominant dynamics. The dune morphodynamics are controlled
394 by spatial variations of the sand flux. The bed is eroded where sand flux increases and aggrades where sand flux de-
395 creases. Those spatial variations can make a dune (i) grow in height, (ii) migrate (propagation perpendicular to crest),
396 and (iii) elongate (increasing the length of the crest line). Growth in height, migration, and elongation correspond
397 to dune dynamics in the three dimensions: vertical, horizontal across-crest, and horizontal along-crest, respectively.
398 Growth in height generally corresponds to an overall increase in dune cross-sectional area so that the dune transverse
399 length increases accordingly. In a multidirectional wind regime, these three dynamics are present to different degrees.
400 Depending on boundary conditions and wind regimes, one of these dynamical processes may prevail and drive the
401 overall dune shape and orientation. The orientation of the dune with respect to the different sand fluxes further
402 determines the secondary dune dynamics, which ultimately set the dune type.

403
404 One first has to choose the dune’s length scale or time period to address, the two being linked (Section 7.1). The
405 tree describes dunes of a given length scale. When different scales are present, each scale can be described by going
406 down the tree separately. The first decision relates to a boundary condition through the question: Is the bed made
407 of mobile sand ? A dune may strongly interact with the surrounding sand bed, *e.g.*, growing from the erosion of
408 the interdune. Conversely, the interaction may be limited like for a barchan, which evolves on a starved floor. Bed
409 mobility has a strong influence on dune dynamics and shapes. We recognize the floor as a mobile sand bed when the
410 granular medium that composes the dune and the bed surface are similar.

411 **3.2.1 Mobilized sand bed – Growth in height prevails**

412 Where the bed is made of loose sand, dunes develop from the destabilization of the bed by the wind. We call this
413 growth mechanism the *bed instability*, and growing in height is the prevailing dynamics. The resulting dunes are long

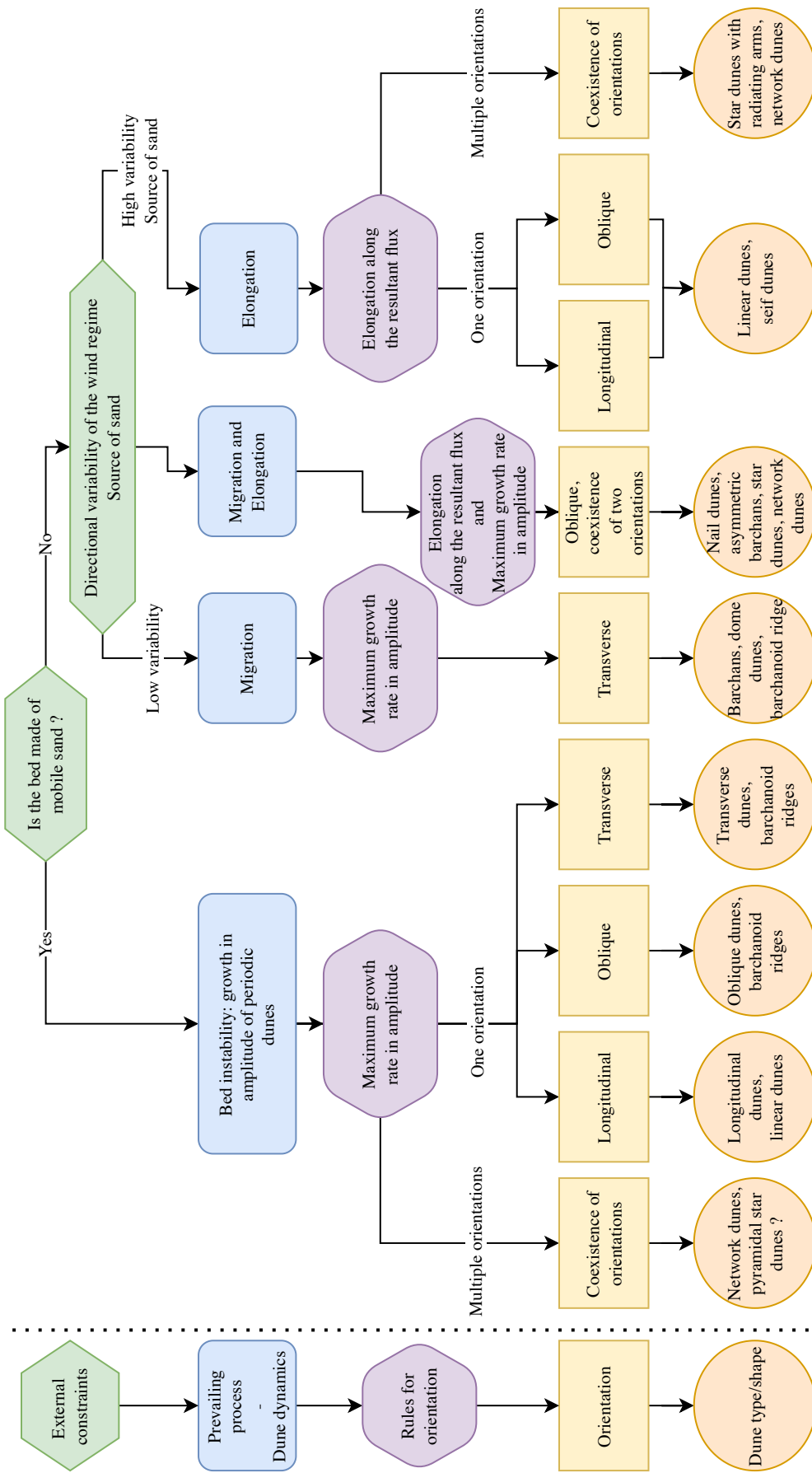


Figure 6: Identification tree for dunes classification based on dune dynamics.

414 ridges generally organized periodically. In accordance with an instability mechanism, the observed pattern, and in
415 particular its orientation, is such that the growth rate in height is maximum [40, 41, 121]. Dunes are built up by
416 normal-to-crest sand fluxes. In a multidirectional wind regime, sand fluxes of opposite directions both contribute to
417 build the dune, so that dune orientation maximizes the *gross bedform-normal transport*, which, depending on the wind
418 regime, can be transverse, oblique, or longitudinal with respect to the resultant drift direction, *i.e.*, the direction of
419 the net transport on a flat bed. For some specific wind regimes, multiple orientations can coexist, resulting in intricate
420 patterns that may be defined as network or star dunes.

421 The shape of dunes is then modulated by the two other processes and by migration in particular. Migration tends
422 to promote the development of sinuous crestlines. In contrast, along-crest sand flux promotes dune straightness [65].

423 3.2.2 Non-mobilized bed – Migration and elongation prevail

424 When large areas of the bed cannot be mobilized by the winds (but sand is still piled up by the winds), the growth
425 in height of a periodic pattern is inhibited due to sand supply limitation. Consequently, *migration* and/or *elongation*
426 prevail in determining the dune type and orientation [41]. Although dunes may be arranged periodically, dunes can
427 be isolated objects unlike in the case of a sand-covered bed. They persist because capture of free sand flux and losses
428 balance out, or slowly evolve such that their shape and size are quasi-static.

429 What controls the relative dominance of migration and elongation remains largely unknown to date. Elongation
430 can be promoted over migration by some boundary conditions that ‘pin’ part of the dune. Generally, the dune would
431 grow from a localised sand source, behind an obstacle that acts as a sand trap for example. A part of the dune may
432 also be so large that it barely moves during the time it takes to develop extensions, such as the core of a star dune
433 [90]. If part of the dune is pinned, the dune cannot propagate but only elongate in the direction of the sediment net
434 transport [41, 64]. We recognize elongation as another growth mechanism. For most wind regimes, a dune would
435 migrate and elongate if not pinned. The boundary condition can force elongation and prevent migration, but only if
436 such dunes are stable. An elongating dune cannot form in a unidirectional transport regime [122]. Although it extends
437 in the direction of transport, its growth in height and sustainability requires that it is subjected to components of wind
438 perpendicular to its crest line. Because the direction of elongation is generally different from that of the bed instability
439 [41, 64], secondary superimposed patterns in the bed instability mode are likely to develop on an elongating dune [41].
440 An isolated dune parallel to the direction of the net transport, therefore, eventually destabilizes into migrating dunes
441 if its growth rate in height is not sufficiently large relative to that of the superimposed patterns in the bed instability
442 mode [122, 64].

443 **Elongation prevails** If elongation prevails, dunes have a ratio of length (along crest) over width (across crest)
444 much larger than unity and extend in the direction of the net transport if they do not migrate [63, 122, 41, 64]. The
445 orientation of such elongating dunes can be longitudinal or oblique to the resultant drift direction (RDD) because the
446 dune topography modifies the transport (Section 7.2.3) [41]. In a multidirectional wind regime, the topography may
447 change the relative magnitudes between the different transport directions, so that the dune experiences a direction of
448 net transport different from that over a flat sand bed, the RDD. Some specific wind regimes allow multiple elongation
449 directions, which may produce star dunes with radiating arms. Arms of adjacent dunes may connect to form network
450 dunes.

451 **Migration prevails** Migration does not directly select dune orientation because only dune growth mechanisms
452 do, *i.e.* growth in height (cross-section) and growth in crest line length, or elongation. However, migration mutes
453 elongation, so when migration prevails, we expect dune crest orientations to maximize growth rate in height, as in
454 the bed instability mode. The archetypical migrating dune is the barchan dune. As variability in wind directions
455 increases, the width of the slipface decreases from the full width of the dune (barchan) to a restricted region, and then
456 finally disappears completely (dome dune) [122, 123]. Although they may correspond to transitional regimes, we also
457 classify isolated barchanoid ridges in this branch. Experiments under water [122] and numerical simulations [124, 125]
458 have shown that an isolated straight-crested transverse dune subjected to a monodirectional flow is unstable when
459 migrating on a non mobile sand bed and without being fed by an incoming free flux. A perturbation in height or in
460 streamwise length leads to a perturbation in the migration velocity and to the redistribution of sand towards where the
461 avalanche face is concave, which amplifies the perturbation. The transverse dune transforms into a barchanoid ridge
462 and eventually breaks into a row of barchans. In light of those studies, barchanoid ridges on beds of non-mobile sand
463 correspond to transient dynamics. Because they might be stable in different conditions and because this instability is
464 driven by migration, we recognize these dunes as being shaped by migration.

465 **Concurrent migration and elongation** If boundary conditions and/or sand supply do not force the development
466 of an elongating dune from a point source, a dune is likely to migrate except in some specific symmetric wind regimes,
467 *e.g.*, two transport directions with an obtuse divergence angle and equal magnitudes [63, 122]. Most multidirectional
468 transport regimes should involve both dynamics. The transition between elongation and migration is not sharp but
469 a continuum. As such, when neither migration nor elongation fully dominates, the orientations and morphologies
470 associated with these two dynamics coexist, leading for example to asymmetric barchans, nail dunes or raked linear
471 dunes [91, 92, 64, 126].

472 3.3 Future refinements

473 Here, we classify dunes in broad classes and determine their orientation according to the prevailing formative dynamical
474 processes. The ideal classification would not only address the dune orientation but also the detailed dune morphology
475 in a quantitative way, *e.g.*, the straightness of linear dunes or the sinuosity of their crest line, the curvature of the
476 avalanche face of barchans, or their asymmetry like the relative size of their horns. Many of those properties should
477 depend on the competition between the three processes, which are all at work at different degrees in multidirectional
478 flow regimes. How much a morphologically long dune migrates relative to its growth in height or elongation should
479 for example control the dune straightness as conceptually proposed in [65]. Growth in height and elongation should
480 favor the spatial coherence of the dune, while migration destabilizes it [122, 124, 125]. Periodic transverse dunes in
481 the bed instability mode being subject to migration are often barchanoid ridges. However, all these morphological
482 properties not only lack unified predictive models but also quantitative measurements. The quantitative prediction of
483 dune orientation is an important step to the quantitative prediction of dune types as it is a prerequisite to evaluate
484 the components of sand flux associated to the different dynamics.

485 Other boundary conditions associated with obstacles or confining factors could be added to the tree, *e.g.*, a dune
486 may interact with a cliff to form an echo dune. Finally, the role of cohesion and vegetation is another challenge that
487 could provide additional understanding if incorporated in a dynamics-based classification.

488
489 In Section 5.2, we apply this dynamics-based classification to various characteristic dune types on Earth, using
490 observed wind regimes to evaluate the dynamical processes in a simple way, and giving rules to select the prevailing
491 dynamics and to calculate crest orientations.

492 4 Classification of aeolian dunes based on fluid mechanics and sediment 493 transport

494 4.1 Purpose and approach

495 Dunes are observed at a variety of scales that differ from one environment to another. The purpose of the following
496 classification, based on fluid and sediment-transport mechanics, is to define the different dune-formation regimes,
497 which determine the range of possible dune sizes. As environmental and boundary conditions vary from one planetary
498 body to another, Earth and extraterrestrial dunes sample these different formation regimes, such that predictions from
499 theory can be compared with observations returned by planetary exploration missions (Section 5.4).

500
501 For simplicity, we consider the case of a steady, unidirectional wind blowing over a cohesionless granular bed with
502 unlimited sediment supply, at wind velocity above the threshold of transport. Under these conditions, a flat bed may
503 become unstable, leading to the growth of a periodic dune pattern with a characteristic wavelength, λ_{\min} . Specifically,
504 dunes form from a positive feedback between fluid flow, sediment transport and bed topography as sketched in
505 Figure 7. Spatial variations of sand flux drive erosion and deposition through conservation of mass (Exner equation
506 [127, 128, 129]). A dune grows with time if sand is, on average, deposited at its top, *i.e.*, if the sand flux peaks
507 upwind of the dune top and decreases further downwind from the top. Conversely, a peak in flux downwind of the
508 dune top leads to dune flattening. The minimum possible dune wavelength corresponds to the size at the threshold
509 between a negative and a positive feedback of the instability, *i.e.*, when loci of maximum sand flux and maximum
510 topography coincide. This minimum wavelength is very close to the wavelength of incipient dunes that would form
511 from an initially flat sand bed, which formally is the wavelength for which the feedback is maximum.

512 The response of the fluid flow to a positive topographic perturbation is characterized by an increase of the fluid
513 velocity (speed-up) above the bump and a space shift between the fluid velocity close to the ground, the shear
514 velocity, and the bed elevation profile. For a sinusoidal elevation profile, the space shift corresponds to a phase shift,

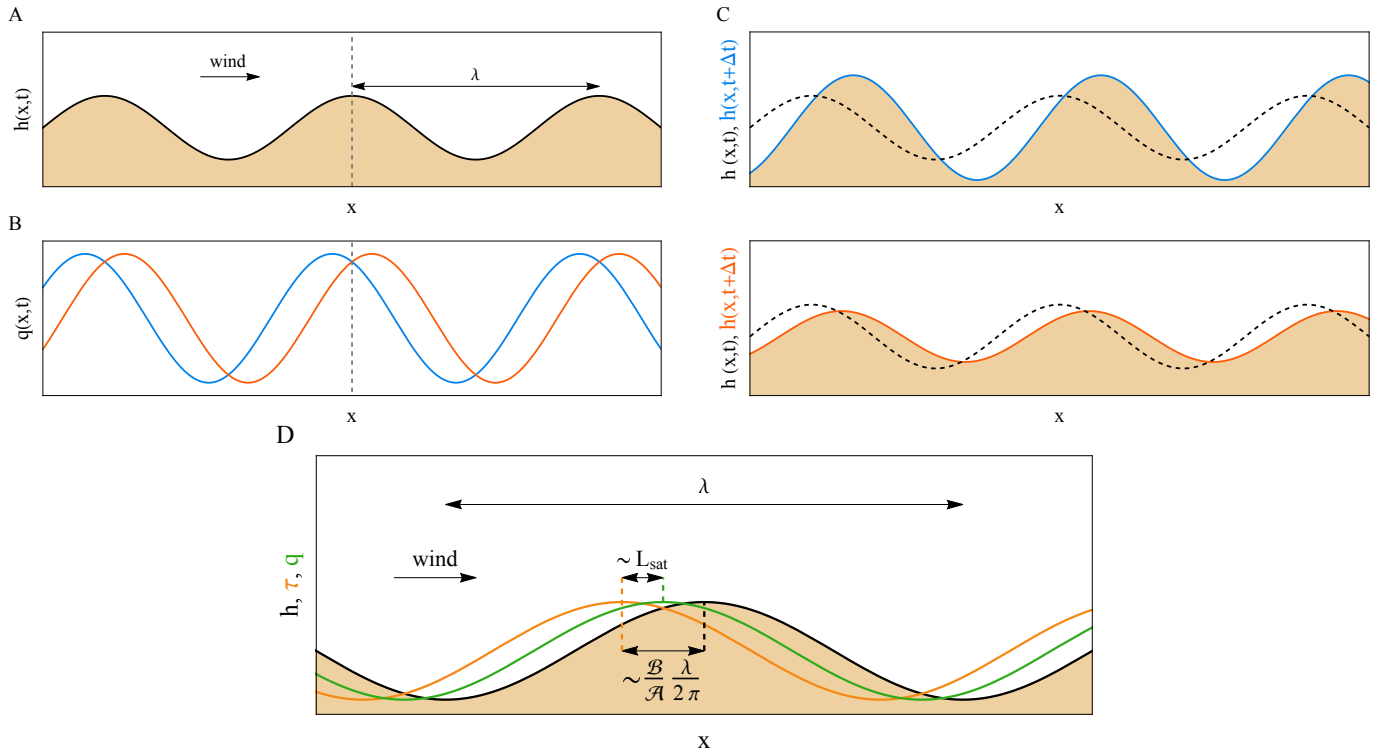


Figure 7: Formation of periodic dunes. A. Sand bed with a sinusoidal elevation profile $h(x)$ of wavelength λ at time t . B. Sand flux q along the bed profile. The sand flux peaks upwind (blue) or downwind (orange) of the topography. C. Evolution of the sand bed depending on how the sand flux is shifted with respect to the bedform. D. Basal shear stress is shifted - here upwind (positive) shift - with respect to bedform by a distance $(\mathcal{B}/\mathcal{A})\lambda/(2\pi)$. Sand flux lags with respect to shear stress by a distance L_{sat} .

515 $\phi = \arctan(\mathcal{B}/\mathcal{A})$ (see sec. 7.4.1). In turn, the maximum possible sand flux, called saturated sand flux, increases
 516 with shear velocity, or equivalently, with wind shear stress on the bed. However, the flux itself lags spatially as it
 517 responds to a change in shear stress. It peaks at some distance downwind of the locus of maximum shear stress. This
 518 relaxation distance is called the saturation length, L_{sat} , and is defined as the characteristic length over which sediment
 519 flux adapts to a change in transport conditions. The spatial shifts between the basal shear stress and the topography,
 520 and between the sand flux and the shear stress are the two key parameters that determine the formation of dunes
 521 from a flat sand bed. If the shear stress is maximum downwind of the peak in topography, dunes cannot develop. If
 522 the shear stress is maximum upwind of the topography, so could be the sand flux. If the saturation length is short
 523 enough for the flux to peak upwind of the crest, a dune grows. Conversely, if the saturation length is so long that the
 524 flux would still increase at the dune top, a dune cannot form.

525 The sign of the phase shift between shear stress and bed topography (positive if shear stress peaks upwind of the
 526 topography, negative otherwise) is controlled by the aerodynamic roughness of the bed. In turn, saturation length
 527 for wind saltation has been measured on Earth in the field [130, 131] and in wind tunnels [132] but still lacks a
 528 robust mechanistic model for predictive capabilities. Current empiricism suggests that it depends on three dimen-
 529 sionless numbers, which compare the density of sand grains with that of the fluid (s), the fluid force to the apparent
 530 weight of grains (the Shields number Θ) and gravitational to viscous effects (the Galileo number \mathcal{G}). The mechanics of
 531 a turbulent flow over a corrugated soil and sand transport are further discussed in sections 7.4.1 and 7.2.1, respectively.

532
 533 Dunes migrate as they develop. This migration forces interactions (such as collisions) between dunes, which leads
 534 to the coarsening of the dune pattern, *i.e.*, an increase in dune wavelength. The mechanism by which this coarsening
 535 process stops, limiting the size of giant dunes, is debated, possibly involving a limit in sand availability of the bed,
 536 flow confinement, or a change in flow regime (Section 7.5).

537
 538 In Figure 8, we describe how the minimum and maximum dune wavelengths are determined depending on fluid and
 539 transport conditions, and discuss the different scenarios below. Dunes have been observed on a variety of planetary

bodies, which we discuss in section 5.4. Several parameters that control dune wavelength are currently unknown, and theoretical knowledge gaps still exist, offering exciting avenues for future research and investigation.

4.2 Tree description – Selection of dune size from flow regime and sediment transport

4.2.1 Turbulent flow above the bed - Aerodynamic roughness

The response of the flow to a topographic perturbation is sensitive to the bed’s *aerodynamic roughness*, which affects the turbulent mixing and the surface drag ([133], fig. 2 in [134]). The roughness length scale, r , is a property of the bed surface at a much shorter length scale than the bedform wavelength, λ . Depending on the specific bed, flow, and transport conditions, r can be controlled by grain size, d , the thickness of the transport layer (fig. 23 in [135]), or by the height of smaller bedforms like ripples. How this length scale compares with the viscous length scale, ν/u_* (where u_* is wind shear velocity and ν the kinematic viscosity of the fluid), has a critical impact on dune formation. Two regimes can be distinguished based on a roughness Reynolds number, $\mathcal{R}_r = ru_*/\nu$ (see Section 7.4.1).

Above $\mathcal{R}_r \simeq 100$, shear stress peaks upwind of maximum topography regardless of dune wavelength. This is the *aerodynamically rough regime*, favorable to the onset of the dune instability.

For \mathcal{R}_r values lower than ~ 10 , the spatial shift between shear stress and topography strongly depends on the wavelength, λ , of the bedform perturbation. This regime is called the *aerodynamically smooth regime* and allows for shear stress to peak downwind of the topography maximum for a range of λ , preventing the formation of dunes at these scales, *i.e.*, a *dune gap* [134, 136, 137]. We refer to this particular response of the flow to the bedform perturbation as the *Hanratty anomaly* [138]. This peculiar flow regime is predicted from extrapolation of a model built on a few measurements (sec. 7.4.1) and is supported by observations of bedforms on Mars [139] as well as dissolution and melting patterns in nature (sec. 7.4.2).

For intermediate \mathcal{R}_r values ($\sim 10 - 100$), some effects of the Hanratty anomaly are detected but the physics essentially remains that of the rough limit [134].

4.2.2 Minimum wavelength selection

Rough flow The first branch of Figure 8 we consider is that of aerodynamically rough conditions. In this case, assuming that the flow is virtually unconfined in the vertical direction, the minimum wavelength, λ_{\min} , is found, to a first approximation, to scale with the saturation length, L_{sat} , multiplied by a prefactor that is controlled by the fluid mechanics, *i.e.*, the spatial shift between the topography and the shear stress, which is roughly constant in this regime, *i.e.*, $\lambda_{\min} \propto (\mathcal{B}/\mathcal{A})L_{\text{sat}}$ [140, 141, 105] (Section 7.4.1).

Smooth flow In the aerodynamically smooth regime, shear stress peaks downwind of the topography for a range of wavelengths that is determined by the viscous length. Dunes with those wavelengths cannot develop, creating a dune gap that extends over an order of magnitude around $\lambda_g \simeq 10^4\nu/u_*$. As a result, the minimum wavelength for dunes can either be set by the viscous length, as for λ_g , or by the saturation length. A saturation length-based Reynolds number $\mathcal{R}_L = L_{\text{sat}}u_*/\nu$, allows for comparisons of these two length scales.

When \mathcal{R}_L is large enough (*i.e.*, $L_{\text{sat}} > \lambda_g$, $\mathcal{R}_L \gtrsim 10^4$), saturation length is the limiting factor like in the rough regime and the most unstable wavelength scales as $\lambda_{\min} \propto (\mathcal{B}/\mathcal{A})L_{\text{sat}}$ [137].

For $\mathcal{R}_L \lesssim 10^3$, *i.e.*, $L_{\text{sat}} < \lambda_g$, bedforms with two distinct populations of wavelengths can coexist, either independently or as a superimposed pattern. They are scale-separated by the dune gap around λ_g . The minimum wavelength is thus either λ_g for dunes above the gap, or is controlled by L_{sat} for dunes under the gap. In the latter case, the minimum wavelength scales with $\lambda_{\min} \propto L_{\text{sat}}/\mathcal{R}_L^\alpha$, where $\alpha \simeq 0.4$ [12, 102, 142, 139, 83, 137].

4.2.3 Maximum wavelength selection

Once initiated, dunes migrate, interact, and coarsen leading to an increase in wavelength. The size of giant dunes in the solar system, however, appears to be limited, with a maximum wavelength, λ_{\max} . The upper limit on dune wavelength could be set either by the boundary conditions, introducing another characteristic length scale, or by the lower bound of the gap (*i.e.*, λ_g) for dunes initiated below the gap in the aerodynamically smooth regime. The smallest of those scales sets the maximum wavelength for dunes.

A first potential size limitation comes from the sediment supply and availability, which can lead to stalling dune growth [143]. Another possible limitation arises from flow confinement. In the case of a non-confined flow, *i.e.*, a deep flow, a dune perturbs the flow above it over a characteristic height proportional to the dune’s wavelength. An

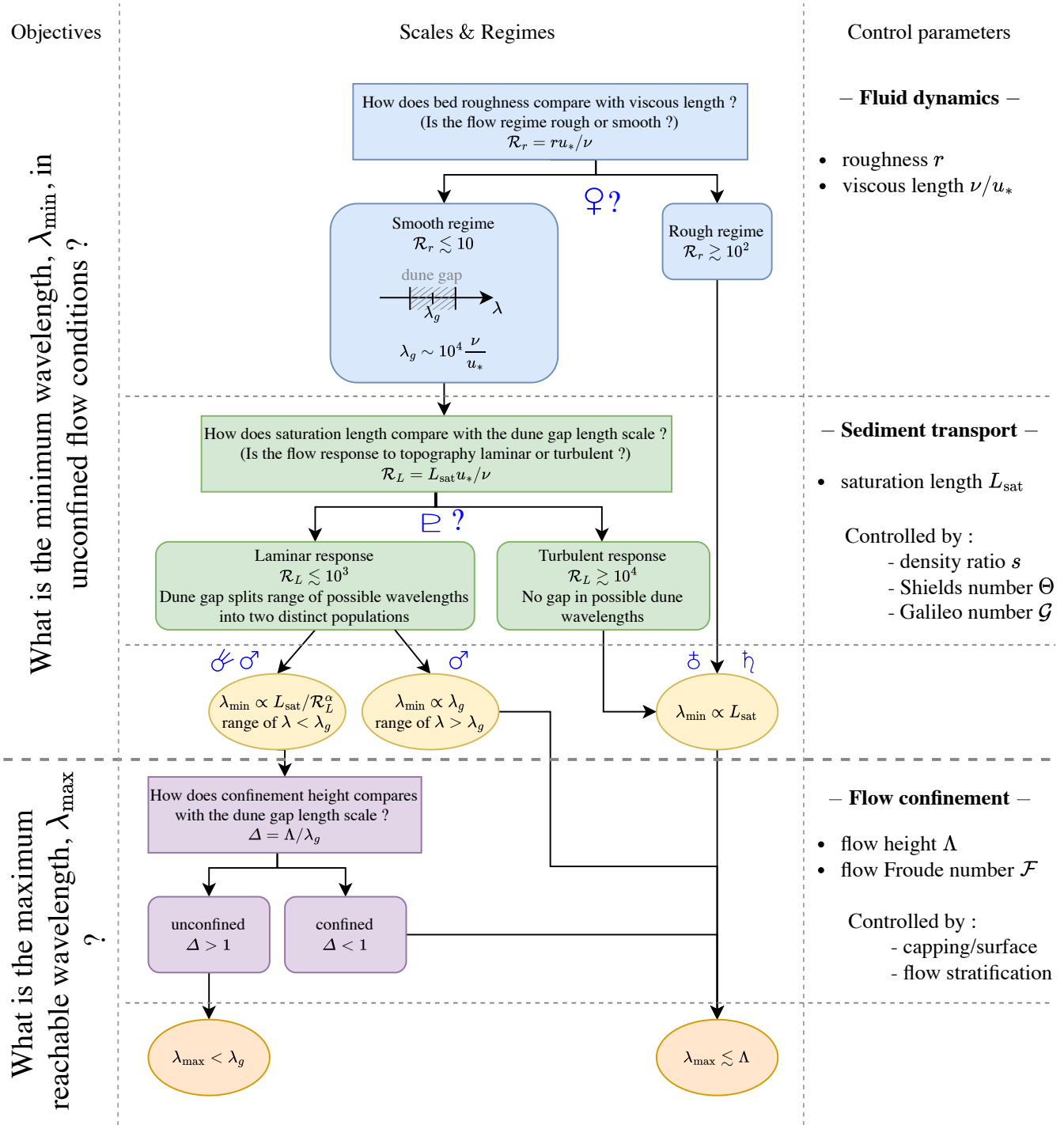


Figure 8: Identification tree for dune classification based on fluid mechanics and sediment transport. Notations: roughness length scale r , shear velocity u_* , fluid kinematic viscosity ν , fluid to particle density ratio $s = \rho_f/\rho_p$; Shields number $\Theta = u_*^2/[(s-1)gd]$ where g is the gravitational acceleration and d is the grain size, Galileo number $\mathcal{G} = \sqrt{(s-1)gd^3}/\nu$. Several Froude numbers can be associated with flow confinement or stratification, e.g. $\mathcal{F} = U/\sqrt{g\Lambda\Delta\rho_f/\rho_f}$ where U is the flow velocity and $\Delta\rho_f$ is the fluid density jump, both evaluated at altitude Λ . Blue symbols: astronomical representations of planetary bodies (♀ Venus, ♂ Earth, ♃ Mars, ♆ Titan, ♇ Pluto, ♄ comets (e.g., 67P/Churyumov-Gerasimenko)).

588 upper wall or a free surface confining the flow to a comparable or thinner height than dune wavelength would affect
589 the flow above the dune and the sediment transport. Whereas the influence of a free surface is clear in rivers, it
590 remains debated for planetary atmospheres, which are density stratified and do not exhibit a sharp interface. It was
591 proposed that the thickness, Λ , of the atmospheric boundary layer (ABL) could act as a confinement scale, stalling
592 dune coarsening at $\lambda_{\max} \simeq \Lambda$ (see sec. 7.5) [144]. This hypothesis, showing a correlation between the size of giant
593 dunes on Earth and calculations of Λ , was challenged by another data set, in which the size of giant dunes on Earth
594 were compared with satellite measurements of Λ [145]. The question of the confining role of the ABL remains debated
595 to date and may lack a good estimate or proxy of the characteristic thickness, Λ , which depends on time and space.

596 5 Case study

597 Here we analyse different dune fields in light of our classifications, demonstrating their complementarity and practical
598 utility to describe and understand the observed patterns and untangle their complexity. We provide keys and methods
599 for approaching pattern interpretation and their evolution through space and time. These also emphasize remaining
600 open questions in fully predictive classifications and models.

601 5.1 Spatial and temporal changes in dune morphology as markers of evolution in ex- 602 ternal forcing and boundary conditions

603 5.1.1 Framework

604 Although dunes are dynamical patterns that interact with each other and continuously grow, migrate, elongate, or
605 reorganize within the field, equilibrium dune types are entirely determined by external forcing and boundary condi-
606 tions, which include bed mobility, sand and vegetation coverage. Some changes in equilibrium dune type occur as
607 discontinuous jumps when a factor varies. Perhaps the best known example is the change associated with the abrupt
608 switch in dune orientation of periodic pattern on a mobile bed from transverse to longitudinal that occurs where
609 the divergence angle between equal bimodal winds changes from less than to more than approximately 90° , as has
610 been studied by experiments, simulations, and theory [40, 63, 122]. Other changes in dune morphology vary along a
611 continuum, such as where increasing dispersion of wind directions produces dome dunes rather than barchans [123].

612 Just as external forcing and bed conditions can cause differing equilibrium dune morphologies—either abrupt or
613 along a continuum—dune type and morphology can also vary spatially or evolve temporally as forcing or bed conditions
614 change through space or time. Typical changes in external forcing include environmental, climatic, and tectonic (*e.g.*,
615 [51]). These manifest as changes in wind regime (speed and direction), sediment supply, substrate type or geology,
616 topography, vegetation cover, or moisture (*e.g.*, water table). In today’s vast active sand seas on Earth, dune types
617 are, in most cases, in approximate equilibrium with current conditions [41] (Figs. 12 - 20). However, shifts in winds
618 on millennial scales change dune morphology [146] and can create generations of superimposed dune morphologies
619 [147, 148]. Larger dunes incorporate so much sand and inherit so much environmental history that to rearrange their
620 morphology could indeed require millennia (*e.g.*, [119, 147]). In a multidirectional wind regime, dune type can also
621 vary with dune size at the same location because they respond with different time scales to a change in wind direction.
622 On the other hand, dunes formed in drier, windier climates may now be fully vegetated because of a climate shift.
623 One example is the Nebraska Sand Hills, which were once characterised by active megabarchans and large barchanoid
624 ridges during the Pleistocene but are now fully vegetated relict dunes [149].

625 Environmental factors and bed conditions typically vary over 100’s km length scales across vast sand seas, or across a
626 few kilometers’ length scales, as with small dune fields that develop from beaches and extend inland. Migrating dunes
627 can experience varying conditions, and the larger dunes may then inherit morphology from upwind, thereby lagging
628 the local conditions in which they are observed.

629
630 As far as equilibrium dune types are concerned, changes in space or time are interchangeable. In transitional zones
631 or periods, however, dunes are likely out of equilibrium. In case of migrating dunes, spatial morphologic and size
632 changes in a dune field could still reveal what would be an autogenic temporal evolution. Ageing and coarsening of
633 dune pattern, the arrangement of dunes within a field and response to changes in external forcing or boundary con-
634 ditions are generally driven by the autogenic process of dune interactions such as dune collision, linking or repulsion,
635 which may result in recognizable dune patterns that are inherently transitional. In the simplest case, merging and
636 linking of barchan dunes results in transient barchanoid ridges (*e.g.*, [115]). Blowouts and, to a lesser extent, migrating
637



Figure 9: The Porto do Mangue, Northeast Brazil coastal transgressive dunefield illustrating downwind development as the areal coverage of sand increases. The middle of the image lies at (5.05°S, 36.83°W). Credit: Maxar Technologies.

638 parabolic dunes may be other examples. Some dune types also correspond to specific boundary conditions, *i.e.*, to a
 639 transitional zone. This is the case for foredunes or lunettes, for example.

640

641 These examples paint a complex picture of dune equilibrium, where understanding the morphologic and dynamic
 642 richness of such dune fields relies first and foremost on recognizing changes in dune type and size along with potential
 643 external factors that influence these changes [50].

644 5.1.2 Field examples

645 Here we first discuss two examples of spatial evolution of dune type within fields where sediment enters the system
 646 from an upwind source (*e.g.*, beach [50]). These examples illustrate the interplay between spatially varying external
 647 drivers and autogenic processes occurring at different stages of dune development. Then, we show how the observation
 648 of dunes in continuously varying environments can be extrapolated to the time evolution of dunes. We use examples
 649 that show evolutionary sequences that result in parabolic dunes, but the trajectory to the parabolic state differs due
 650 to varying boundary conditions within the fields.

651

652 Where sediment enters a dune system from a margin (*i.e.*, not from excavation of substrate alone) such as a beach
 653 or a playa, dune morphological evolution generally follows a characteristic morphologic evolutionary sequence that
 654 largely reflects a decrease in flux that drives an increase in sand cover and dune growth through dune-dune interactions.
 655 Boundary conditions including water table and feedbacks from vegetation growth can also play a role.

656

657 Figure 9 illustrates a coastal example from NE Brazil where sand sheets and stringers evolve downwind across
 658 a deflation plain into barchans and barchanoids, and then sinuous transverse dunes. This change in dune types is
 659 correlated to a change in bed surface cover, which suggests that the wind is weakening inland.

660 It is also common to observe active mobile dunes such as barchans, barchanoids, and transverse dunes evolve into
 661 parabolic dunes downwind in both continental and coastal environments [29, 26, 120, 150, 151] as observed in the
 662 White Sands dunefield shown in Figure 10. White Sands is a gypsum dune system in New Mexico, which formed in
 663 the late Pleistocene to early Holocene ($\sim 9 - 12$ kya), and which provides a type-example of spatial changes in dune

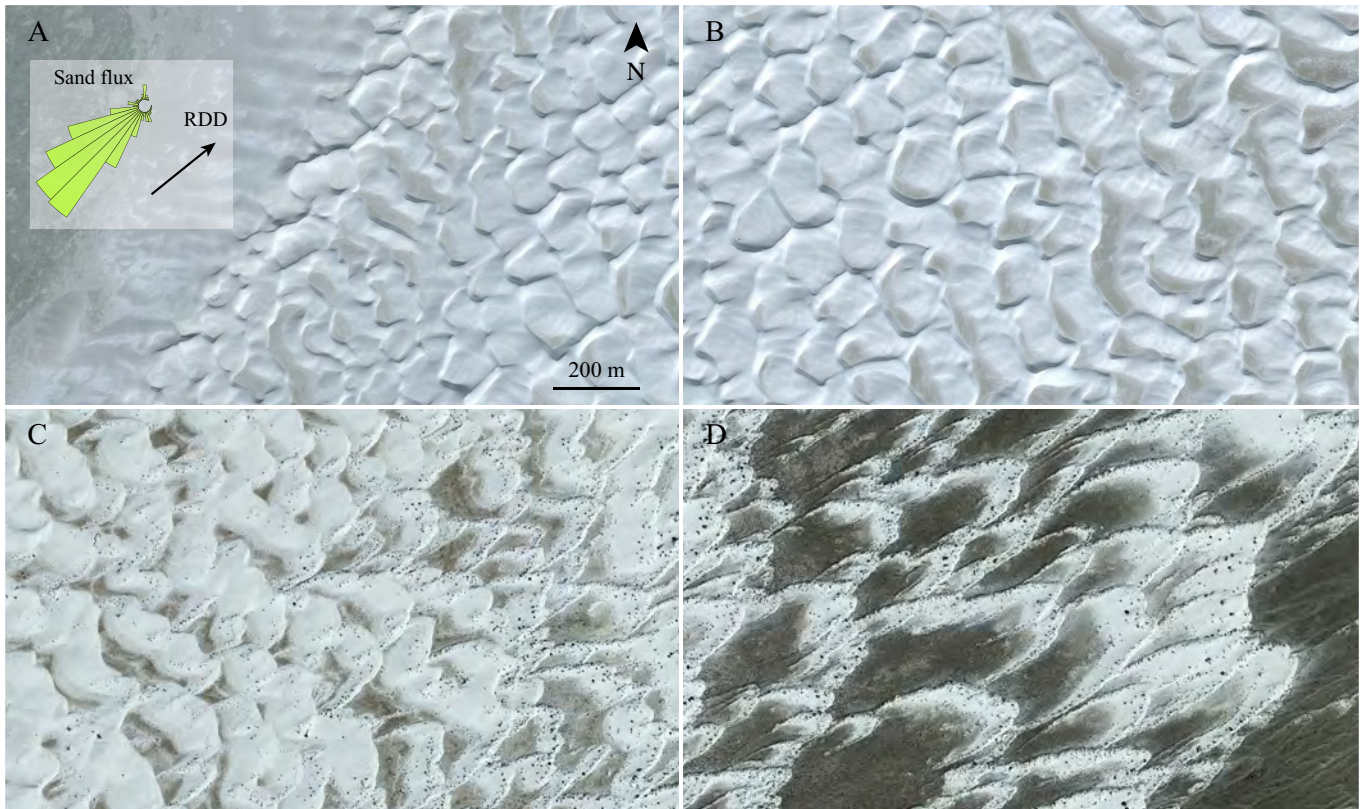


Figure 10: A west-to-east along-RDD (30°N of East) evolutionary sequence across the White Sands dunefield starting at $(32.802^\circ\text{N}, 106.302^\circ\text{W})$ and ending at $(32.848^\circ\text{N}, 106.204^\circ\text{W})$, which corresponds to a distance of approximately 11 km. Sand sheets and proto dunes (A) evolve into transverse barchanoid dunes (B) and downwind into parabolic dunes (D). Scale is the same for all snapshots. Sand flux rose points upstream. See Sections 7 for details of calculations. After the publication [105], we used a density $\rho_s = 2300 \text{ kg}\cdot\text{m}^{-3}$ and a grain diameter $d = 670 \mu\text{m}$ in the calculations for this particular case. Credit: Maxar Technologies.

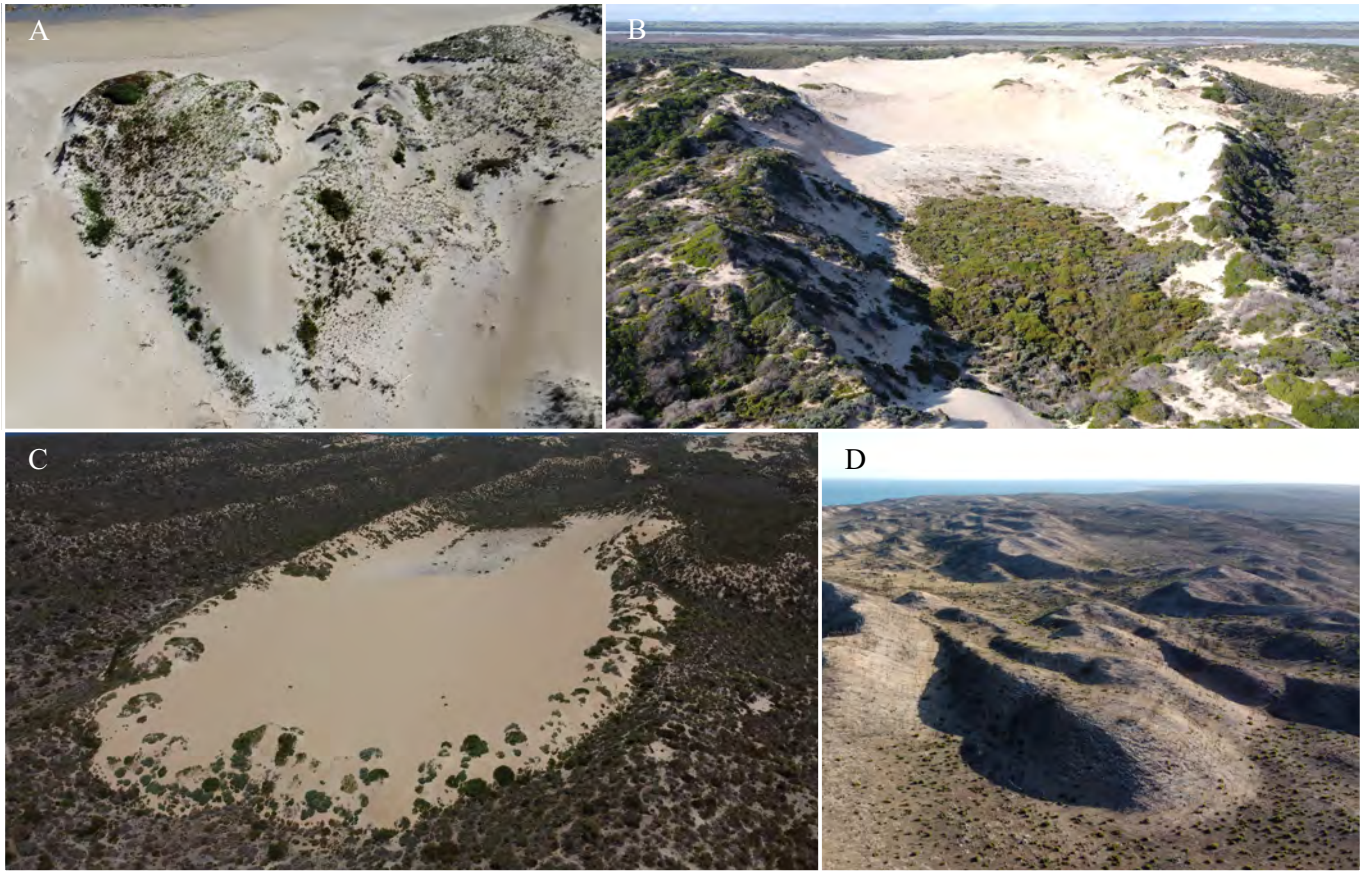


Figure 11: A: Blowout at Maurpetuis Bay, Kangaroo Island, South Australia. B: Parabolic dune near Cantara, Younghusband Peninsula, South Australia. C: Long-walled parabolic dune near Port Lincoln, South Australia. D: Nested, long-walled parabolic dunes at Snake lagoon, Kangaroo island, South Australia. Credits: P. Hesp.

664 morphology from an upwind sediment source [152, 153, 154, 155, 156]. Sand sheets and proto dunes emerging from the
 665 playa margin (Fig. 10 A; [157, 105]) develop into sinuous transverse dunes (also termed crescentic dunes), then more
 666 complex barchanoids (Fig. 10 B), to sub-parabolic and then parabolic dunes (Figs. 10 C and D) as vegetation takes
 667 root and expands [158, 50, 159, 160]. The evolution from incipient dunes at the playa margin to stagnant parabolic
 668 dunes downwind occurs over a distance of approximately 10 km. This transition has been explained by changes in
 669 field-scale aerodynamics and a reduction in groundwater salinity [160, 161]. In the aerodynamic model, vegetation
 670 growth becomes more favorable downwind as the dune roughness relative to the playa reduces near-bed wind speed
 671 and therefore sediment transport [160, 162]. The barchan-parabolic transition occurs 8 km downstream of the field
 672 margin over a characteristic distance of 2 km and is correlated with an increase in plant density (number per dune)
 673 and a drop in the groundwater table [160]. The change in morphology begins with vegetation stabilizing the margins
 674 and horns of the mobile barchan dunes. The edges being held while the central and larger volumetric portion of dunes
 675 continue to advance, the dunes switch to a parabolic morphology [159, 158].

676
 677 Different from the transition described above, parabolic dunes may instead develop directly from blowouts forming
 678 above the backshore of beaches or within continental dune fields where vigorous vegetation growth is supported
 679 [163, 164, 74]. The driving external factors affecting the transition are sediment transport rate and supply, and climate
 680 (*e.g.*, rainfall) [21, 70]. Figure 11 shows dunes at different stages of interaction with vegetation, suggesting what
 681 might be the typical evolution of a dune from its initiation on a foredune to its complete stabilization by vegetation
 682 as it migrates inland. This scenario may apply to the coastal dunefield in South Australia shown in Figure 4 I, for
 683 example. In Figure 11 A, a 50-m long blowout is beginning to develop incipient trailing ridges (note the vegetation
 684 stabilizing the edges of the depositional lobe), and if the depositional lobe continues to migrate downwind will evolve
 685 into a parabolic dune. Figure 11 B shows a large 750-m long parabolic dune with a well vegetated deflation basin. In

686 semi-arid environments, nebkha often form on the depositional-lobe margins as shown in this photograph. Figure 11
687 C illustrates an older, long-walled (2.3-km long) parabolic dune with pronounced nebkha development around the lobe
688 and trailing ridges. Trailing ridges are eventually stabilized by nebkha development and vegetation cover of the inside
689 trailing ridges. Deflation basins and plains are colonised by vegetation once the deflation has occurred down to a base
690 level fixed by the water table (*e.g.*, Figure 11 B), or hard surfaces (*e.g.* Pleistocene calcrete in Figure 11 C). Parabolic
691 dunes eventually stabilise due to running out of sediment or due to climate changes (*e.g.*, higher rainfall, lower wind
692 speeds). Figure 11 D illustrates an example where a later parabolic dune phase has created a nested parabolic dune
693 system.

694
695 The changes in dune morphology described herein present a basis for interpreting environmental conditions and
696 processes through space and time. The characteristic spatial changes in morphology from an upwind source reveal
697 the direction of the source area, sediment flux gradient, bed conditions, and the presence, absence, and feedbacks
698 of vegetation in the system. The trajectory of the morphologies toward vegetated parabolic dunes from differing
699 initial states highlights the strong control vegetation has on dune morphology. These systems also show the time-
700 varying characteristics of dune morphological evolution that has only been revealed through long-term observations,
701 field-scale experiments, and scaled laboratory experiments [122, 165, 157]. Understanding these as time-varying
702 morphological characteristics provides a foundation to interpret the recent and future climatic changes that could
703 affect dune morphology [148].

704 5.2 Revealing dune patterns on Earth from dynamical processes

705 The dynamic-based classification aims to quantify dune morphodynamics according to environmental forcing. Dune
706 type and orientation result from the competition between three dynamical processes, which are growth in height,
707 elongation, and migration. The prevailing dynamics and the balance between the three depend on boundary conditions
708 and on the regime of sand transport the dune experiences, *e.g.* the distribution of sand flux orientations, the magnitude
709 and the sequence of transport events. In order to quantify this concept, three different sand flux components are defined
710 in the reference frame of the dune crest. (i) The crest-normal component of sand flux, Q_{\perp} , is associated with dune
711 migration. (ii) The crest-parallel component of sand flux, Q_{\parallel} , is associated with dune elongation. (iii) The growth in
712 height is quantified through the gross bedform-normal transport, *i.e.*, the absolute value of crest-normal component
713 of sand flux, $|Q_{\perp}|$. Following the mass conservation principle, the rates of growth in height, of migration, and of
714 elongation correspond to the divergences of $|Q_{\perp}|$, Q_{\perp} , and Q_{\parallel} , respectively. Growth in height depends on gross
715 transport (absolute value) rather than net transport, because transports in both directions across the crest contribute
716 to dune growth [40, 41]. Before presenting analysis for various common dune types on Earth, we review the different
717 parameters that allow us to evaluate the dynamical processes and calculate their relative importance from wind data.

718 5.2.1 Parameters for the characterization of the sand transport regime and dune dynamics

719 **Sand transport regime** Using wind data from the last decade [166, 167], we calculate the instantaneous shear
720 velocities, u_* , with Equation 6, and saturated sand flux vectors on a flat sand bed, \vec{Q}_0 , with transport law Equation
721 2 using a transport threshold velocity, u_t (Appendix Section 7). We then characterize the sand transport regime at
722 a given location using the time-averaged $\langle u_* \rangle$, $\langle \vec{Q}_0 \rangle$, and $\langle \|\vec{Q}_0\| \rangle$. The mean shear velocity, $\langle u_* \rangle$, is averaged over
723 the duration of active transport, *i.e.* when $u_* > u_t$, so that the ratio $\langle u_* \rangle / u_t$ is a proxy for the average intensity of
724 transport when it occurs. Because norms of individual sand flux vectors, \vec{Q}_0 , correspond to the saturated values of
725 sand fluxes, the norm of the time averaged sand flux is equivalent to the resultant drift potential, $\|\langle \vec{Q}_0 \rangle\| = \text{RDP}$, and
726 its direction is the resultant drift direction, RDD. The time averaged norm of instantaneous sand flux is equivalent
727 to the drift potential, $\langle \|\vec{Q}_0\| \rangle = \text{DP}$, so that the ratio RDP/DP is a measure of the directional variability of sand
728 transport [168, 169]. $\text{RDP}/\text{DP} \rightarrow 1$ where sand transport tends to be unidirectional; $\text{RDP}/\text{DP} \rightarrow 0$ indicates high
729 directional variability resulting in a small net transport compared with what it would be if all transport events were
730 in the same direction.

731
732 **Sand transport over dunes** Dune dynamics are evaluated with the time-averaged components of the characteristic
733 sand flux over dunes, $\langle \vec{Q}_c \rangle$, in the reference frame of dune crest: $\langle Q_{\perp} \rangle$, $\langle Q_{\parallel} \rangle$, and $\langle |Q_{\perp}| \rangle$, which we calculate using the
734 same transport law as for Q_0 , but including the effect of *wind speed-up*, *i.e.*, the increase in wind shear velocity induced
735 by dune topography. For this purpose, we reduce dunes to straight and symmetrical ridges of constant orientation

736 and cross-section, *i.e.*, the cross section is constant along the dune and neither the size nor the slopes of the dune
737 change over time. This simplification is valid within the limit of very large dunes that fully integrate the wind regime
738 (Appendix Section 7.1). Following the approach of Jackson and Hunt [170], we assume that the relative increase in
739 wind velocity is proportional to the dune slope (increase in elevation with distance) along a wind streamline. It is
740 maximum when wind is perpendicular to dune crest. Wind speed-up modulates the norm of each instantaneous sand
741 flux, $\|\vec{Q}_c\|$, as a function of the angle between the wind and the dune crest. Therefore, the direction and magnitude
742 of the time averaged sand flux the dune experiences, $\langle\vec{Q}_c\rangle$, depend on dune orientation in a multidirectional transport
743 regime. $\langle Q_\perp\rangle$, $\langle Q_\parallel\rangle$, and $\langle|Q_\perp|\rangle$ can be evaluated for any potential dune orientation, α . The functions $\langle Q_\perp\rangle(\alpha)$,
744 $\langle Q_\parallel\rangle(\alpha)$, and $\langle|Q_\perp|\rangle(\alpha)$ allow us to predict the orientation of dunes according to the prevailing dynamical process.

746 **Evaluating dune orientation** When growth in height prevails, as in the bed instability mode where dunes develop
747 from a loose sand bed, the selected orientation of dunes, α_H , maximizes the growth rate in height, σ , or equivalently
748 the gross bedform-normal transport, $\langle|Q_\perp|\rangle$ (Section 7.3.3). In a multidirectional wind regime, this orientation can
749 be transverse, oblique or parallel to the RDD. α_H is also the expected crest orientation when migration prevails.
750 When elongation prevails, and when dunes elongate without migrating, dune crest orientation, α_E , corresponds to the
751 direction of elongation, which in this case is also the direction of the mean sand flux on the dune, $\langle\vec{Q}_c\rangle$, such that
752 $\langle Q_\perp\rangle(\alpha_E) = 0$ (no migration), and $\langle Q_\parallel\rangle(\alpha_E) > 0$ (elongation).

753 **Sand flux depending on dune orientation** We use these predicted dune orientations to determine the averaged
754 characteristic sand flux experienced by dunes in the two modes of orientation: $\langle\vec{Q}_H\rangle = \langle\vec{Q}_c(\alpha_H)\rangle$ of direction θ_H
755 for dunes whose orientation maximizes growth in height, and $\langle\vec{Q}_E\rangle = \langle\vec{Q}_c(\alpha_E)\rangle$ of direction $\theta_E = \alpha_E$ for dunes that
756 elongate without migrating. Except in singular situations, dunes whose orientation maximizes growth in height migrate
757 according to the resultant crest-normal component of sand flux of norm $\|\langle\vec{Q}_M\rangle\| = |\langle Q_\perp(\alpha_H)\rangle| = \|\langle\vec{Q}_H\rangle\| |\sin(\theta_H - \alpha_H)|$
758 and direction θ_M , which is perpendicular to the dune orientation, α_H . The characteristic migration velocity equals
759 $\|\langle\vec{Q}_M\rangle\|/H$, where H is dune height. Similarly, the elongation rate of a non-migrating dune can be evaluated with
760 $\|\langle\vec{Q}_E\rangle\|/H$ [171, 126].

761 **Competition between dynamics** We argue that the orientation of dunes depends on the prevailing growth mech-
762 anism and that dune type depends on the competition (relative balance) between the three dynamical processes. The
763 main factor selecting the prevailing growth mechanism for dune orientation appears to be the mobility of the sand
764 bed and the boundary conditions. The variety of shapes within a given regime of orientation, as the balance between
765 migration and elongation on a starved bed, must depend on the wind regime alone when all boundary conditions
766 are fixed. Aiming at a fully predictive phase diagram of dune type and equilibrium shape, we propose and review
767 from previous studies several dimensionless ratios built from the different sand fluxes just discussed to assess the
768 competition between the three dynamics. For each of the two dune orientations, the three sand fluxes, $|\langle Q_\parallel\rangle|$, $|\langle Q_\perp\rangle|$,
769 and $\langle|Q_\perp|\rangle$, associated with the three dynamics, elongation, migration, and growth in height, can be calculated. Be-
770 cause dunes in the elongation mode do not migrate ($\langle Q_\perp\rangle(\alpha_E) = 0$), this makes 5 different fluxes ($2 \times 3 - 1$), with
771 which we can build 4 independent dimensionless ratios, *i.e.*, that cannot be built as a combination of the others. We
772 chose 3 ratios that allow us to compare dynamics of dunes in a given mode of orientation: $|\langle Q_\parallel(\alpha_H)\rangle/\langle Q_\perp(\alpha_H)\rangle|$
773 and $|\langle|Q_\perp(\alpha_H)|\rangle/\langle Q_\perp(\alpha_H)\rangle|$ for dunes whose orientation maximizes growth in height, and $|\langle|Q_\perp(\alpha_E)|\rangle/\langle Q_\parallel(\alpha_E)\rangle|$ for
774 dunes in the elongation mode of orientation. In addition, we select one ratio that compares the two different modes of
775 orientation: $\langle|Q_\perp(\alpha_E)|\rangle/\langle|Q_\perp(\alpha_H)|\rangle = \sigma_E/\sigma_H$, the ratio between growth rates in height of dunes under the two modes
776 of orientation. In the simplified framework we developed, these 4 ratios fully define the parameter space associated
777 with wind forcing.

778
779 σ_E/σ_H varies between 0 and 1, and has been proposed to evaluate the stability of elongating dunes when they
780 develop from a point source [64]. The stability of dunes elongating on a starved bed from a sand source in bimodal flow
781 regimes was numerically studied in [64]. Depending on the parameters (divergence angle and transport ratio between
782 alternate flow directions), an elongating linear dune or a train of propagating asymmetric barchans is observed. The
783 ratio σ_E/σ_H was found to discriminate between the two morphologies ; elongating dunes are observed when σ_E/σ_H is
784 larger than 0.6. This result reflects the fact that perpendicular flows are required to build a dune, regardless of its
785 orientation. Note that in experiments under water with a bimodal flow regime, a linear dune elongating from a point

Figure	12	13	14	15	16	17	18	19	20		
Bed surface	CB	CB	CB	SB	SB	SB	SB	SB	PSB		
Shear velocity											
$\langle u_* \rangle$ (m s ⁻¹)	0.243	0.24	0.229	0.218	0.227	0.288	0.238	0.236	0.226		
u_t (m s ⁻¹)	0.153	0.153	0.153	0.153	0.153	0.153	0.153	0.153	0.153		
$\langle u_* \rangle / u_t$	1.585	1.56	1.498	1.425	1.484	1.879	1.56	1.544	1.47		
Flux on a flat sand bed											
DP = $\langle \ \vec{Q}\ \rangle$ (m ² yr ⁻¹)	39.1	37.2	31.8	25.1	43.8	118.9	39	30.2	29.5		
RDP = $\ \langle \vec{Q} \rangle\ $ (m ² yr ⁻¹)	33.8	8.3	14.9	14.4	19.2	100.5	28.7	15.4	2.5		
RDP/DP	0.86	0.22	0.47	0.57	0.44	0.85	0.74	0.51	0.08		
RDD (deg., mod 360°)	48.6	294.9	337.5	214.1	226.6	253.1	88.5	219.4	62.7		
Dune orientations and flux at the crest											
α_H (deg., mod 360°)	136.6	51.6	148.2	144.2	111.4	162.9	168.9	110.2	127.7		
α_E (deg., mod 360°)	45.2	272.5	328.8	202.9	237.4	251.5	90.3	222.3	30.8	119.7	253
$\ \langle \vec{Q}_H \rangle\ = \ \langle \vec{Q}_c(\alpha_H) \rangle\ $ (m ² yr ⁻¹)	151.0	29.4	42.9	68.1	87.6	393.7	142.1	67.5	9.1		
θ_H (deg., mod 360°)	48.2	292.4	329.3	223.8	220.1	253.3	86.4	212.2	112		
$\ \langle \vec{Q}_M \rangle\ = \langle Q_\perp(\alpha_H) \rangle $ (m ² yr ⁻¹)	150.1	25.7	0.87	67	82.9	393.7	140.8	66.1	2.45		
$\ \langle \vec{Q}_E \rangle\ = \ \langle \vec{Q}_c(\alpha_E) \rangle\ $ (m ² yr ⁻¹)	57.1	18.5	42.9	41.4	44.2	144.3	46.7	37.4	11.6	7.7	3.8
$\Delta\alpha = \alpha_H - \alpha_E$ (deg.)	88.6	40.9	0.7	58.7	53.9	88.7	78.5	67.9	86.9	10	54.7
$\Delta\theta = \theta_H - \alpha_E$ (deg.)	3.0	19.9	0.5	20.9	-17.3	1.84	3.9	-10.2	81.2	-7.7	-141
Competition between dynamics											
σ_E / σ_H	0.18	0.67	1	0.61	0.67	0.18	0.23	0.4	0.23	0.98	0.48
$ \langle Q_\parallel(\alpha_H) \rangle / \langle Q_\perp(\alpha_H) \rangle $	0.03	0.56	49.4	0.18	0.34	0.01	0.13	0.21	0.03		
$ \langle Q_\perp(\alpha_H) \rangle / \langle Q_\perp(\alpha_H) \rangle $	1.09	5.9	115	1.54	1.88	1.07	1.18	1.83	54.05		
$ \langle Q_\perp(\alpha_E) \rangle / \langle Q_\parallel(\alpha_E) \rangle $	0.51	5.5	2.31	1.51	2.38	0.53	0.81	1.31	2.7	17	16.8

Table 3: Wind velocity, sand flux, and dune orientation calculated from wind data at the different locations shown in Figs. 12 - 20. Orientations are with respect to the east. Bed surface: CB, SB, and PSB stand for sand covered bed, starved bed, and partially starved bed, respectively. α_H is the dune orientation that maximizes growth rate in height. α_E is the dune orientation in the elongation mode. θ_H is the direction of mean sand flux at dune crest for dunes whose orientation is α_H .

786 source on a starved bed turns into a highly asymmetric barchan with an oblique arm if the sand source shuts off so
 787 that the dune is free to migrate [93]. The sand source helps to mute dune migration. As such, the ratio σ_E/σ_H cannot
 788 be used to distinguish the prevailing dynamics for dune orientation under different boundary conditions, including bed
 789 mobility. This could be the case if the boundary conditions were specifically taken into account in the calculations,
 790 which is not the case here.

791 When dunes develop on a sand covered bed, growth in height prevails and the orientation of the dune crest is
 792 α_H . Within given boundary conditions, the variety of shapes should then depend on the two dimensionless ratios
 793 $|\langle Q_{\parallel}(\alpha_H) \rangle / \langle Q_{\perp}(\alpha_H) \rangle|$ and $|\langle |Q_{\perp}(\alpha_H)| \rangle / \langle Q_{\perp}(\alpha_H) \rangle|$. The ratio between the crest-parallel and the crest-normal com-
 794 ponents of sand flux, $|\langle Q_{\parallel}(\alpha_H) \rangle / \langle Q_{\perp}(\alpha_H) \rangle|$, was conceptually proposed to quantify the impact of a wind regime on
 795 straightness and sinuosity of morphologically long dunes [65]. The crest-parallel component should favor straight-
 796 ness whereas the crest-normal component, causing migration, should favor sinuosity. We note, however, that crest-
 797 straightness could be favored by the gross crest-parallel component instead of the crest-parallel component of resultant,
 798 so that the ratio $|\langle |Q_{\parallel}(\alpha_H)| \rangle / \langle Q_{\perp}(\alpha_H) \rangle|$ could be more appropriate. Growth in height also building the spatial co-
 799 herence of a dune, the ratio $|\langle |Q_{\perp}(\alpha_H)| \rangle / \langle Q_{\perp}(\alpha_H) \rangle|$ may be relevant as well to evaluate dunes straightness.

800 For dunes migrating on a starved bed, which also have an orientation that maximizes growth in height, the
 801 migrating direction, θ_M , does not necessarily correspond to the resultant transport direction, θ_H . This asymmetry in
 802 sand fluxes must drive an asymmetry in dune morphology that the ratio $|\langle Q_{\parallel}(\alpha_H) \rangle / \langle Q_{\perp}(\alpha_H) \rangle|$ may account for.

803 For dunes elongating without migrating, the crest-normal component of sand flux is constant and null, so that the
 804 ratio $|\langle |Q_{\perp}(\alpha_E)| \rangle / \langle Q_{\parallel}(\alpha_E) \rangle|$ alone should allow us to distinguish between morphologies.

805 Finally, since the very existence of a dune relies on the ability of winds to raise it in height, the competition between
 806 migration and elongation for selecting dune orientation on a starved bed could be assessed by comparing the two ratios
 807 $|\langle |Q_{\perp}(\alpha_H)| \rangle / \langle Q_{\perp}(\alpha_H) \rangle|$ and $|\langle |Q_{\perp}(\alpha_E)| \rangle / \langle Q_{\parallel}(\alpha_E) \rangle|$. A mode should be promoted when it favors the growth in height
 808 over migration or elongation.

809
 810 Although only a few of these parameters have been tested in previous studies, and that their calculations are based
 811 on numerous simplifications, they formalize a framework for studying dune morphology based on dune dynamics. We
 812 already remarked that fully predictive parameters should take into account boundary conditions, which affect sand
 813 supply and the divergence of sand flux at the dune scale. Some parameters, such as free flux, may be difficult to infer
 814 remotely. For example, the migration velocity of an isolated dune like a barchan decreases and its size may increase
 815 when it captures an incoming free flux. More importantly, useful predictive parameters should involve the temporal
 816 sequence of transport directions and the relative size of dune to transport capacity of individual wind events. For
 817 example, experiments in water and simulations have shown that a dune initially turns over before migrating when
 818 subjected to an abrupt change in flow direction [172, 41, 173, 174, 175]. Furthermore, in bidirectional wind regime,
 819 numerical simulations show that the ratio between the time period of wind reversal and the characteristic turnover
 820 time of dune controls the crest sinuosity [63, 176].

821 5.2.2 Example analyses for various characteristic free dune types on Earth

822 Different types of free dunes from the geomorphological classification 2 are shown in Figures 12-20. The sand trans-
 823 port parameters for characterization of their formative processes and orientation according to the dynamics-based
 824 classification are gathered in Table 3. Transport parameters are calculated from wind data provided by the ECMWF
 825 ERA5-Land reanalysis [166, 167] from January 1, 2010 to December 31, 2020.

826 **Periodic long-crested dunes on sand covered beds** Periodic long-crested dunes on sand covered beds can form
 827 with orientations that are transverse (Fig. 12), oblique (Fig. 13), or longitudinal (Fig. 14) with respect to the RDD.
 828 The expected prevailing process on a sand bed is the growth in height of a periodic pattern (bed instability) with dunes
 829 forming in the orientation that maximizes the gross bedform-normal transport, *i.e.* $\langle |Q_{\perp}| \rangle$, which is well verified in
 830 these three examples. Where the wind regime tends to be unidirectional, long transverse dune patterns develop with
 831 an asymmetric shape showing a clear difference between gentle stoss side and steep lee face (Fig. 12). Under bimodal
 832 wind regimes, long-crested dune patterns with more symmetric slopes are observed. As predicted from theory, they are
 833 usually oblique in orientation (Fig. 13), and only tend to be longitudinal where the transport ratio between the two
 834 main wind directions is close to one (Fig. 14) [65, 41]. Transverse dune patterns in Fig. 12 show a range of sizes with
 835 smaller dunes either being superimposed on larger ones or spatially separated. Subjected to an almost unidirectional
 836 flux regime, they must migrate as predicted by the net sand flux associated with migration, $\langle Q_M \rangle = \langle |Q_{\perp}(\alpha_H)| \rangle$.
 837 In these examples, the ratios $|\langle Q_{\parallel}(\alpha_H) \rangle / \langle Q_{\perp}(\alpha_H) \rangle|$ and $\langle |Q_{\perp}(\alpha_H)| \rangle / \langle Q_{\perp}(\alpha_H) \rangle|$ increase by more than two orders of
 838 magnitude as the predicted dune orientation goes from transverse, to oblique, to longitudinal. This may explain why

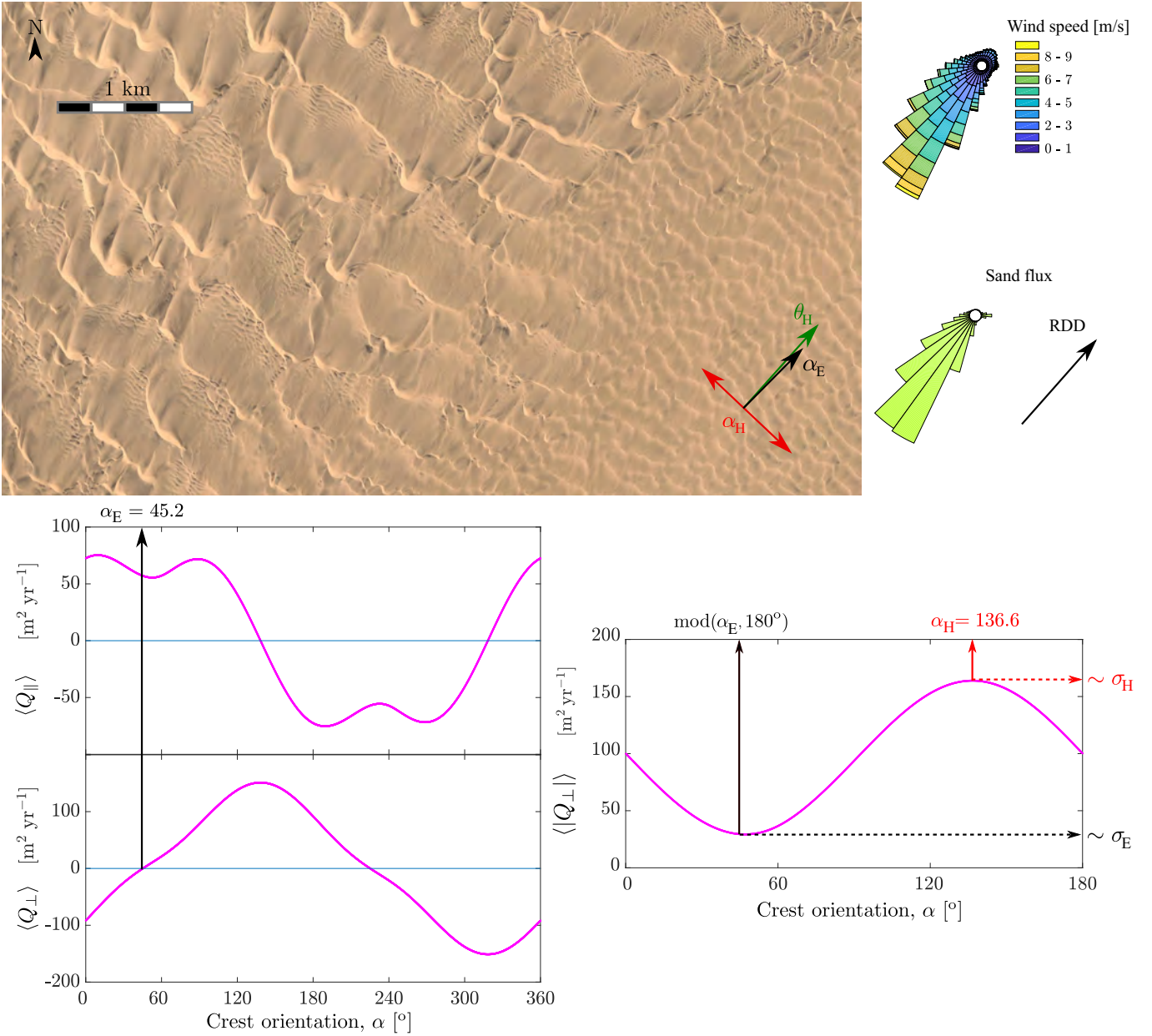


Figure 12: Transverse long-crested dunes on sand-covered bed in the Moçâmedes desert in Angola (16.6°S, 11.9°E). Roses, which point upstream, show the wind regime and the corresponding sand flux regime from 2010 to 2020. See Sections 5.2 and 7 for an explanation of parameters and details of calculations. Graphics show the calculated characteristic sand flux components $\langle Q_{\perp} \rangle$, $\langle Q_{\parallel} \rangle$ and $\langle |Q_{\perp}| \rangle$ over a dune with orientation α . Angles are measured relatively to east direction. The dune orientation α_I that maximizes $\langle |Q_{\perp}| \rangle$, the dune direction α_E for elongation ($\langle Q_{\parallel} \rangle > 0$) without migration ($\langle Q_{\perp} \rangle = 0$) and the direction of the sand flux θ_I over a dune of orientation α_I are shown by arrows in the image.

839 transverse dunes in Fig. 12 resemble barchanoid ridges, whereas oblique dunes in Fig. 13 are straighter. Although
 840 these ratios are even larger for longitudinal dunes in Fig. 14, crests are not straight. However, we observe that in
 841 this region, sand transport directions are much more widely distributed. Here, the amplitude of variation of gross
 842 bedform-normal transport as a function of crest orientation, $\langle |Q_{\perp}| \rangle(\alpha)$, varies by a factor 1.4, compared to a factor
 843 3 for transport regimes in Figures 12 and 13. This smaller value may lead to a weaker selection of a well defined
 844 orientation, especially at small scale, explaining the greater variability in crest alignment observed for the longitudinal
 845 dunes in Fig. 14.

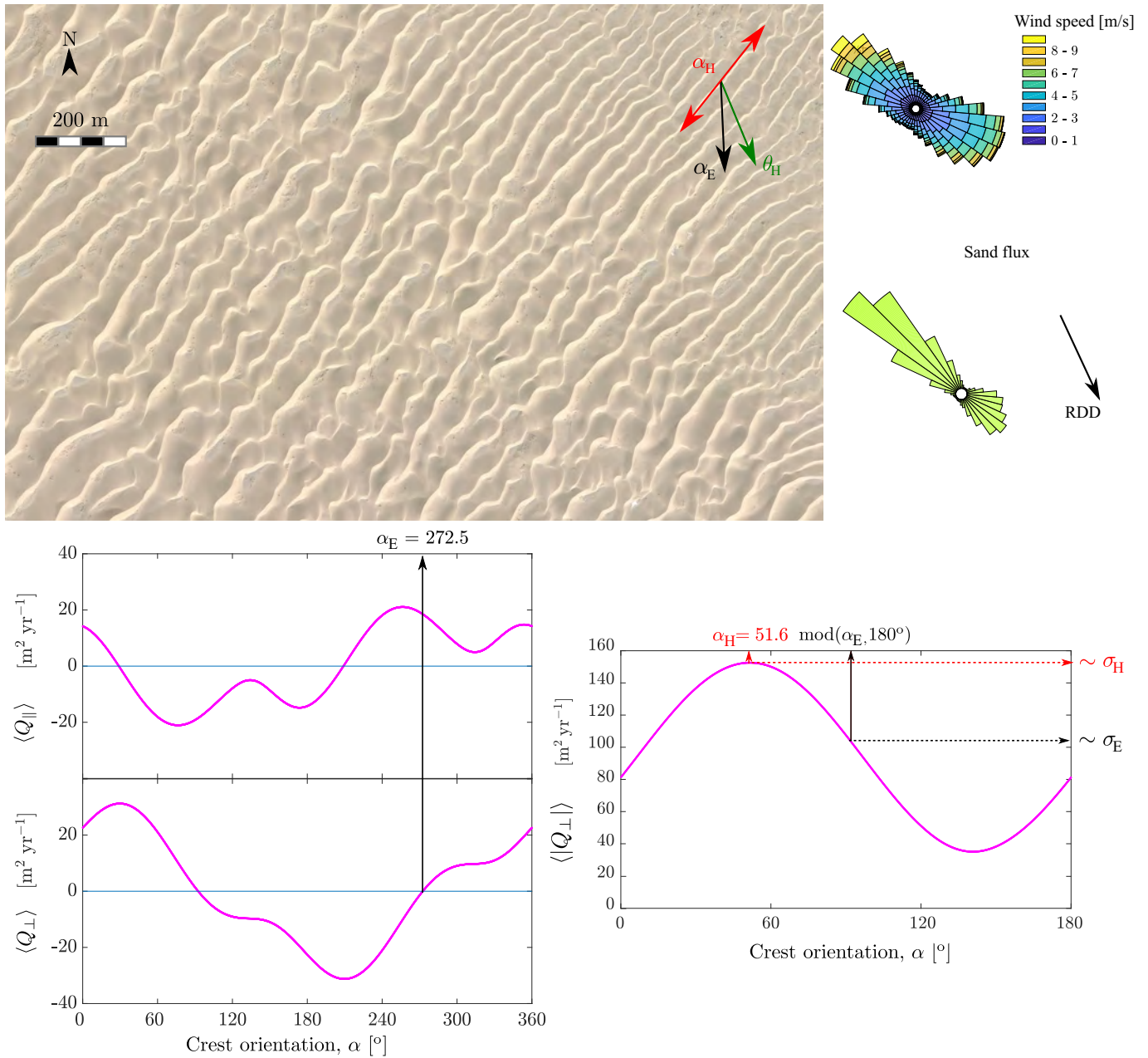


Figure 13: Oblique long-crested dunes in the Tengger desert in China (37.7°N, 105.1°E). See caption of fig. 12 for an explanation of parameters and plots.

846 **Dunes on starved beds** Starved interdunes should promote elongation or migration. Elongation without migration
 847 can be favoured by boundary conditions such as a localized sand source, which inhibits migration. Extended lee dunes
 848 are good candidates, like the isolated linear longitudinal dune extending downwind from a large obstacle shown in
 849 Figure 15. The predicted orientation for elongation without migration, α_E , corresponds to the observed orientation
 850 of this dune and is significantly different from the oblique alignment predicted for dunes whose orientation maximizes
 851 growth in height, α_H . In a different bimodal wind regime, the larger periodic linear longitudinal dunes shown in
 852 Figure 16 also exhibit an orientation that corresponds to the elongation mode. Longitudinal linear dunes, because
 853 they can elongate along the sand flow paths, are the most likely to extend over long distances. These massive dunes
 854 follow a major sand flow path that extends from southwestern Algeria to Mauritania. Unlike the bed instability,
 855 experiments and numerical simulations have shown that the elongation process does not directly determine dune size
 856 or wavelength [41, 177]. The periodicity and sizes are instead imposed by boundary and initial conditions, which,

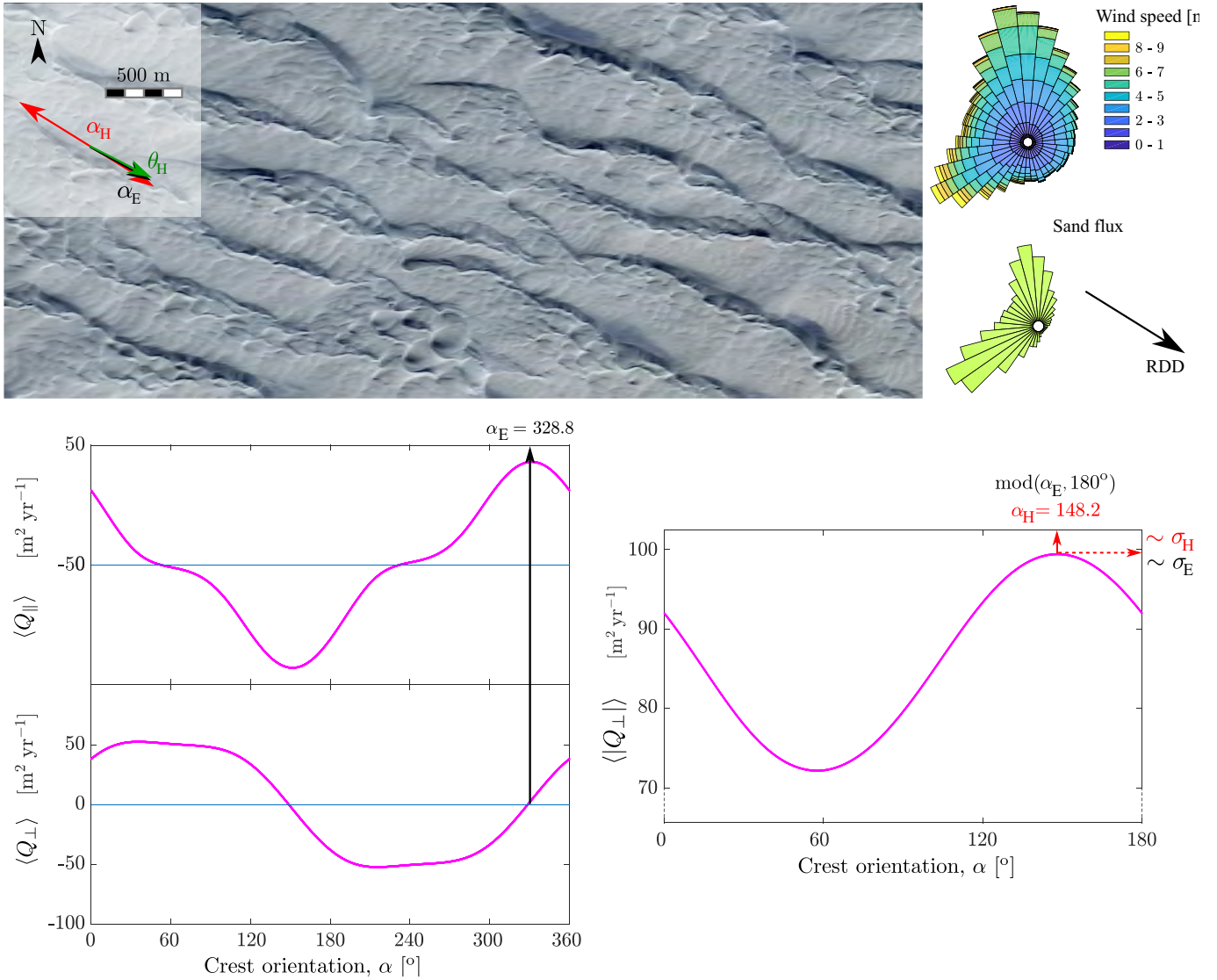


Figure 14: Longitudinal long-crested dunes on sand-covered bed in the Northern Sinai in Egypt (30.9°N, 33.1°E). See caption of fig. 12 for an explanation of parameters and plots.

857 with self-organization, are possible explanations for the periodic pattern in the elongation mode observed in this field.
 858 However, we note that the ratio σ_E/σ_H is quite large (0.67 when the largest possible value is 1), and, according to [64],
 859 this enables dunes in this region to be oriented in the elongating mode, as shown by the isolated linear dune shown in
 860 Fig. 15 ($\sigma_E/\sigma_H = 0.61$).

861 Barchan dunes, such as those observed in the Western Sahara (Fig. 17), are the archetype of migrating dunes.
 862 Here, and most often for symmetric barchans, the sand flux regime is unidirectional. Barchan slip faces are oriented
 863 perpendicular to the transport direction, which, for a unidirectional transport regime, is also the predicted orientation
 864 when growth in height prevails.

865 **Towards complex patterns** The recognizable shape and ubiquity of barchan dunes on Earth and Mars have made
 866 them a popular subject of study, so much so that models have been developed to explain ‘seif dunes’, *i.e.*, isolated
 867 longitudinal linear dunes on a starved bed, as morphological evolutions of this elemental dune [32, 178, 94, 179]. This
 868 approach is certainly partly motivated by the widespread observation of asymmetric barchans with an elongated arm.
 869 Four causes for barchan asymmetry have been identified [180, 92, 181]: dune collision, asymmetry of influx, inclined
 870 topography, and bidirectional winds. The autogenic processes of dune collision and asymmetry of influx are very
 871 common in barchan dune fields [115, 116] but only cause transitional dune asymmetry, localized in space and time.

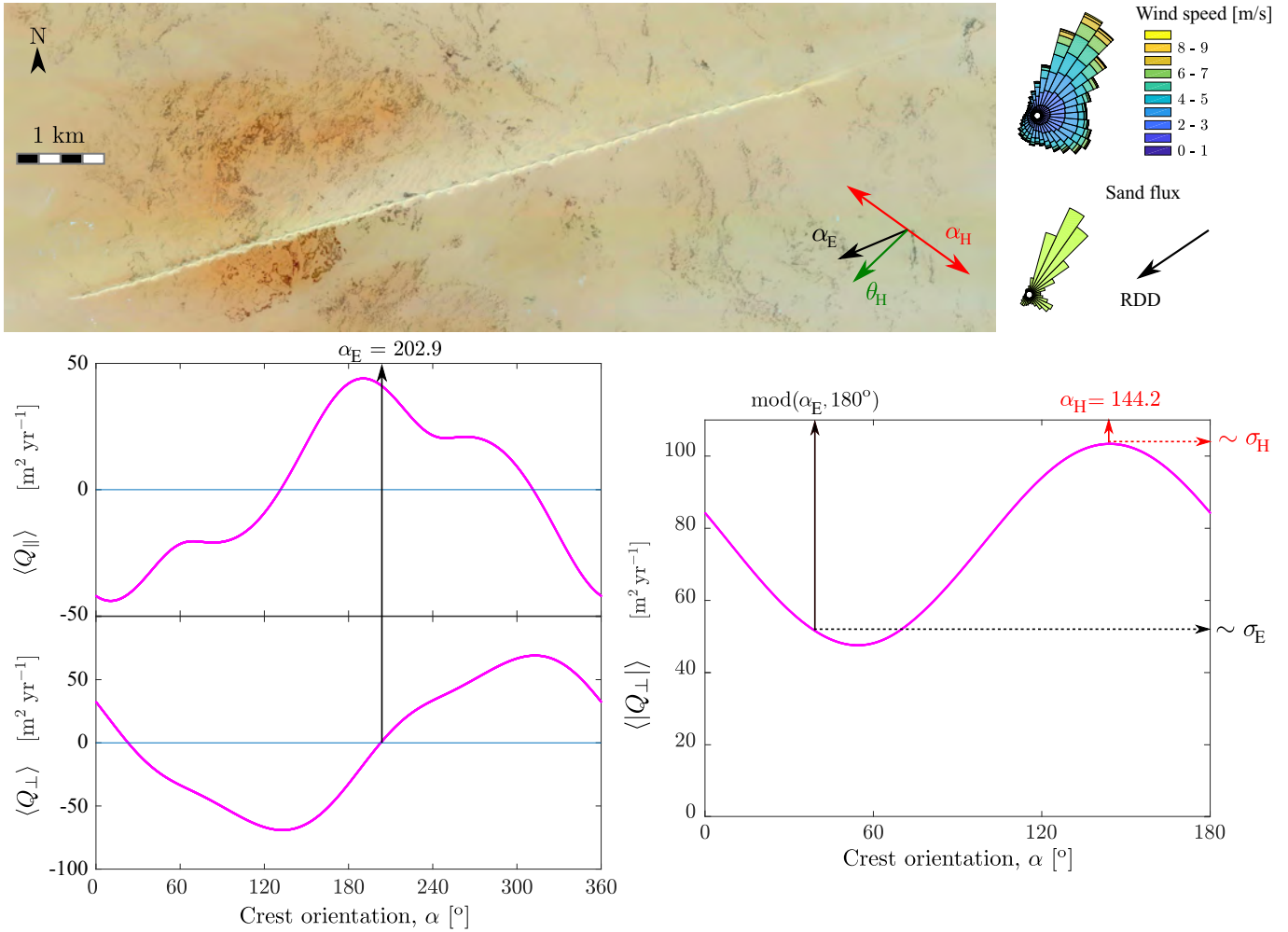


Figure 15: Isolated longitudinal linear dune on a starved bed in the Ténéré in Niger (18.8°N, 12.6°E). It extends downwind from an obstacle.

See caption of fig. 12 for an explanation of parameters and plots. It is worth noting that the direction of elongation does not generally correspond to the direction for which sand flux along crest is maximum but to the one for which crest-normal component of sand flux vanishes (no migration) and crest-parallel component is positive. There, the derivative of the crest-normal component is negative, which makes the elongation direction a stable equilibrium.

872 Along the Skeleton coast of Namibia, asymmetric barchans always have an elongated arm on the same side, which is
 873 ascribed to the bidirectional sand flux regime (Fig. 18). Only larger dunes have an elongated arm. This might be
 874 because smaller dunes are not large enough to integrate the entire period of wind reorientation. In this bidirectional
 875 regime of sand flux, the direction of sand flux for a dune that maximizes growth in height, θ_H , does not correspond
 876 to the direction of migration, θ_M . Here, the two directions differ by 7.5°, compared with 0.4° in the symmetric
 877 barchans example (Fig. 17). We believe this small difference drives an asymmetric sand redistribution between the
 878 two arms, which in this case leads to the elongation of the arm that is fed by the crest-parallel component of sand
 879 flux. In this example, the ratio $|\langle Q_{\parallel}(\alpha_H) \rangle / \langle Q_{\perp}(\alpha_H) \rangle|$ equals 0.13, which is significantly larger than the value for the
 880 symmetric barchans example (0.01, Fig. 17). Such asymmetric barchans with an elongated arm can also be seen as
 881 combinations of a migrating dune and an elongating dune at different scales. The barchan dune (large scale) has little
 882 sediment supply but is a source of sand for the elongated arm (smaller scale). As a result of dune migration, the
 883 orientation of the elongated arm is shifted in the opposite direction of the migration from that which would prevail
 884 for elongation without migration. Conversely, an elongating dune can be seen as an extremely asymmetric barchan,
 885 whose crescentic base is prevented from migrating. This picture is supported for example by the dune termination
 886 of one of the large linear longitudinal dunes on a partially starved bed observed in the bottom right corner of Fig.
 887 16. The ratio $|\langle Q_{\parallel}(\alpha_H) \rangle / \langle Q_{\perp}(\alpha_H) \rangle|$ increases consistently with dune asymmetry in the three examples of symmetric

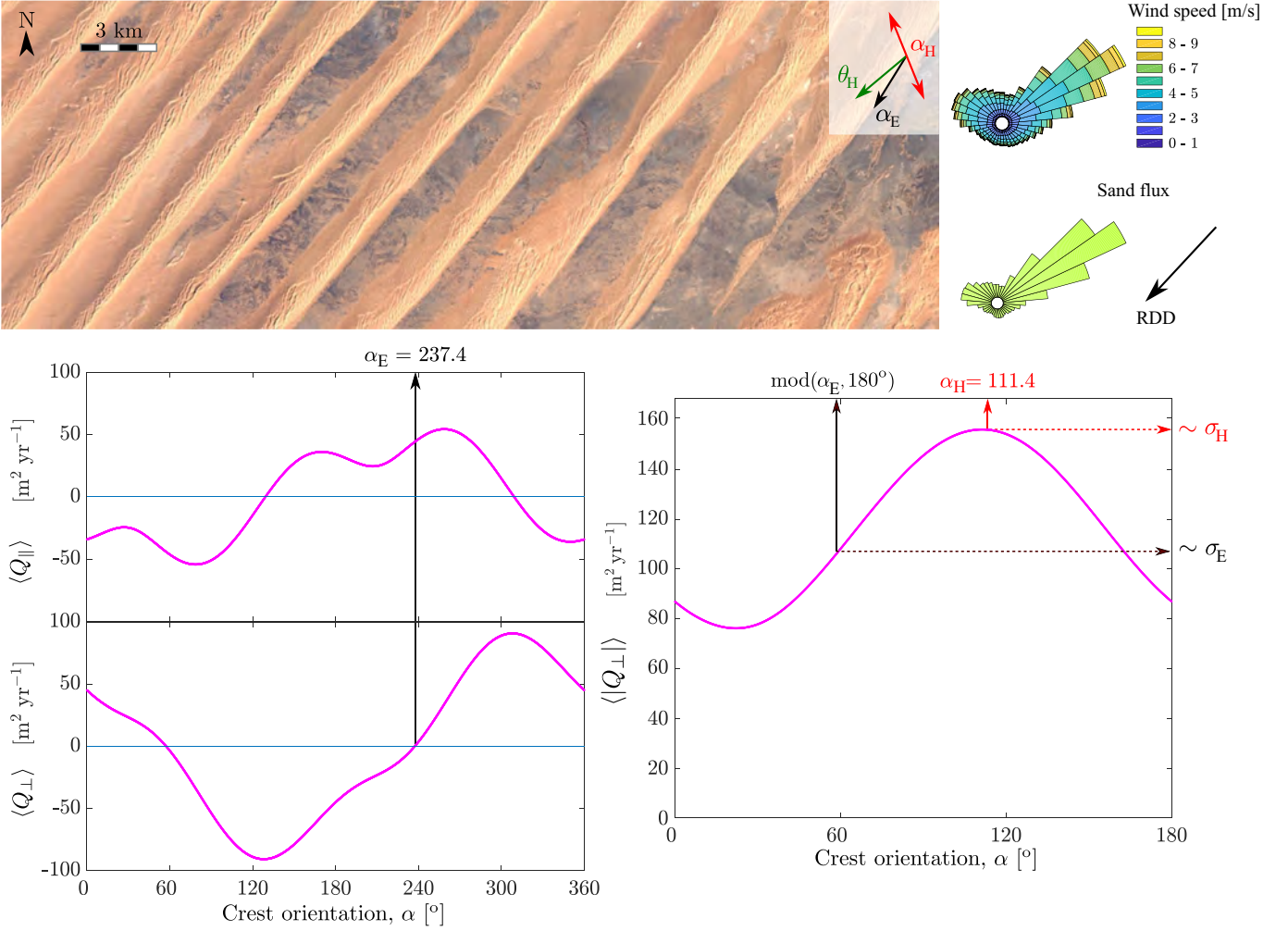


Figure 16: Periodic longitudinal linear dunes with partially starved interdune area in the Chech desert in Namibia (24.3°N, 4.4°E).

See caption of fig. 12 for an explanation of parameters and plots.

barchans, asymmetric barchans and large longitudinal linear dunes.

Asymmetric barchans are isolated structures, which clearly exemplify a coexistence of the different formative processes. Meanwhile, most sand seas exhibit complex dune patterns that may also result from such a coexistence. ‘Raked linear dunes’, such as those shown in Figure 19, are another example of this coexistence. These dunes, which lie on an armored bed composed of coarse grains, have a constant orientation for considerable distances and a marked asymmetry between a periodic pattern of semi-crescentic structures on one side and a continuous slope on the other. This semi-crescentic periodic pattern has been described as resulting from the development of superimposed dunes that grow in height and migrate in the bed instability mode on a linear dune that elongates [126]. The coexistence of these processes produces coupled primary and secondary patterns with similar height but with different shapes and orientations, which are oblique to each other. The orientation of primary linear ridges corresponds to the predicted orientation for dunes in the elongating mode, albeit σ_E/σ_H has an intermediate value. The tri-directional transport regime also enables the secondary raked pattern to develop on the leeward side of the primary linear ridges, in consistency with the predicted migration direction for dunes in the bed instability mode. Quantitatively predicting the occurrence of this pattern still remains challenging.

Finally, we examine an example of star dunes (Fig. 20), one of the most emblematic intricate dune patterns. In this example in the Namib desert, the bed is partially starved and, starting from dune summits, we observe three main directions for the arms: (i) towards the northeast, (ii) the northwest, and (iii) the southeast. Star dunes should be observed in zones where the migration process is muted and where elongation and/or growth in height may be promoted

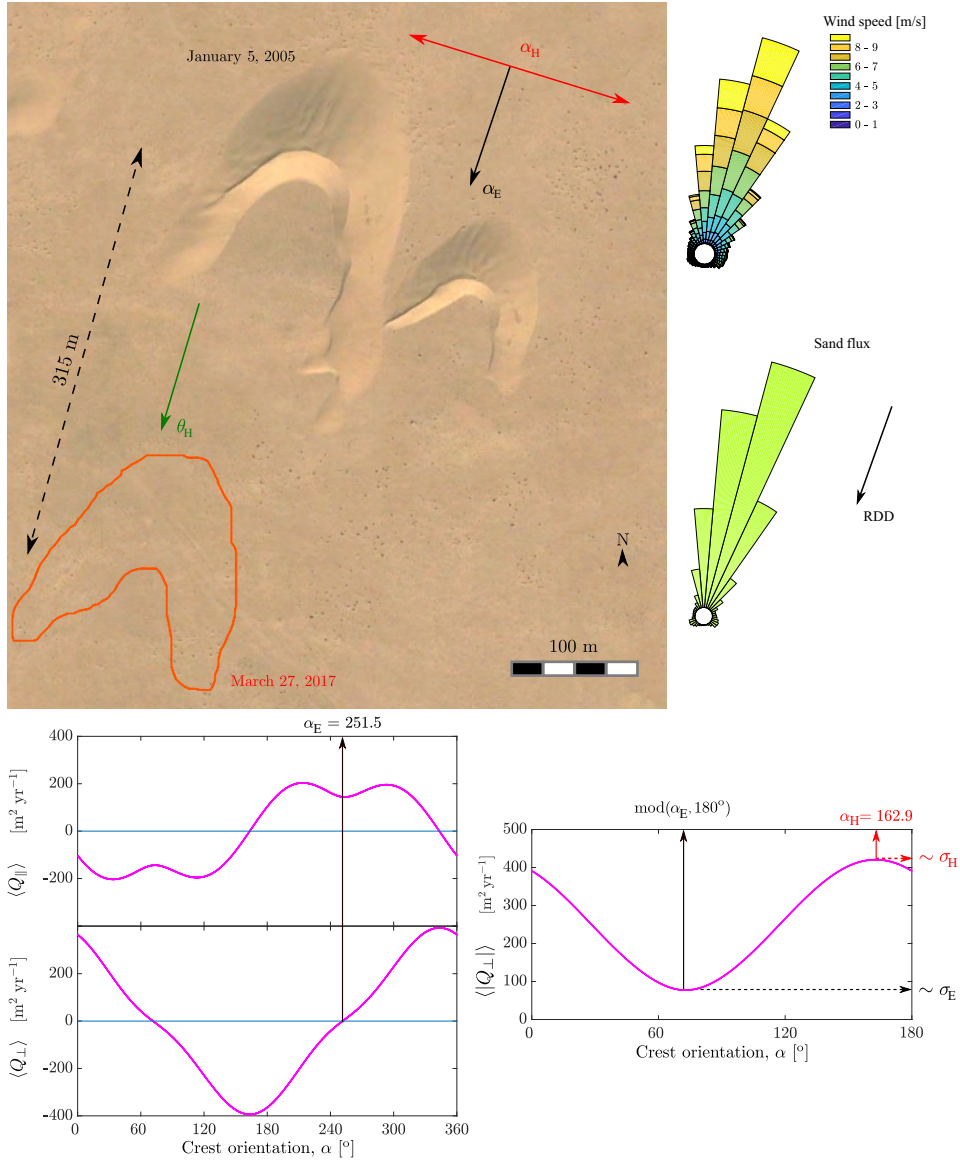


Figure 17: Barchan dune on a starved bed in Western Sahara in Morocco (27.5°N, 13.1°W).

The red contour shows location of the main barchan 12 years after the snapshot was taken. See caption of fig. 12 for an explanation of parameters and plots.

906 in multiple directions. In this specific area, the ratio RDP/DP is close to zero so that dunes should barely migrate.
 907 The model shows three directions for elongation, which is made possible by taking into account the wind speed-up:
 908 towards the northeast ($\alpha_{E,1}$), the northwest ($\alpha_{E,2}$), and the southwest ($\alpha_{E,3}$). The predicted orientation for dunes that
 909 maximizes growth in height is very close to the northwest direction of elongation, $\alpha_{E,2}$. For this direction of elongation,
 910 the ratio σ_E/σ_H is very close to one. This is clearly the prevalent crest orientation in the field. However, an orientation
 911 close to $\alpha_{E,1}$ is observed, although the ratio $\sigma_E/\sigma_H = 0.23$ is significantly smaller than the theoretical threshold value
 912 of 0.6. On the other hand, the predicted orientation $\alpha_{E,3}$ is not observed, while the ratio σ_E/σ_H is larger (but still
 913 below the theoretical threshold value). Instead, we observe a direction towards the southeast, opposite to the other
 914 prevailing direction, which is towards the northwest. Both orientations could therefore correspond to orientation
 915 that maximizes growth in height, which would be promoted by migration. We note that $|\langle Q_{\perp}(\alpha_H) \rangle|/\langle Q_{\perp}(\alpha_H) \rangle|$ is
 916 much larger than any of the values of $|\langle Q_{\perp}(\alpha_E) \rangle|/\langle Q_{\parallel}(\alpha_E) \rangle|$. Such a pattern with several orientations is still difficult
 917 to predict correctly. This could be partly due to the fact that the pattern is significantly different from the symmetric
 918 linear ridges assumed in calculations.

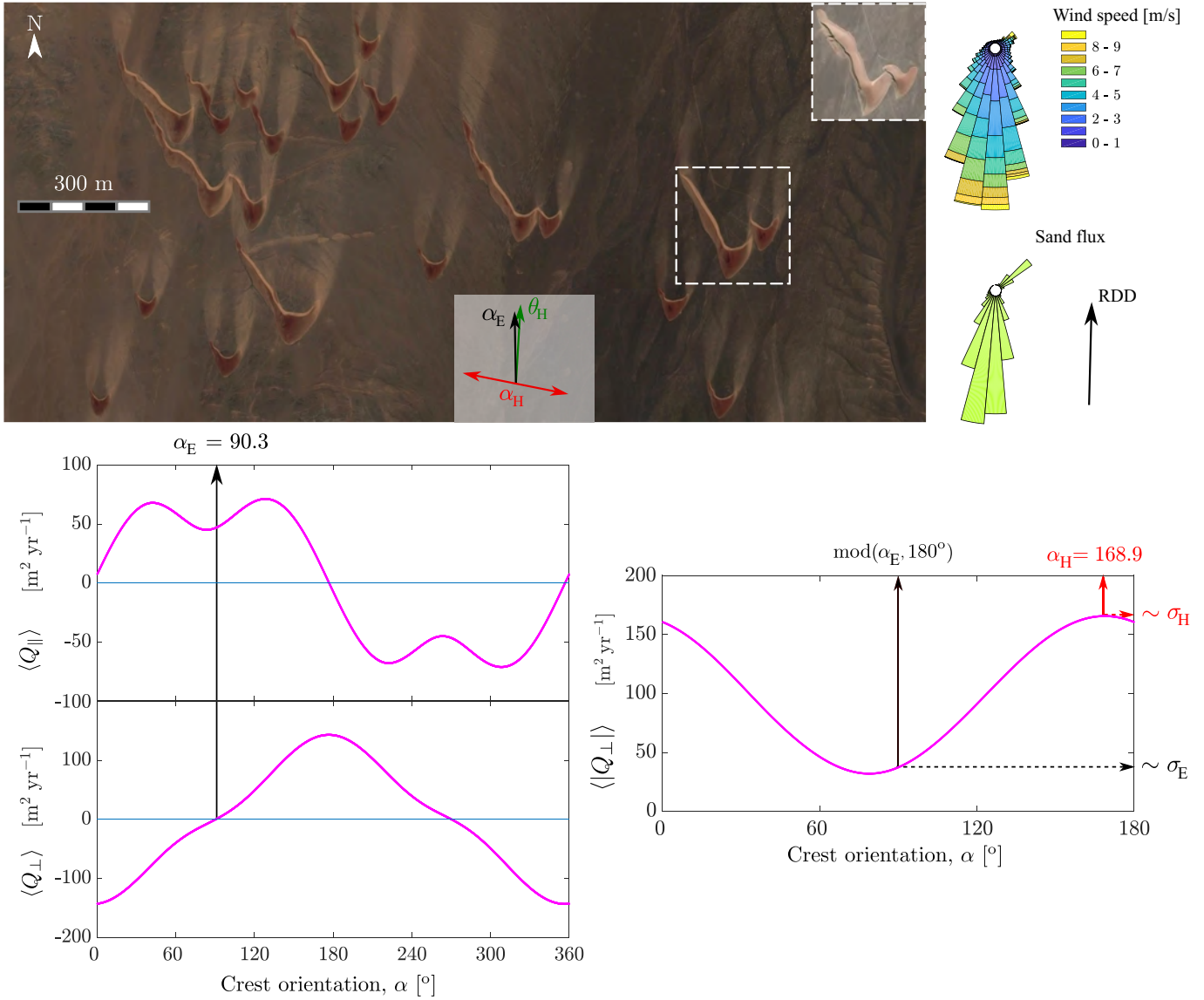


Figure 18: Asymmetric barchan dune with an elongated horn on a starved bed on the Skeleton coast in Namibia (20.1°S, 13.3°E).

The inset shows the dunes in the white dashed box after a secondary wind event. See caption of fig. 12 for an explanation of parameters and plots.

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As demonstrated in the examples above, the proposed framework for dynamics-based dune classification appears promising. In most cases, it correctly predicts dune orientation. However, the definitive phase diagram of dune morphologies is still far from being developed, and the relevant parameters are only very partially determined. Such a comprehensive phase diagram requires a more exhaustive study than the few examples discussed here, for which we note that the values of the various dimensionless parameters are correlated. The model may also gain predictive power by further taking into account the coupling between dune morphology and sand transport. A first step could be to include the larger sand flux that occurs during crest reversals, as observed over the entire height of the elongated arm (inset in Fig. 18). Such a refinement of sand flux calculations requires consideration of the variation of the upwind slope with the time series of wind directions.

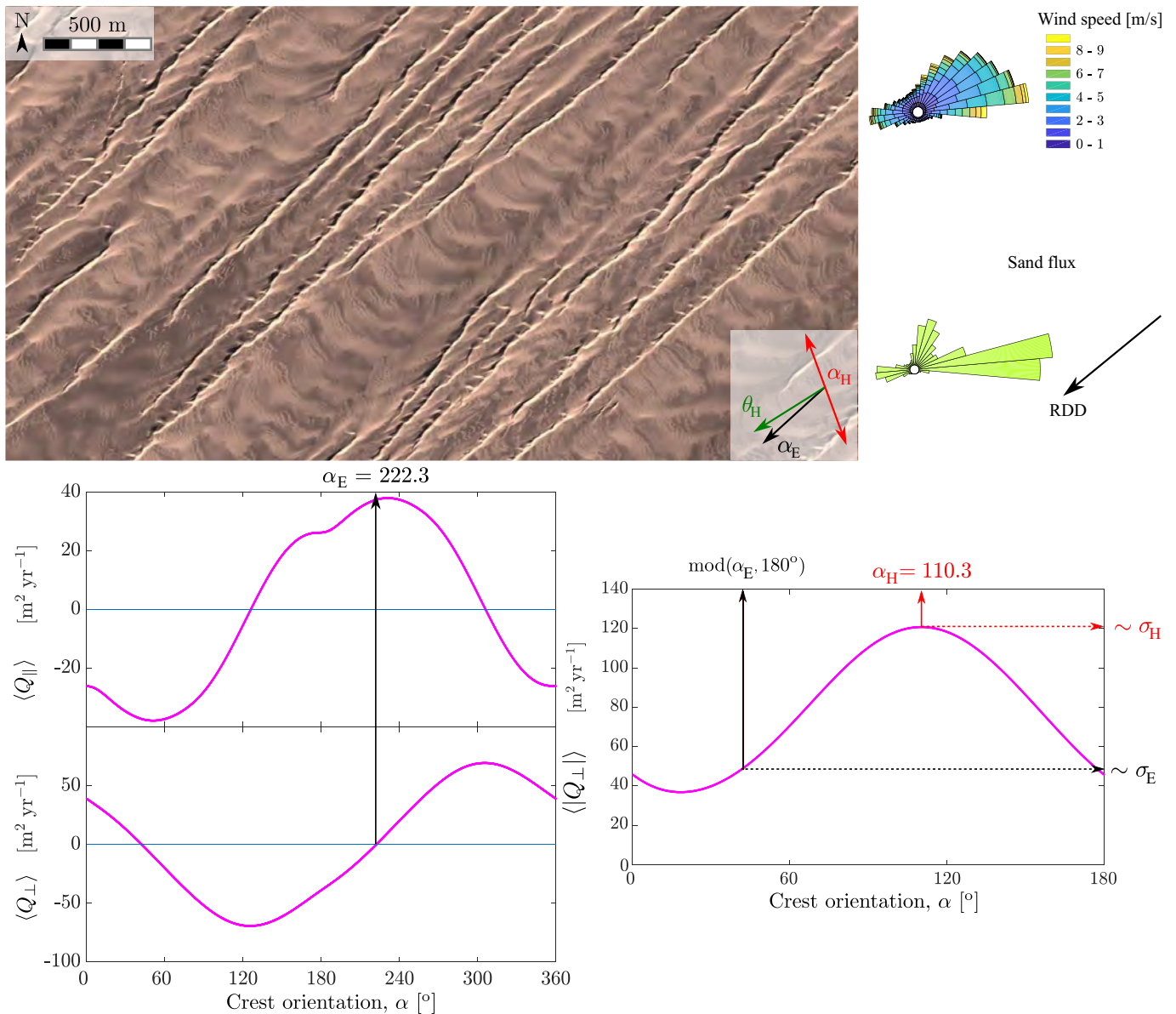


Figure 19: Raked linear dunes on an armored bed of coarse grains in Kumtagh desert in China (40.3°N, 92.7°E). See caption of fig. 12 for an explanation of parameters and plots.

5.3 Determining dune orientation relative to sand transport direction from morphology

The preceding discussion considers classification of dunes and characterization of formative mechanisms where wind measurements or sand transport measurements are available. As discussed in Section 2.3.2, however, some dynamic properties can be inferred even without such measurements. The long-crested examples in Figures 12-16, 19, and 21 can be used to test the approach of inferring dune orientation relative to the net transport direction based solely on morphology, as proposed in Table 2.3.2. In some of these examples dune morphology/asymmetry is unclear unless the image is rotated, so that the illumination comes from the top of the image. Except for Figure 16—in which the resolution is insufficient to determine the cross-sectional asymmetry of the main dunes and superimposed dunes—the examples are consistent with Table 2.3.2.

The dunes in Figures 12 and 21 are known from wind measurements to be transverse dunes, and in both cases this interpretation is demonstrable from morphology alone. The main dunes are asymmetrical, and the superimposed lee-side spurs are approximately symmetrical in cross-section (the spurs lack a steep lee side and gentle stoss side),

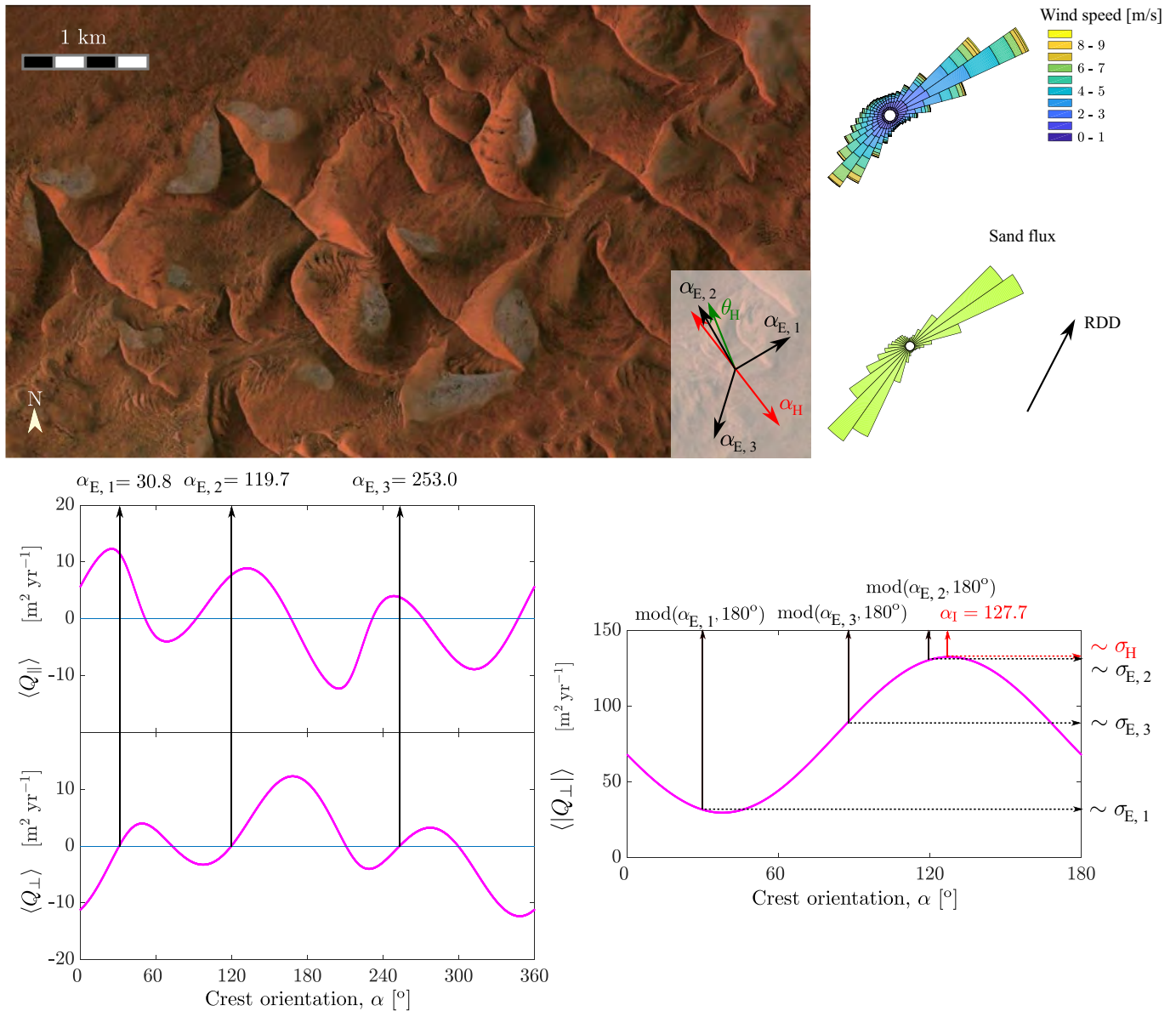


Figure 20: Star dunes on a partially starved bed in the Namib desert in Namibia (26.9°S, 15.9°E). See caption of fig. 12 for an explanation of parameters and plots.

942 demonstrating a lack of systematic migration and net transport parallel to the crests of the main dunes.
 943 The dunes in Figure 13 are oblique to the resultant transport direction, and this can also be demonstrated from
 944 morphology. The main dunes have cross-sectional asymmetry indicating across-crest transport toward the southeast.
 945 Most of the dunes in the figure lack sufficient superimposed features to determine their migration direction, but
 946 the superimposed leeside spurs in the lower right corner of the image have slipfaces that preferentially dip in a
 947 direction parallel to the crests of the main dunes. This combination of across-crest transport and along-crest transport
 948 demonstrates obliquity of the main dunes, at least in this corner of the image.
 949 Wind data show that the dunes in Figures 14, 15, and 19 are longitudinal dunes, which also can be demonstrated
 950 from morphology alone. The main dunes are relatively symmetrical in cross-section, and the superimposed dunes (Figs.
 951 14 and 19) or sinuosities (Fig. 15), have consistent asymmetric cross-sections, thereby demonstrating systematic
 952 transport in an along-crest direction over the main dunes. The steeper lee sides of the superimposed features indicate
 953 net transport parallel to the crests of the main dunes toward the east southeast in Figure 14, and toward the west
 954 southwest in Figure 15. In Figure 19, the superimposed features resemble half-barchans migrating southwest along
 955 the lengths of the main dunes [126].

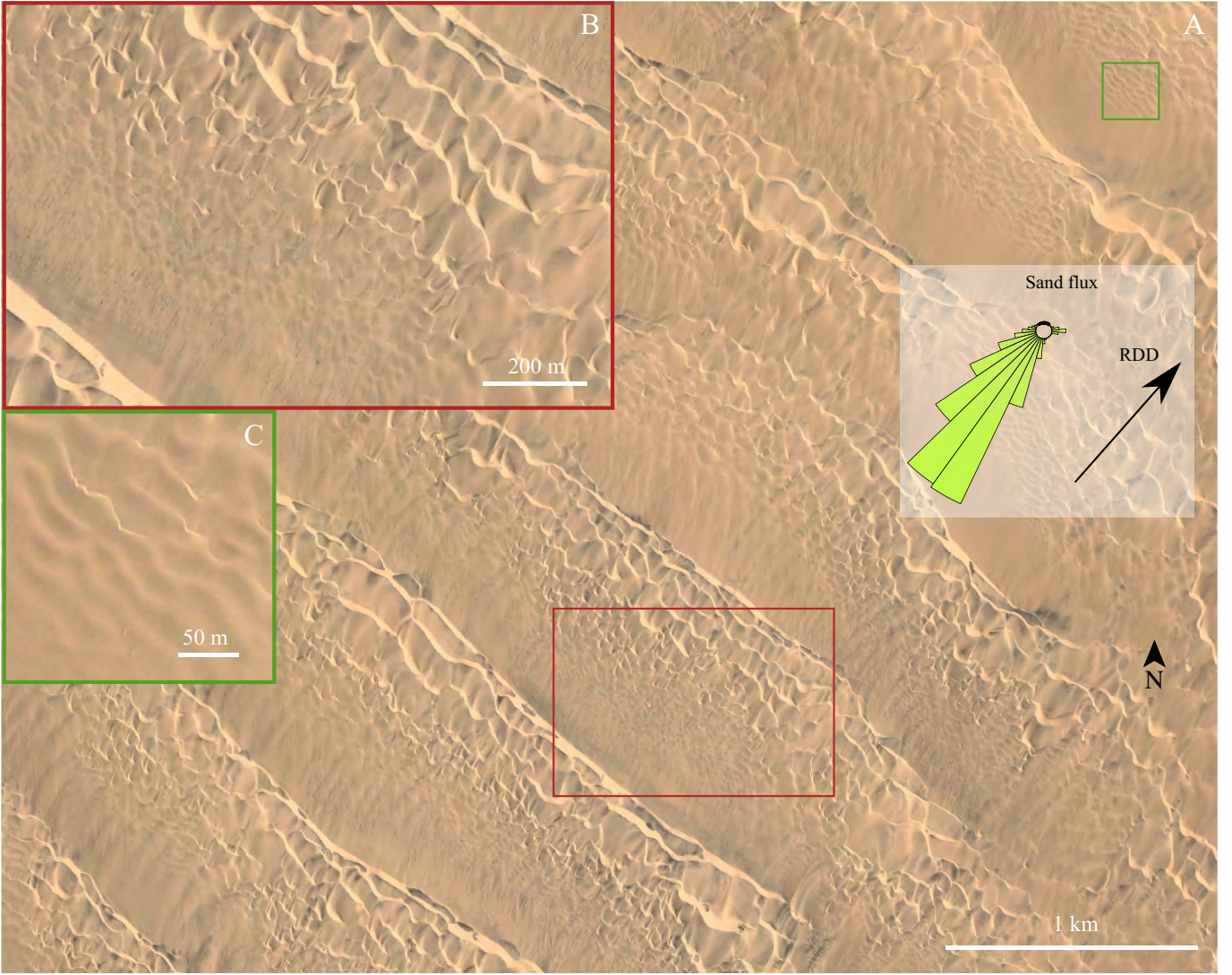


Figure 21: Illustration of the rough case with terrestrial dunes ($\mathcal{R}_r \simeq 80$). Photographs of a coastal dune field in southwest Angola (16.4°S , 11.9°E). Local winds are approximately unidirectional (see flux rose), and the dunes, with fairly straight crests perpendicular to the resultant drift direction (RDD), can be denominated as ‘linear transverse’. Sand availability is high (no cohesive interdune). The dunes form at the elementary scale $\lambda_{\min} \simeq 20$ m (panel c, green frame), interact and coarsen in the course of their migration downwind (panel b, red frame) and the large-scale dune pattern has a wavelength $\lambda_{\max} \simeq 850$ m (panel a). Credit: Maxar Technologie (2022).

956 In summary, the migration directions superimposed topographic features relative to the main dunes can be used to
 957 determine whether the main dunes are transverse, oblique, or longitudinal and similarly to constrain the direction of
 958 net sand transport relative to the dunes. These inferences are qualitative and do not necessarily match the quantitative
 959 definition of Hunter et al. [38]. Nevertheless, this approach may be useful in interpretations of dunes where wind data
 960 or measurements of sand transport or dune migration are not available. This is often the case for other planets and
 961 is always the case for interpreting deposits of dunes in the stratigraphic record, the purpose for which this approach
 962 was initially developed (as discussed in Section 2.3.2).

963 5.4 Expected dune sizes in the solar system from fluid and sediment-transport me- 964 chanics

965 As environmental and boundary conditions significantly vary from one planetary body to another, dunes on their
 966 surfaces sample the different formation regimes.

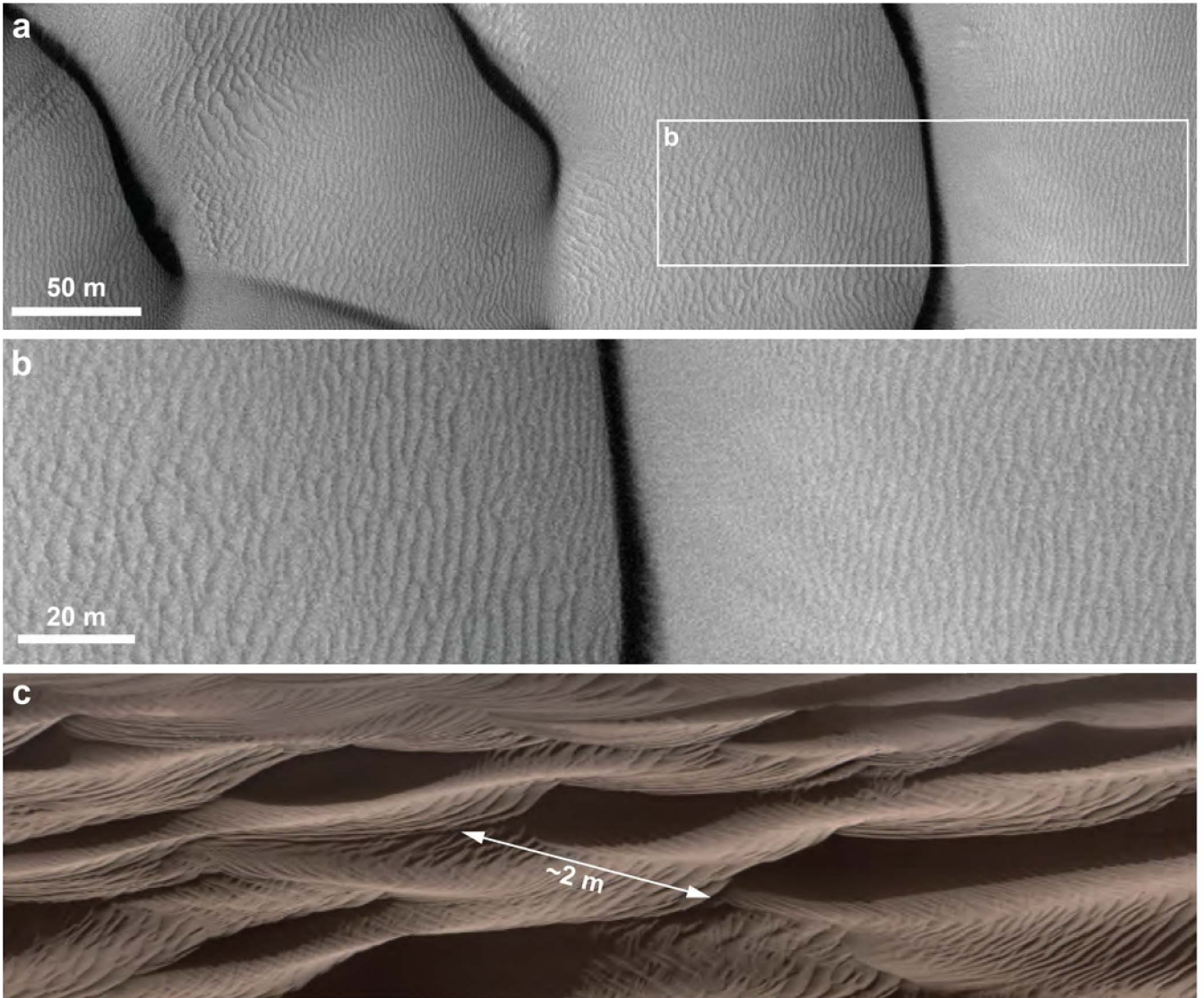


Figure 22: Illustration of dunes forming under aerodynamically smooth conditions, showing large martian ripples and dunes ($\mathcal{R}_r \simeq 1$). Dunes in Lyot (a,b) and Gale (c) craters. Martian bedforms display distinct scales. Dunes with hundred-meter-scale wavelengths (a) are mantled with meter-scale ripples (b). No bedforms with wavelengths between those two scales are observed. A closer look at a dune’s surface with the Curiosity rover (here on ‘Namib dune’ within the Bagnold Dune Field) reveals smaller decimeter-scale bedforms, construed as impact ripples migrating on top of large ripples. Credit: (a-b), HiRISE image ESP_017605_2295; (c), NASA/JPL-Caltech/MSSS/Thomas Appere.

967 On Earth, typical grain sizes are about $200 \mu\text{m}$, such that the thickness of the saltation layer is of the order of
 968 $20 d \simeq 4 \text{ mm}$ [182]. The height of aeolian impact ripples that mantle the surface of sand dunes is of similar, millimeter-
 969 scale. For an aerodynamic roughness, r , of that scale, and a typical shear velocity of the order of 0.4 m/s , one finds
 970 $\mathcal{R}_r \simeq 100$ (with $\nu = 1.5 \times 10^{-5} \text{ m}^2/\text{s}$), indicating that terrestrial saltation occurs in the rough regime. A smooth regime
 971 could possibly be observed for smaller grains and weak winds just above the transport threshold, for example. The
 972 typical value of the density ratio on Earth is $s \simeq 2200$ for quartz grains ($\rho_s \simeq 2650 \text{ kg/m}^3$) and air ($\rho_f \simeq 1.2 \text{ kg/m}^3$),
 973 which justifies the use of the transport law given in Eq. 2. The typical saturation length measured on Earth is of the
 974 order of 1 m , which yields a minimum dune wavelength in the order of 10 m , consistent with observations in nature
 975 [105, 183] and in a landscape scale field experiment [131].

976 Fig. 21 illustrates dunes forming in the aerodynamically rough regime, showing a coastal dune field in southwest
 977 Angola. The environment there closely matches the assumptions behind our classification, with a fairly unidirectional

wind blowing over a loose sand bed. Incipient dunes form with a scale $\lambda_{\min} \simeq 20$ m. Then, dunes coarsen as evidenced by a range of dune sizes, and display a consistent large-scale wavelength of the dune pattern, $\lambda_{\max} \simeq 850$ m. There, using ERA5-Land reanalysis data based on global atmospheric models ([167]), the thickness of ABL, Λ , is directly calculated as 300 m, or as 2 km using the same proxy as in [144].

For the low-pressure (*i.e.*, low-density, $\rho_f \simeq 0.02$ kg/m³) atmosphere of Mars, atmospheric kinematic viscosity is larger than on Earth, $\nu \simeq 10^{-3}$ m²/s and although uncertain, typical shear velocities are thought to be of the order of $u_* \simeq 1$ m/s [12, 137, 184] so that $\mathcal{R}_r \sim 1$ for a millimeter-scale aerodynamic roughness. As a result, saltation occurs under an aerodynamically smooth regime on Mars, and two distinct ranges of dune sizes are expected if the ratio between the saturation length and the viscous length is smaller than about 1000. Saturation length has not been measured on Mars to date. With a larger density ratio than on Earth ($s \simeq 1.5 \times 10^5$ for basalt grains, $\rho_s \simeq 3000$ kg/m³), some models also predict a larger saturation length on Mars than on Earth [185]. Two distinct scales of bedforms larger than impact ripples are observed in monodisperse sand on Mars, as shown in fig. 22. Meter-scale ripples (distinct from smaller, decimeter-scale impact ripples) migrate on top of large hundred-meter-scale dunes [12, 102, 142, 186]. In that sense, large martian ripples and dunes are analogous to ripples superimposed on dunes in a subaqueous environment [12, 187, 137], where a smooth regime is expected as well. An alternative interpretation was proposed for these meter-scale ripples, under which they would simply be large impact ripples, growing to meter-scale wavelengths from their initial, decimeter-scale wavelengths [89]. However, that model cannot explain the absence of bedforms with wavelengths in the $\sim 20 - 80$ cm range in relatively well sorted sand [142]. Furthermore, the observed gap in bedform wavelengths between a large ripples and Mars' smallest dunes ($\lambda \sim 80$ m) matches the predictions of the Hanratty anomaly [139].

Dune formation regimes, and thus dune scales, on other planetary bodies remain highly uncertain (owing, *e.g.*, to large uncertainties in the materials that make up dune sand). On Titan, grains have been proposed to be made of complex hydrocarbons, water ice, or a combination of both; a density of $\rho_s \simeq 1000$ kg/m³ was proposed for porous organics that would form from a photochemical haze [188]. Although many uncertainties remain, a value of $s \simeq 200$ seems reasonable given current knowledge. For $u_* \simeq 0.1$ m/s, $\nu \simeq 10^{-6}$ m²/s and $r \simeq 10^{-3}$ m [189], one finds that $\mathcal{R}_r \simeq 100$, indicating that dunes on Titan would form under an aerodynamically rough regime, similar to terrestrial dunes. Linear dunes on Titan are observed around the equatorial region with a kilometer-scale, similar in scale to large linear dunes found on Earth [4]. Finally, on Venus, we consider the case of $u_* \simeq 5 \cdot 10^{-2}$ m/s, $\nu \simeq 5 \cdot 10^{-7}$ m²/s and $r \simeq 10^{-4}$ m [190], such that $\mathcal{R}_r \simeq 10$, *i.e.*, a value near the upper bound of the smooth regime.

6 Conclusion

Dunes are common landforms throughout the solar system that can teach us about past and current environmental conditions on Earth as well as on a variety of planetary bodies including Venus, Mars, Titan, and Pluto. Building a fundamental understanding of dune morphodynamics as well as predictive capabilities and mitigation strategies will require scientific collaboration across traditional field boundaries, and thus, a common language to describe sand dunes. Here, we synthesized existing terminology and distilled it into three independent but complementary dune classification schemes. First, we proposed a unifying description of dunes based on their morphology. This classification builds on a long legacy of existing terminology and solely requires observations of dune shapes as can be gathered from the ground, or from aerial or satellite imagery. Terminology can be further refined within that classification scheme when information about winds and dune migration is available. Second, we synthesized state-of-the-art models for dune morphodynamics, tying specific morphologic types with their formative dynamics as selected by wind regimes and boundary conditions. Third, we presented a classification of the various fluid dynamics regimes under which dunes may form, leading to different controls on initial and equilibrium dune sizes. This last scheme was developed for transverse dunes only, and encapsulates the complexity of variable boundary and environmental conditions that may be found across planetary bodies of the solar system. Together, these three classification schemes allow for variable levels of descriptive detail depending on available data, and encompass dune shape, dynamics, and scale as linked to their environmental conditions. Importantly, they offer a complete and unified framework, anchored in the mechanics of dune formation, for future studies to describe dunes on Earth and other planets.

1027 7 Appendix – Concepts, models and methods

1028 7.1 Dune size and timescale of wind-regime integration

1029 We define the dune length scale as the square root of the dune cross-section, $\sqrt{HL/2}$, where H and L are the dune
1030 height and length. This approximates the dune profile to a triangle. To this dune size is associated a time scale
1031 $HL/(2Q)$ called the *dune turnover time*, which is the characteristic time to completely reshape the dune with a sand
1032 flux of norm Q [32]. Q is a volumetric flux per unit width (*i.e.*, it has the dimension of a length squared per time) and
1033 takes into account the dune compactness. This timescale sets the time duration over which one should consider the
1034 wind regime so that the dune integrates the whole complexity of the wind regime and is in a dynamical equilibrium
1035 over the considered period of time. A complex wind regime consists in a succession of wind events with different
1036 directions and strengths. We consider a flow sequence of duration τ_1 between significant changes of sand flux direction.
1037 In a ideal scenario, the various flow sequences repeat periodically, defining a flow regime. The dunes integrate the
1038 wind regime if none of individual flow sequence completely reshapes the dune [62]. On the other hand one can expect
1039 that the dune has reached a dynamical equilibrium if the full considered period of time, τ_{tot} , is long enough to shape
1040 the dune. Thus, the size of the dune sets the period of time to consider (and vice versa) such that

$$\int_{\tau_1} Q(t)dt \ll HL/2 \leq \int_{\tau_{\text{tot}}} Q(t)dt. \quad (1)$$

1041 The boundary conditions should also not significantly change over the considered period of time. Note that Equation
1042 1 does not take into account the direction of the sand flux because two winds with opposite directions contribute
1043 to build dunes. In principle, only the component of fluxes in line with the cross-section could be considered. In
1044 practice, it is difficult to obtain wind data for longer than the time required to build very large dunes. Shorter periods
1045 may be appropriate, but the duration should not be less than one period of the wind regime. Considering a shorter
1046 time period also enables to study whether the wind regime has changed [191, 192, 146]. The integral of the sand
1047 flux over a period of time T , $\int_T Q(t)dt$, where T is the duration of the wind regime, also defines the characteristic
1048 minimum cross-sectional area of dunes that integrates the wind regime. This concept of turnover time is derived from
1049 dimensional analysis and has never been systematically studied. It only sets a time scale as in principle, a dune can
1050 be fully reshaped without having to move all the sand it contains [99].

1051 7.2 Sediment transport

1052 7.2.1 Characterization of aeolian sediment transport

1053 Sediment transport is coupled to fluid flow in the transport layer. It is characterized by a threshold (minimum)
1054 shear velocity (or basal shear stress), a saturated (maximum) sediment flux in steady state, the saturated flux, and
1055 a saturation length [193]. The saturation length corresponds to the spatial lag of the response of flux to a change
1056 in transport conditions, *i.e.*, how far downwind from a change in transport conditions (such as an increase of wind
1057 velocity) the sediment flux equilibrates with the new conditions. Three main parameters control sediment transport
1058 and saltation in particular.

1059 First, the particle-to-fluid density ratio, $s = \rho_s/\rho_f$, encapsulates the reduced weight of particles in the fluid. Second,
1060 a commonly used parameter is the Shields number, $\Theta = u_*^2/[(s-1)gd]$ (where u_* , g , and d are the wind shear velocity,
1061 the gravitational acceleration and the grains diameter, respectively), which allows for comparisons of the basal shear
1062 stress on grains ($\rho_f u_*^2$) relative to the grains' apparent weight per unit of surface ($\sim (\rho_s - \rho_f)gd$). Third, the Galileo
1063 number, $\mathcal{G} = \sqrt{(s-1)gd^3/\nu}$ (where ν is the fluid viscosity), can be envisioned as the ratio of gravitational and viscous
1064 effects. It is the square root of a Reynolds number in which the length scale is taken as the grain size and the velocity
1065 scale as the Stokes limit of the grains' settling velocity. This parameter is useful to compare different environments
1066 regardless of flow velocity.

1067 The dynamic threshold for sediment transport is defined by the minimum shear stress exerted by the fluid on the
1068 granular bed that can sustain transport in saltation. It is associated to a critical value of the Shields number, Θ_t . Θ_t
1069 varies with grain and flow properties, and is typically expressed as a function of grain size and apparent density or
1070 equivalent dimensionless parameters such as \mathcal{G} and s [194, 195, 196, 184, 197].

1071 The saturated sand flux at steady state, Q_{sat} , is proportional to the difference between the actual Shields number
1072 and its critical value. Such transport laws (*e.g.*, Eq. 2) are valid for saltation transport over a loose sand bed, when
1073 $s \gtrsim 100$. Under such conditions, the velocity of grains in saltation is independent of wind shear velocity, u_* (fig. 13
1074 in [198]) due to coupling between the wind flow and sediment transport. If transport laws are fairly well established

1075 and calibrated in steady and homogeneous conditions [198, 197, 199, 32, 200, 201, 202, 203, 204], threshold Shields
 1076 numbers are much less constrained under extraterrestrial conditions, including on Mars [184, 205].

1077 Some experimental data are available for the saturation length, L_{sat} , under terrestrial saltation (with L_{sat} typically
 1078 between 0.5 and 1 m for sand grains in the range [100, 200 μm], [32, 206, 130, 132, 131]) as well as under subaqueous
 1079 suspension [207], but significant knowledge gaps remain [185, 208, 209, 210] rendering predictions for saltation on
 1080 Mars or Titan challenging. These knowledge gaps are particularly significant given that L_{sat} is a key parameter in
 1081 dune-instability analysis [193, 141]. As the minimum length required to generate an effective coupling between the
 1082 wind flow and sand transport, L_{sat} sets the characteristic wavelength of incipient dunes in most cases (Fig. 8).

1083 7.2.2 Transport law – Saturated sand flux and onset of transport

1084 There are many transport laws, which are more or less phenomenological, each with their own validity regime [182].
 1085 Here, we compute the saturated sand flux, Q_{sat} , (*i.e.*, maximum sand flux over a flat sand bed) using the relationship
 1086 proposed in [211, 135], such that

$$Q_{\text{sat}} = \begin{cases} a_q \frac{\rho_f}{\rho_s g} u_t (u_*^2 - u_t^2) \equiv a_q \frac{s-1}{s} du_t (\Theta_*^2 - \Theta_t^2) & \text{if } u_* > u_t, \\ 0 & \text{otherwise,} \end{cases} \quad (2)$$

1087 where the dimensionless prefactor, $a_q \approx 8.33$, was calibrated in [135] and takes into account the dune compactness
 1088 (~ 0.6) so that sand fluxes can be used directly for dunes dynamics. Equation 2 is a volumetric flux per unit width,
 1089 *i.e.*, it has the dimension of length squared per time. In this formula, u_t is the threshold shear velocity for transport,
 1090 which corresponds to the critical value of the Shields number, Θ_t , such that

$$u_t = \sqrt{\Theta_t} \sqrt{\frac{\rho_s - \rho_f}{\rho_f} g d}. \quad (3)$$

1091 We take $\Theta_t^{1/2} = 0.082$ [212], so that u_t accounts for the dynamic (or impact) threshold, *i.e.*, the lowest shear velocity
 1092 that can sustain saltation once it has been initiated [32]. For calculations, we use a constant threshold velocity of
 1093 $u_t = 0.153$ m/s, which corresponds to quartz sand grains of diameter $d \simeq 160$ μm in air on Earth.

1094 The transport law Eq. 2 is valid for transport in saltation over a mobile sand bed for moderate wind velocities
 1095 above transport threshold. In such conditions, sand transport occurs in a few centimeters thick surface layer above the
 1096 bed [203, 213]. In this law, first proposed by Ungar and Haff [201], the amount of transported grains is proportional
 1097 to the difference between the basal shear stress and its threshold value, and the mean velocity of grains corresponds to
 1098 the transport threshold wind velocity. Grains in saltation slow the wind in the transport layer, down to the transport
 1099 threshold value when flux is saturated [135, 182].

1100 7.2.3 Wind speed-up over a dune

1101 The first source of complexity in morphodynamics of dunes relies in the coupling between wind flow, sand transport,
 1102 and dune topography. Positive topography deflects the air flow and the compression of streamlines causes the wind
 1103 speed close to the ground to be stronger over the obstacle than over a flat bed away from any topography, so that the
 1104 sand flux over a dune depends on its shape. Usually defined as the *speed-up*, this is a critical ingredient that couples
 1105 the dune topography and sand transport at all stages of the development of dunes, from the formation of incipient
 1106 bedforms to the dynamics of major dune systems. It is often described in terms of a *fractional speed-up ratio* (or
 1107 *relative speed-up ratio*), δ , which is the relative speed-up between the wind velocity at some elevation above the dune
 1108 profile (u) and the wind velocity at the same elevation above a flat bed away from any topography (u_0) [170]:

$$\delta = \frac{u - u_0}{u_0}. \quad (4)$$

1109 The value of δ varies with position along the dune profile, elevation above the bed, dune aspect ratio and aerodynamic
 1110 roughness [170, 214]. In the field, measurements should be made in the transport layer to be relevant for flux variation
 1111 and, if possible, under transport conditions because transport affects aerodynamic roughness. When measured close
 1112 to the bed for different types of dunes on Earth, δ -values are reported to typically range between 0.4 and 1 for the
 1113 maximum value, or for the one measured at the top of the dune [71, 215, 216, 217, 218]. In general, the wind velocity
 1114 is not expected to be maximum at the top of the dune when measured sufficiently close to the ground, but slightly
 1115 upwind (Section 7.4.1). However, on mature dunes the difference between the maximum wind velocity and wind

1116 velocity at the dune top should be small.

1117

1118 Jackson and Hunt analyzed the theoretical response in 2D of a turbulent air flow in aerodynamically rough regime
 1119 to the perturbation induced by a symmetric hump with small aspect ratio and curvature (Sections 4 & 7.4.1) [170].
 1120 They found that in this case the fractional speed-up, taken at its maximum or at top of the hump, is independent on
 1121 wind velocity, depends only weakly on aerodynamic roughness, and is almost proportional to the hump aspect ratio
 1122 [170] :

$$\delta \simeq \beta \frac{H}{L}, \quad (5)$$

1123 where L and H are the characteristic length and height of the hump, and β a dimensionless coefficient that takes into
 1124 account the aerodynamic roughness, position along the hump profile, and elevation above the bed. For typical values
 1125 of aerodynamic roughness encountered in dunes on Earth, $\beta \simeq 6$ when considering the maximum of shear velocity, u_*
 1126 [170, 141, 219] (Section 7.4.1).

1127

1128 The increase in wind shear velocity leads to an increase in saturated sand flux up the dune slope. Equivalently to
 1129 the relative speed-up ratio, one can define a relative flux-up ratio [41, 219], $\gamma = (Q - Q_0)/Q_0$, which can be calculated
 1130 for saturated flux with Equation 2 given a value for δ . Because of transport threshold in Equation 2, γ depends on
 1131 wind velocity even if δ does not as in Equation 5, *e.g.*, γ goes to infinity if unperturbed wind velocity, u_0 , is just below
 1132 u_t [64].

1133 The variation of wind shear velocity along the dune profile, which leads to variation of sand flux, is a key parameter
 1134 to understand dune growth and dynamics (Section 4 & 7.4.1). Its dependency on the dune streamwise aspect ratio,
 1135 H/L , which has been observed in the field for reversing dunes [220], also has a strong impact on dune dynamics in
 1136 multidirectional wind regimes (Sections 5.2 & 7.3.3). The 3D shape of the dune and the orientation of wind relative to
 1137 the dune (*e.g.*, the wind may not blow perpendicularly to crest line) can also deflect the sand flux direction above the
 1138 dune [221, 222, 71, 223, 217, 224] by combined effects of wind streamline deflection and of gravity, because the dune
 1139 topography gradient has components in both parallel and normal to (non-deflected) wind. The sand flux deflection by
 1140 topography explains the 3D shapes of a barchan [225, 122] or the transverse instability [124, 125] in a unidirectional
 1141 flow. Here, we do not take into account the sand flux deflection in calculations but only the primary effect of dune
 1142 topography, which is wind-speed-up and its dependency on apparent dune aspect ratio.

1143 7.3 Sand flux from wind data

1144 7.3.1 Sand flux on a flat sand bed from wind data

1145 We compute sand fluxes using the surface wind data from the ECMWF ERA5-Land reanalysis [166, 167]. This global
 1146 weather forecasting model based on data assimilation aims to include all available measurements from weather stations,
 1147 radiosondes, ships, and satellites. It provides numerical extrapolation of many parameters from the beginning of 1979
 1148 up to now (2023) with a horizontal spatial resolution of $0.1^\circ \times 0.1^\circ$ (about 11 km at the equator) and a time resolution
 1149 of 1 hour. We denote t_i , $i \in [1; N]$ the different times, which are regularly spaced. We extract from data the wind
 1150 direction, θ , relative to the east and wind speed, u , at a height $z = 10$ m and calculate the shear velocity, u_* , over a
 1151 flat sand bed using the law of the wall:

$$u_*(t_i) = u(z, t_i) \frac{\kappa}{\ln(z/z_0)}, \quad (6)$$

1152 where κ is the von-Kármán constant ($\kappa = 0.41$) and z_0 the aerodynamic roughness length scale, which is modified by
 1153 grains in saltation in transport conditions. We take $z_0 = 1$ mm for all calculations on Earth.

1154 We define the mean shear velocity $\langle u_* \rangle$ as the shear velocity averaged over the time periods when transport is
 1155 occurring:

$$\langle u_* \rangle = \frac{1}{\sum_{i=1}^N H_u} \sum_{i=1}^N H_u u_* \quad \text{where} \quad H_u = \begin{cases} 1 & \text{when } u_* > u_t, \\ 0 & \text{else.} \end{cases} \quad (7)$$

1156 Using the calculated shear velocity, $u_*(t_i)$, and extracted wind direction, $\theta(t_i)$, we compute the saturated sand flux
 1157 over a flat sand bed, \vec{Q}_0 , using the transport law in Eq. 2:

$$\vec{Q}_0(t_i) = \begin{pmatrix} Q_{\text{sat}}(t_i) \cos[\theta(t_i)] \\ Q_{\text{sat}}(t_i) \sin[\theta(t_i)] \end{pmatrix}, \quad (8)$$

1158 which we integrate over the entire time period to calculate the resultant sand flux:

$$\langle \vec{Q}_0 \rangle = \frac{1}{N} \sum_{i=1}^N \vec{Q}_0. \quad (9)$$

1159 7.3.2 Sand flux over dunes from wind data

1160 To evaluate the characteristic growth, migration, and elongation rates of dunes in the field, we need to calculate the
 1161 characteristic wind shear velocity and corresponding sand flux over dunes. This characteristic wind shear velocity
 1162 takes into account the wind speed-up and is typically the maximum value, or the one at the top of the dune, which
 1163 we take to be the same. Wind fractional speed-up ratio, δ (Eq. 4), depends on streamwise dune aspect ratio (Eq. 5).
 1164 Considering a symmetric linear dune of uniform and steady aspect ratio H/L in cross section and orientation α , a
 1165 wind of direction θ experiences an apparent dune aspect-ratio $H |\sin(\theta - \alpha)| / L$, so that the characteristic wind shear
 1166 velocity, u_c , depends on the angle between wind direction and dune orientation:

$$u_c(\alpha) = u_* (1 + \delta |\sin(\theta - \alpha)|), \quad \text{where} \quad \delta = \beta \frac{H}{L}. \quad (10)$$

1167 In Eq. 10, u_* is the shear velocity over a flat sand bed (Eq. 6). For the calculations in field examples of Section 5.2,
 1168 we use $\delta = 0.8$. Future improvements could include unsteady aspect ratios, which are probably important to better
 1169 model reversing dunes, but here we only take into account the effect of dune orientation with respect to winds.

1170 The corresponding characteristic sand flux over the dune, which we assume to be saturated, can be decomposed
 1171 into crest-parallel and crest-normal components of transport:

$$\vec{Q}_c(\alpha) = \begin{pmatrix} Q_{\parallel}(\alpha) \\ Q_{\perp}(\alpha) \end{pmatrix} = \begin{pmatrix} Q_{\text{sat}}(u_c) \cos(\theta - \alpha) \\ Q_{\text{sat}}(u_c) \sin(\theta - \alpha) \end{pmatrix}, \quad (11)$$

1172 where Q_{sat} is calculated with Eq. 2 using u_c (Eq. 10). The effect of the topography on the saturated sand flux due to
 1173 the wind speed-up is not directly integrated into the sand flux but into wind shear velocity to better account for the
 1174 transport threshold. Sand can be transported at the dune top while shear velocity is below the threshold for transport
 1175 where the bed is flat. However, our analysis assumes that wind velocity is above threshold where the bed is flat. \vec{Q}_c ,
 1176 is the characteristic sand flux over the dune, which can be assimilated to the maximum value of the sand flux or the
 1177 value at the dune top. In our analysis, we assume that the sand flux varies in line with the topography along the
 1178 dune profile (and wind boundary layer detaches in the lee side), so that the averaged sand flux over a streamwise dune
 1179 section is assumed to be proportional to Q_c , and the characteristic divergence of the sand flux is Q_c/L , where L is the
 1180 streamwise length of the dune. This is not valid if transport occurs only at the dune top.

1181 Then, the time-averaged components of these fluxes allow us to predict the orientation of dunes according to the
 1182 prevailing growth mechanism.

1183 7.3.3 Orientation of dunes in multidirectional flow regimes depending on the prevailing growth mech- 1184 anism

1185 **Growth rate in height** The growth rate in height of a dune is directly related to the sand deposition rate at
 1186 the crest and erosion rate at the trough, which correspond to divergences of the sand flux at the crest and the
 1187 trough, respectively. For long straight dunes with an avalanche face or large aspect ratio that are much larger than
 1188 the minimum dune size (Section 4), we assume that the sand flux typically varies over a length proportional to the
 1189 streamwise dune length, L , between a maximum value Q_{max} upwind of the crest and zero downwind of the crest,
 1190 because the wind boundary layer detaches in the lee side. Thus, the characteristic growth rate is:

$$\sigma = \frac{Q_{\text{max}}}{HL}, \quad (12)$$

1191 where H is the dune height. Dunes that exhibit smaller streamwise lengths have larger growth rate.

1192 When growth in height is the prevailing growth mechanism, like in the bed instability mode, the orientation of the
 1193 dune crestline is such that the growth rate is maximum. In a multidirectional wind regime, a dune should then have an
 1194 orientation α_H that minimizes the different streamwise lengths, so that it experiences the *Maximum Gross Bedform-
 1195 Normal Transport* (MGBNT) [40], which we consider to be equivalent to the maximum of $\langle |Q_{\perp}| \rangle$. The absolute value
 1196 reflects that transport in both directions across the crest contribute to dune growth [40, 41]. Using signed values of
 1197 transport subtracts transport in opposing directions, even though they both contribute to dune growth.

1198 MGBNT yields good agreement with experiments on wind ripples [40], centimetric and decimetric sand bedforms
 1199 under water [62, 122, 41], numerical simulations [119, 63, 64], and field measurements for bimodal flow regimes [165, 131]
 1200 in full mobilized bed conditions.

1201 **Elongation without migration** The direction of elongation is the direction of sand net transport as experienced
 1202 by the dune. The direction α_F of an elongating dune correspond to a null perpendicular to crest net transport ($\langle Q_{\perp} \rangle = 0$)
 1203 and a positive along crest net transport ($\langle Q_{\parallel} \rangle > 0$). This is the predicted orientation for dunes on a
 1204 starved bed if they do not migrate. Prediction of several directions of elongation within a same wind regime is made
 1205 possible by taking into account the dependency of wind speed-up on dune orientation.

1206 The predicted orientation for elongating dunes is in good agreement with experiments on centimetric sand bedforms
 1207 under water [41], numerical simulations [64] and field measurements [226] for bimodal flow regimes. This rule also
 1208 correctly predicts the multiple directions of elongation observed for star dunes on a starved bed in numerical simulations
 1209 when the flow regime is multidirectional with a null RDP [90].

1210 7.4 Wind flow over incipient free dunes

1211 7.4.1 Turbulent flow model

1212 Wind flow over a flat bottom or topography with a small aspect ratio can be divided in different regions above the
 1213 bed (or layers) in which the transfer of fluid momentum is dominated by different processes. Sediment transport
 1214 takes place in an *inner layer* where the flow is turbulent (high Reynolds number) and is characterized by velocity
 1215 fluctuations and intermittent flow structures that transport the fluid momentum to the bed, where the flow is slowed
 1216 by wall friction. In the outer layer, the fluid is comparable to a perfect fluid. In a sublayer, just above the bed, viscous
 1217 effects are dominant.

1218 Turbulent stress in the inner layer can be described by means of an *eddy viscosity* characterized by a *mixing length*.
 1219 There is no characteristic length scale intrinsic to turbulence and the mixing length is a local quantity, usually similar
 1220 to the distance to the obstacle that is causing the momentum transfer. Under this model for a stationary outer flow,
 1221 the time-averaged velocity profile, $u(z)$, inside the inner layer above a flat wall is known as the law of the wall:

$$u(z) = \frac{u_*}{\kappa} \ln \left(\frac{z}{z_0} \right), \quad (13)$$

1222 where u_* is the shear velocity, $\kappa \simeq 0.41$ the von Kármán constant, z the distance from the wall, and z_0 is the distance
 1223 at which the velocity as given by Eq. (13) goes to zero and relates to *aerodynamic roughness*. The shear velocity is
 1224 defined from the basal shear stress τ , such that $\tau = \rho_f u_*^2$, where ρ_f is the fluid density. The law of the wall is well
 1225 supported by experimental data [227, 228, 133]. The length z_0 is either set by the characteristic roughness length
 1226 scale of the wall, r , or by the thickness of the viscous sublayer when it is larger than roughness. The characteristic
 1227 thickness of the viscous sublayer is ν/u_* , where ν is the fluid kinematic viscosity. The roughness Reynolds number,
 1228 $\mathcal{R}_r = ru_*/\nu$, compares that two length scales. When $\mathcal{R}_r > 100$, flow is aerodynamically rough ; conversely, when
 1229 \mathcal{R}_r is smaller than a few units, flow is aerodynamically smooth, and the thickness of the viscous sublayer sets the
 1230 characteristic length z_0 [133].

1231
 1232 For a wavy bed, the thicknesses of the boundary layers are modulated by topography. They are thinner where
 1233 flow velocity spatially increases, and thicker where flow velocity decreases. In order to study the formation of incipient
 1234 dunes from a flat sand bed, we consider the linear response of the flow to a small sinusoidal perturbation of the bed
 1235 topography of wavelength λ and low amplitude h_0 ($h_0/\lambda \ll 1$), such that

$$h(x) = h_0 \sin(2\pi x/\lambda). \quad (14)$$

1236 In this case, the perturbation of the basal shear stress is proportional to the aspect ratio of the topographic pertur-
 1237 bation, and can be expressed as the sum of two components: one that is in phase with topography, with a weight
 1238 denoted \mathcal{A} , and one that is in phase quadrature, with a weight denoted \mathcal{B} , *i.e.*,

$$\tau(x) = \tau_0 \left[1 + \frac{2\pi h_0}{\lambda} (\mathcal{A} \sin(2\pi x/\lambda) + \mathcal{B} \cos(2\pi x/\lambda)) \right], \quad (15)$$

1239 where τ_0 is the unperturbed (flat bed) shear stress. In principle, coefficient \mathcal{A} accounts for dune migration, whereas
 1240 coefficient \mathcal{B} accounts for dune growth [193, 229, 219]. Within this framework, the ratio of the relative increase in

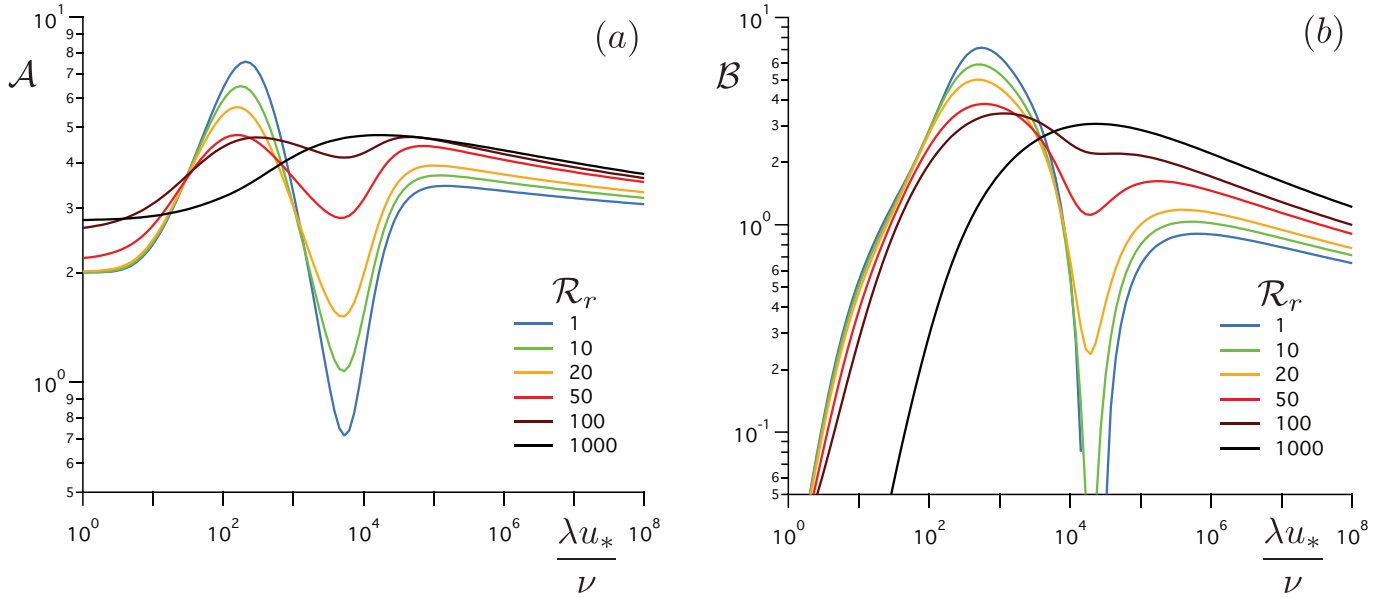


Figure 23: Response of the flow to a sinusoidal perturbation of wavelength λ for different boundary Reynolds numbers \mathcal{R}_r , *i.e.*, r (adapted from [134]). The coefficients \mathcal{A} (a) and \mathcal{B} (b) are the weight coefficients of the perturbation of the basal shear stress for the in-phase and in-phase-quadrature components, respectively. They are plotted as a function of wavelength, normalized by the viscous length, *i.e.*, a bedform-scale Reynolds number.

1241 shear-stress at the tops of the sinusoidal bed equals $\mathcal{A}(2\pi h_0/\lambda)$. In the limit of small perturbations, the relative
 1242 speed-up of shear velocity is half of that (because $\tau \propto u_*^2$) and the fractional speed-up ratio is $\delta = (\mathcal{A}/2)(2\pi h_0/\lambda)$ at
 1243 the tops of the bed. However, wind shear velocity and shear stress are not maximum at the tops of the bed but are
 1244 phase-shifted. The phase shift, $\phi = \arctan(\mathcal{B}/\mathcal{A})$, is positive if τ is maximum upwind of the top of the bed. There,
 1245 the maximum fractional speed-up ratio of shear velocity is $\delta = (\sqrt{\mathcal{A}^2 + \mathcal{B}^2}/2)(2\pi h_0/\lambda)$.
 1246

1247 Under the above framework, the flow response to a small perturbation was calculated by Jackson and Hunt [170]
 1248 in the rough case (constant z_0 , $\mathcal{R}_r \rightarrow \infty$). In that limit, \mathcal{A} and \mathcal{B} weakly (logarithmically) vary with the wavelength
 1249 of the bed perturbation (Fig. 23). More importantly, in rough conditions, \mathcal{B} and the phase shift are always positive,
 1250 favoring the development of the dune instability at all wavelengths [193, 230].

1251 This positive shift can be explained as follows. In the outer region of the flow, well above the bed, the flow
 1252 behaves as a potential flow, in which inertia and pressure gradient balance each other. Therefore, the flow is in
 1253 phase with the topography, such that wind velocity is maximum over topographic highs and minimum over troughs.
 1254 According to Bernoulli's principle, pressure is in turn minimal over topographic highs and maximum over troughs.
 1255 Thus, along a streamline, the pressure gradient is minimum (and negative) where the velocity increases the most,
 1256 above the maximum slope. It is in phase quadrature with topography. In contrast, flow velocity is reduced due
 1257 to friction in the inner turbulent layer, and fluid inertia is therefore reduced. In the inner layer, the flow is driven
 1258 by the transverse turbulent transport of momentum from the outer region to the inner layer (in phase) and by the
 1259 longitudinal (streamwise) pressure gradient inherited from the outer flow and that is maintained in the inner layer
 1260 (in phase quadrature). Therefore, fluid velocity and shear stress peak upwind of the maximum of topography. This
 1261 positive offset was measured in the field on dunes with small aspect ratios [218, 131].
 1262

1263 Under aerodynamically smooth conditions, both the turbulent and viscous boundary layers are modulated by
 1264 topography and coupled. Regarding the turbulent mixing, the usual correction relative to the rough conditions over a
 1265 flat bed is modeled as a decrease in both the mixing length and the turbulent viscosity near the bed, with a characteristic
 1266 length scale proportional to the thickness of the viscous boundary layer [231]. Such corrections are important when
 1267 considering, *e.g.*, the development of dissolution and melting patterns (Section 7.4.2). More importantly for dunes, the
 1268 shear stress on the bed is controlled by the viscous boundary layer. A larger shear stress goes with a thinner viscous
 1269 boundary layer, which thins out when the (negative) pressure gradient decreases. To model unusual experimental
 1270 measurements of the shear stress on a wavy bed in a water channel [232, 233], Hanratty assumed a spatial lag between
 the effective pressure gradient inside the viscous boundary layer and the outer pressure gradient proportional to the

1271 characteristic thickness of the viscous boundary layer [138, 234]. Under this assumption, it follows that the factors
 1272 \mathcal{A} and \mathcal{B} in particular abruptly decrease for a range of wavelengths (fig. 23). This sharp transition can explain the
 1273 dissolution pattern [235, 134]. Although not directly evidenced by experimental data, this model interestingly predicts
 1274 a negative phase shift between shear stress and topography for a range of wavelengths set by the characteristic thickness
 1275 of the viscous boundary layer (Fig. 23). If this prediction is correct, dunes with these wavelengths can not develop
 1276 [141, 137]. This prediction is consistent with the general absence of dunes with wavelengths from a few to a few
 1277 tens of meters on Mars, where the kinematic viscosity of the atmosphere is larger than on Earth [187, 102, 137].
 1278 The fundamental origin of this anomalous flow regime is an active area of research, and is likely associated with a
 1279 laminar-to-turbulent transition [232, 234].

1280 7.4.2 Dissolution and melting patterns as evidence for the Hanratty anomaly

1281 Transverse patterns may form on an ice sheet under deep flows when the ice sublimates in air or melts under water.
 1282 Similar patterns are also observed in soluble rocky substrates dissolving under water flow. These erosion patterns can
 1283 only appear if the rate of substrate erosion (sublimation, melting, or dissolution) is greater in the troughs than on
 1284 the bumps. Dissolution increases the concentration of solutes at the solid/fluid interface, and sublimation or melting
 1285 extracts heat from the fluid, which decreases temperature at the solid/fluid interface. To sustain the melting or disso-
 1286 lution process, heat or solute concentration must be transported towards or away from the interface. This transport
 1287 may be buoyancy-driven by melting or dissolution themselves, leading to density stratification into the fluid caused
 1288 by gradients in temperature or concentration [236, 237, 238, 239, 240]. In the case of a stable density stratification,
 1289 differential erosion between bumps and troughs is explained by the coupling between flow and topography, which
 1290 modulates the thicknesses of boundary layers (analogously to the case of dunes discussed above) and the turbulent
 1291 transport of heat and concentration [235, 134]. The existence of dissolution patterns suggests that the turbulent
 1292 viscosity and diffusion of heat or concentration are more significant in troughs than over bumps.

1293 Such patterns were reproduced experimentally in laboratory flumes over plaster [241, 242, 243] or ice [244, 245, 246]
 1294 beds and under turbulent flows. Based on these studies, the initial and mature patterns have wavelengths that scale
 1295 with the inverse of flow velocity, *i.e.*, that are proportional to the characteristic thickness of the viscous boundary
 1296 layer [247, 248]. Furthermore, mature patterns reach a stationary state with constant amplitude and wavelength. This
 1297 pattern saturation seems to coincide with a smooth-to-rough regime transition, at which point bedform amplitude
 1298 becomes much greater than the thickness of the boundary layer. These experimental observations support the existence
 1299 of a particular turbulent flow regime in the smooth case for a range of wavelengths, the Hanratty anomaly.

1300 7.5 Possible confinement by the atmospheric boundary layer

1301 A planetary atmosphere, like Earth's, displays a vertical structure. As radiation from the sun heats the ground, it
 1302 leads to an unstable density stratification (with warmer, less dense air at the bottom), which in turn drives convection
 1303 in the atmospheric boundary layer (ABL). Above the ABL, in the free atmosphere, density stratification is stable
 1304 (*i.e.*, air density decreases with altitude). The height of the ABL, Λ , depends on the heat flux from the ground
 1305 to the atmosphere and varies spatially due to, *e.g.*, lateral variations in ground albedo, and in time, *e.g.*, due to
 1306 seasonal changes in sun radiative flux. Such fluctuations make it challenging to determine a characteristic height of
 1307 the ABL that is relevant to dune formation, especially as giant dunes integrate flows over a long period, but only
 1308 under conditions that are conducive to active sediment transport [145]. Shear stresses in excess of threshold values
 1309 for sand transport tend to occur during strong diurnal convection within the boundary layer over dune fields—this
 1310 convection is associated with high Λ values [249].

1311 Another challenge lies in modeling wind flow over a dune under ABL confinement. In the subaqueous case, there is a
 1312 coupling between the bottom topography and the water free surface, which can deform and where waves can propagate
 1313 due to the gravitational restoring force. In particular, this coupling causes a transition between downstream-moving
 1314 dunes and upstream-moving antidunes when the flow over the dune exceeds the velocity of water waves. In this case,
 1315 flow regime is controlled by the Froude number, a dimensionless number that relates flow velocity to speed of shallow
 1316 water waves. Like in the subaqueous case, gravity waves can also propagate at the interface between the ABL and
 1317 the free atmosphere, where a density jump, $\Delta\rho_f$, occurs. One can thus define a Froude number, $\mathcal{F} = U/\sqrt{g\Lambda\Delta\rho_f/\rho_f}$,
 1318 where U is the flow velocity at altitude Λ [250, 251, 252, 253].

1319 A second Froude number, \mathcal{F}_N , can be defined from the strength of the vertical density stratification in the free
 1320 atmosphere, $\partial_z\rho_f$. Stable stratification in the free atmosphere sets a characteristic oscillation frequency of fluid
 1321 particles, the Brunt-Väisälä frequency, $N = \sqrt{-g\partial_z\rho_f/\rho_f}$, such that $\mathcal{F}_N = U/(\Lambda N)$ [250, 251, 252, 253]. The coupling
 1322 between dune dynamics and the ABL, and its dependence on these Froude numbers, remain largely unexplored. In

1323 addition to the ABL, flow confinement can result from other phenomena, such as katabatic winds. Each specific
1324 confinement scenario is likely to be characterized by different controls on Λ . Cross-wind or secondary dune patterns
1325 could possibly be induced by such confinement effects [101].

1326 Acknowledgement

1327 This work began during a funded workshop at the Domaine des Treilles, a wonderful place in the south of France for
1328 scientific exchange, creativity and reflection. The authors would like to thank several collaborators who contributed
1329 to the work reviewed in this paper, namely, in alphabetical order: B. Andreotti, F. Charru, S. Douady, O. Durán, M.
1330 Génois, P. Jia, M. Louge, P. Lü, E. Reffet, S. Rodriguez, O. Rozier.

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