

**1 Local wind regime induced by giant linear dunes:  
2 comparison of ERA5-Land reanalysis with surface  
3 measurements**

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**10 Abstract**

11 Emergence and growth of sand dunes results from the dynamic interaction  
12 between topography, wind flow and sediment transport. While feedbacks be-  
13 tween these variables are well studied at the scale of a single and relatively  
14 small dune, the average effect of a periodic large-scale dune pattern on atmo-  
15 spheric flows remains poorly constrained, due to a lack of data in major sand  
16 seas. Here, we compare local measurements of surface winds to the predictions  
17 of the ERA5-Land climate reanalysis at four locations in Namibia, within and  
18 outside the giant linear dune field of the Namib sand sea. In the desert plains  
19 to the north of the sand sea, observations and predictions agree well. This  
20 is also the case in the interdune areas of the sand sea during the day. Dur-  
21 ing the night, however, an additional wind component aligned with the giant  
22 dune orientation is measured, in contrast to the easterly wind predicted by

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the ERA5-Land reanalysis. For the given dune orientation and measured wind regime, we link the observed wind deviation (up to 50°) to the daily cycle of the turbulent atmospheric boundary layer. During the night, a shallow boundary layer induces a flow confinement above the giant dunes, resulting in large flow deviations, especially for the lower easterly winds. During the day, the feedback of the giant dunes on the atmospheric flow is much weaker due to the thicker boundary layer and higher wind speeds. Finally, we propose that the confinement mechanism and the associated wind deflections induced by giant dunes could explain the development of smaller-scale secondary dunes, elongating obliquely in the interdune areas of the primary dune patterns.

**Keywords** Atmospheric boundary layer · Sand dunes · Flow over hills

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**34 1 Introduction**

35 The description of turbulent flows over complex topography is relevant for  
36 a large variety of different environmental systems (Sherman 1978; Walmsley  
37 et al. 1982; Baines 1995; Wood 2000; Venditti et al. 2013; Finnigan et al.  
38 2020). For example, the flow over hills is of primary interest for wind power,  
39 meteorological and air pollution phenomena (Taylor et al. 1987). The proper-  
40 ties of these flows are also key to the understanding of geophysical phenom-  
41 ena, including the formation of wind-driven waves on the ocean surface (Sulli-  
42 van and McWilliams 2010), dissolution bedforms (Claudin et al. 2017; Guérin  
43 et al. 2020), or sedimentary ripples and dunes (Bagnold 1941; Charru et al.  
44 2013; Courrech du Pont 2015). Importantly, the troposphere presents a vertical  
45 structure, with a lower convective boundary layer, of typical kilometer-scale  
46 thickness, capped by a stably stratified region (Stull 1988). The largest topo-  
47 graphic obstacles, such as mountains, can therefore interact with this upper  
48 region and lead to internal wave generation or significant wind disturbances,  
49 such as lee-side downslope winds (Durran 1990).

50 Compared to hills and mountains, aeolian sand dunes offer idealized ele-  
51 vation profiles for the study of atmospheric turbulent flow over topographies,  
52 due to their smooth shape, free of canopies. Besides, dunes provide a rather  
53 wide range of scales, from decameters to kilometers, and very often come in  
54 a fairly regular pattern, which further simplifies the flow structure analysis.  
55 Past studies have highlighted two important topographic feedbacks on the  
56 wind flow close to the dune/hill surface. First is the effect on wind speed, with  
57 documented flow acceleration on upwind slopes (Weaver and Wiggs 2011) and  
58 deceleration on downwind slopes (Baddock et al. 2007), where the speed-up  
59 factor is essentially proportional to the obstacle aspect ratio (Jackson and Hunt  
60 1975). Under multidirectional wind regimes with frequent wind reversals, this  
61 speed-up effect induces large differences in the amplitude and orientation of  
62 the resultant sediment transport between flat sand beds and the dune crests  
63 (???). In addition, the maximum in wind velocity is typically shifted upwind  
64 of the obstacle crest (Jackson and Hunt 1975; Claudin et al. 2013). This be-  
65 haviour has been theoretically predicted by means of asymptotic analysis of  
66 a neutrally stratified boundary-layer flow over an obstacle of vanishing aspect  
67 ratio (Jackson and Hunt 1975; Mason and Sykes 1979; Sykes 1980; Hunt et al.  
68 1988; Belcher and Hunt 1998; Kroy et al. 2002). Experiments in flumes (Zilker  
69 et al. 1977; Zilker and Hanratty 1979; Frederick and Hanratty 1988; Poggi et al.  
70 2007; Bristow et al. 2022), in wind tunnels (Gong and Ibbetson 1989; Finnigan  
71 et al. 1990; Gong et al. 1996) and in field conditions at all scales (Taylor and  
72 Teunissen 1987; Claudin et al. 2013; Fernando et al. 2019; Lü et al. 2021),  
73 have also documented this effect. Interestingly, a similar behaviour exists for  
74 the pressure perturbation, but with a slight downwind shift for the pressure  
75 minimum (Claudin et al. 2021).

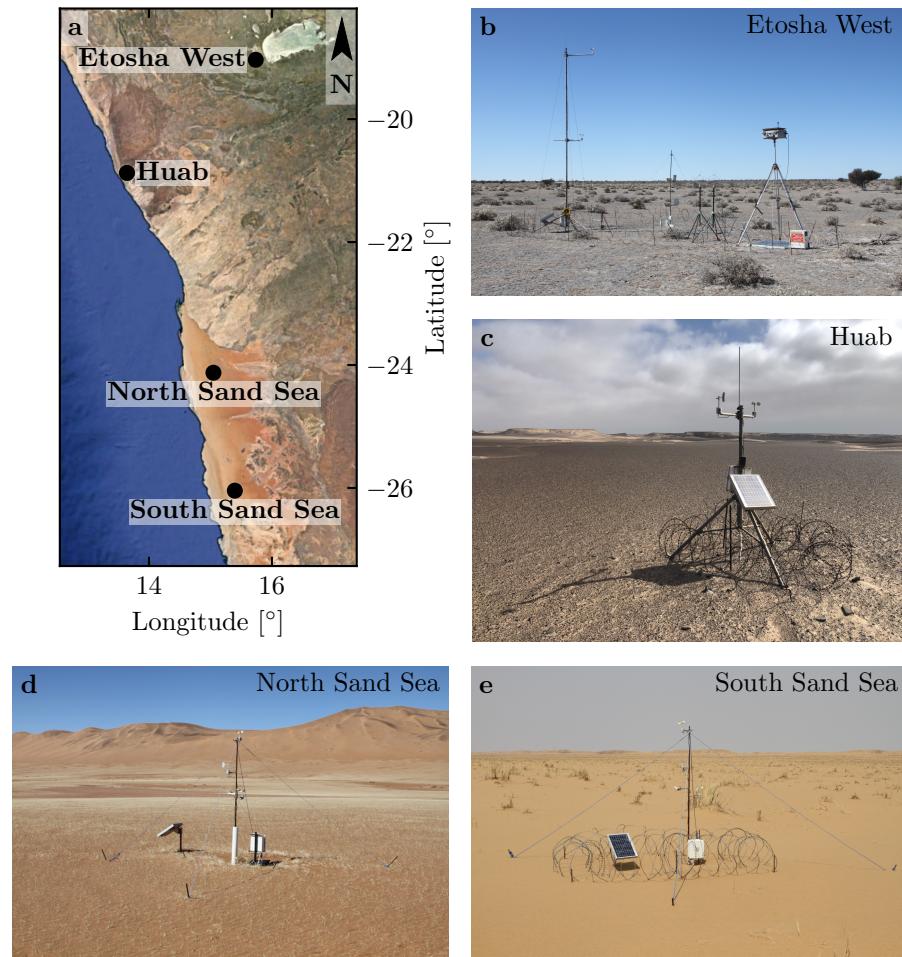
76 The second effect, much less studied, is the flow deflection that occurs when  
77 the incident wind direction is not perpendicular to the ridge crest. While pre-  
78 dicted to be small (less than 10°) in the linear regime valid for shallow topogra-

79 phy (Gadal et al. 2019), significant flow steering has been reported in the field  
80 on the downwind side of steep enough obstacles, such as well-developed sand  
81 dunes (Tsoar and Yaalon 1983; Sweet and Kocurek 1990; Walker and Nickling  
82 2002; Smith et al. 2017) and in particular coastal foredunes (e.g. Hunter et al.  
83 1983; Rasmussen 1989; Walker et al. 2006, 2009; Hesp et al. 2015; Walker  
84 et al. 2017; de Winter et al. 2020), mountain ranges (Kim et al. 2000; Lewis  
85 et al. 2008; Fernando et al. 2019), and valley topographies (Wiggs et al. 2002;  
86 Garvey et al. 2005).

87 Wind measurements over sand dunes has been mainly performed over small  
88 bedforms, typically a few meters high (corresponding to several tens of meters  
89 long) (e.g. Mulligan 1988; Hesp et al. 1989; Lancaster et al. 1996; Mckenna  
90 Neuman et al. 1997; Sauermann et al. 2003; Andreotti et al. 2002; Walker and  
91 Nickling 2002; Weaver and Wiggs 2011). For practical reasons, fewer studies  
92 performed similar measurements on giant dunes (Havholm and Kocurek 1988),  
93 with kilometer-scale wavelengths and heights of tens of meters. However, such  
94 large dunes provide a choice configuration for the study of turbulent flows over  
95 a complex topography. First, one expects larger wind disturbances for larger  
96 obstacles. Secondly, their large size can make them interact with the vertical  
97 structure of the atmosphere (Andreotti et al. 2009). Third, they usually form  
98 large patterns in sand seas and thus behave as rather clean periodic perturba-  
99 tions, in contrast with isolated dunes. Finally, because the morphodynamics of  
100 aeolian bedforms is strongly dependent on the local wind regime (Livingstone  
101 and Warren 2019), one can expect to see the consequences of windflow dis-  
102 turbance by large dunes on neighbouring small dunes (Brookfield 1977; Ewing  
103 et al. 2006). A similar effect is observed on the properties of impact ripple  
104 patterns due to the presence of dunes (Howard 1977; Hood et al. 2021).

105 Atmospheric flows have been much studied at the desert-scale with climate  
106 reanalyses based on global atmospheric models (Blumberg and Greeley 1996;  
107 Livingstone et al. 2010; Ashkenazy et al. 2012; Jolivet et al. 2021; Hu et al.  
108 2021; Gunn et al. 2021b), such as ERA-40, ERA-Interim or ERA5 (Uppala  
109 et al. 2005; Dee et al. 2011; Hersbach et al. 2020). However, the spatial resolu-  
110 tion of these reanalyses (tens of kilometers) implies average quantities that do  
111 not resolve the smaller scales of interest, which range from individual dunes  
112 to small mountains (Livingstone et al. 2010). Recently, the release of ERA5-  
113 Land has partly resolved this limitation by providing up to 70 years of hourly  
114 wind predictions at a 9 km spatial resolution (Muñoz-Sabater et al. 2021).  
115 However, its validity remains to be studied, especially in remote desert areas  
116 where assimilation of measured data is very low.

117 In this work, we compare local wind speeds and directions measured by  
118 meteorological stations at four different locations inside and north of the giant  
119 linear dune field of the Namib sand sea to the regional predictions of the ERA5-  
120 Land climate reanalysis. Where the meteorological stations are surrounded by  
121 a relatively flat environment, we show that local measurements and regional  
122 predictions agree well. The agreement is also good in the interdune areas of  
123 the sand sea, except for some weak winds blowing at night, which exhibit an  
124 additional component aligned with the giant dune orientation. These winds

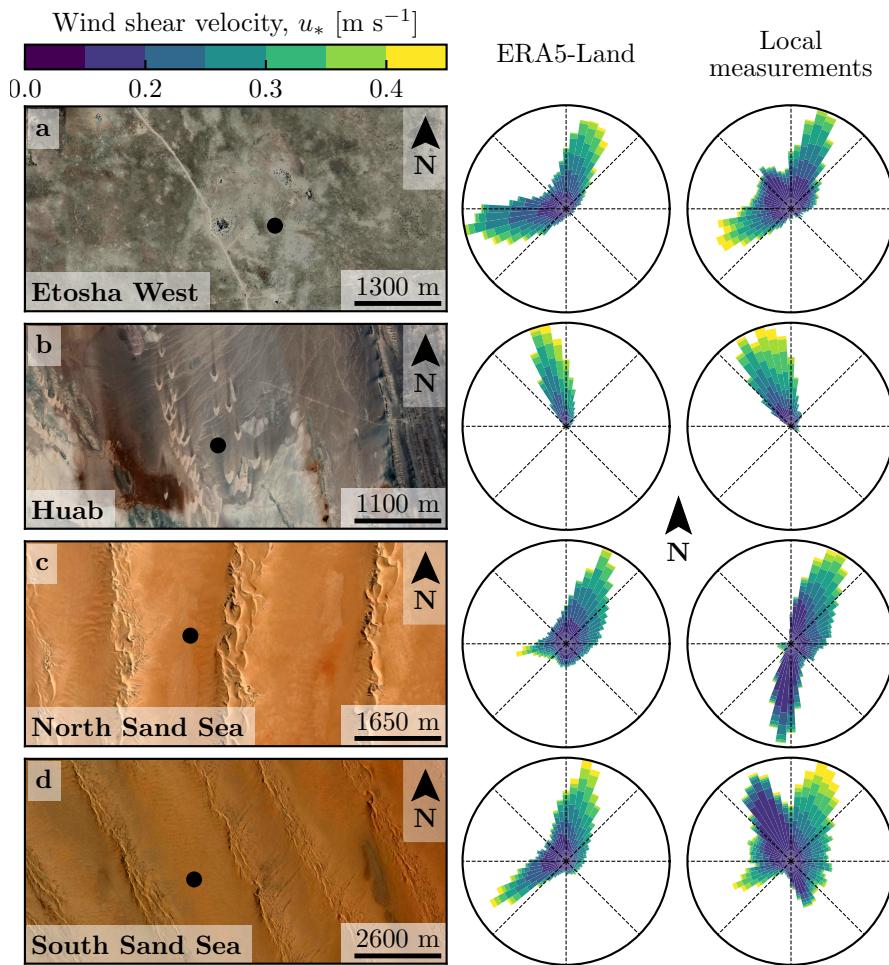


**Fig. 1** Studied field sites. **a:** Location of the different sites in Namibia. **b–e:** Photographs of the meteorological stations.

125 are not predicted by the ERA5-Land reanalysis (section 2). Further, we are  
 126 able to link the magnitude of these differences to the circadian cycle of the  
 127 atmospheric boundary layer (section 3). Finally, we draw implications for the  
 128 wind disturbances on smaller-scale dunes (section 4), suggesting a possible  
 129 origin for crossing dunes.

## 130 2 Wind regimes across the Namib Sand Sea

131 We measured the wind regime at four different locations in Namibia, represen-  
 132 tative of various arid environments across the Namib desert (Fig. 1, Fig. 2).  
 133 The Etosha West station was located at the Adamax waterhole to the west



**Fig. 2** Wind data used in this study. Satellite images of the different environments (Google-Earth, Maxar Technologies, CNES/Airbus) are shown on the left. The black dots show the location of the wind measurements stations. On the right of the photos, the corresponding wind roses representing the data from the ERA5-Land climate reanalysis and the local wind stations are displayed. Note: the graphical convention for the wind roses is that the bars show the direction towards which the wind blows (see color bar for velocity scale).

of Etosha Pan in northern Namibia, in a sparsely vegetated area. The Huab station was near the coast on a hyper-arid flat gravel plain lying north the ephemeral Huab river. Here, barchan dunes up to a few meters in height develop from the sediment blowing out of the river valley (Nield et al. 2017; Hesp and Hastings 1998). These two stations were both located in relatively flat environments. In contrast, the North Sand Sea and South Sand Sea stations were located in the interdunes between linear dunes with kilometer-scale wavelengths, hectometer-scale heights and superimposed patterns. In this sec-

<sup>142</sup> tion, we describe and compare winds from local measurements and climate  
<sup>143</sup> reanalysis predictions.

<sup>144</sup> 2.1 Wind and elevation data

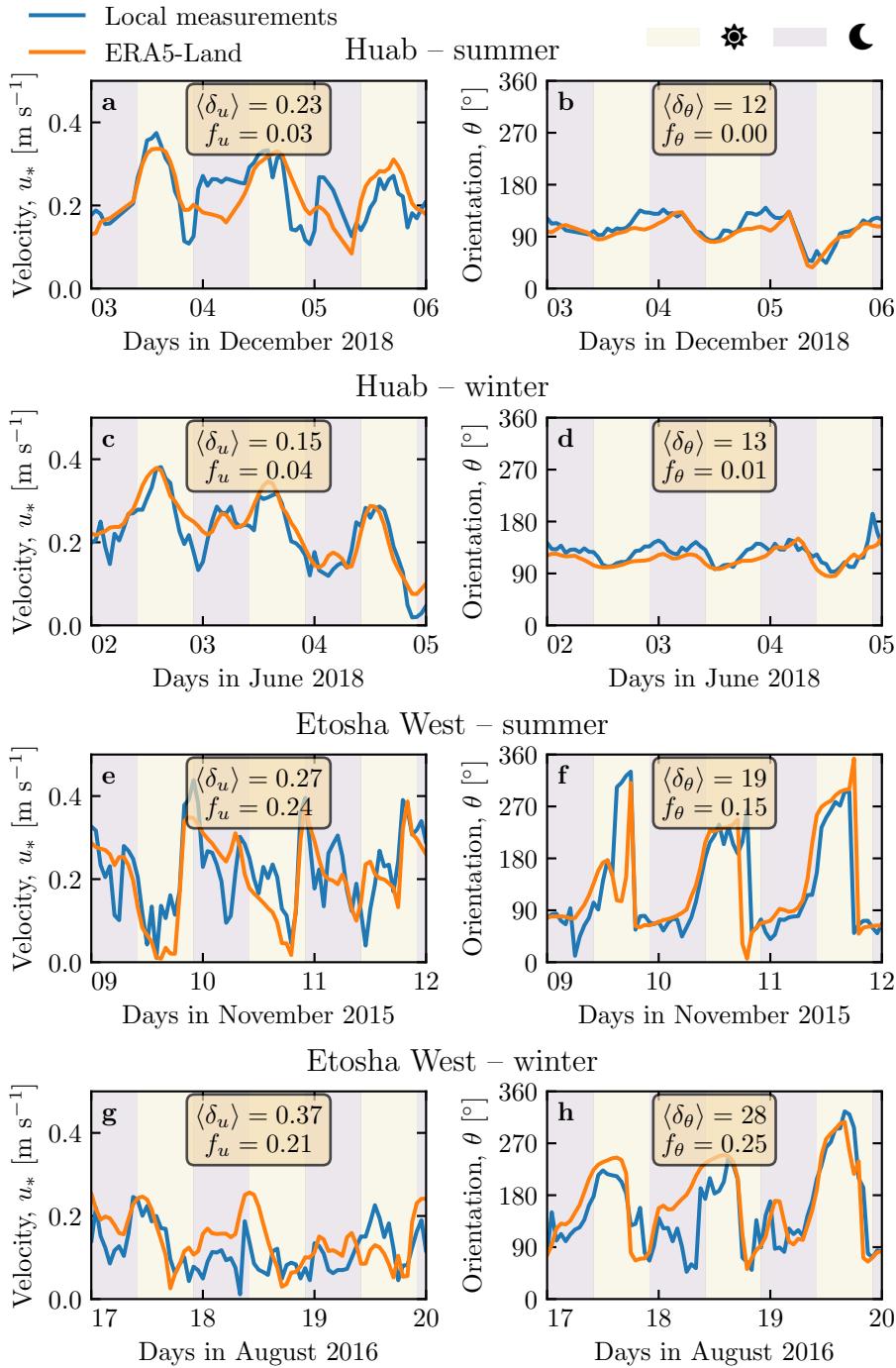
<sup>145</sup> At each meteorological station (Fig. 1), wind speed and direction were sampled  
<sup>146</sup> every 10 minutes using cup anemometers (Vector Instruments A100-LK) and  
<sup>147</sup> wind vanes (Vector Instruments W200-P) at a single height, which was between  
<sup>148</sup> 2 m and 3 m depending on the station. The available period of measurements at  
<sup>149</sup> each station ranged from 1 to 5 discontinuous years distributed between 2012  
<sup>150</sup> and 2020 (Online Resource Fig. S1). We checked that at least one complete  
<sup>151</sup> seasonal cycle was available for each station. Regional winds were extracted  
<sup>152</sup> at the same locations and periods from the ERA5-Land dataset, which is a  
<sup>153</sup> replay at a smaller spatial resolution of ERA5, the latest climate reanalysis  
<sup>154</sup> from the ECMWF (Hersbach et al. 2020; Muñoz-Sabater et al. 2021). This  
<sup>155</sup> dataset provided hourly predictions of the 10-m wind velocity and direction  
<sup>156</sup> at a spatial resolution of  $0.1^\circ \times 0.1^\circ$  ( $\simeq 9$  km in Namibia).

<sup>157</sup> To enable direct comparison, the local wind measurements were averaged  
<sup>158</sup> into 1-hr bins centered on the temporal scale of the ERA5-Land estimates  
<sup>159</sup> (Online Resource Fig. S2). As the wind velocities of both datasets were pro-  
<sup>160</sup> vided at different heights, we converted them into shear velocities  $u_*$  (Online  
<sup>161</sup> Resource section 1), characteristic of the turbulent wind profile. Wind roses  
<sup>162</sup> in Fig. 2 show the resulting wind data.

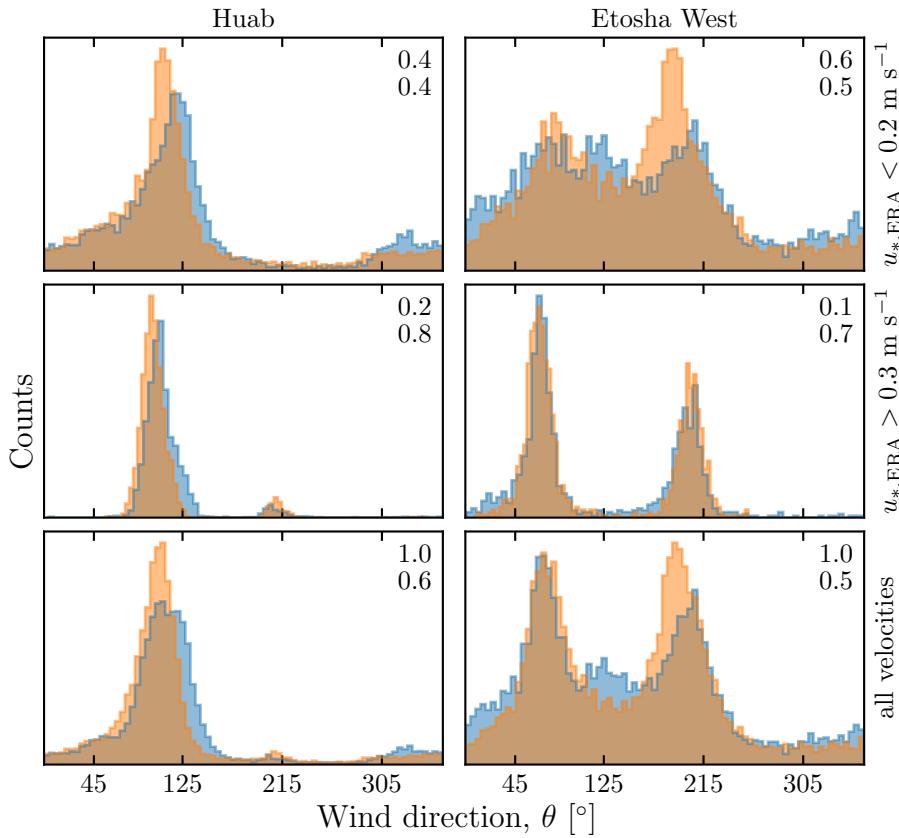
<sup>163</sup> Dune properties were computed using autocorrelation on the 30-m Digital  
<sup>164</sup> Elevation Models (DEMs) of the shuttle radar topography mission (Farr et al.  
<sup>165</sup> 2007). For the North and South Sand Sea stations, we obtain, respectively,  
<sup>166</sup> orientations of  $85^\circ$  and  $125^\circ$  with respect to the North, wavelengths of 2.6 km  
<sup>167</sup> and 2.3 km and amplitudes (or half-heights) of 45 m and 20 m (Online Resource  
<sup>168</sup> Fig. S4 for more details). This agrees with direct measurements made on site.

<sup>169</sup> 2.2 Comparison of local and regional winds

<sup>170</sup> The measured and predicted wind regimes are shown in Fig. 2. In the Namib,  
<sup>171</sup> the regional wind patterns are essentially controlled by the sea breeze, result-  
<sup>172</sup> ing in strong northward components (sometimes slightly deviated by the large  
<sup>173</sup> scale topography) present in all regional wind roses (Lancaster 1985). These  
<sup>174</sup> daytime winds are dominant during the period October-March (Fig. 3f and  
<sup>175</sup> Online Resource Fig. 4f). During April-September, an additional (and often  
<sup>176</sup> nocturnal) easterly component can also be recorded, induced by the combina-  
<sup>177</sup> tion of katabatic winds forming in the mountains, and infrequent ‘berg’ winds,  
<sup>178</sup> which are responsible for the high wind velocities observed (Lancaster et al.  
<sup>179</sup> 1984). The frequency of these easterly components decreases from inland to  
<sup>180</sup> the coast. As a result, bidirectional wind regimes within the Namib Sand Sea  
<sup>181</sup> and at the west Etosha site (Fig. 2a,c,d) and a unidirectional wind regime on  
<sup>182</sup> the coast at the outlet of the Huab River (Fig. 2b) are observed.



**Fig. 3** Temporal comparison between the wind data coming from the ERA5-Land climate reanalysis (orange lines) and from the local measurements (blue lines). Coloured swathes indicate day (between 10.00 UTC and 22.00 UTC) and night (before 10.00 UTC or after 22.00 UTC). Numbers in legends indicate the average flow deflection  $\delta_\theta$  and relative wind modulation  $\delta_u$  over the displayed period (see section 3.2 for their definitions), as well as the percentage  $f_\theta$  and  $f_u$  of occurrence of extreme events ( $\delta_\theta > 50^\circ$ ,  $|\delta_u| > 0.6$ ). **a–b**: Huab station in summer. **b–c**: Huab station in winter. **d–e**: Etosha West station in summer. **f–g**: Etosha West station in winter. Time series of the two other stations are shown in Fig. 5.



**Fig. 4** Distributions of wind direction at Huab and Etosha West stations for the ERA5-Land climate reanalysis (orange) and the local measurements (blue). In each subplot, both distributions are plotted from the same time steps, selected for different ranges of the wind wind velocity (rows) in the ERA5-Land dataset. The numbers at the upper right corners give the percentage of time steps selected in each sub-range (top), as well as the percentage of them corresponding to the day – defined between 10.00 UTC and 22.00 UTC (bottom).

In the case of the Etosha West and Huab stations, the time series of wind speed and direction from the regional predictions quantitatively match those corresponding to the local measurements (Figs. 3, 4 and Online Resource Fig. S5). For the North Sand Sea and South Sand Sea stations within the giant linear dune field, we observe that this agreement is also good, but limited to the October-March time period (Fig. 4a, b, e, f). However, the field-measured wind roses exhibit additional wind components aligned with the dune orientation, as evidenced on the satellite images (Fig. 2c,d).

More precisely, during the April-September period, the local and regional winds in the interdune match during daytime only, i.e. when the southerly/southwesterly sea breeze dominates (Figs. 5c,d,g,h and 6). In the late afternoon and during the night, when the easterly ‘berg’ and katabatic winds blow, measurements and predictions differ. In this case, the angular wind distribution of the lo-

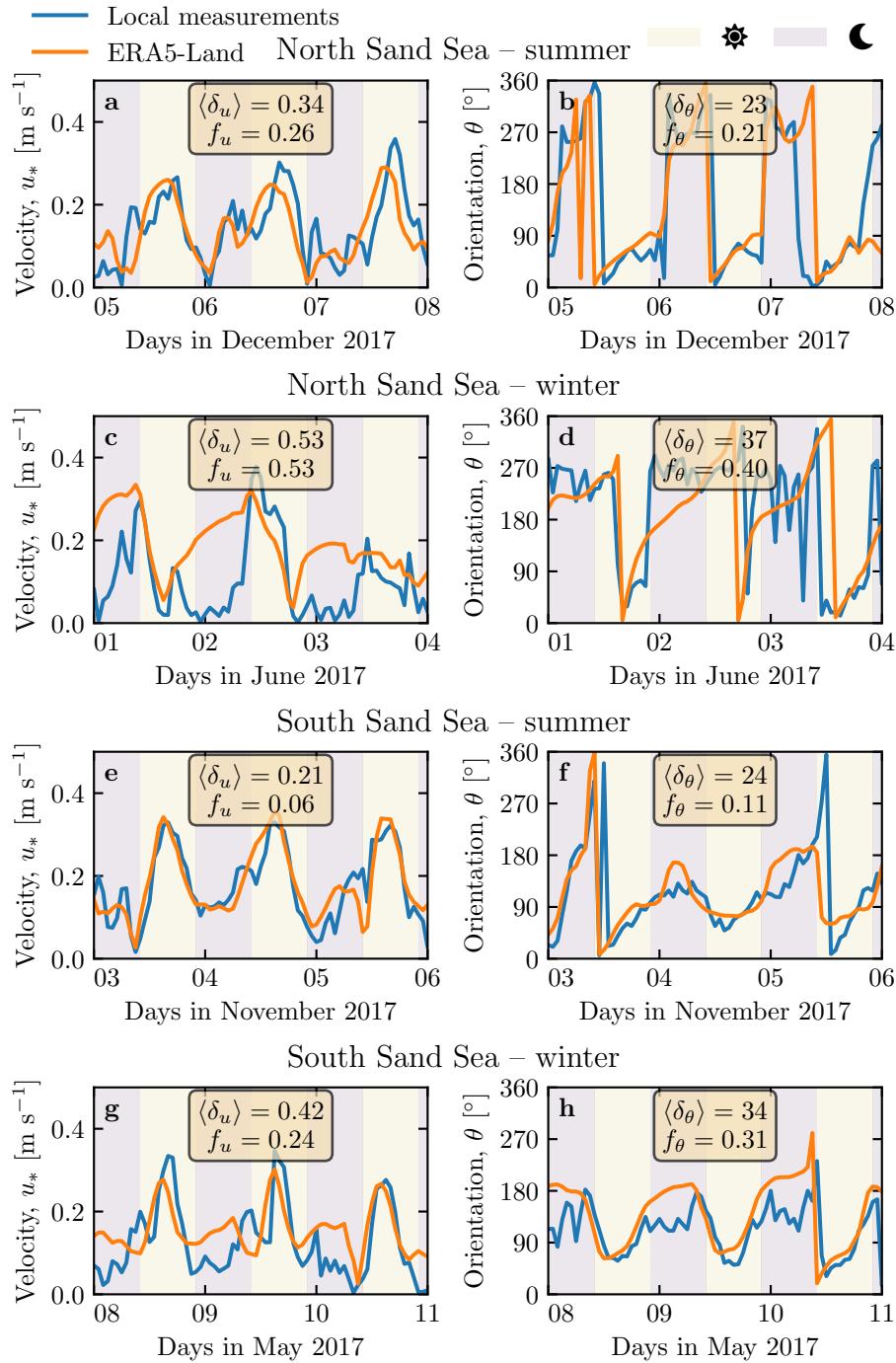
cal measurements exhibits two additional modes corresponding to reversing winds aligned with the dune orientation (purple frame in Fig. 6, Online Resource Fig. S6). This deviation is also associated with a general attenuation of the wind strength (Online Resource Fig. S7). Remarkably, all these figures show that these wind reorientation and attenuation processes occur only at low velocities of the regional wind, typically for  $u_*^{\text{ERA5-Land}} \lesssim 0.2 \text{ m s}^{-1}$ . For shear velocities larger than  $u_*^{\text{ERA5-Land}} \simeq 0.3 \text{ m s}^{-1}$ , the wind reorientation is not apparent. Finally, for intermediate shear velocities, both situations of wind flow reoriented along the dune crest and not reoriented can be successively observed (Online Resource Fig. S6). Importantly, these values are not precise thresholds (and certainly not related to the threshold for sediment transport), but indicative of a crossover between regimes, whose physical interpretation is discussed in the next section.

### 209 3 Influence of wind speed and circadian cycle on the atmospheric 210 boundary layer

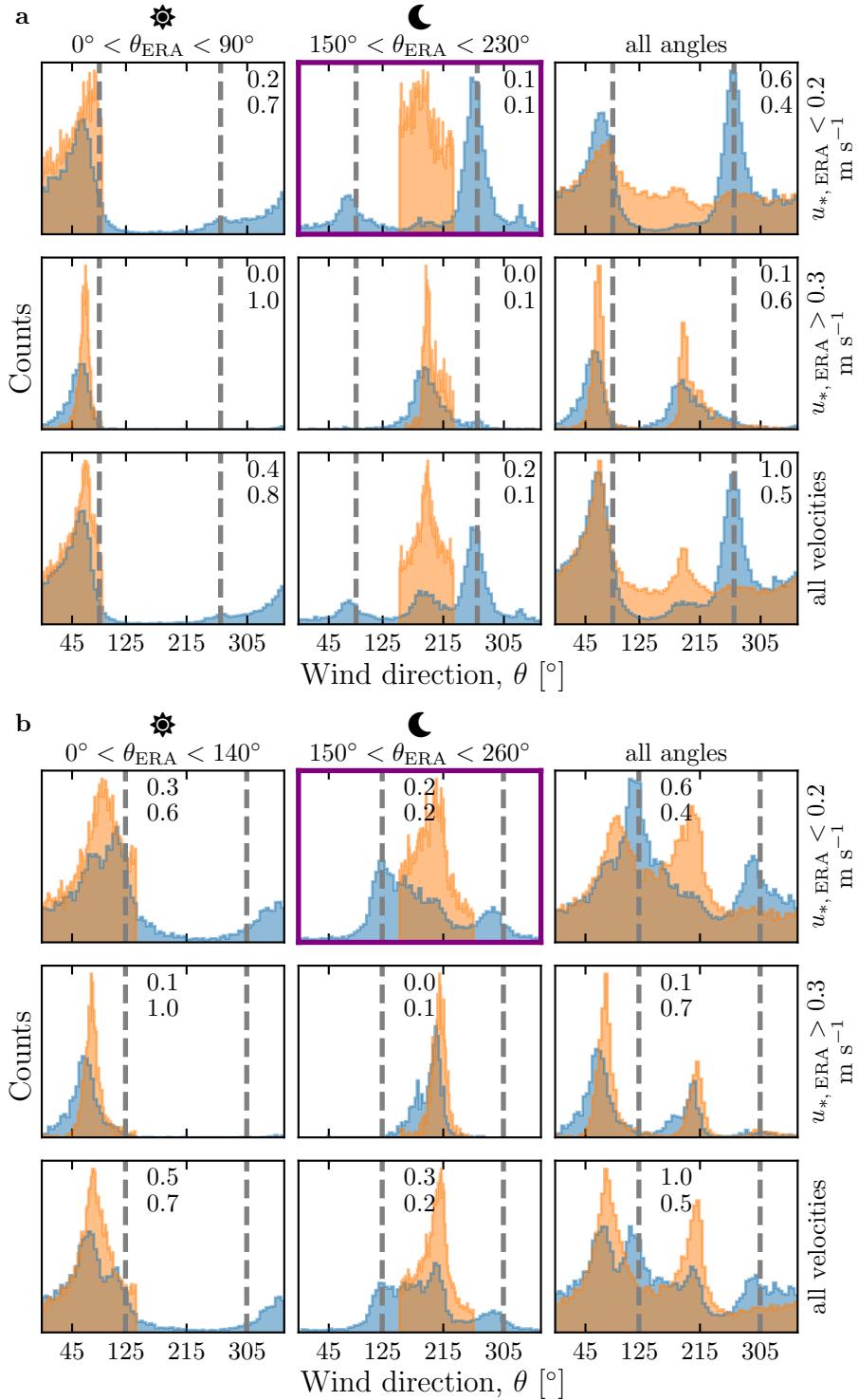
211 The wind deflection induced by dunes has previously been related to the in-  
212 cident angle between wind direction and crest orientation, with a maximum  
213 deflection evident for incident angles between  $30^\circ$  and  $70^\circ$  (Walker et al. 2009;  
214 Hesp et al. 2015). In the data analysed here, the most deflected wind at both  
215 the North and South Sand Sea stations is seen to be where the incident angle  
216 is perpendicular to the giant dunes (Figs. 2 and 6). It therefore appears that  
217 in our case, the incident wind angle is not the dominant control on maximum  
218 wind deflection. Further, and as shown in Fig. 6, winds of high and low veloc-  
219 ities show contrasting behaviour in characteristics of deflection. This suggests  
220 a change in hydrodynamical regime between the winds. In this section, we  
221 discuss the relevant parameters associated with the dynamical mechanisms  
222 that govern the interactions between the atmospheric boundary layer flow and  
223 giant dune topographies. This analysis allows us to provide a physics-based  
224 interpretation of our measured wind data.

#### 225 3.1 Flow over a modulated bed

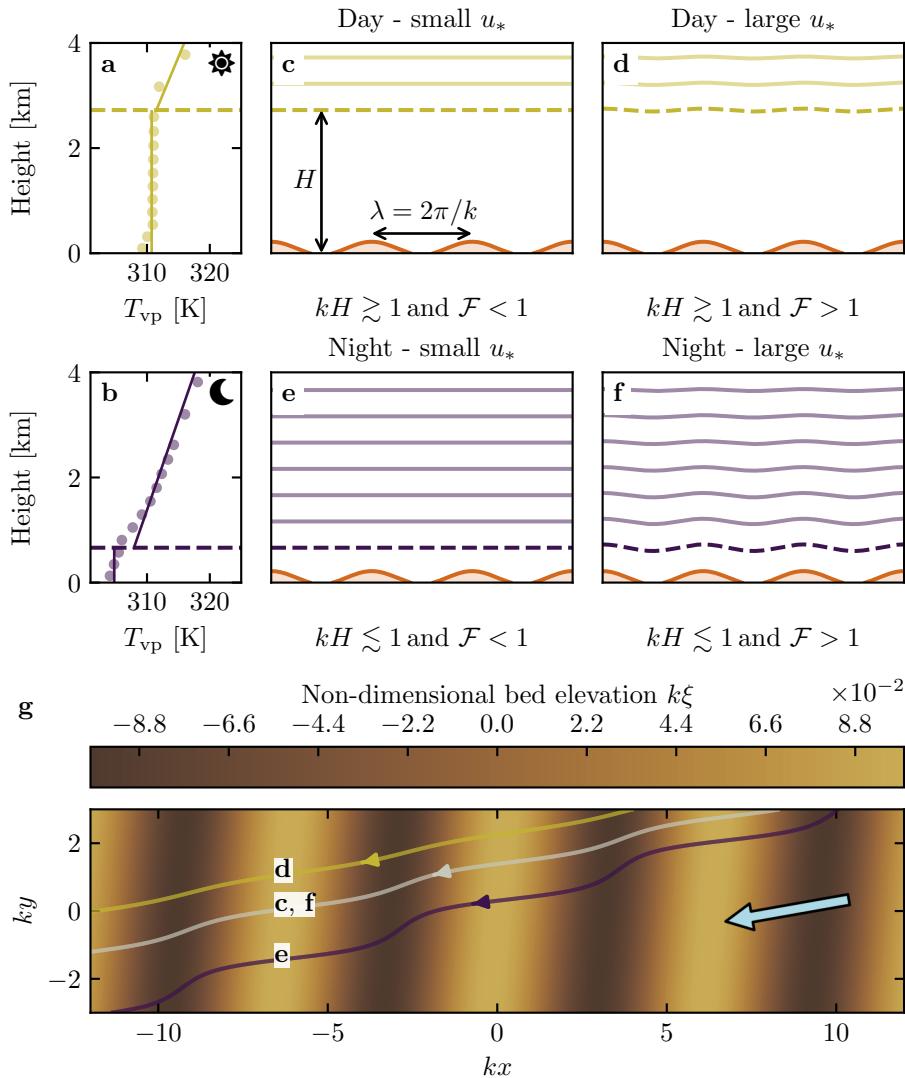
226 Taking as a reference the turbulent flow over a flat bed, the general framework  
227 of our study is understanding and describing the flow response to a bed mod-  
228 ulation (e.g. a giant dune). Without loss of generality, we can consider in this  
229 context an idealised bed elevation in the form of parallel sinusoidal ridges, with  
230 wavelength  $\lambda$  (or wavenumber  $k = 2\pi/\lambda$ ) and amplitude  $\xi_0$ , and where the ref-  
231 erence flow direction makes a given incident angle with respect to the ridge  
232 crest (Andreotti et al. 2012). Part of this response, on which we focus here,  
233 is the flow deflection by the ridges. In a simplified way, it can be understood  
234 from the Bernoulli principle (Hesp et al. 2015): as the flow approaches the  
235 ridge crest, the compression of the streamlines results in larger flow velocities,



**Fig. 5** Same as Fig. 3 for North Sand Sea station in summer (a–b), North Sand Sea station in winter (b–c), South Sand Sea station in summer (d–e) and South Sand Sea station in winter (f–g).



**Fig. 6** Same as Fig. 4 but for North Sand Sea (**a**) and South Sand Sea (**b**) stations. Here, subplots correspond to different ranges for the wind direction (columns) and wind velocity (rows) of the ERA5-Land dataset. The grey vertical dashed lines indicate the main dune orientation. In contrast with observations at the Huab and Etosha West stations (Fig. 4), histograms do not match well at low wind velocities, and the purple frame highlights the regime (low wind velocities, nocturnal easterly wind) in which the data from both datasets differ most.



**Fig. 7** **a–b:** Vertical profiles of the virtual potential temperature  $T_{vp}$  at two different time steps (day - 03/11/2015 - 12.00 UTC, night - 01/13/2013 - 09.00 UTC) at the North Sand Sea station. Dots: data from the ERA5 reanalysis. Dashed lines: boundary layer height given by the ERA5 reanalysis (Online Resource section 2). Plain lines: vertical (boundary layer) and linear (free atmosphere) fits to estimate the stratification properties. **c–f:** Sketches representing the interaction between the giant dunes and the atmospheric flow for different meteorological conditions. **g:** Streamlines over a sinusoidal topography  $\xi(x, y)$  qualitatively representing the effect of low, medium and strong flow confinement, in relation to the above panels (see Appendix 1 for more details). The blue arrow indicates the undisturbed wind direction.

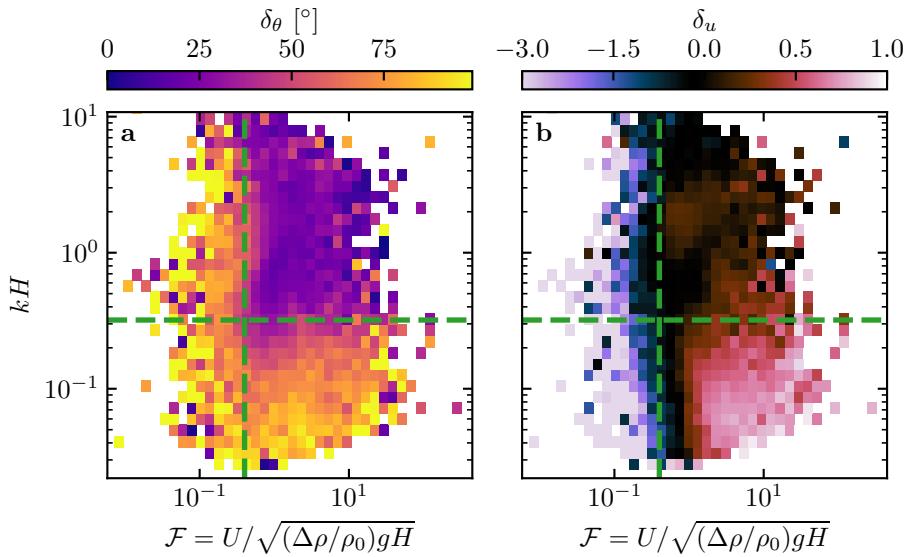
and thus lower pressures (Jackson and Hunt 1975). An incident flow oblique to the ridge is then deflected towards lower pressure zones, i.e towards the crest. Turbulent dissipation tends to increase this effect downstream, resulting in wind deflection along the crest in the lee side (Gadal et al. 2019).

Flow confinement below a capping surface, which enhances streamline compression, has a strong effect on the hydrodynamic response and typically increases flow deflection. This is the case for bedforms forming in open channel flows such as rivers (Kennedy 1963; Chang and Simons 1970; Mizumura 1995; Colombini 2004; Fourrière et al. 2010; Andreotti et al. 2012; Unsworth et al. 2018). This is also relevant for aeolian dunes as they evolve in the turbulent atmospheric boundary layer (ABL) capped by the stratified free atmosphere (FA) (Andreotti et al. 2009). Two main mechanisms, associated with dimensionless numbers must then be considered (Fig. 7). First, topographic obstacles typically disturb the flow over a characteristic height similar to their length. As flow confinement is characterised by a thickness  $H$ , the interaction between the dunes and the wind in the ABL is well captured by the parameter  $kH$ . The height  $H$  is directly related to the sensitive heat flux from the Earth surface. It is typically on the order of a kilometre, but significantly varies with the circadian and seasonal cycles. Emerging and small dunes, with wavelengths in the range 20 to 100 m, are not affected by the flow confinement, corresponding to  $kH \gg 1$ . For giant dunes with kilometer-scale wavelengths, however, their interaction with the FA can be significant (Andreotti et al. 2009). This translates into a parameter  $kH$  in the range 0.02–5, depending on the moment of the day and the season. A second important mechanism is associated with the existence of a thin intermediate so-called capping layer between the ABL and the FA. It is characterised by a density jump  $\Delta\rho$ , which controls the ‘rigidity’ of this interface, i.e. how much its deformation affects streamline compression. This is usually quantified using the Froude number (Vosper 2004; Stull 2006; Sheridan and Vosper 2006; Hunt et al. 2006; Jiang 2014):

$$\mathcal{F} = \frac{U}{\sqrt{\frac{\Delta\rho}{\rho_0}} g H}, \quad (1)$$

where  $U$  is the wind velocity at the top of the ABL and  $\rho_0$  its average density. The intensity of the stratification, i.e. the amplitude of the gradient  $|\partial_z \rho|$  in the FA, also impacts the ability to deform the capping layer under the presence of an underlying obstacle, and thus affects the influence of flow confinement. This can be quantified using the internal Froude number (Vosper 2004; Stull 2006; Sheridan and Vosper 2006; Hunt et al. 2006; Jiang 2014)  $\mathcal{F}_I = kU/N$ , where  $N = \sqrt{-g\partial_z \rho / \rho_0}$  is the Brunt-Väisälä frequency (Stull 1988). Both Froude numbers have in practice the same qualitative effect on flow confinement (a smaller Froude corresponding to a stiffer interface), and we shall restrict the main discussion to  $\mathcal{F}$  only.

With this theoretical framework in mind, and in the context of the measured wind data in the North and South Sand Sea stations, the smallest wind disturbances are expected to occur during the day, when the ABL depth is



**Fig. 8** Regime diagrams of the wind deviation  $\delta_\theta$  (a) and relative attenuation/amplification  $\delta_u$  (b) in the space  $(\mathcal{F}, kH)$ , containing the data from both the North Sand Sea and South Sand Sea stations. The green dashed lines empirically delimit the different regimes. The point density in each bin of the diagrams is shown in Online Resource Fig. S10 – 95% of the data occur in the range  $-1 < \delta u < 1$ . Similar regime diagrams in the spaces  $(\mathcal{F}_I, kH)$  and  $(\mathcal{F}_I, \mathcal{F})$  are shown in Online Resource Fig. S11.

278 the largest and comparable to the dune wavelength ( $kH \gtrsim 1$ ), which corre-  
 279 sponds to a weak confinement situation (Fig. 7c,d). In contrast, large wind  
 280 disturbances are expected to occur during the night, when the confinement  
 281 is mainly induced by a shallow ABL (Fig. 7e). However, this strong confine-  
 282 ment can be somewhat reduced in the case of strong winds, corresponding to  
 283 large values of the Froude number and a less ‘rigid’ interface (Fig. 7f). This  
 284 is in qualitative agreement with the transition from deflected to non-deflected  
 285 winds related to low and high velocities observed in our data (Sec. 2.2).

### 286 3.2 Data distribution in the flow regimes

287 We can go one step further and analyse how our data quantitatively spread  
 288 over the different regimes discussed above. For that purpose, one needs to  
 289 compute  $kH$  and  $\mathcal{F}$  from the time series.  $H$ ,  $U$  and the other atmospheric  
 290 parameters can be deduced from the various vertical profiles (temperature,  
 291 humidity) available in the ERA5 climate reanalysis (Online Resource section  
 292 2). We quantify the flow deflection  $\delta_\theta$  as the minimal angle between the wind  
 293 orientations comparing the local measurements and the regional predictions.  
 294 We also compute the relative velocity modulation as

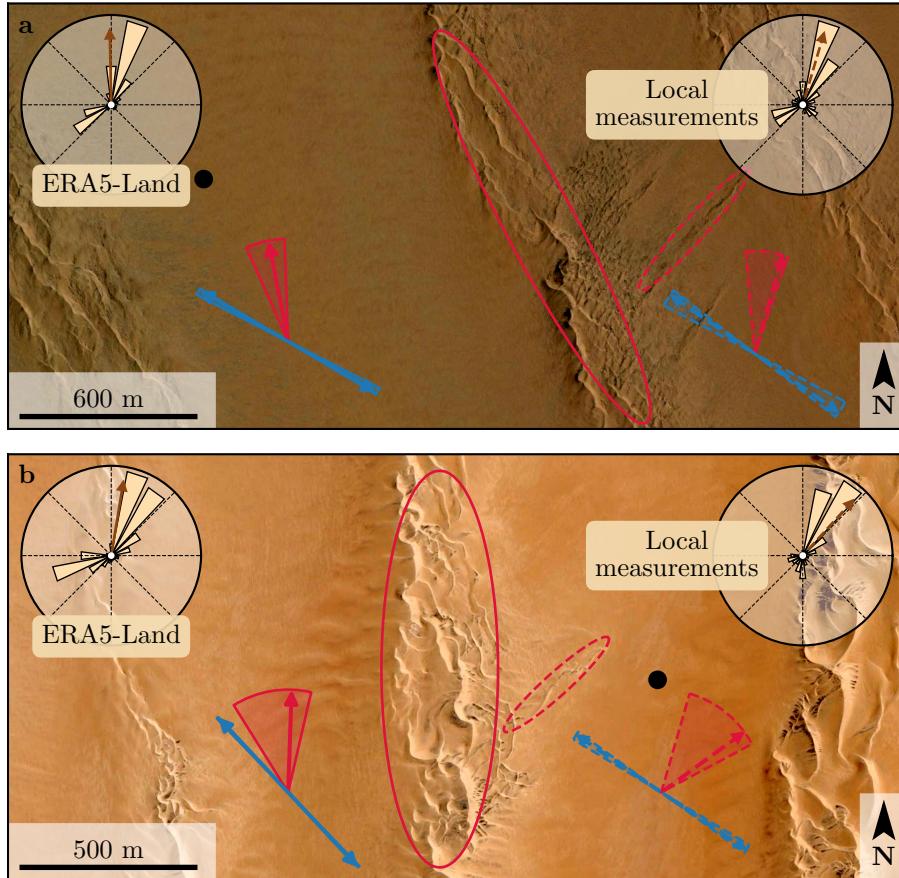
$$\delta_u = \frac{u_*^{\text{ERA5-Land}} - u_*^{\text{Local mes.}}}{u_*^{\text{ERA5-Land}}}. \quad (2)$$

These two quantities are represented as maps in the plane ( $\mathcal{F}$ ,  $kH$ ) (Fig. 8a,b), and one can clearly identify different regions in these graphs. Small wind disturbances (small  $\delta_\theta$  and  $\delta_u$ ) are located in the top-right part of the diagrams, corresponding to a regime with low-interaction as well as low-confinement ( $kH$  and  $\mathcal{F}$  large enough, Fig. 7d). Lower values of  $kH$  (stronger interaction) or of Froude number (stronger confinement) both lead to an increase in wind disturbances, both in terms of orientation and velocity. Below a crossover value  $kH \simeq 0.3$ , wind disturbance is less sensitive to the  $\mathcal{F}$ -value. This is probably due to enhanced non-linear effects linked to flow modulation by the obstacle when confinement is strong (e.g. wakes and flow recirculations). The Froude number also controls a transition from damped to amplified wind velocities in the interdune, with a crossover around  $\mathcal{F} \simeq 0.4$  (Fig. 8b). Such an amplification is rather unexpected. Checking the occurrence of the corresponding data, it appears that these amplifications are associated with the southerly sea breeze, and occur dominantly during the October-March period, when the other easterly wind is not present (Online Resource Fig. S12a–b). Furthermore, they occur less frequently during the afternoon, and more frequently at the end of the day (Online Resource Fig. S12c). This effect may be linked to a change in the flow behaviour in the lee side of the obstacles but further measurements are needed in order to assess the different possibilities (Baines 1995; Vosper 2004).

As the hydrodynamic roughness  $z_0$  determine the magnitude of wind shear velocities, Froude number  $\mathcal{F}$  and relative velocity modulation  $\delta_u$ , it is important to discuss the sensitivity of the results to the  $z_0$ -values chosen for both the ERA5-Land and the field data (see Online Resource section 4). Other quantities associated with wind direction are independent of this choice. Considering the possible range of realistic roughness values, the uncertainty on velocities estimated using the law of the wall is at most 30 %. A similar maximum uncertainty applies to the Froude number. This uncertainty also propagates to  $\delta_u$ , for which Figure S14 shows that the choice of roughness has little influence on its temporal variations, even if it can induce a global increase or decrease of its values. Hence, the choice of the  $z_0$ -values will not qualitatively affect the overall aspect of the regime diagram presented in Figure 8b. It may only change the value of  $\delta_u$  for which the transition between regimes is observed (dashed green lines in Figure 8b). Our conclusions are thus robust with respect to the somewhat arbitrary choice of the hydrodynamic roughness values.

#### 4 Discussion and conclusion

The feedback of the giant dunes on the wind flow has important implications for smaller scales bedforms. As illustrated in Fig. 9, small linear dunes ( $\sim 50$  m wide) are often present in the 1–2 km interdune between giant linear dunes in the Namib Sand Sea (Livingstone et al. 2010). These smaller dunes do not exhibit the same orientation as the large ones, and are sometimes named ‘crossing dunes’ (Chandler et al. 2022). Whilst differences between large and small



**Fig. 9** Implications for smaller scale patterns in (a) the South Sand Sea and (b) North Sand Sea. The ellipses indicate the different types of elongating dunes, at large (plain line) and small (dashed line) scales. Dune orientation are predicted using the model of Courrech du Pont et al. (2014) from the sand flux angular distributions, shown here (roses) along with the resultant transport direction (brown arrow) for typical values (grain size  $180 \mu\text{m}$ , flux-up ratio of 1.6). Code for corresponding arrows: use of ERA5-Land data (plain line), use of local measurements (dashed line), prediction for the bed instability growth mechanism (blue), prediction for the elongation growth mechanism (red). Wedges show the uncertainty on the orientation calculation when parameters are varied. The black dots indicate the location of the meteorological stations in the interdune. See Appendix 2 for additional details.

338 scale dune patterns are observed ubiquitously, they are usually attributed to  
 339 the presence of two different dune growth mechanisms, leading to two dif-  
 340 ferent dune patterns (orientations and/or morphologies) for the same wind  
 341 regime (Courrech du Pont et al. 2014; Runyon et al. 2017; Lü et al. 2017; Song  
 342 et al. 2019; Gadal et al. 2020; Hu et al. 2021). Here, however, our arguments  
 343 enable the development of differing orientations for the small and giant lin-  
 344 ear dunes governed by the same dune growth mechanism (elongating mode).  
 345 Figure 9 shows how the orientations for the small and giant dunes can be

346 derived from the locally measured and regionally predicted winds respectively  
 347 (red arrows in Fig. 9). These predictions require to specify the threshold of  
 348 aeolian sand transport. Importantly, its value (a shear velocity estimated at  
 349  $u_{th} \simeq 0.15 \text{ m s}^{-1}$ ) can be reached in periods during which winds are deflected  
 350 (recall that the stronger winds, responsible for most of the sediment transport  
 351 and associated dune morphodynamics, are not deflected – see Fig. 6). The  
 352 feedback of the giant dunes on the wind described in this study, through wind  
 353 deflection and attenuation, thus provides a potential explanation for the exis-  
 354 tence of these small linear dunes elongating across the interdune, a dynamic  
 355 which has remained unresolved to date. These crossing dunes could provide  
 356 additional constraints for the inference of local winds from bedforms, similarly  
 357 to that currently performed on Mars using ripple orientations (Liu and Zim-  
 358 belman 2015; Hood et al. 2021). Further work is needed to investigate these  
 359 processes in more detail, including measurements of sediment transport and  
 360 flow on the top of dunes.

361 This study presents the evidence that wind flow patterns around giant  
 362 dunes are influenced by the atmospheric boundary layer, particularly during  
 363 nocturnal conditions. However, we do not address here the question of the  
 364 limitation of their pattern coarsening, and leave open the debate as to whether  
 365 the size of giant dunes is controlled by the depth of this layer (Andreotti et al.  
 366 2009), in contrast to sediment supply limited and ever-slower growth with  
 367 size (Werner and Kocurek 1999; Gunn et al. 2021a). More field evidence is  
 368 definitively needed from additional dune fields, but this mechanism would  
 369 allow for the inference of the ABL depth from giant bedform wavelengths  
 370 where measurements are not feasible or available, such as Titan (Lorenz et al.  
 371 2010).

372 To conclude on conditions under which the ERA5-Land reanalysis data can  
 373 reliably be used to study dune morphodynamics, we summarise the compari-  
 374 son of local (direct measurements) and regional (climate reanalysis) wind data  
 375 as follows. In flat areas, the agreement between the two confirms the ability of  
 376 the ERA5-Land climate reanalysis to predict the wind regime down to scales  
 377  $\sim 10 \text{ km}$ , i.e the model grid. When smaller scale topographies are present  
 378 (giant dunes in our case), locally measured winds can significantly differ from  
 379 the regionally predicted ones. This is the case when the disturbances induced  
 380 by the dunes interact with the lower part of the ABL vertical structure, which  
 381 presents circadian variations. During the day, when the capping layer is typi-  
 382 cally high, this interaction is small, and the ERA5-Land predictions are also  
 383 quantitatively consistent with the local data. During the night, however, the  
 384 presence of a shallow atmospheric boundary layer induces a strong confine-  
 385 ment of the flow, and is associated with large wind deflection by the dunes.  
 386 Importantly, we find that this effect can be counterbalanced for large wind  
 387 velocities, which are capable of deforming the capping layer, thus decreasing  
 388 the influence of the confinement.

389 The theoretical computation of the wind disturbances induced by sinu-  
 390 soidal ridges under flow confinement has been performed in the linear limit  
 391 (Andreotti et al. 2009, 2012), i.e. when the aspect ratio of these ridges is small

( $k\xi_0 \ll 1$ ). These models are able to qualitatively reproduce the observed wind deflection (Appendix 1, Online Resource Figs. S11 and S13), and thus provide the physical support for the interpretation we propose here based on hydrodynamic regimes. However, these models cannot quantitatively predict the magnitude of our observations, probably due to the presence of expected non-linearities in high confinement situations linked to strong flow modulations. Besides, these linear calculations only predict wind attenuation in the interdune, in contrast with the observed enhanced velocities associated with particular evening winds from the South during the period October-March (Online Resource Fig. S12). Some other models predict different spatial flow structures in response to a modulated topography, such as lee waves and rotors (Baines 1995; Vosper 2004). However, our measurements are located at a single point in the interdune, and we are thus unable to explore these types of responses. Data at different places along and across the ridges are needed to investigate and possibly map such flow structures, and for further comparisons with the models.

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All data used in this study can be found in Gadali et al. (2022). Note that it contains modified Copernicus Climate Change Service Information (2021). Neither the European Commission nor ECMWF is responsible for any use that may be made of the Copernicus Information or Data it contains. Documented codes used in this study to analyse this data are available at <https://github.com/Cgadali/GiantDunes> (will be made public upon acceptance of this manuscript for publication).

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#### Appendix 1: Linear theory of wind response to topographic perturbation

Following the work of Fourrière et al. (2010), Andreotti et al. (2012) and Andreotti et al. (2009), we briefly describe in this appendix the framework for the linear response of a turbulent flow to a topographic perturbation of

<sup>439</sup> small aspect ratio. As a general bed elevation can be decomposed into Fourier  
<sup>440</sup> modes, we focus here on a sinusoidal topography:

$$\xi = \xi_0 \cos [k (\cos(\alpha)y - \sin(\alpha)x)], \quad (3)$$

<sup>441</sup> which is also a good approximation for the giant dunes observed in the North  
<sup>442</sup> Sand Sea and South Sand Sea Station (Fig. 2 and Online Resource Fig. S4).  
<sup>443</sup> Here,  $x$  and  $y$  are the streamwise and spanwise coordinates,  $k = 2\pi/\lambda$  the  
<sup>444</sup> wavenumber of the sinusoidal perturbation,  $\alpha$  its crest orientation with respect  
<sup>445</sup> to the  $x$ -direction (anticlockwise) and  $\xi_0$  its amplitude. The two components  
<sup>446</sup> of the basal shear stress  $\tau = \rho_0 u_* \mathbf{u}_*$ , constant in the flat bottom reference  
<sup>447</sup> case, can then be generically written as:

$$\tau_x = \tau_0 \left( 1 + k\xi_0 \sqrt{\mathcal{A}_x^2 + \mathcal{B}_x^2} \cos [k (\cos(\alpha)y - \sin(\alpha)x) + \phi_x] \right), \quad (4)$$

$$\tau_y = \tau_0 k\xi_0 \sqrt{\mathcal{A}_y^2 + \mathcal{B}_y^2} \cos [k (\cos(\alpha)y - \sin(\alpha)x) + \phi_y], \quad (5)$$

<sup>448</sup> where  $\tau_0$  is the reference basal shear stress on a flat bed. We have defined  
<sup>449</sup> the phase  $\phi_{x,y} = \tan^{-1}(\mathcal{B}_{x,y}/\mathcal{A}_{x,y})$  from the in-phase and in-quadrature hy-  
<sup>450</sup> drodynamical coefficients  $\mathcal{A}_{x,y}$  and  $\mathcal{B}_{x,y}$ . They are functions of  $k$  and of the  
<sup>451</sup> flow conditions, i.e the bottom roughness, the vertical flow structure and the  
<sup>452</sup> incident flow direction, and the theoretical framework developed in the above  
<sup>453</sup> cited papers proposes methods to compute them in the linear regime.

<sup>454</sup> Following Andreotti et al. (2012), the effect of the incident wind direction  
<sup>455</sup> can be approximated by the following expressions:

$$\mathcal{A}_x = \mathcal{A}_0 \sin^2 \alpha, \quad (6)$$

$$\mathcal{B}_x = \mathcal{B}_0 \sin^2 \alpha, \quad (7)$$

$$\mathcal{A}_y = -\frac{1}{2} \mathcal{A}_0 \cos \alpha \sin \alpha, \quad (8)$$

$$\mathcal{B}_y = -\frac{1}{2} \mathcal{B}_0 \cos \alpha \sin \alpha, \quad (9)$$

<sup>456</sup> where  $\mathcal{A}_0$  and  $\mathcal{B}_0$  are now two coefficients independent of the dune orienta-  
<sup>457</sup> tion  $\alpha$ , corresponding to the transverse case ( $\alpha = 90^\circ$ ). In the case of a fully  
<sup>458</sup> turbulent boundary layer capped by a stratified atmosphere, these coefficients  
<sup>459</sup> depend on  $kH$ ,  $kz_0$ ,  $\mathcal{F}$  and  $\mathcal{F}_I$  (Andreotti et al. 2009). For their computation,  
<sup>460</sup> we assume here a constant hydrodynamic roughness  $z_0 \simeq 1$  mm (Online Re-  
<sup>461</sup> source section 1). For the considered giant dunes, this leads to  $kz_0 \simeq 2 \cdot 10^{-6}$ ,  
<sup>462</sup> as their wavelength is  $\lambda \simeq 2.4$  km (or  $k \simeq 2 \cdot 10^{-3}$  m $^{-1}$ ). Values of  $z_0$  extracted  
<sup>463</sup> from field data indeed typically fall between 0.1 mm and 10 mm (Sherman and  
<sup>464</sup> Farrell 2008; Field and Pelletier 2018). Importantly,  $\mathcal{A}_0$  and  $\mathcal{B}_0$  do not vary  
<sup>465</sup> much in the corresponding range of  $kz_0$  (Fourrière et al. 2010), and the results  
<sup>466</sup> presented here are robust with respect to this choice.

<sup>467</sup> With capping layer height and Froude numbers computed from the ERA5-  
<sup>468</sup> Land time series, the corresponding  $\mathcal{A}_0$  and  $\mathcal{B}_0$  can be deduced, as displayed  
<sup>469</sup> in Online Resource Fig. S13. Interestingly, it shows similar regimes as in the

470 diagrams of Fig. 8 and Online Resource Fig. S11a,b, supporting the underlying  
 471 physics. However, the agreement is qualitative only. Further, the linearity  
 472 assumption of the theoretical framework requires  $(|\tau| - \tau_0)/\tau_0 \ll 1$ , which  
 473 translates into  $k\xi\sqrt{\mathcal{A}_0^2 + \mathcal{B}_0^2} \ll 1$ . In our case, the giant dune morphology  
 474 gives  $k\xi_0 \simeq 0.1$ , which means that one quits the regime of validity of the  
 475 linear theory when the coefficient modulus  $\sqrt{\mathcal{A}_0^2 + \mathcal{B}_0^2}$  becomes larger than a  
 476 few units. In accordance with the theoretical expectations, these coefficients  
 477 present values on the order of unity ( $\mathcal{A}_0 \simeq 3$  and  $\mathcal{B}_0 \simeq 1$ ) in unconfined sit-  
 478 uations (Claudin et al. 2013; Lü et al. 2021). In contrast and as illustrated  
 479 in Online Resource Fig. S13a,b, larger values are predicted in case of strong  
 480 confinement, which does not allow us to proceed to further quantitative com-  
 481 parison with the data.

482 Finally, the linear model is also able to reproduce the enhancement of the  
 483 flow deflection over the sinusoidal ridges when  $\sqrt{\mathcal{A}_0^2 + \mathcal{B}_0^2}$  is increased (Online  
 484 Resource Fig. S13). Here, using  $k\xi_0 \simeq 0.1$  to be representative of the amplitude  
 485 of the giant dunes at the North Sand Sea station, the coefficient modulus is  
 486 bounded to 10.

## 487 Appendix 2: Sediment transport and dune morphodynamics

488 We summarise in this appendix the sediment transport and dune morphody-  
 489 namics theoretical framework leading to the prediction of sand fluxes and dune  
 490 orientations from wind data.

491 *Sediment transport* — The prediction of sand fluxes from wind data has been  
 492 a long standing issue in aeolian geomorphological studies (Fryberger and Dean  
 493 1979; Pearce and Walker 2005; Sherman and Li 2012; Shen et al. 2019). Based  
 494 on laboratory studies in wind tunnels (Rasmussen et al. 1996; Iversen and  
 495 Rasmussen 1999; Creyssels et al. 2009; Ho et al. 2011), as well as physical  
 496 considerations (Ungar and Haff 1987; Andreotti 2004; Durán et al. 2011; Pähzt  
 497 and Durán 2020), it has been shown that the steady saturated saltation flux  
 498 over a flat sand bed depends linearly on the shear stress:

$$\frac{q_{\text{sat}}}{Q} = \Omega \sqrt{\Theta_{\text{th}}} (\Theta - \Theta_{\text{th}}), \quad (10)$$

499 where  $\Omega$  is a proportionality constant,  $Q = d\sqrt{(\rho_s - \rho_0)gd/\rho_0}$  is a character-  
 500 istic flux,  $\Theta = \rho_0 u_*^2 / (\rho_s - \rho_0)gd$  the Shields number, and  $\Theta_{\text{th}}$  its threshold  
 501 value below which saltation vanishes.  $\rho_s = 2.6 \text{ g cm}^{-3}$  and  $d = 180 \mu\text{m}$  are  
 502 the grain density and diameter, and  $g$  is the gravitational acceleration. The  
 503 shear velocity, and consequently the Shields number as well as the sediment  
 504 flux, are time dependent.

505 Recently, Pähzt and Durán (2020) suggested an additional quadratic term  
 506 in Shields to account for grain-grain interactions within the transport layer at  
 507 strong wind velocities:

$$\frac{q_{\text{sat}, t}}{Q} = \frac{2\sqrt{\Theta_{\text{th}}}}{\kappa\mu} (\Theta - \Theta_{\text{th}}) \left( 1 + \frac{C_M}{\mu} [\Theta - \Theta_{\text{th}}] \right), \quad (11)$$

508 where  $\kappa = 0.4$  is the von Kármán constant,  $C_M \simeq 1.7$  a constant and  $\mu \simeq 0.6$  is  
 509 a friction coefficient, taken to be the avalanche slope of the granular material.  
 510 The fit of this law to the experimental data of Creyssels et al. (2009) and Ho  
 511 et al. (2011) gives  $\Theta_{\text{th}} = 0.0035$ . The fit of Eq. 10 on these same data similarly  
 512 gives  $\Omega \simeq 8$  and  $\Theta_{\text{th}} = 0.005$ . The sand flux angular distributions and the  
 513 dune orientations in Fig. 9 are calculated using this law (11). We have checked  
 514 that using the ordinary linear relationship (10) instead does not change the  
 515 predicted dune orientations by more than a few degrees.

516 *Dune orientations* — Dune orientations are predicted with the dimensional  
 517 model of Courrech du Pont et al. (2014), from the sand flux time series com-  
 518 puted with the above transport law. Two orientations are possible depending  
 519 on the mechanism dominating the dune growth: elongation or bed instabil-  
 520 ity. The orientation  $\alpha$  corresponding the bed instability is then the one that  
 521 maximises the following growth rate (Rubin and Hunter 1987):

$$\sigma \propto \frac{1}{H_d W_d T} \int_0^T q_{\text{crest}} |\sin(\theta - \alpha)| dt, \quad (12)$$

522 where  $\theta$  is the wind orientation measured with respect to the same reference  
 523 as  $\alpha$ , and  $H_d$  and  $W_d$  are dimensional constants respectively representing the  
 524 dune height and width. The integral runs over a time  $T$ , which must be repre-  
 525 sentative of the characteristic period of the wind regime. The flux at the crest  
 526 is expressed as:

$$q_{\text{crest}} = q_{\text{sat}} [1 + \gamma |\sin(\theta - \alpha)|], \quad (13)$$

527 where the flux-up ratio  $\gamma$  has been calibrated to 1.6 using field studies, under-  
 528 water laboratory experiments and numerical simulations. Predictions of the  
 529 linear analysis of Gadal et al. (2019) and Delorme et al. (2020) give similar  
 530 results.

531 Similarly, the dune orientation corresponding to the elongation mechanism  
 532 is the one that verifies:

$$\tan(\alpha) = \frac{\langle q_{\text{crest}}(\alpha) e_\theta \rangle \cdot e_{WE}}{\langle q_{\text{crest}}(\alpha) e_\theta \rangle \cdot e_{SN}}, \quad (14)$$

533 where  $\langle \cdot \rangle$  denotes a vectorial time average. The unitary vectors  $e_{WE}$ ,  $e_{SN}$  and  
 534  $e_\theta$  are in the West-East, South-North and wind directions, respectively.

535 The resulting computed dune orientations, blue and red arrows in Fig. 9,  
 536 then depend on a certain number of parameters (grain properties, flux-up ratio,  
 537 etc.), for which we take typical values for aeolian sandy deserts. Due to the lack  
 538 of measurements in the studied places, some uncertainties can be expected. We  
 539 therefore run a sensitivity test by calculating the dune orientations for grain  
 540 diameters ranging from  $100 \mu\text{m}$  to  $400 \mu\text{m}$  and for a speed-up ratio between  
 541 0.1 and 10 (wedges in Fig. 9).

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888      **Local wind regime induced by giant linear dunes**  
 889      — Supplementary Material —

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896    **1. Shear velocity and calibration of the hydrodynamical roughness**

897    As the regionally predicted and locally measured velocities are available at  
 898    different heights, we can not compare them directly. We therefore convert  
 899    all velocities into shear velocities  $u_*$ , characteristic the turbulent logarithmic  
 900    velocity profile (Spalding 1961; Stull 1988):

$$\frac{u(z)}{u_*} = \frac{1}{\kappa} \ln \left( \frac{z}{z_0} \right), \quad (15)$$

901    where  $z$  is the vertical coordinate,  $\kappa = 0.4$  the von Kármán constant and  $z_0$  the  
 902    hydrodynamic roughness. Note that, strickly speaking, this logarithmic profile  
 903    is valid for a neutrally stratified ABL only. Vertical density gradients occurring  
 904    in other conditions may thus induce large discrepancies (Monin and Obukhov  
 905    1954; Garratt 1994; Dyer 1974). However, as our wind measurements are in  
 906    the flow region close enough to the surface, where these effects are negligible,  
 907    this logarithmic wind profile remains a fairly good approximation in all conditions  
 908    (Gunn et al. 2021b). Several measurements of hydrodynamic roughnesses  
 909    are available (Raupach 1992; Bauer et al. 1992; Brown et al. 2008; Nield et al.  
 910    2014). In the absence of sediment transport, it is governed by the geometric  
 911    features of the bed (Flack and Schultz 2010; Pelletier and Field 2016). When  
 912    aeolian saltation occurs, it is rather controlled by the altitude of Bagnold's  
 913    focal point (Durán et al. 2011; Valance et al. 2015), which depends on the  
 914    wind velocity and grain properties (Sherman and Farrell 2008; Zhang et al.  
 915    2016; Field and Pelletier 2018). Whether associated with geometric features  
 916    or with sediment transport, its typical order of magnitude is the millimetre  
 917    scale on sandy surfaces.

918    We do not have precise velocity vertical profiles to be able to deduce an  
 919    accurate value of  $z_0$  in the various environments of the meteorological stations  
 920    (vegetated, arid, sandy). Our approach is to rather select the hydrodynamic  
 921    roughness which allows for the best possible matching between the regionally  
 922    predicted and locally measured winds, i.e. minimising the relative difference  $\delta$   
 923    between the wind vectors of both datasets:

$$\delta = \frac{\sqrt{\langle \| \mathbf{u}_{*,\text{era}} - \mathbf{u}_{*,\text{station}} \|^2 \rangle}}{\sqrt{\langle \| \mathbf{u}_{*,\text{era}} \|^2 \rangle \langle \| \mathbf{u}_{*,\text{station}} \|^2 \rangle}}, \quad (16)$$

where  $\langle \cdot \rangle$  denotes time average. This parameter is computed for values of  $z_0$  in ERA5-Land analysis ranging from  $10^{-5}$  m to  $10^{-2}$  m for the four different stations. Note that for the North Sand Sea and South Sand Sea stations, where the giant dunes feedback presumably affect the wind, we take into account the non-deflected winds only in the calculation of  $\delta$  (with a  $15^\circ$  tolerance).

As shown in Online Resource Fig. S3, the minimum values of  $\delta$  in the space  $(z_0^{\text{ERA5Land}}, z_0^{\text{local}})$  form a line. We thus set the roughness in the ERA5-Land analysis to the typical value  $z_0 = 10^{-3}$  m, and deduce the corresponding ones for the local stations. It leads to 2.7, 0.8, 0.1 and 0.5 mm for the Etosha West, North Sand Sea, Huab and South Sand Sea stations, respectively. Importantly, this approach somewhat impacts the calculation of the shear velocities, but not that of the wind directions. As such, most of our conclusions are independent of such a choice. However, it may affect the magnitude of the wind velocity attenuation/amplification in flow confinement situations.

## 2. Computation of the ABL characteristics

The estimation of the non-dimensional numbers associated with the ABL requires the computation of representative meteorological quantities. In arid areas, the vertical structure of the atmosphere can be approximated by a well mixed convective boundary layer of height  $H$ , topped by the stratified free atmosphere (Stull 1988; Shao 2008). In this context, one usually introduces the virtual potential temperature  $T_{\text{vp}}$ , which is a constant  $T_0$  inside the boundary layer, and increases linearly in the FA (Online Resource Fig. S8a):

$$T_{\text{vp}}(z) = \begin{cases} T_0 & \text{for } z \leq H, \\ T_0 \left(1 + \frac{\Delta T_{\text{vp}}}{T_0} + \frac{N^2}{g}(z - H)\right) & \text{for } z \geq H, \end{cases} \quad (17)$$

where  $\Delta T_{\text{vp}}$  is the temperature discontinuity at the capping layer and  $N = \sqrt{g\partial_z T_{\text{vp}}/T_0}$  is the Brunt-Väisälä frequency, characteristic of the stratification. Note that, under the usual Boussinesq approximation, temperature and air density variations are simply related by  $\delta T_{\text{vp}}/T_0 \simeq -\delta\rho/\rho_0$  (see Online Resource of Andreotti et al. (2009)), so that  $N$  can equivalently be defined from the density gradient as next to (1).

The ERA5 dataset provides vertical profiles of the geopotential  $\phi$ , the actual temperature  $T$  and the specific humidity  $\eta$  at given pressure levels  $P$ . The vertical coordinate is then calculated as:

$$z = \frac{\phi R_t}{g R_t - \phi}, \quad (18)$$

where  $R_t = 6371229$  m is the reference Earth radius and  $g = 9.81$  m s $^{-2}$  is the gravitational acceleration. One also computes the virtual potential temperature as:

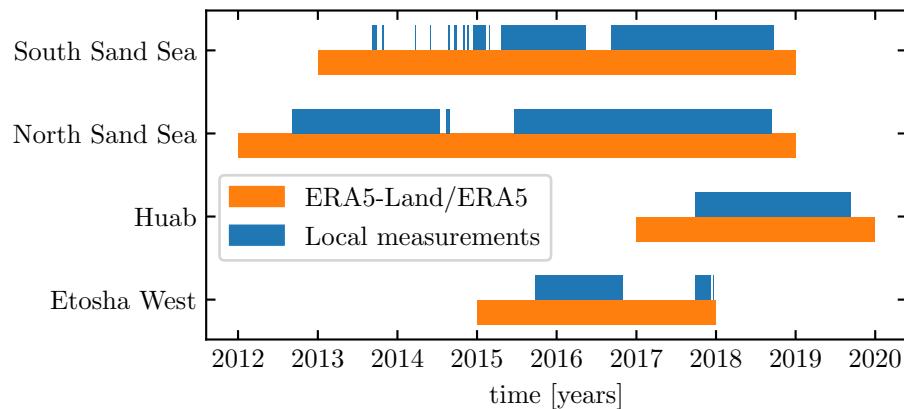
$$T_{\text{vp}} = T \left[1 + \left(\frac{M_d}{M_w} - 1\right) \eta\right] \left(\frac{P_0}{P}\right)^{R/C_p}, \quad (19)$$

where  $P_0 = 10^5$  Pa is the standard pressure,  $R = 8.31$  J/K is the ideal gas constant,  $C_p \simeq 29.1$  J/K is the air molar heat capacity, and  $M_w = 0.018$  kg/Mol and  $M_d = 0.029$  kg/Mol are the molecular masses of water and dry air respectively. The specific humidity is related to the vapour pressure  $p_w$  as

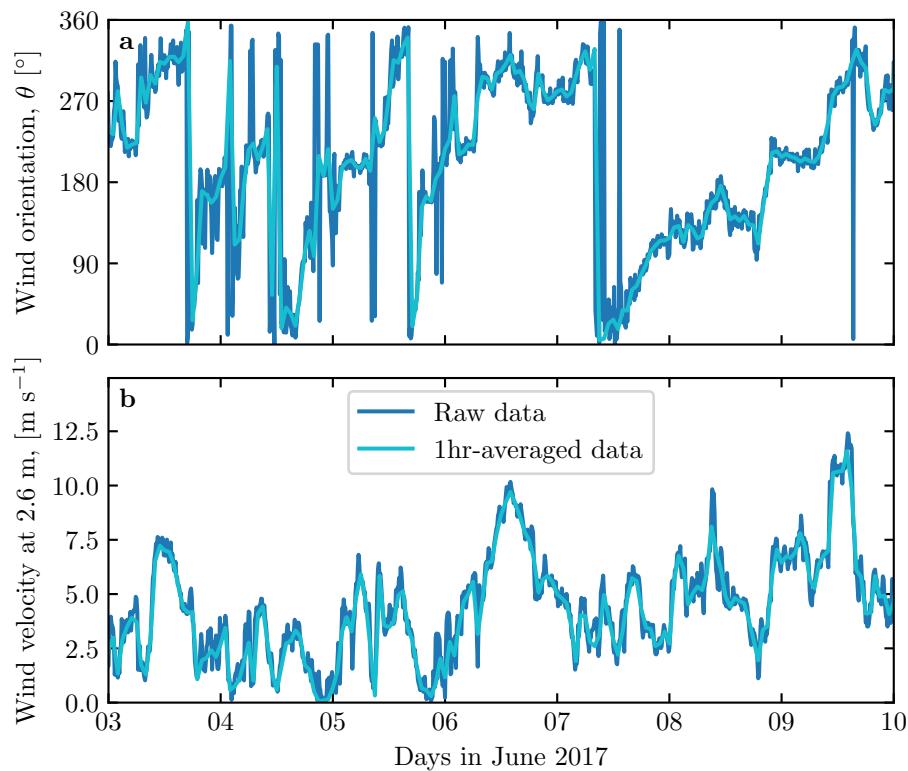
$$\eta = \frac{\frac{M_w}{M_d} p_w}{p - \left(1 - \frac{M_w}{M_d}\right) p_w}. \quad (20)$$

The ERA5 dataset also provides an estimate of the ABL depth  $H$ , based on the behaviour of the Richardson vertical profile. This dimensionless number is defined as the ratio of buoyancy and flow shear terms, and can be expressed as  $\text{Ri} = N^2 / (\partial_z u)^2$ . It vanishes in the lower well-mixed layer where  $T_{\text{vp}}$  is constant, and increases in the stratified FA. Following the method and calibration of Vogelegang and Holtlag (1996); Seidel et al. (2012), the value  $\text{Ri}(z) \simeq 0.25$  has been shown to be a good empirical criterion to give  $z \simeq H$  within a precision varying from 50% for the shallower ABL (e.g. at night) to 20% for situations of stronger convection.

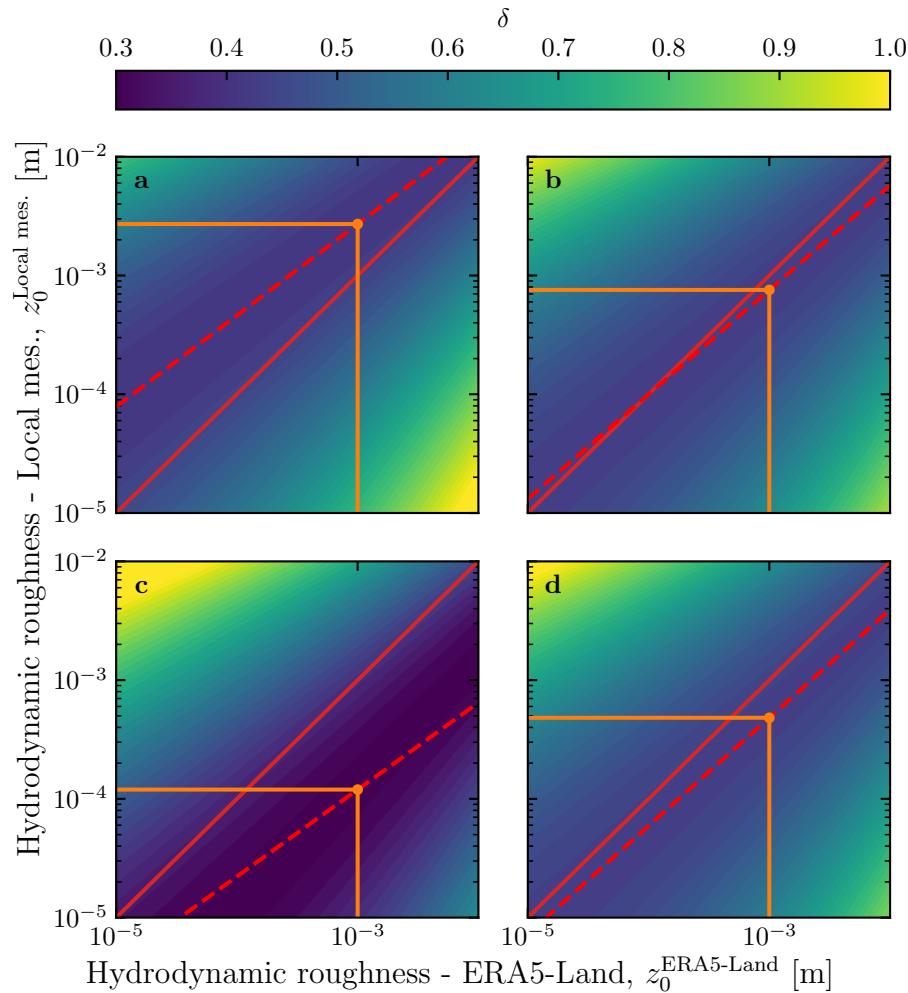
Examples of vertical profiles of the virtual potential temperature deduced from ERA5 are shown in Online Resource Fig. S8a. For each of them, an average temperature is computed below the ABL depth ( $z < H$ ), and a linear function is fitted above, allowing us to extract the temperature jump  $\Delta T_{\text{vp}}$ . Importantly, some profiles display a vertical structure that cannot be approximated by the simple form (17) used here (Online Resource Fig. S8b). In practice, we removed from the analysis all of those leading to the unphysical case  $\Delta T_{\text{vp}} < 0$ . We have noticed that these ‘ill-processed’ profiles dominantly occur in winter and are evenly spread across the hours of the day. Importantly, they represent  $\simeq 12\%$  of the data only (Online Resource Fig. S8c,d), and we are thus confident that this data treatment does not affect our conclusions.



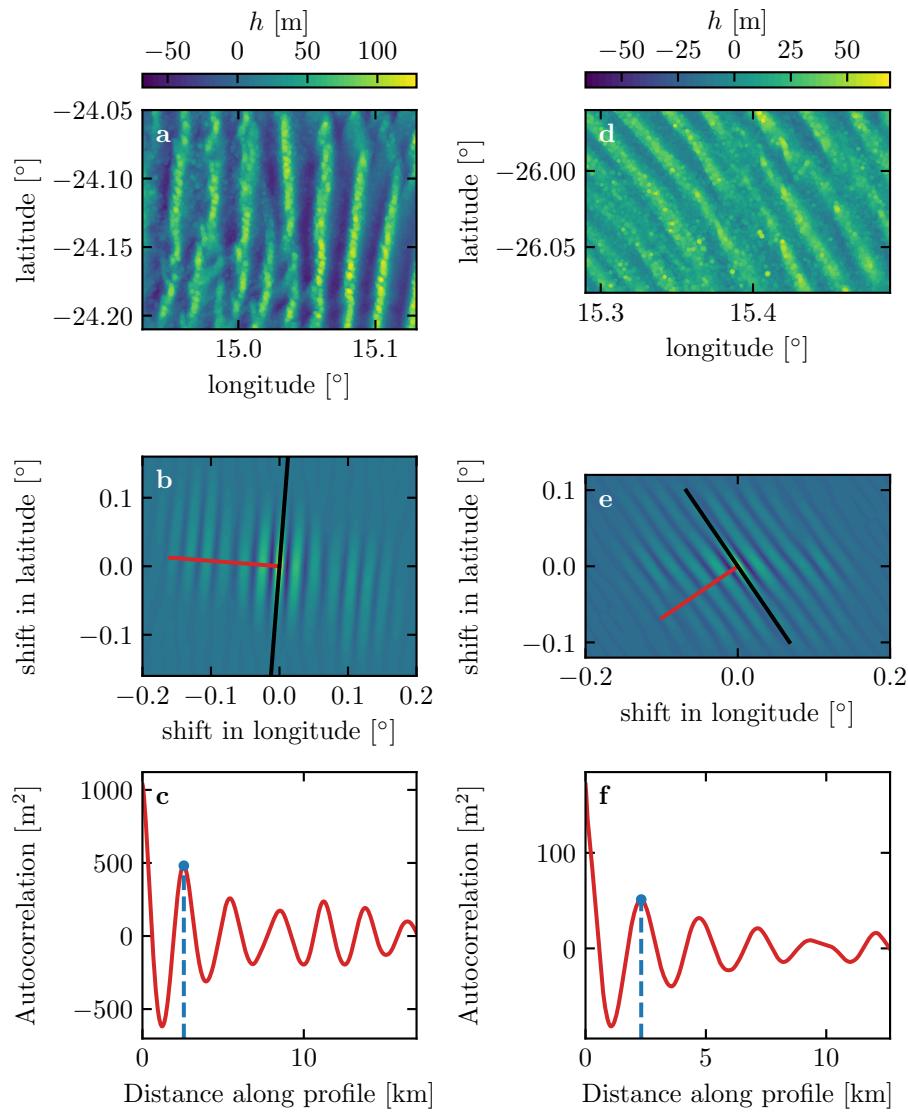
**Fig. S1** Gant chart representing the valid time steps for the two data sets, for all stations.



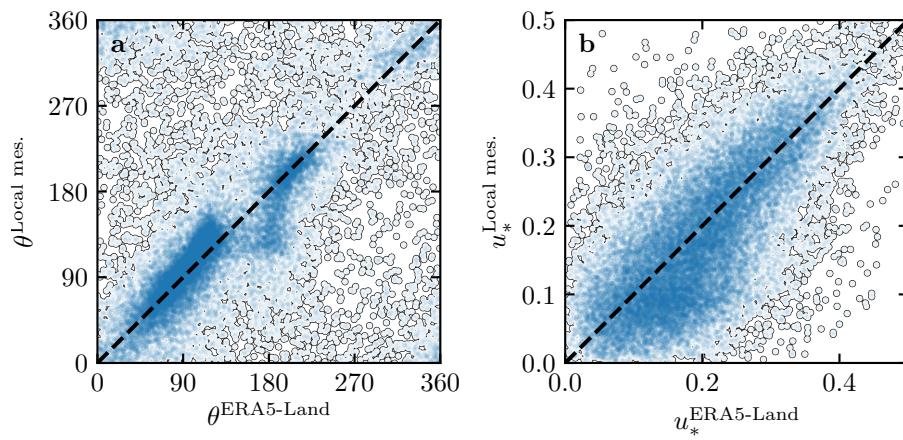
**Fig. S2** Local wind measurements: comparison between raw (blue) and hourly-averaged (light blue) data from South Sand Sea station. **a:** wind direction. **b:** wind velocity at height 2.6 m.



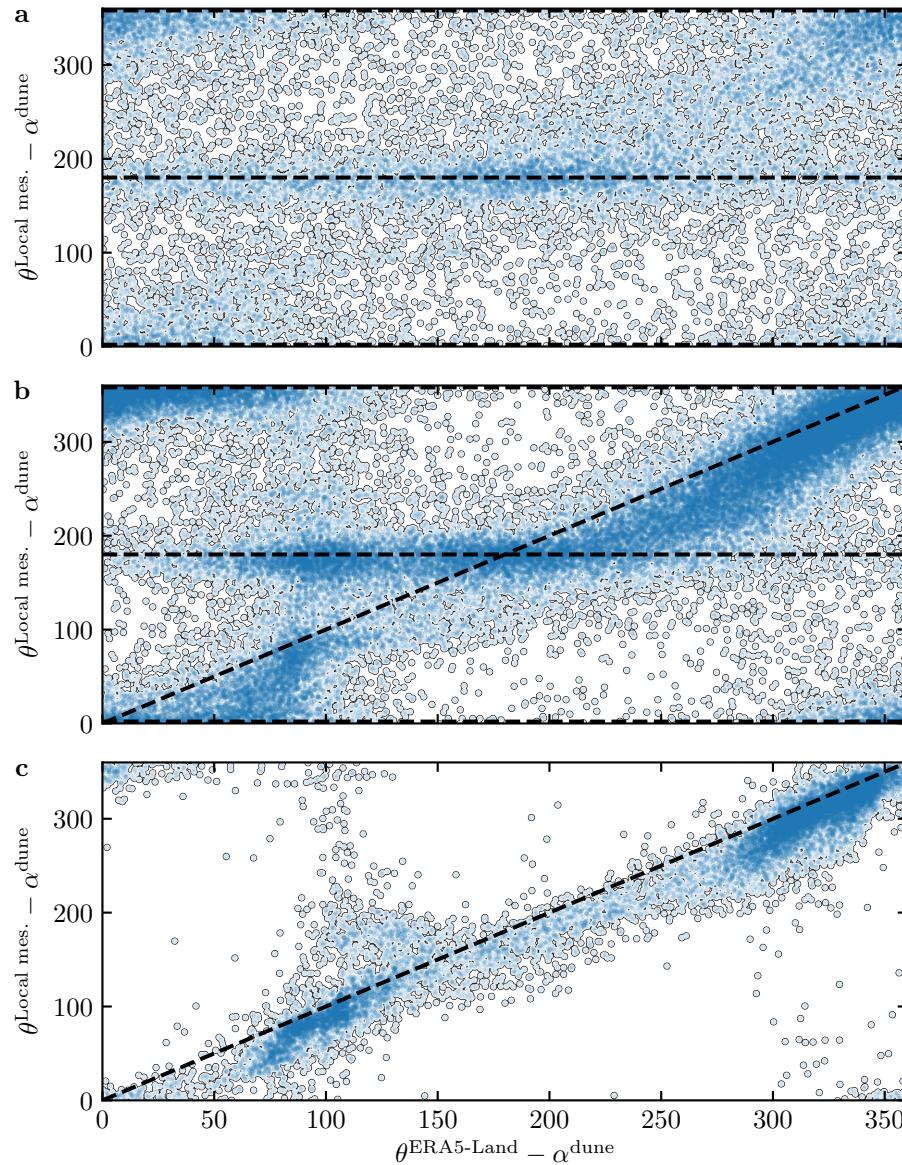
**Fig. S3** Calibration of hydrodynamic roughness. The parameter  $\delta$  (16) quantifying the difference between local and predicted winds is shown in color scale as a function of the hydrodynamic roughnesses chosen for the ERA5-Land and for local winds, for the (a) Etosha West, (b) North Sand Sea, (c) Huab and (d) South Sand Sea stations. The red dashed and plain lines shows the minima of  $\delta$  and the identity line, respectively. The orange lines and dots highlight the chosen hydrodynamic roughnesses for the local winds deduced from setting  $z_0^{\text{ERA5-Land}} = 1 \text{ mm}$ .



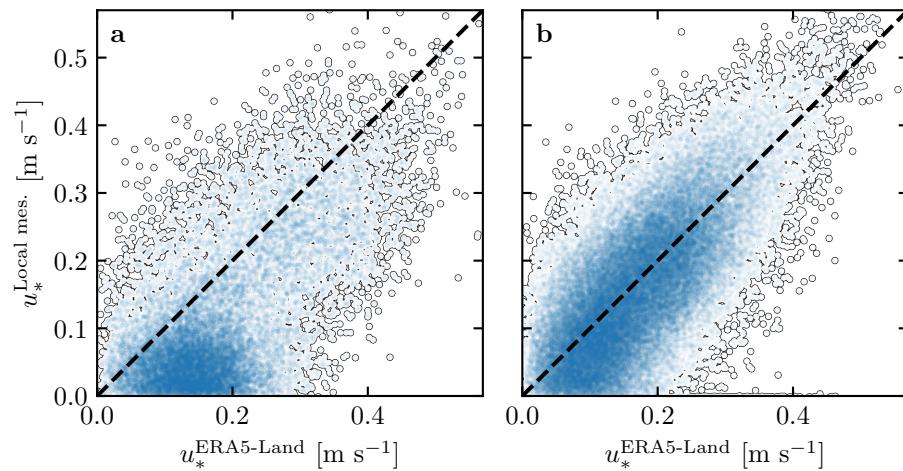
**Fig. S4** Analysis of the DEMs of the North Sand Sea (left column – panels **a**, **b**, **c**) and South Sand Sea (right column – panels **d**, **e**, **f**) stations. **a–d**: Bed elevation detrended by a fitted second order polynomial base-line. **b–e**: Autocorrelation matrix shown in color scale. The black line shows the detected dune orientation, and the red line represents the autocorrelation profile along which the dune wavelength is calculated, displayed in **c–f**. The blue lines and dots show the first peak of the autocorrelation profile, whose abscissa gives the characteristic wavelength of the dune pattern.



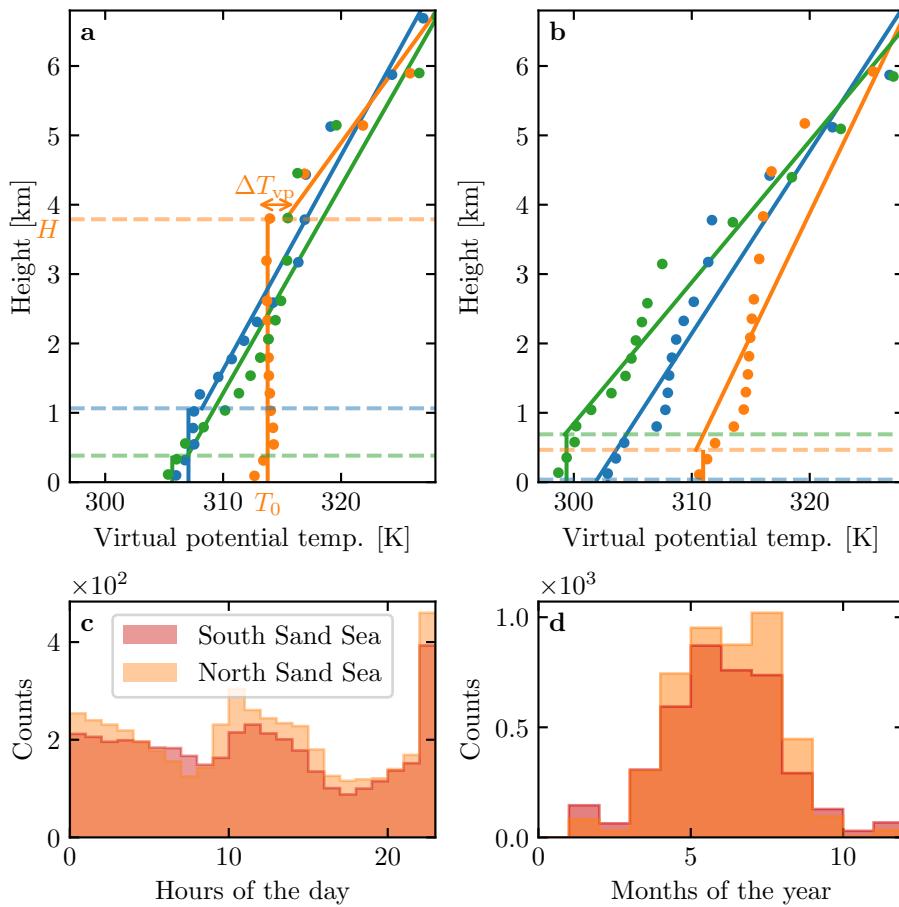
**Fig. S5** Statistical comparison of the wind orientation (a) and velocity (b) between the ERA5-Land dataset and the local measurements for the Huab and Etosha West stations. Data point clustering around identity lines (dashed and black) provide evidence for agreement of the two sets.



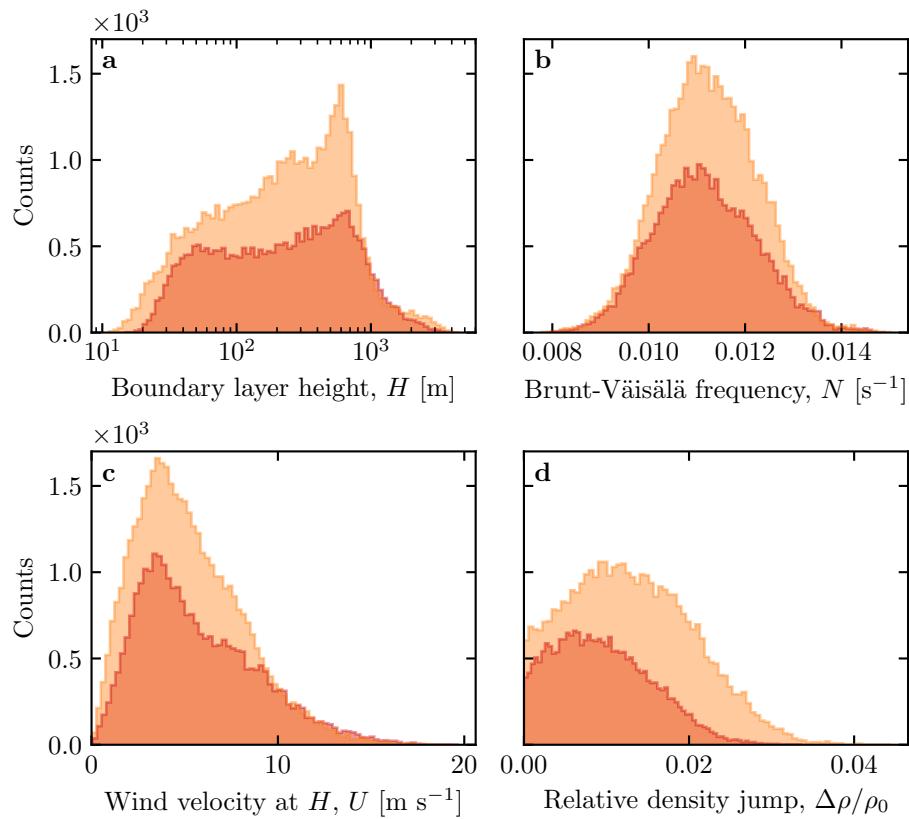
**Fig. S6** Statistical comparison of the wind orientation between the ERA5-Land dataset and the local measurements for the South Sand Sea and North Sand Sea stations, for different velocity ranges. **a:**  $u^{\text{ERA5-Land}}_* < 0.1 \text{ m s}^{-1}$ . **b:**  $0.1 < u^{\text{ERA5-Land}}_* \leq 0.25 \text{ m s}^{-1}$ . **c:**  $u^{\text{ERA5-Land}}_* \geq 0.25 \text{ m s}^{-1}$ . The measured dune orientations are subtracted to the wind orientation, which allows us to plot both stations on the same graph. Black dashed lines indicate locally measured orientations aligned with the dune crests (here  $0^\circ$ ,  $180^\circ$  and  $360^\circ$  – panels **a**, **b**), as well as the identity lines (panels **b**, **c**).



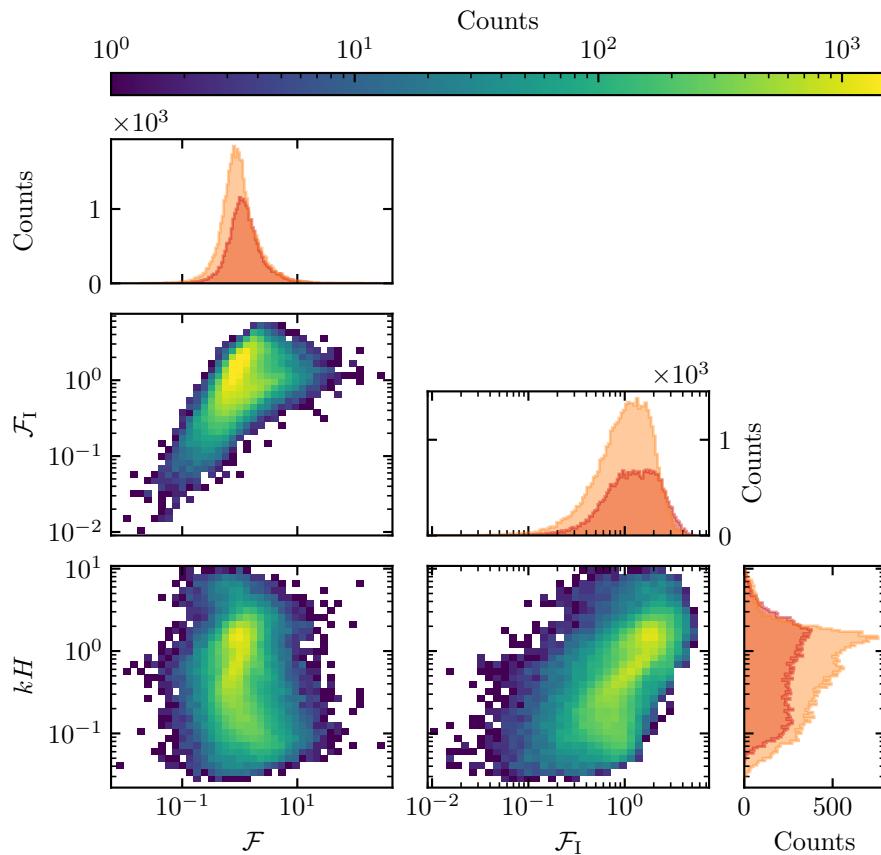
**Fig. S7** Statistical comparison of the wind velocity between the ERA5-Land dataset and the local measurements for the South Sand Sea and North Sand Sea stations. **a:** Nocturnal summer easterly wind. **b:** Diurnal southerly wind. Black dashed lines are identity lines. The angle ranges used to select diurnal and nocturnal summer winds are the same as those in Figs. 4 and Figs. 6 of the main article.



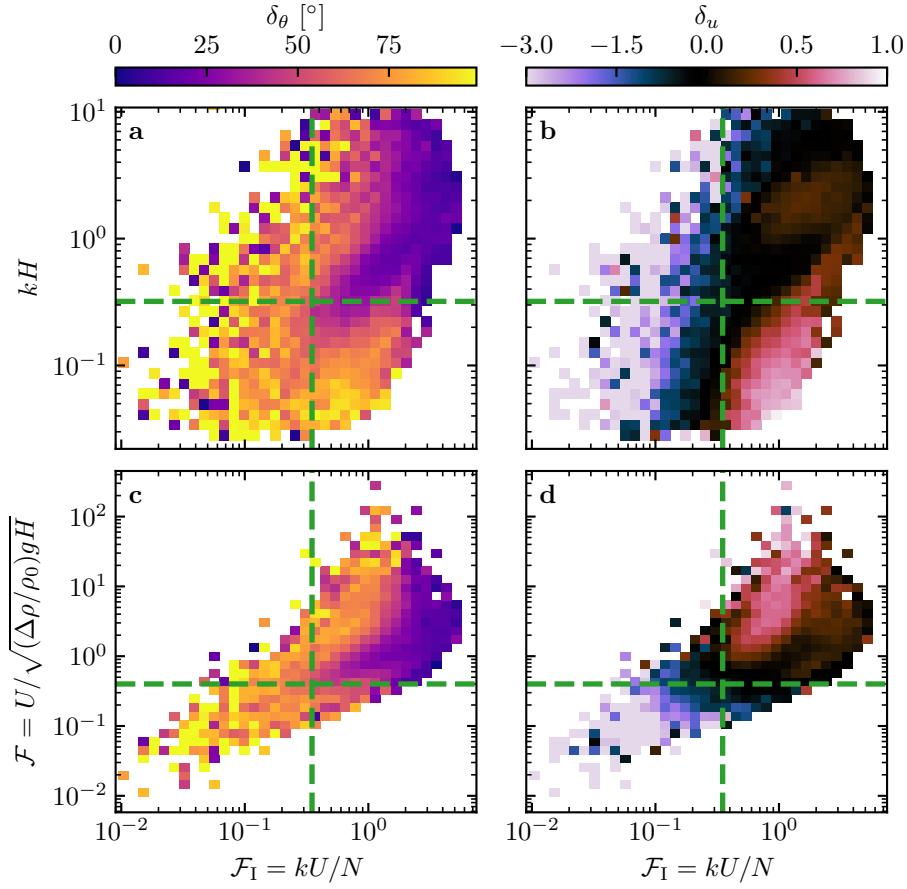
**Fig. S8** **a:** Vertical profiles of the virtual potential temperature at three different times (blue: 29/11/2012 - 11.00 UTC, orange: 21/03/2017 - 12.00 UTC, green: 21/03/2017 - 20.00 UTC) at the South Sand Sea station. Dots: data from the ERA5 reanalysis. Dashed lines: boundary layer height given by the ERA5 reanalysis. Plain lines: vertical (ABL) and linear (FA) fits to estimate the quantities displayed in Online Resource Fig. S9. **b:** Examples of ill-processed vertical profiles at three different times (blue: 2/12/2013 - 23.00 UTC, orange: 20/03/2017 - 00.00 UTC, green: 14/07/2017 - 14.00 UTC) at the South Sand Sea station. Distribution of ill-processed vertical profiles at South (orange) and North (light orange) Sand Sea station: hourly **(c)** and monthly **(d)** counts.



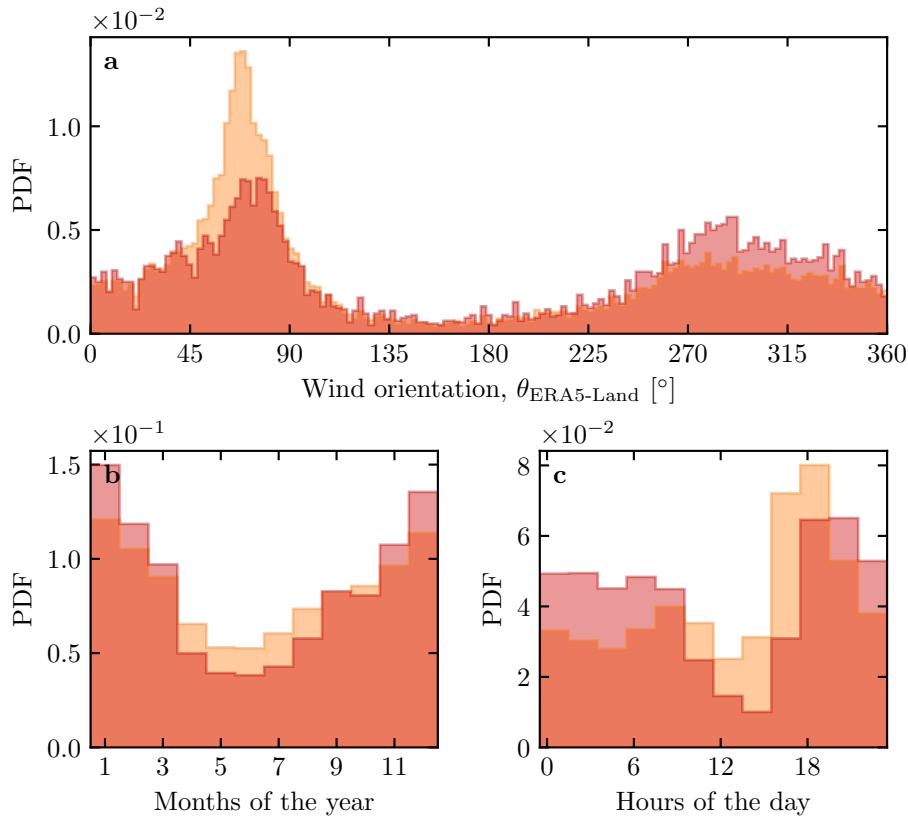
**Fig. S9** Distributions of the meteorological parameters resulting from the processing of the ERA5-Land data for the South Sand Sea (orange) and the North Sand Sea (light orange) stations.



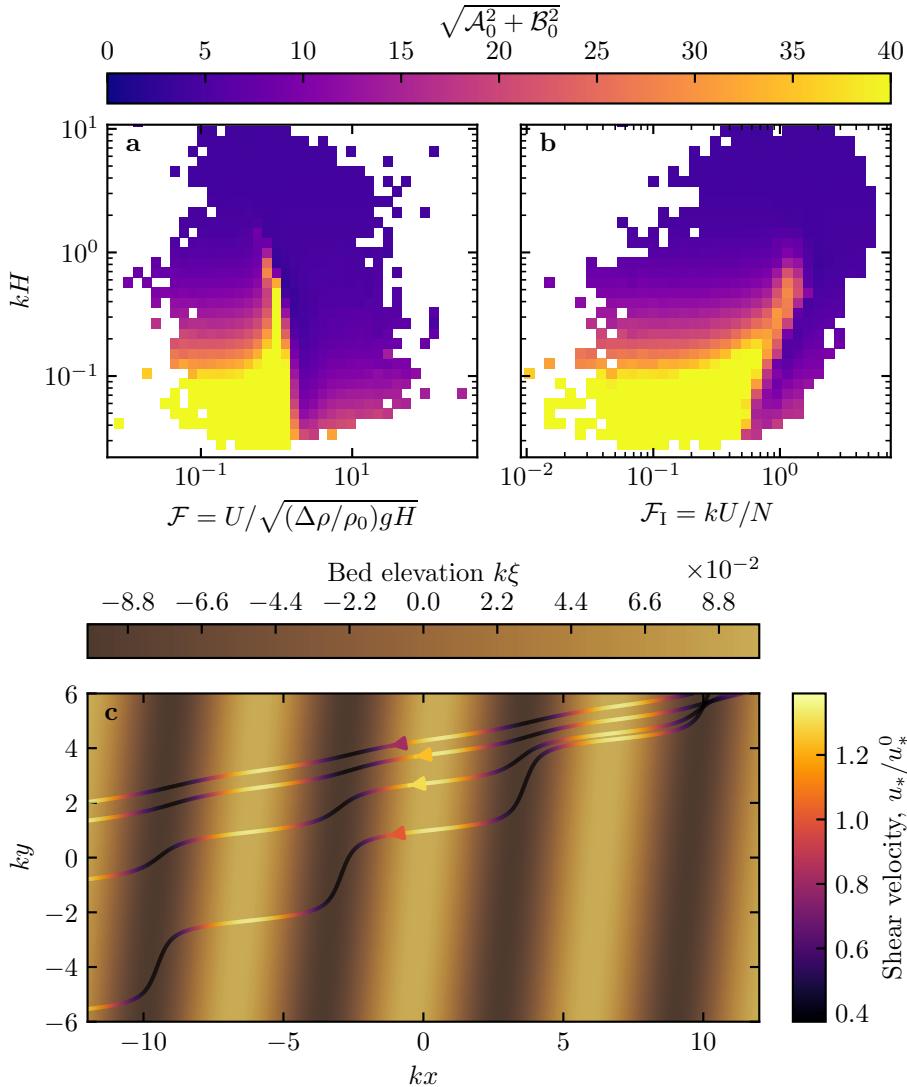
**Fig. S10** Non-dimensional parameters distributions. For the marginal distributions, the light orange corresponds to the South Sand Sea station, and the orange to the North Sand Sea station.



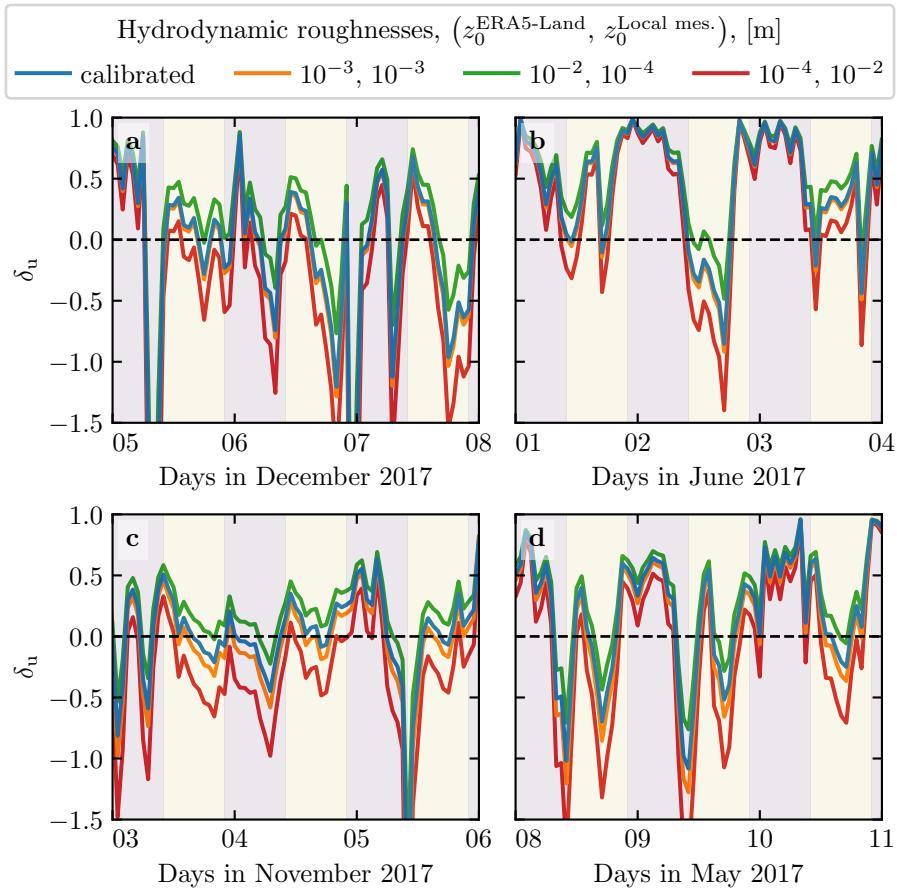
**Fig. S11** Regime diagrams of the wind deviation  $\delta_\theta$  and relative attenuation/amplification  $\delta_u$  in the spaces  $(\mathcal{F}_I, kH)$  and  $(\mathcal{F}_I, \mathcal{F})$ , containing the data from both the North Sand Sea and South Sand Sea stations. Green dashed lines empirically delimit the different regimes. The point density in each bin of the diagrams is shown in Online Resource Fig. S10 – 95% of the data occur in the range  $-1 < \delta u < 1$ . The similar regime diagrams in the space  $(\mathcal{F}, kH)$  are shown in Fig. 8.



**Fig. S12** Normalized distributions of amplified velocities for the North sand Sea (light orange:  $\delta_u < 0$ , orange:  $\delta_u < -0.5$ ). **a:** Angular distributions. **b:** Monthly distributions. **c:** Hourly distributions.



**Fig. S13** Computation of the flow disturbance with the linear model of Andreotti et al. (2009). **a–b:** Magnitude of the hydrodynamic coefficients  $A_0$  and  $B_0$ , calculated from the time series of the non-dimensional numbers corresponding to the ERA5-Land wind data and ERA5 data on vertical pressure levels. **c** Shear velocity streamlines over sinusoidal ridges of amplitude  $k\xi_0 = 0.1$  and for increasing values of  $\sqrt{A_0^2 + B_0^2}$ . From the upper to the lower streamline, values of  $(kH, \mathcal{F}, \mathcal{F}_I, A_0, B_0, \sqrt{A_0^2 + B_0^2})$  are  $(1.9, 0.6, 1.5, 3.4, 1.0, 3.5)$ ,  $(1.5, 0.3, 0.4, 4.8, 1.4, 5.0)$ ,  $(0.1, 3.5, 1.0, 8.6, 0.1, 8.6)$ ,  $(0.5, 0.05, 0.04, 9.6, 2.5, 9.9)$ .



**Fig. S14** Time series of the relative velocity disturbance  $\delta_u$  corresponding to Fig. 5, for different values of the hydrodynamic roughnesses. **a:** North Sand Sea – summer, **b:** North Sand Sea – winter, **d:** South Sand Sea – summer, **e:** South Sand Sea – winter. Note that  $\delta_\theta$  is independent of the choice of  $z_0^{\text{ERA5-Land}}$  and  $z_0^{\text{Local mes.}}$ .