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**1 Significant wind disturbances induced by giant  
2 dunes.**

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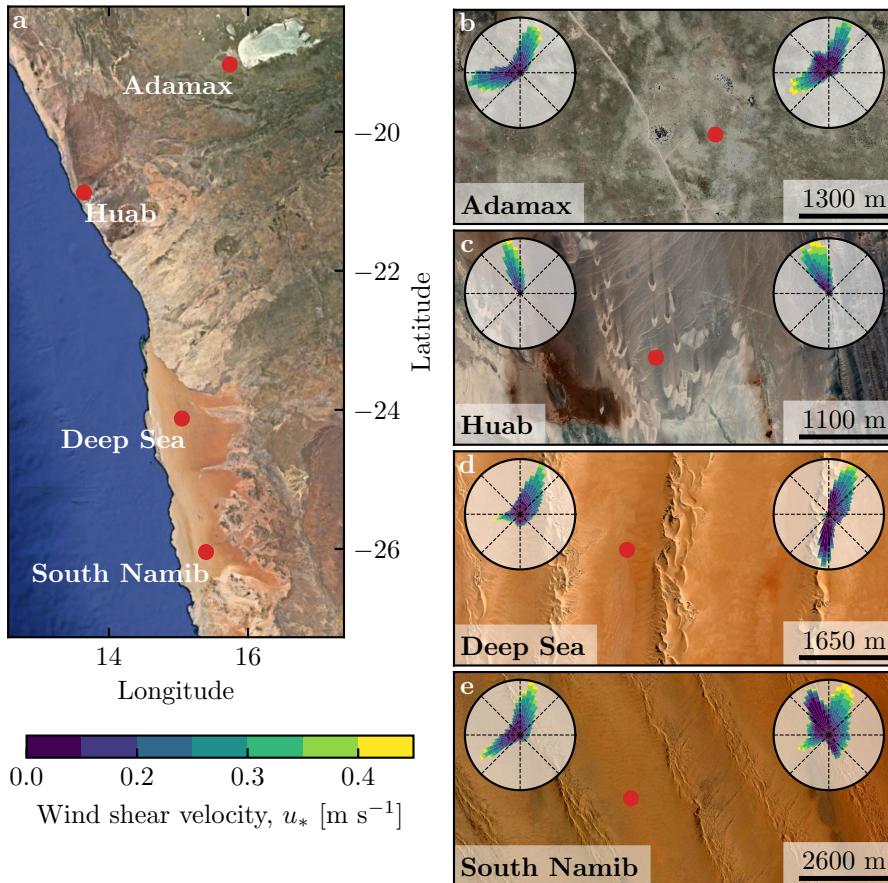
**13 1 Introduction**

14 Whenever a flow encounters an obstacle, different interactions can arise de-  
15 pending on the different time and length scales involved. In the case of atmo-  
16 spheric flows, this mainly depends on the part of the atmosphere, schematically  
17 composed of a turbulent boundary layer topped by a part free of turbulence,  
18 with which the obstacle interacts (Stull 1988). At the largest scale, the feed-  
19 back of mountains on the stratified flow of the free atmosphere results in wave  
20 generation as well as significant wind disturbances, such as downslope winds in  
21 the lee side (Durran 1990). Inside the boundary layer, the interaction between  
22 a turbulent flow and hilly surfaces is for example key to the understanding  
23 ocean surface wind-driven waves, or eolian bedforms in desert (Belcher and  
24 Hunt 1998; Sullivan and McWilliams 2010; Courrech du Pont 2015).

25 Indeed, eolian sand dunes typically emerge from the feedback of the topog-  
26 raphy on the turbulent flow, which speeds up close to the dune crest (Rubin  
27 and Hunter 1987; Charru et al. 2013; Courrech du Pont et al. 2014). Later  
28 on, when the dune reaches an intermediate size, it may also induce signifi-  
29 cant wind deflections. For example, this can affect the sediment pathways of  
30 coastal systems (Hesp et al. 2015), or the collective behavior of dune popu-  
31 lations, through long-range interactions due to flow disturbances induced by  
32 each individual (Smith et al. 2017; Bacik et al. 2020). As the dunes increase  
33 in size by collisions and coarsening, they sometimes reach a giant size compa-  
34 rable to the boundary layer depth (Andreotti et al. 2009). However, the wind  
35 disturbances induced by these giant dunes have never been quantified.

36 The study of the feedback of obstacles on atmospheric flows allows its  
37 incorporation into numerical meteorological models. The latter then become  
38 mainly limited by the accuracy of the included topographical data, as well  
39 as by the spatial grid of the model. For example, the ERA5-Land climate  
40 reanalysis is limited by its 9 km spatial resolution, while including the data 30-  
41 m Digital Elevation Models (DEMs) of the shuttle radar topography mission  
42 (Farr et al. 2007; Muñoz-Sabater et al. 2021). As such, it can not reproduce  
43 the flow disturbances induced by giant dunes, which have a typical length scale  
44  $\sim 1$  km.

45 Here, we compare the wind speed and direction of the ERA5-Land dataset  
46 to local measurements in four different places across the Namib desert. Where  
47 no significant topographies smaller than the model grid are present, we show  
48 that the two wind datasets agree with each other. On the contrary, in places  
49 with giant dunes, we show that they may differ for some specific meteorological  
50 conditions, that we link to the circadian cycle of the atmospheric boundary  
51 layer. In doing so, we highlight the importance of medium-scale topographies  
52 for local wind regimes and, in the case of sand seas, draw implications for  
53 smaller-scale eolian bedforms.



**Fig. 1** Wind data used in this study **a**: Location of the studied sites. **b–e**: Satellite images of the studied sites (Google-Earth, Maxar Technologies, CNES/Airbus). In each subplot, the left and right wind roses represent the data from the ERA5Land climate reanalysis and the local wind stations, respectively. Note that the bars show the direction towards which the wind blows. The red dots show the location of local wind stations.

## 54 2 Wind regimes across the Namib Sand Sea

55 In this study, we focus on four places across and nearby the Namib desert,  
 56 highlighting different environments (see Fig. 1). The Adamax station is lo-  
 57 cated near the Adamax salt pan, in a highly vegetated area. The Huab station,  
 58 located on the coast at the outlet of the Huab river is an arid environment  
 59 exhibiting 60-m scale barchan dunes. While these two stations are in environ-  
 60 ments with no mid-scale topography, this is not the case for the Deep Sea  
 61 and South Namib stations. Both are located in the interdune between giant  
 62 linear dunes with kilometric wavelengths and superimposed patterns. In this  
 63 section, we describe and compare the wind regimes resulting from the available  
 64 datasets in each station.

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65    2.1 Datasets

66    Two wind datasets are used in this study. First, local winds are provided by  
 67    stations situated in the four different places (see Fig. 1). The wind strength and  
 68    direction are measured every 10 minutes by cup anemometers and wind vanes,  
 69    at heights between 2 m and 3 m depending on the station. The available period  
 70    of measurements ranges from 1 to 5 discontinuous years distributed between  
 71    2012 and 2020 (see Fig. S1). We checked that at least one complete seasonal  
 72    cycle is available at each station. Then, regional winds are extracted at the  
 73    same locations and periods from the ERA5-Land dataset, which is a replay  
 74    at a smaller spatial resolution of ERA5, the latest climate reanalysis from  
 75    the ECMWF (Hersbach et al. 2020; Muñoz-Sabater et al. 2021). It provides  
 76    hourly estimates of the 10-m wind velocity and direction at a spatial resolution  
 77    of  $\sim 9$  km ( $0.1^\circ \times 0.1^\circ$ ).

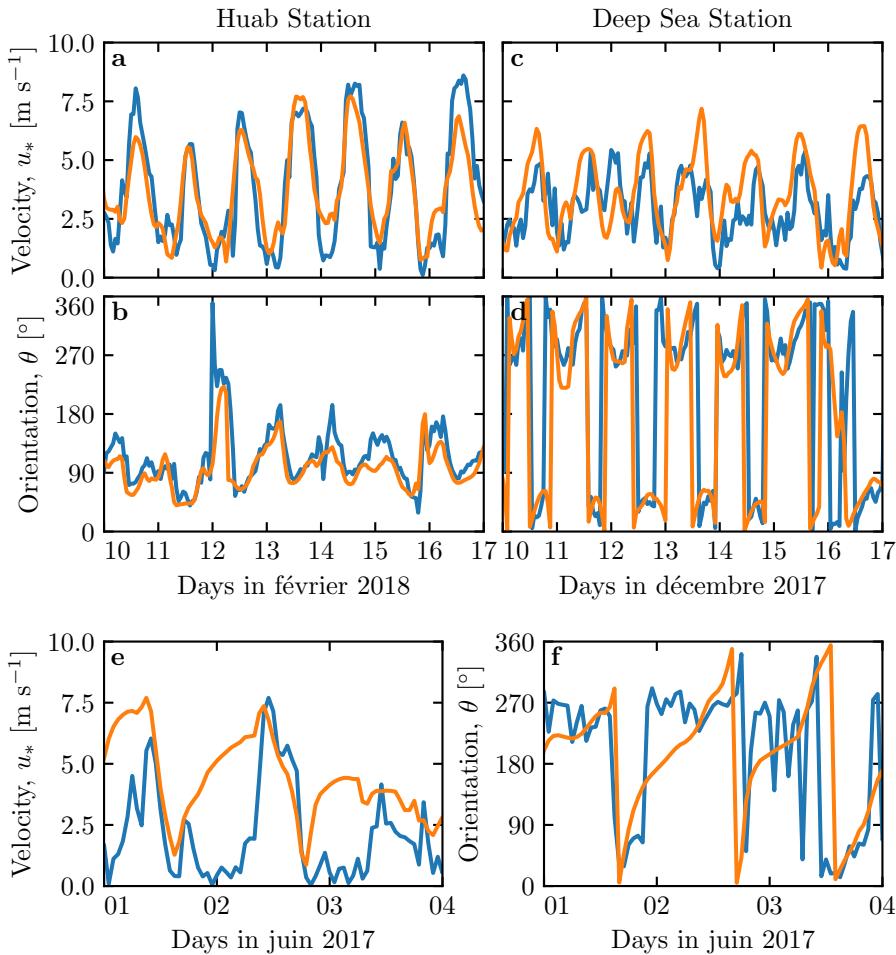
78    For comparison, the local measurements are averaged into 1-hr bins cen-  
 79    tered on the temporal scale of the ERA5-Land estimates (see Fig. S2). As the  
 80    wind velocities of both datasets are provided at different heights, we convert  
 81    them into shear velocities (see SI section 1), characteristic of the whole turbu-  
 82    lent wind profile within the atmospheric boundary layer, which are then used  
 83    together with the wind direction for further analysis. The resulting wind data  
 84    are shown on the wind roses of Fig. 1(b–e).

85    Finally, the dune properties are computed using autocorrelation on the 30-  
 86    m Digital Elevation Models (DEMs) of the shuttle radar topography mission  
 87    (Farr et al. 2007). For the South Namib and Deep Sea stations, we obtain  
 88    respectively orientations of  $85^\circ$  and  $125^\circ$ , wavelengths of 2.6 km and 2.3 km  
 89    and amplitudes of 45 m and 20 m (see Fig. S4 for more details).

90    2.2 Agreement between local and regional winds

91    The obtained wind regimes are shown in figure 1. In the Namib, the regional  
 92    wind patterns are essentially controlled by the sea breeze, resulting in strong  
 93    northward components (sometimes slightly deviated by the large scale topogra-  
 94    phy) present in all regional wind roses (Lancaster 1985). These daily winds are  
 95    dominant during the second-half of the year (Septembre-January). In winter,  
 96    an additional easterly component can be recorded during the night, induced by  
 97    the combination of katabatic winds forming on the mountains, and infrequent  
 98    ‘berg’ winds, which are responsible of the high wind velocities observed (Lan-  
 99    caster 1984). The frequency of these easterly components decreases from the  
 100   inland to the coast, resulting in bidirectional wind regimes within the Namib  
 101   Sand Sea and at the Adamax salt pan (Fig. 1b, 1d and 1e) and a unidirectional  
 102   wind regime on the coast at the outlet of the Huab River (Fig. 1c).

103   In the case of the Adamax and Huab stations, the regional wind roses  
 104   qualitatively match those corresponding to the local in situ measurements.  
 105   However, for the Deep Sea and South Namib stations, the local wind roses  
 106   exhibit additional components aligned with the giant dune orientation visible

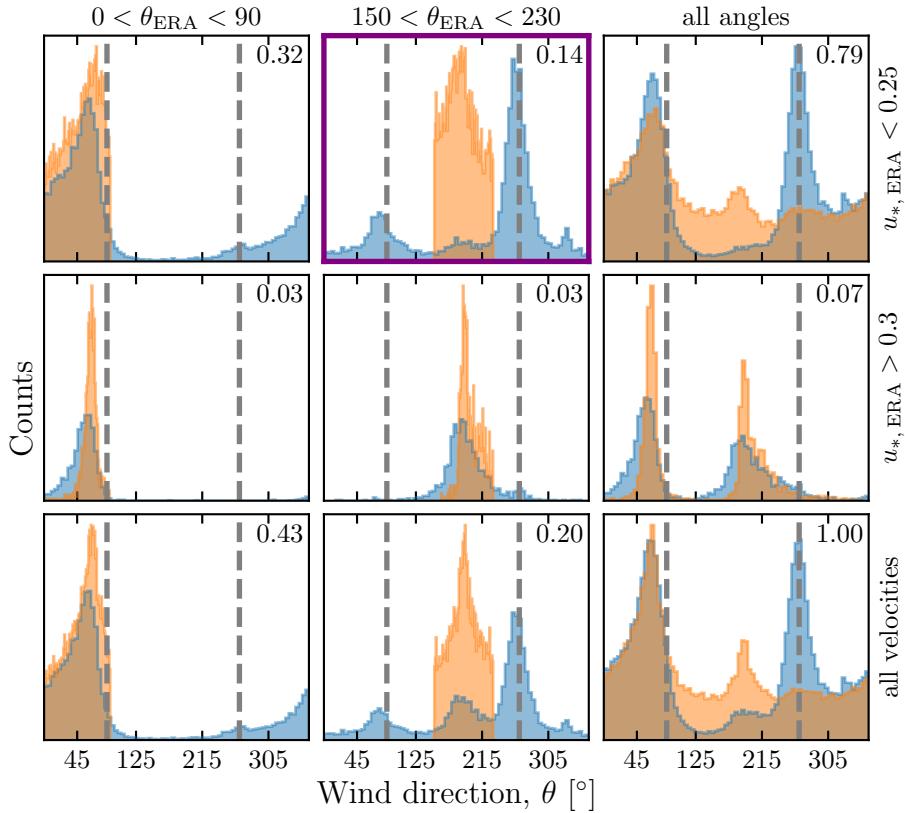


**Fig. 2** Temporal comparison between the wind data coming from the Era5Land climate reanalysis (orange lines) and from the local measurements (blue lines). **a–b:** Huab station. **c–d:** Deep Sea station in winter. **e–f:** Deep Sea station in summer.

on the satellite images (Fig. 1c–d). Indeed, the analysis of the wind speed and direction time series shows that the agreement between the local and regional datasets is always verified when no mid-scale topography are present (Fig. 2a–b) and Fig. S5). In contrast, for the stations within the giant dune field, we observe that this agreement is limited to the Septembre–Janvier time periods (Fig. 2c–d).

### 113 2.3 Influence of the giant dunes on local wind regimes

When giant dunes are present, in the February–August period, the local and regional winds match only during the morning, i.e when the southerly/southwesterly



**Fig. 3** Distributions of wind direction at the Deep Sea Station for the Era5Land climate reanalysis (orange) and the local measurements (blue). In each subplot, both distributions are plotted from the same time steps, selected with constraints on the wind direction (columns) and/or wind velocity (rows) of the Era5Land dataset. The gray vertical dashed lines indicate the dune orientation, and the top right numbers the percentage of the total number of time steps selected in each subplot. The purple frame highlights the regime (small wind velocities, nocturnal summer wind) in which the data from both datasets differs. A similar figure can be obtained for the Deep Sea station (see Fig. S6).

116 sea breeze dominates (see Fig. 2(e–f), Fig. 3 and Fig. S6). In the late afternoon  
 117 and during the night, when the northwesterly ‘berg’ and katabatic winds  
 118 blow, the two datasets differ. In this case, the angular wind distribution of  
 119 the local measurements exhibits two additional modes separated of  $\simeq 180^\circ$ ,  
 120 each corresponding to the giant dune alignment (purple frame in Fig. 3 and  
 121 Fig. S6, as well as Fig. S7). This deviation is also associated with a global  
 122 attenuation of the wind strength (Fig. S8). Remarkably, all these figures show  
 123 that this process occurs for low wind velocities, typically for  $u_* < 0.1 \text{ m s}^{-1}$ .  
 124 For shear velocities larger than  $0.25 \text{ m s}^{-1}$ , this wind reorientation does not  
 125 occur. Finally, for intermediate shear velocities, both reorientation along the  
 126 dune crest and no reorientation are observed (Fig. S7).

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127 **3 Influence of the circadian cycle of the atmospheric boundary  
128 layer**

129 In the case of linear ridges, dune-induced flow disturbances have mainly been  
130 related to the incident wind direction (Walker et al. 2009; Hesp et al. 2015).  
131 In our case, it is unlikely to be the dominant parameter, as the most deflected  
132 wind for both stations is the most perpendicular, where it should be winds  
133 with incident directions between 30° and 70° (Hesp et al. 2015). An important  
134 observation is the difference in behavior between low and high wind velocities,  
135 which suggests a change in the hydrodynamical regime.

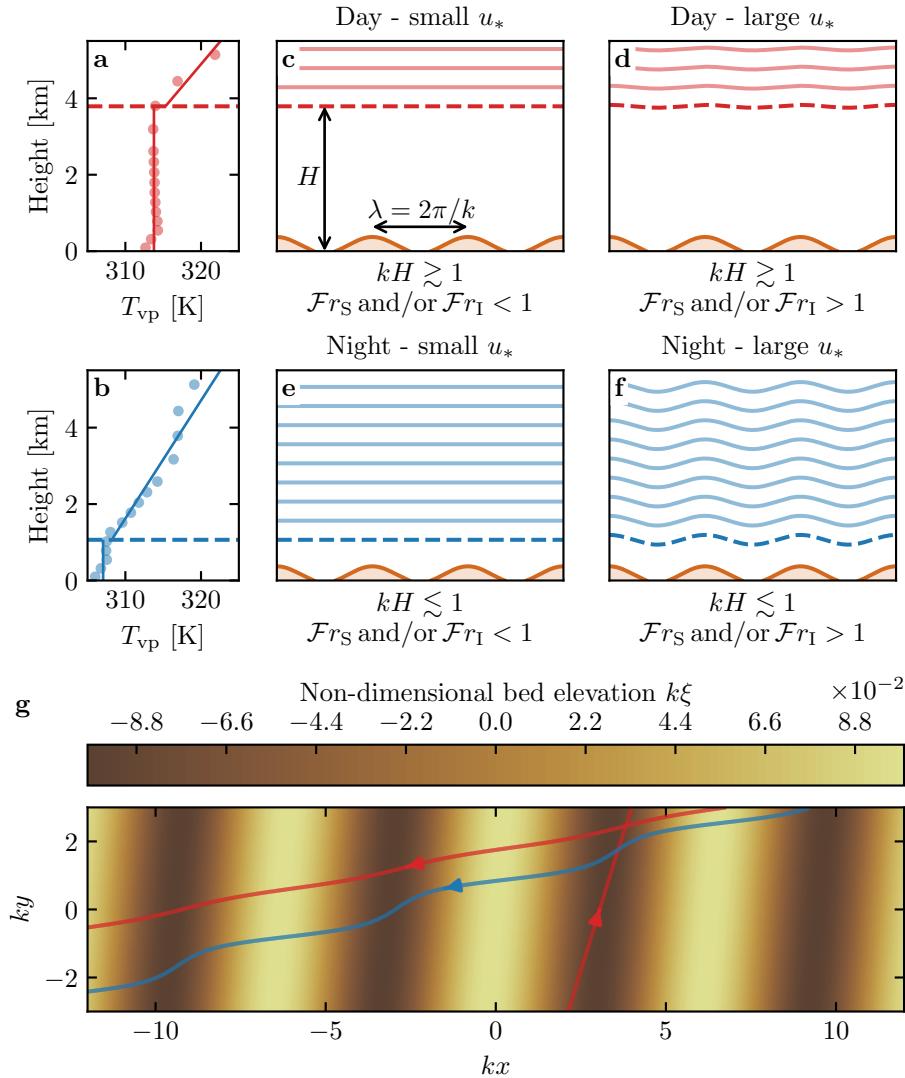
136 Previous studies have linked atmospheric flow around and over topographical  
137 obstacles to the vertical structure of the atmosphere (Stull 1988). More  
138 particularly, dunes evolves in its lower part, the turbulent atmospheric bound-  
139 ary layer (ABL), typically characterized by a logarithmic wind profile and a  
140 vertically constant potential temperature. Above, the free atmosphere (FA) is  
141 a stably stratified zone in which turbulence is negligible, and where the flow  
142 is usually considered as incompressible and inviscid. In the middle, a transi-  
143 tional layer, also known as entrainment zone, is characterized by a sharp  
144 increase of the potential temperature, which traps the turbulence resulting  
145 from the surface friction below it.

146 In the following, we sum-up the dominant numbers leading to different  
147 hydrodynamical interactions with topographical obstacles, and interpret the  
148 data with respect to the corresponding physical mechanisms.

149 **3.1 Relevant non-dimensional parameters and physical modeling**

150 Flow deflection over ridges can be simplistically understood from a balance  
151 between inertia and pressure gradients (Hesp et al. 2015). As the flow ap-  
152 proaches the ridge crest, the compression of the streamlines results in larger  
153 flow velocities, and thus lower pressures (Rubin and Hunter 1987). An incident  
154 flow oblique to the ridge is then deflected towards lower pressure zones, i.e to-  
155 wards the crest. Turbulent dissipation at the bottom and non-linearities tends  
156 to increase this effect downstream, resulting in along the crest wind deflection  
157 in the lee side (Hesp et al. 2015; Gadal et al. 2019).

158 Another way to increase the flow deflection is its confinement below a  
159 capping surface, that results in further streamline compression. This happens  
160 when the flow disturbance induced by the obstacle reaches the surface. As  
161 obstacles typically disturb flow over a characteristic height similar to their  
162 width, the potential of interaction between the dunes and the overlying atmo-  
163 spheric structure is well captured by the parameter  $kH$ , where  $k = 2\pi/\lambda$  is  
164 the wavenumber and  $H$  the ABL depth. Here, the giant dunes have kilometric  
165 wavelengths, such that  $0.02 \lesssim kH \lesssim 5$ , and they interact most of the time  
166 with the capping layer and the stratified free atmosphere above (Andreotti  
167 et al. 2009).



**Fig. 4** **a–b:** Vertical profiles of the virtual potential temperature at 2 different time steps (blue - 29/11/2012 - 1100 UTC, red - 21/03/2017 - 1200 UTC) at the Deep Sea station. Dots: data from the ERA5 reanalysis. Transparent dashed lines: boundary layer height given by the ERA5 reanalysis, calculated from the bulk Richardson number (Seidel et al. 2012). 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 qualitatively representing the effect of flow confinement, in the case of the Deep Sea station. The red lines are calculated from the unconfined case similar to **d**. The sinuous blue line represents a confined case similar to **f**. For details on the streamline derivation, see Appendix 4.

168 Note that the ability of the capping layer and stratification to accommodate  
 169 a perturbation induced by the topography directly impacts the strength of  
 170 this confinement effect (Fig. 4). This is typically quantified using surface and  
 171 internal Froude numbers (Vosper 2004; Stull 2006; Sheridan and Vosper 2006;  
 172 Hunt et al. 2006; Jiang 2014):

$$\mathcal{Fr}_S = \frac{U}{\sqrt{\frac{\Delta\rho}{\rho} g H}}, \quad \mathcal{Fr}_I = \frac{kU}{N}, \quad (1)$$

173 where  $U$  is the wind velocity at the top of the ABL,  $\rho$  its average density,  $\Delta\rho$   
 174 the density jump between the ABL and the FA and  $N$  is the Brunt-Väisälä  
 175 frequency, characteristic of the stratification.

176 The smallest wind disturbances are expected during the day, when the  
 177 ABL depth is comparable to the dune wavelength ( $kH \gtrsim 1$ ) and for large  
 178 wind velocities, which correspond to a weak confinement situation (Fig. 4d).  
 179 On the contrary, large wind disturbances are expected to occur during the  
 180 night, when the confinement is mainly induced by shallow ABL (Fig. 4e–f).  
 181 Note that this strong confinement can be somewhat reduced in the case of  
 182 strong winds (corresponding to large Froude numbers, see Fig. 4f), explaining  
 183 the threshold in velocity observed in the data (see section 2.3).

### 184 3.2 Flow regime diagrams

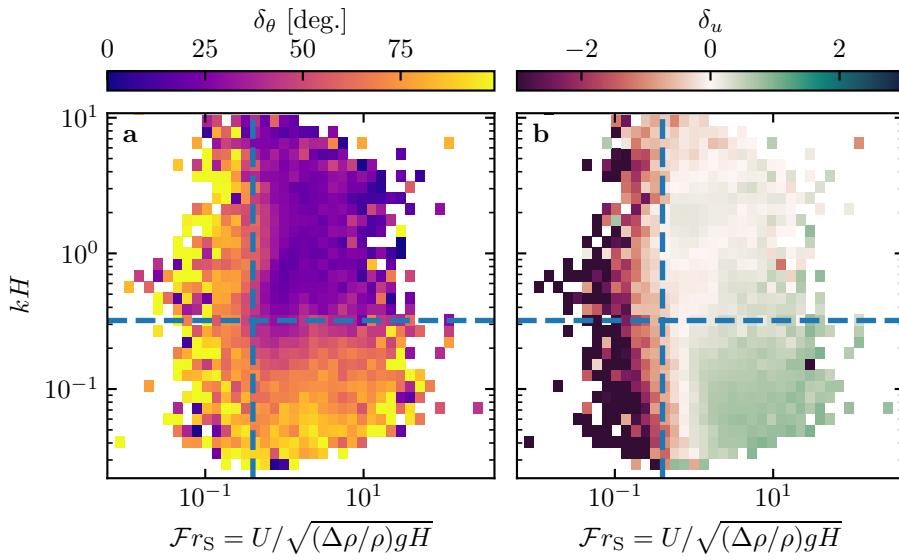
185 To highlight these different regimes from our data, we compute wind dis-  
 186 turbance diagrams in the space defined by the three relevant non-dimensional  
 187 numbers presented above, ( $kH$ ,  $\mathcal{Fr}_S$ ,  $\mathcal{Fr}_I$ ). Those are calculated from the time  
 188 series of the geopotential, temperature and specific humidity vertical profiles  
 189 available in the ERA5 climate reanalysis (see SI section 2). Flow deviation is  
 190 computed as the minimal angle between the wind orientations from the two  
 191 datasets:

$$\delta_\theta = |\min([\theta_{\text{ERA}} - \theta_{\text{station}}] \bmod 360, [\theta_{\text{station}} - \theta_{\text{ERA}}] \bmod 360)|. \quad (2)$$

192 The relative velocity modulation is computed as

$$\delta_u = \frac{u_*^{\text{ERA}} - u_*^{\text{station}}}{u_*^{\text{ERA}}}. \quad (3)$$

193 As both Froude numbers have qualitatively the same impact on the flow  
 194 confinement, we first leave aside the internal Froude number, and focus on  
 195 the space ( $kH$ ,  $\mathcal{Fr}_S$ ). When representing the two variables  $\delta_\theta$  and  $\delta_u$  in this  
 196 space, different regime emerges (Fig. 5). Small wind disturbances ( $\delta_\theta \rightarrow 0$ ,  
 197  $\delta_u \rightarrow 0$ ) are located in the top-right part of the diagrams, corresponding  
 198 to a regime mixing low-interaction and low-confinement ( $kH$  and  $\mathcal{Fr}_S$  large  
 199 enough, Fig. 4d). Lower values of  $kH$  (stronger interaction) or Froude numbers  
 200 (stronger confinement) then both leads to an increase in wind disturbances,



**Fig. 5** Regime diagrams of the wind deviation  $\delta_\theta$  and relative attenuation/amplification  $\delta_u$  in the space  $(\mathcal{Fr}_S, kH)$ , containing the data from both the Deep Sea and South Namib stations. Blue dashed lines empirically delimit the different regimes. The point density in each bin of the diagrams is shown in Fig. S11. The regime diagrams in the spaces  $(\mathcal{Fr}_I, kH)$  and  $(\mathcal{Fr}_I, \mathcal{Fr}_S)$  are shown in Fig. S12.

both in terms of orientation and velocity. Below a threshold value of  $kH \simeq 0.3$ , wind disturbance occurs independently of the Froude numbers value. Furthermore, this also seems to control a transition between damped to amplified wind velocities within the interdune (Fig. 5b), for which we do not have an explanation. Note that the same interpretation can be done with the diagrams including the internal Froude number  $\mathcal{Fr}_I$ , as shown by Fig. S12.

Interestingly, the limit of no-interactions between the topography and the boundary layer structure ( $kH \gg 1$ ), in which the properties of the capping layer and the stratification become irrelevant, is never reached here, in the case of giant dunes.

#### 4 Discussion

The comparison of local (direct measurements) and regional (climate reanalysis) wind data reveals the giant dunes feedback on the flow. In flat areas, the matching between both datasets confirms the ability of the ERA5Land climate reanalysis to predict the wind flow down to scales  $\sim 10$  km, i.e. the grid model. When smaller scale topographies are present (giant dunes in our case), locally measured wind regimes may significantly differ from the regional ones. Furthermore, we link these disturbances induced by the dunes to their interaction with the lower part of the atmospheric vertical structure, and more specifically to its circadian variability. During the night, the presence of a shallow

atmospheric boundary layer (ABL) induces a strong flow confinement, associated with large wind deviation and acceleration or deceleration. During the day, the capping layer is high enough to prevent its interaction with the giant dunes, resulting in a low flow confinement, and thus smaller wind disturbances. Interestingly, we also found that this effect could be counterbalanced by the presence of large wind velocities, capable of deforming the capping layer and/or the FA stratification, thus decreasing the confinement effect.

Simple linear models such as the one of Andreotti et al. (2009) also suggests that larger wind disturbances occur under strong flow confinement such as described above. However, they are unable to reproduce the magnitude of the observed deviations, probably due to the presence of hydrodynamical non-linear effects, negligible in low confinement situations, but not otherwise (see Fig. S12 and Appendix 1). Additionally, note that the spatial flow structures associated with the dune feedback on the wind can not be studied with our single-point measurements. Measurements in different places on and near the same topographical obstacle are then needed for further comparisons with models.

This study highlights the interaction between giant dunes and the atmospheric boundary layer, supporting as well the way the capping layer acts as a bounding surface limiting dune growth (Andreotti et al. 2009; Gunn et al. 2021). This interaction also have implications at smaller scales, where bed-forms then develop from the wind disturbed by the giant dunes, instead of the regional wind regime. Differences between larger and smaller scale dune patterns are observed ubiquitously. As larger dunes are often older than smaller ones, sometimes this discrepancy have been previously attributed to climatic changes in wind regimes (?). Here, using this feedback mechanism, we suggest that current winds are also able to explain dune patterns at all scales. For example, this seems to be the case for the linear dunes ( $\sim 50$  m -wide) elongating within the interdune between two giant linear dunes ( $\sim 2$  km -wide) in the Namib Sand Sea (see Fig. 6).

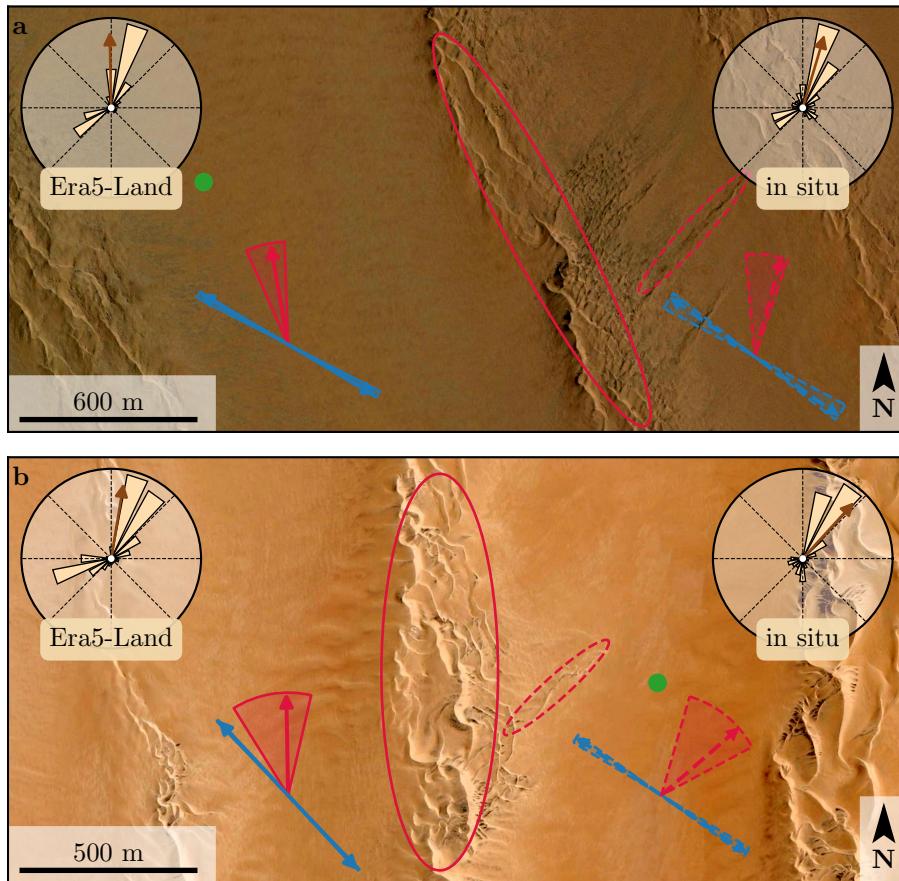
**Acknowledgements** These should follow the concluding section of the paper and precede the References and any appendices, if they are present. The acknowledgements section does not require a section number.

#### Appendix 1: ABL turbulent wind model

Following the work of Fourriere et al. (2010) and Andreotti et al. (2012), we briefly expose in this section the linear response of a turbulent flow to a small aspect ratio perturbation of the topography  $\xi$ . As this topography can be decomposed into several sinusoidal modes, we focus on the response to a sinusoidal topography as:

$$\xi = \xi_0 \cos [k (\cos(\alpha)x + \sin(\alpha)y)], \quad (4)$$

which is also a good approximation to the giant dunes observed in the Deep Sea and South Namib Station (see Fig 1 and Fig S4). Here,  $x$  and  $y$  are



**Fig. 6** Implications for smaller scale patterns in (a) the South Namib and (b) Deep Sea. The ellipses indicates the different types of elongating dunes, at large (plain) and small scale (dashed). The dune orientations are calculated using the model of Currech du Pont et al. (2014) from the sand flux angular distributions, shown here for typical sand quartz grains of  $180 \mu\text{m}$ . The double blue and single red arrows correspond to the two possible dune growth mechanisms, bed instability and elongation, respectively. Likewise, plain arrows are calculated from the ERA5-Land datasets, and dashed arrows from the in situ measurements. Wedges show the uncertainty on the orientation calculation, and the arrows correspond to typical parameters found in the literature, i.e a grain diameter of  $180 \mu\text{m}$  and a flux-up ratio of 1.6. The green dots indicate the position of the measurement stations. See Appendix 2 for additional details.

262 the streamwise and spanwise coordinates,  $k = 2\pi/\lambda$  the wavenumber of the  
 263 sinusoidal perturbation, and  $\alpha$  its crest orientation, calculated with respect to  
 264 the  $y$ -direction.

<sup>265</sup> In terms of basal shear stress  $\tau = \rho u_*^2$ , the flow response can then generally  
<sup>266</sup> be written in as:

$$\tau_x = \tau_0 \left( 1 + k\xi_0 \sqrt{\mathcal{A}_x^2 + \mathcal{B}_x^2} \cos [k(\cos(\alpha)x + \sin(\alpha)y) + \phi_x] \right), \quad (5)$$

$$\tau_y = \tau_0 k \xi_0 \sqrt{\mathcal{A}_y^2 + \mathcal{B}_y^2} \cos [k(\cos(\alpha)x + \sin(\alpha)y) + \phi_y], \quad (6)$$

<sup>267</sup> where  $\tau_0$  is the basal shear stress on a flat bed, and  $\phi_{x,y} = \tan^{-1}(\mathcal{B}_{x,y}/\mathcal{A}_{x,y})$ .  
<sup>268</sup> The in-phase and in-quadrature hydrodynamical coefficients  $\mathcal{A}_{x,y}$  and  $\mathcal{B}_{x,y}$   
<sup>269</sup> are functions of the flow conditions, i.e the bottom roughness, the free surface  
<sup>270</sup> or the incident flow direction (Fourriere et al. 2010; Andreotti et al. 2009, 2012;  
<sup>271</sup> Charru et al. 2013).

<sup>272</sup> Andreotti et al. (2012) have shown that the impact of the incident wind  
<sup>273</sup> direction can be well approximated by the following expressions:

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

$$\mathcal{B}_x = \mathcal{B}_0 \cos^2 \alpha, \quad (8)$$

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

$$\mathcal{B}_y = \frac{1}{2} \mathcal{B}_0 \cos \alpha \sin \alpha, \quad (10)$$

<sup>274</sup> where  $\mathcal{A}_0$  and  $\mathcal{B}_0$  are now two coefficients independent of the dune orientation  
<sup>275</sup>  $\alpha$ . In the case of a fully turbulent boundary layer capped by a free atmosphere  
<sup>276</sup> capping, they now only depend on  $kH$ ,  $kz_0$ ,  $\mathcal{Fr}_I$  and  $\mathcal{Fr}_S$ , as detailed by  
<sup>277</sup> Andreotti et al. (2009). More specifically, their variation in the marginal spaces  
<sup>278</sup> ( $kH$ ,  $\mathcal{Fr}_S$ ) and ( $kH$ ,  $\mathcal{Fr}_I$ ) are shown in Fig. S12.

<sup>279</sup> Typical values for the unconfined case are therefore  $\mathcal{A}_0 = 3.4$  and  $\mathcal{B}_0 = 1$ .  
<sup>280</sup> In our case of giant dunes with  $k\xi_0 \sim 0.1$ , significant wind disturbances are  
<sup>281</sup> then expected when  $\sqrt{\mathcal{A}_0^2 + \mathcal{B}_0^2} \sim 10$ . However, this is also the limit of the  
<sup>282</sup> linear regime where this theoretical model is applicable, as hydrodynamical  
<sup>283</sup> non-linearities become significant when  $k\xi_0 \sqrt{\mathcal{A}_0^2 + \mathcal{B}_0^2} \sim 1$ .

## <sup>284</sup> Appendix 2: Sediment transport and dune morphodynamics

<sup>285</sup> Here, we briefly detail the sediment transport and dune morphodynamics theo-  
<sup>286</sup> retorical framework leading to the prediction of sand fluxes and dune orientations  
<sup>287</sup> from wind data.

<sup>288</sup> The sediment fluxes can been directly linked to the wind basal shear stress  
<sup>289</sup> at each time steps  $t$  from transport laws, whose exact forms depends on the  
<sup>290</sup> sediment transport mechanisms taken into account. In this work, we following  
<sup>291</sup> the recent work of Pähntz and Durán (2020), where the sediment flux  $q_{\text{sat}}$  on a  
<sup>292</sup> flat bed made of loose sand can be expressed as:

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

where  $\kappa = 0.4$  is the von Kármán constant,  $C_M = 1.7$  a constant,  $Q = d\sqrt{(\rho_s - \rho)gd/\rho}$  is a characteristic flux, with  $\rho_s = 2.6 \text{ g cm}^{-3}$  and  $d = 180 \mu\text{m}$  the grain density and diameter, and  $g$  the gravitational acceleration. The friction coefficient  $\mu$  is taken to be the avalanche slope of the granular material, i.e.  $\sim 0.6$ . Finally, the Shields number is defined as  $\Theta = \rho u_{*,t}^2 / (\rho_s - \rho)gd$ , and its threshold value for incipient sediment transport as been calibrated using laboratory experiments to  $\Theta_{\text{th}} = 0.0035$ .

The dune orientations are then predicted from the dimensional model of Courrech du Pont et al. (2014). The orientation  $\alpha$  corresponding the bed instability is then the one that maximizes the following growth rate:

$$\sigma \propto \frac{1}{HWT} \int_t q_{\text{crest},t} |\sin(\theta_t - \alpha)|, \quad (12)$$

where  $H$  and  $W$  are dimensional constants representing the dune height and width, respectively. The flux at the crest is expressed as:

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

where the flux-up ratio  $\gamma$  has been calibrated to 1.6 using field studies, underwater laboratory experiments and numerical simulations. Similarly, the dune orientation corresponding to the elongation mechanism is the one that verifies:

$$\tan(\alpha) = \frac{\langle q_{\text{crest},t}(\alpha) \mathbf{e}_{\theta_t} \rangle \cdot \mathbf{e}_{WE}}{\langle q_{\text{crest},t}(\alpha) \mathbf{e}_{\theta_t} \rangle \cdot \mathbf{e}_{SN}}, \quad (14)$$

where  $\langle . \rangle$  denotes a vectorial time average. The unitary vectors  $\mathbf{e}_{WE}$ ,  $\mathbf{e}_{SN}$  and  $\mathbf{e}_{\theta_t}$  are in the West–East, South–North and wind direction, respectively.

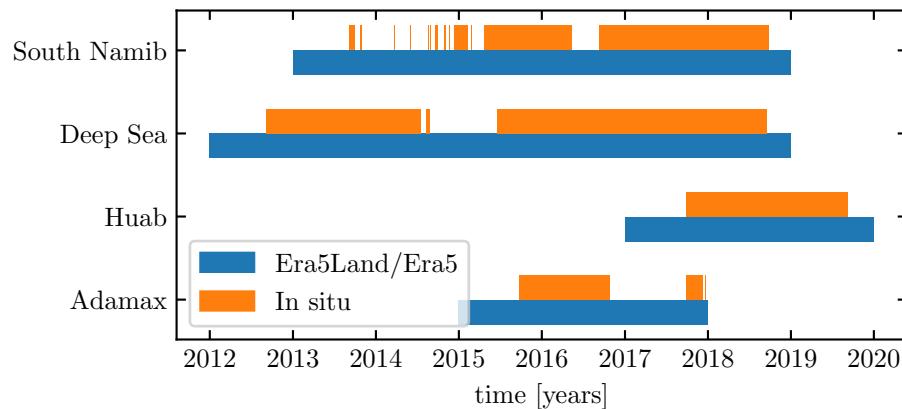
The computed dune orientations, blue and red arrows in figure 6, are however depending on a large number of parameters, for which we took typical values for eolian desert on Earth. We therefore run a sensibility test by calculating the dune orientations for grain diameters ranging from  $100 \mu\text{m}$  to  $400 \mu\text{m}$  and the speed-up ratio from 0.1 to 10 (wedges on figure 6). We also checked the sensibility the transport law by repeating the process with the quadratic transport also used for comparison in Pähzt and Durán (2020), which led to no more than  $n\%$  of variation with respect to (11).

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- 398



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

399 **Supplementary Material for *Boundary-Layer Meteorology* Sample  
400 Paper: Instructions for Authors**

401 **First Author\* · Second Author · Third Author**

402  
403 \*Affiliation and email address for the corresponding author only (note that  
404 the corresponding author does not need to be the first author).

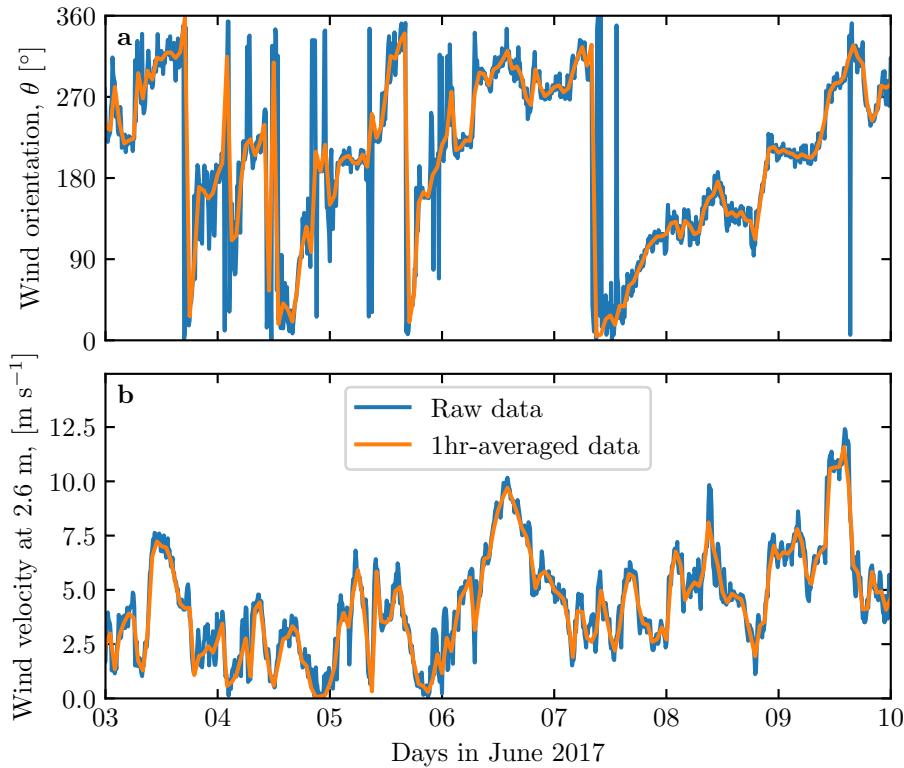
405 **1. Shear velocity and calibration of the hydrodynamical roughness**

406 For each station, the hydrodynamic roughness is calibrated by finding the  
407 one that minimizes the relative difference  $\delta$  between the wind vectors of both  
408 datasets:

$$\delta = \frac{\sqrt{\langle \|u_{*,\text{era}} - u_{*,\text{station}}\|^2 \rangle_t}}{\sqrt{\langle \|u_{*,\text{era}}\| \rangle_t \langle \|u_{*,\text{station}}\| \rangle_t}} \quad (15)$$

409 This  $\delta$ -parameter is computed for hydrodynamic roughness values ranging  
410 from  $10^{-5}$  m to  $10^{-2}$  m for the different stations. As shown by figure S3,  
411 the minimum of  $\delta$  in the space ( $z_0, \text{Era}, z_0, \text{in situ}$ ) forms a line. We thus take  
412 the roughness of the Era5Land dataset as the typical value when sediment  
413 transport occurs,  $10^{-3}$  m, corresponding to the thickness of the transport  
414 layer (?). It leads for the Adamax, Deep Sea, Huab and South Namib stations  
415 values of 2.7 mm, 0.76 mm, 0.12 mm and 0.48 mm, respectively.

416 The choice of the hydrodynamic roughness values only impacts the cal-  
417 culated shear velocities, but note the wind directions. As such, most of our  
418 conclusions are then independent of such a choice, and only the magnitude of  
419 the wind velocity attenuation in confined situation might be affected.



**Fig. S2** Comparison between raw local wind measurements, and centered averaged data over one hour for the South Namib station. **a:** wind direction. **b:** wind velocity at the measurement height, 2.6 m.

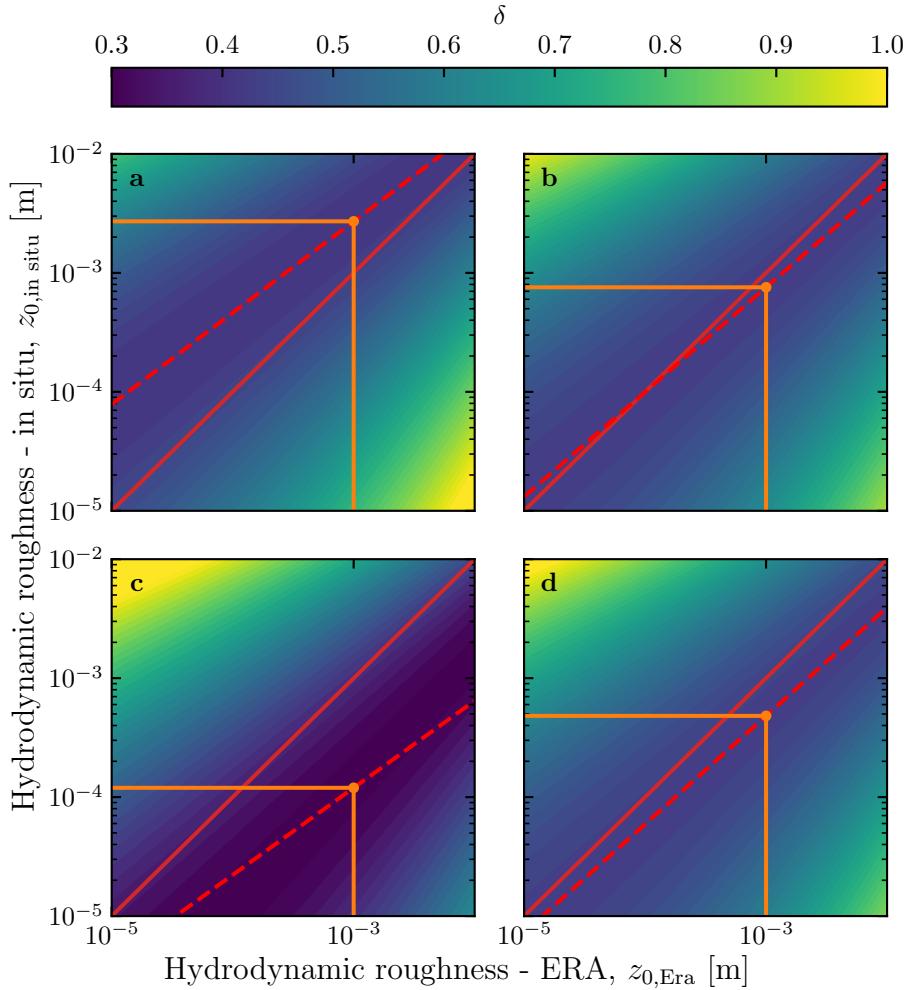
## 420 2. Extraction of the ABL properties

421 In order to estimate the relevant non-dimensional numbers, one need to es-  
 422 timate in addition to the wind and dune properties some parameters of the  
 423 ABL. The Era5 dataset provides a direct bulk estimate of the ABL depth  $H$   
 424 from a bulk Richardson number calculation, as well as vertical profiles of the  
 425 geopotential  $\phi$ , temperature  $T$  and specific humidity  $e_w$  at given pressure lev-  
 426 els  $P$ . From these quantities, the virtual potential temperature, which takes  
 427 into account the vertical pressure and humidity changes, can be calculated as:

$$T_{vp} = T (1 + [R_M - 1] e_w) \left( \frac{P_0}{P} \right)^{P_c(1-0.24e_w)}, \quad (16)$$

428 where  $P_0 = 10^5$  Pa is the standard pressure,  $P_c = 0.2854$  the Poisson coefficient  
 429 for dry air and  $R_M = 1.61$  is the ratio between the molecular masses of dry  
 430 air and water. The vertical coordinates are calculated as:

$$z = \frac{\phi R_t}{g R_t - \phi}, \quad (17)$$

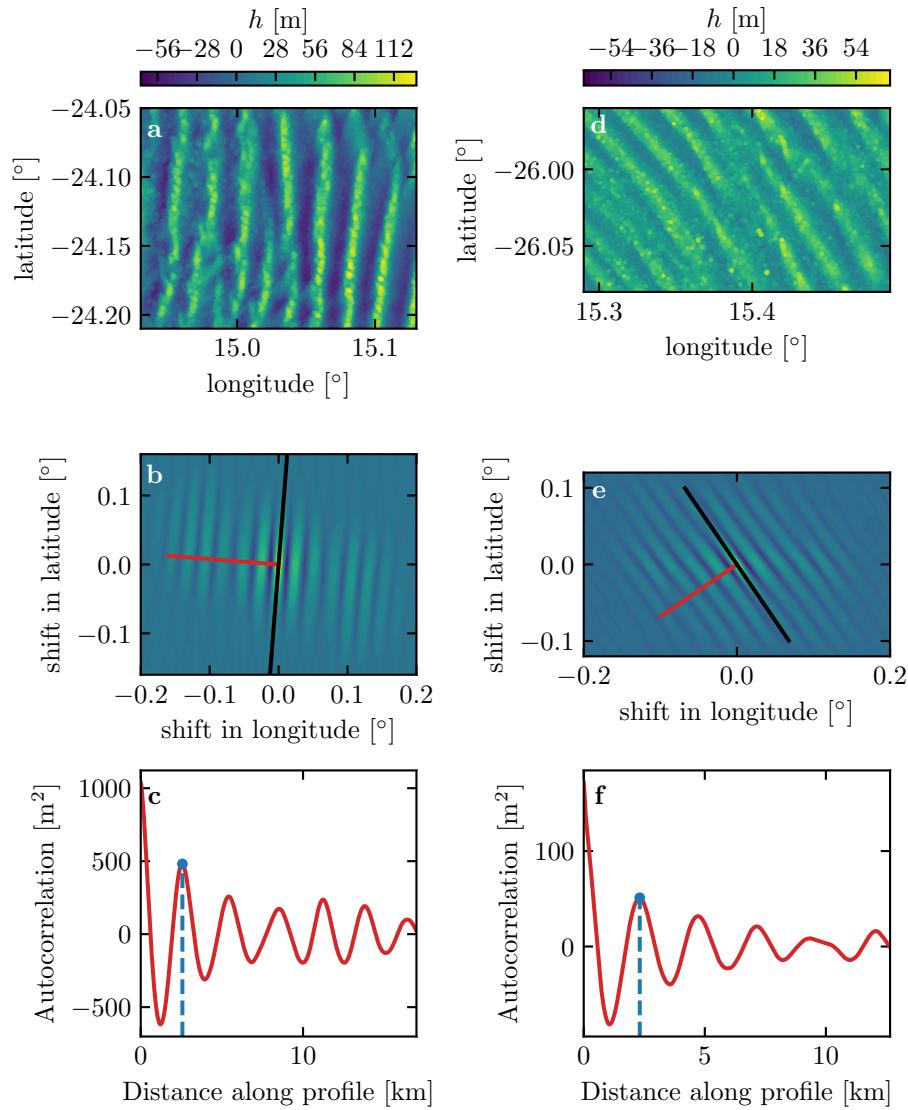


**Fig. S3** Calibration of the hydrodynamic roughnesses. The metric  $\delta$  defined in (15) is represented in colorscale as a function of the hydrodynamic roughnesses chosen for the Era5-Land and in situ datasets, for the Adamax (a), Deep Sea (b), Huab (c) and South Namib (d) Stations. The red dashed and plain lines shows the minima of  $\delta$  and the identity line. The orange lines and dots highlight the chosen hydrodynamic roughnesses for the local datasets, by imposing  $z_{0,\text{ERA}} = 1 \text{ mm}$ , leading for each station to 2.7 mm, 0.76 mm, 0.12 mm and 0.48 mm, respectively.

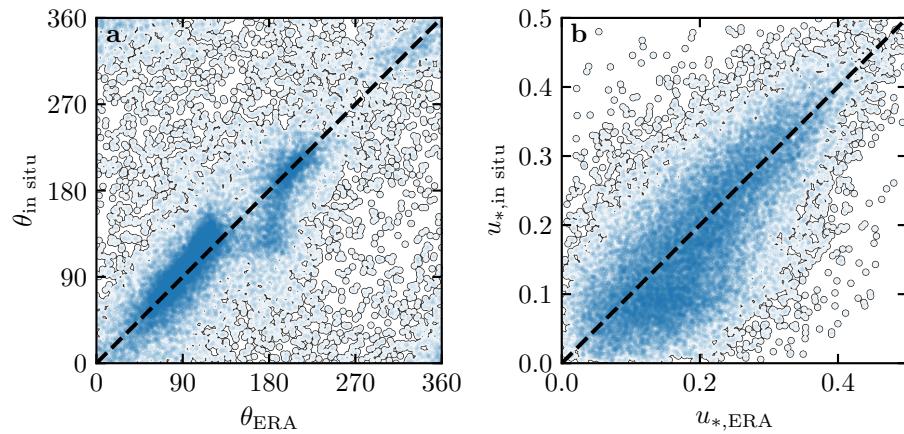
431 where  $R_t = 6356766 \text{ m}$  is the average Earth radius, and  $g = 9.81 \text{ m s}^{-2}$  the  
432 gravitational acceleration.

433 Example of obtained vertical profiles of the virtual potential temperature  
434 are shown in Fig. S9. On each of them, an average is computed below the ABL  
435 depth given by the Era5 dataset, and a linear function is fitted above.

436 Under the Boussinesq approximation, the temperature variations are as-  
437 sumed to induce most of those of the density, leading to  $\Delta\rho/\rho \simeq \Delta T_{\text{vp}}/T_{\text{vp}}$ .



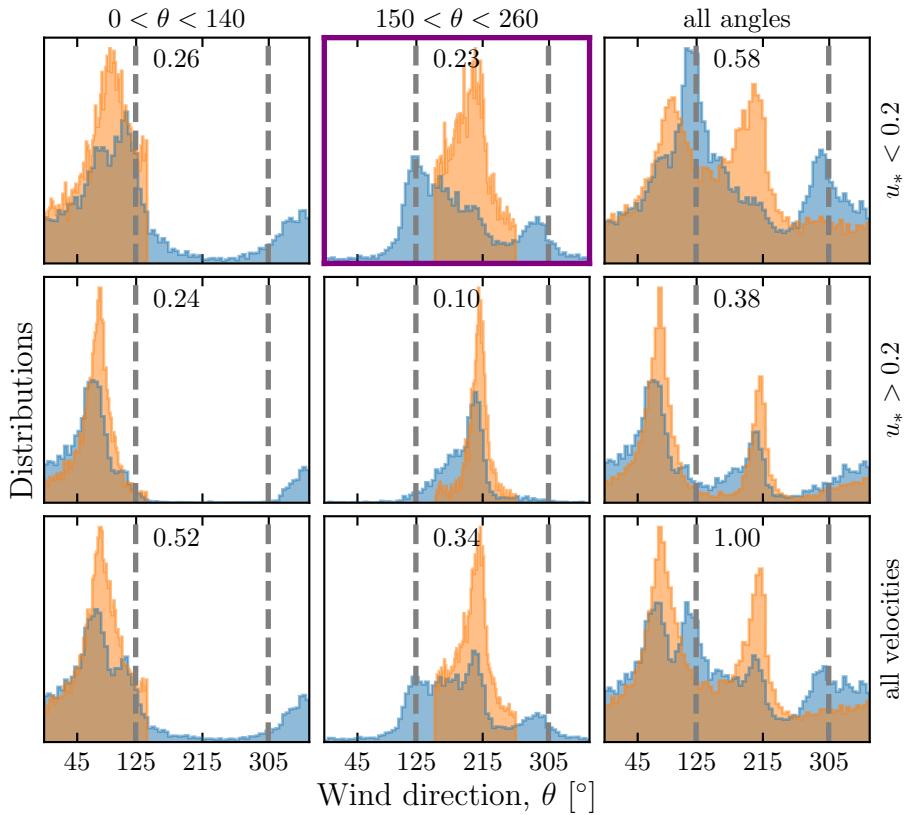
**Fig. S4** Analysis of the DEMs of the Deep Sea (left column – **a**, **b**, **c**) and South Namib (right column – **d**, **e**, **f**) stations. **a–d**: Detrended topography (a second order polynomial is first fitted and then removed). **b–e**: autocorrelation matrix shown in colorscale. The black line shows the detected orientation, and the red line the profile along which the wavelength is calculated, shown 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.



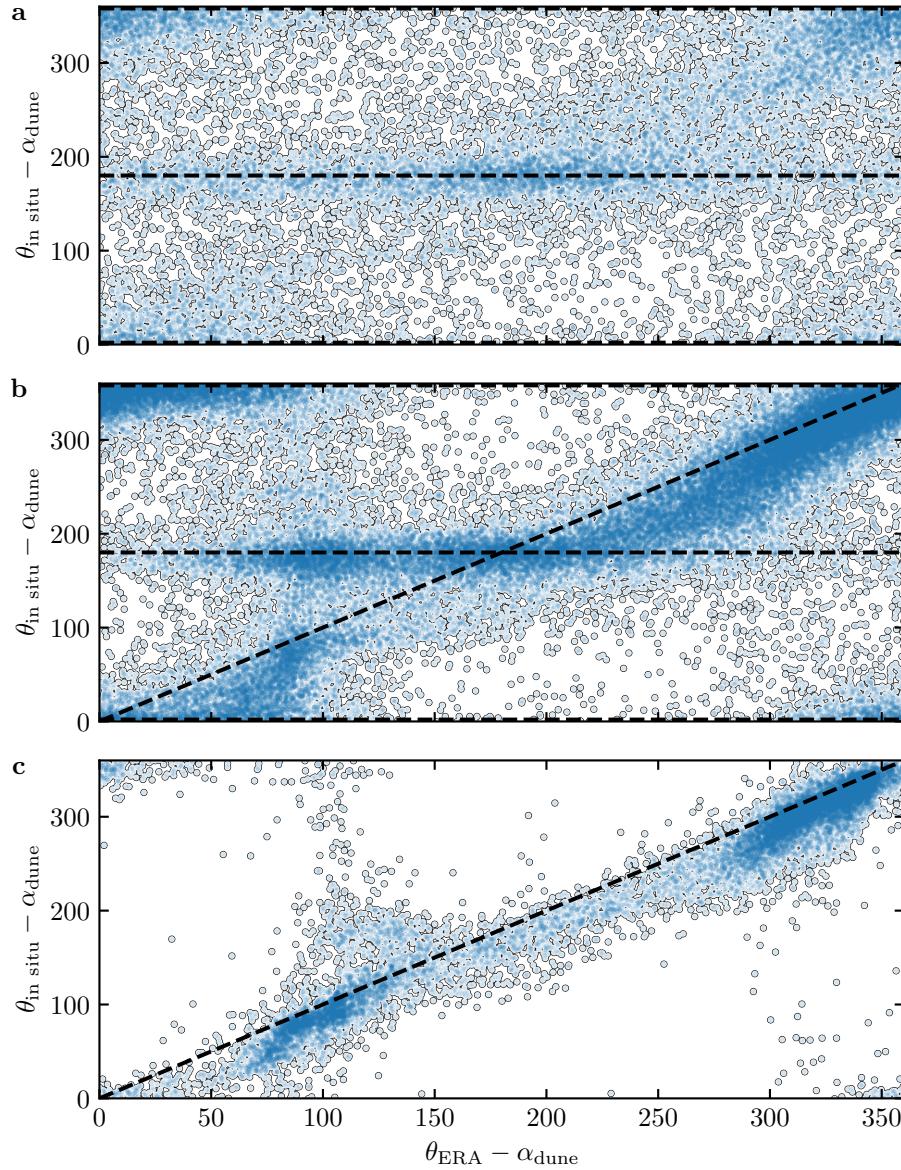
**Fig. S5** Statistical agreement of the wind orientation (a) and velocity (b) between the Era5Land dataset and the local measurements for the Huab and Adamax stations. Note how the points are clustered around identity lines (dashed and black).

438 Here,  $T_{\text{vp}}/T_{\text{vp}}$  is the relative virtual potential temperature jump at the cap-  
 439 ping, directly measured on the vertical profiles.

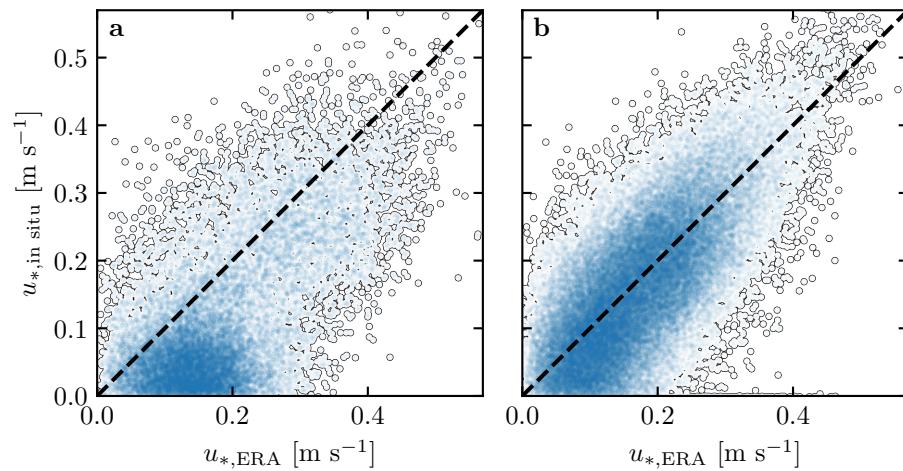
440 Following Tritton (2012), the relative density jump at the capping layer



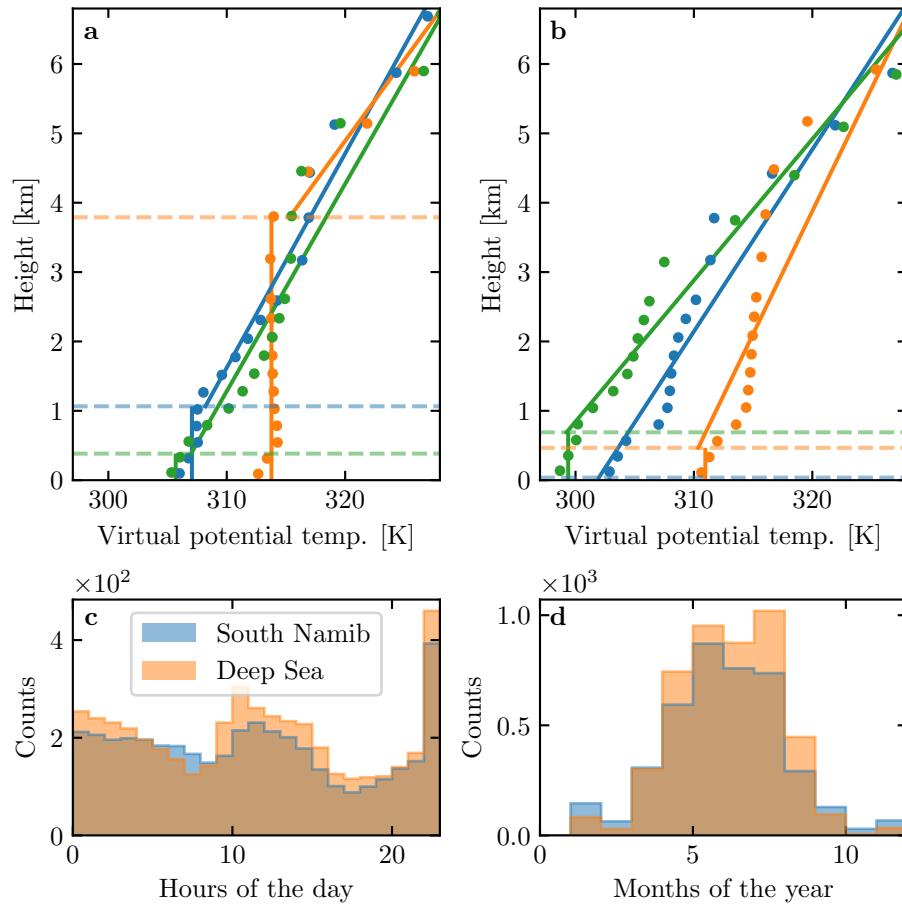
**Fig. S6** Distributions of wind direction at the South Namib Station for the Era5Land climate reanalysis (orange) and the in situ measurements (blue). In each subplot, both distributions are plotted from the same time steps, selected with constraints on the wind direction (columns) and/or wind velocity (rows) in the Era5Land dataset. The grey dashed vertical lines indicate the dune orientation. The purple frame highlights the regime (small wind velocities, nocturnal summer wind) during which the wind data from both datasets differs. A similar figure can be obtained for the South Namib station (see Fig. 3).



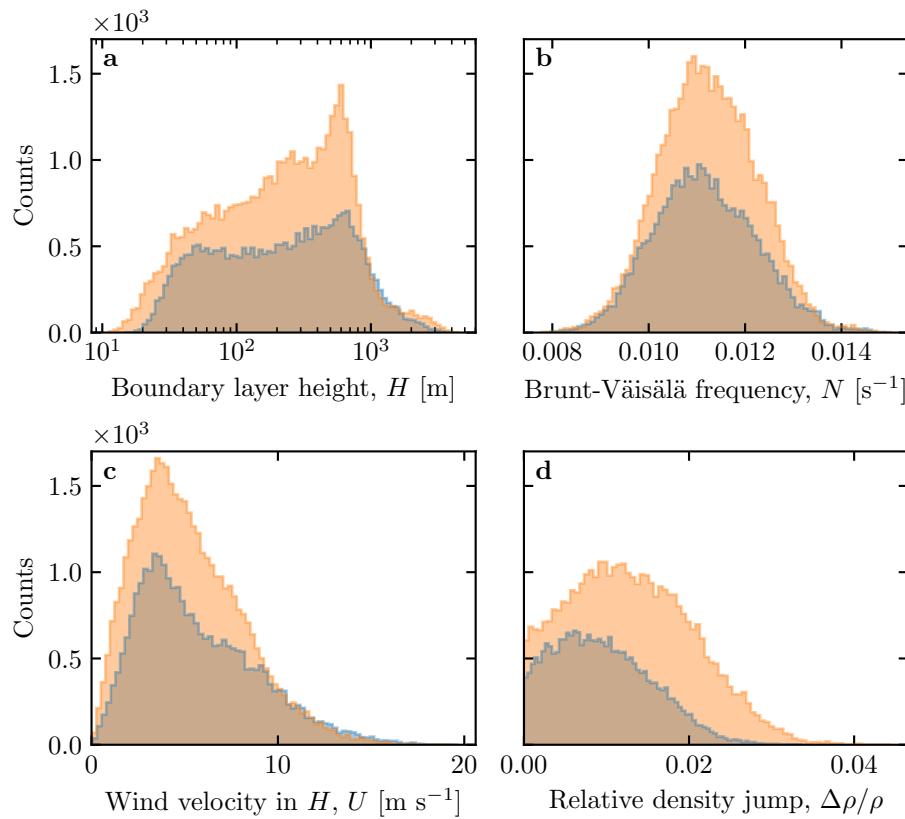
**Fig. S7** Statistical comparison of the wind orientation between the Era5Land dataset and the local measurements for the South Namib and Deep Sea stations, for different velocity ranges. **a:**  $u_{*,\text{ERA}} < 0.1 \text{ m s}^{-1}$ . **b:**  $0.1 < u_{*,\text{ERA}} \leq 0.25 \text{ m s}^{-1}$ . **c:**  $u_{*,\text{ERA}} \geq 0.25 \text{ m s}^{-1}$ . Note that the measured dune orientations are subtracted to the wind orientation, which allows 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  $360^\circ$  – **a, b**), as well as the identity lines (**b, c**).



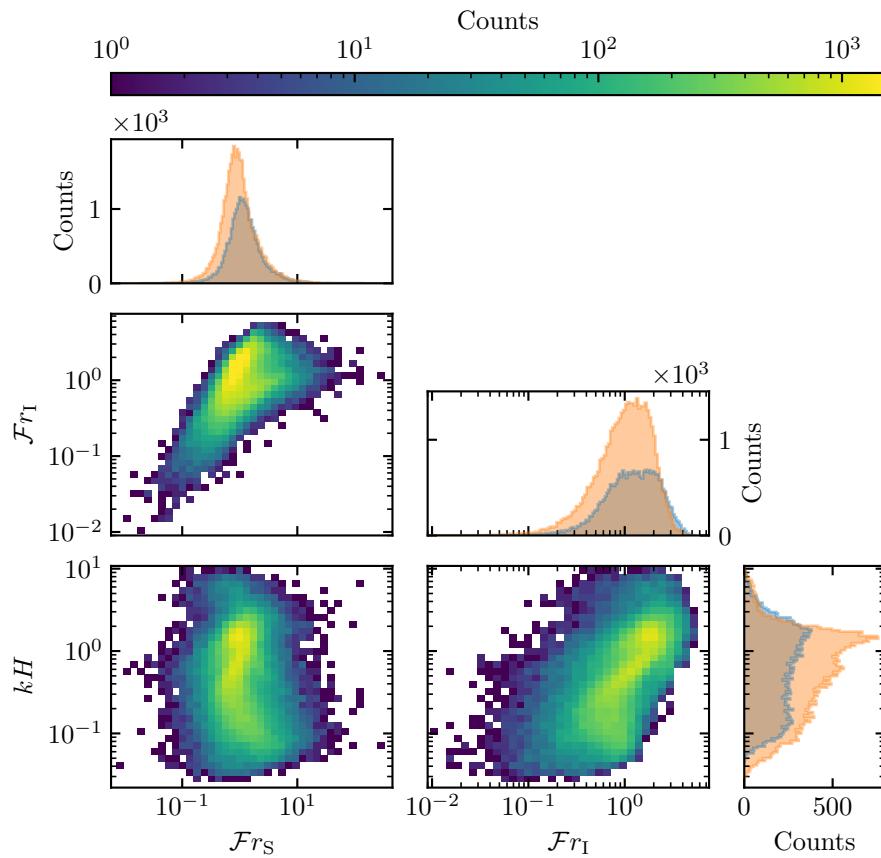
**Fig. S8** Statistical comparison of the wind velocity between the Era5Land dataset and the local measurements for the South Namib and Deep Sea stations. **a:** Nocturnal summer easterly wind. **b:** Diurnal southerly wind. Black dashed lines are identity lines. The angle ranges used to select diurnal and nocturnal summer winds are those taken in Fig. 3 and Fig. S6.



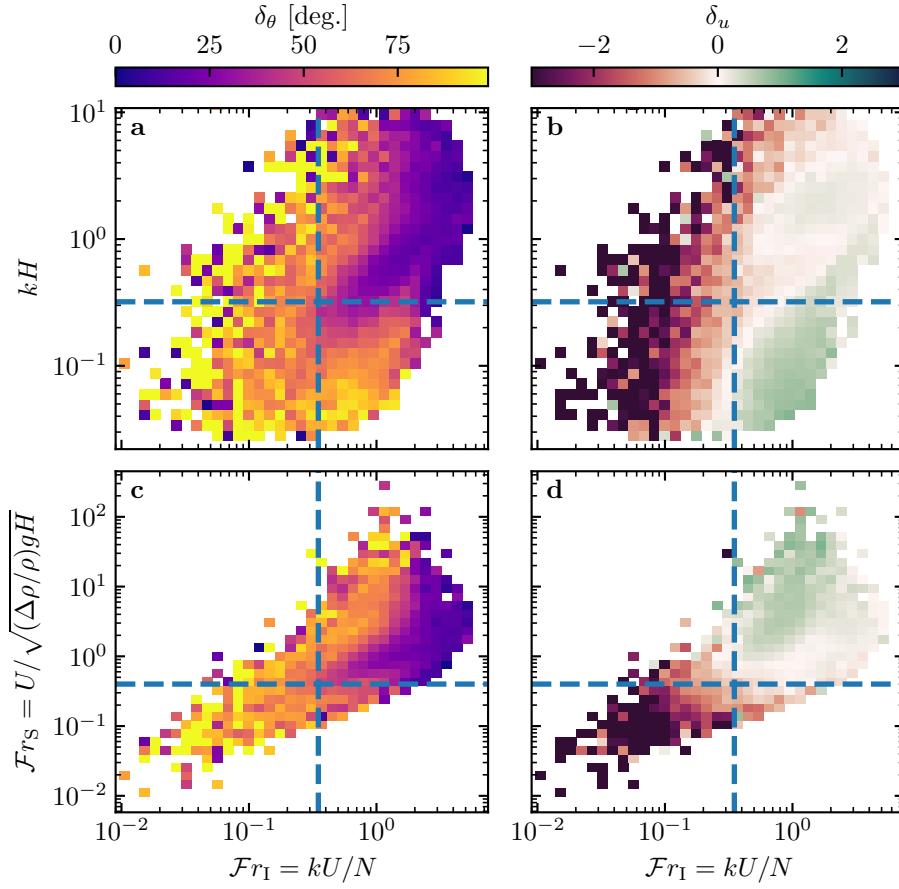
**Fig. S9** **a:** Vertical profiles of the virtual potential temperature at 3 different time steps (blue - 29/11/2012 - 1100 UTC, orange - 21/03/2017 - 1200 UTC, green - 21/03/2017 - 2000 UTC) at the South Namib station. Dots: data from the ERA5 reanalysis. Transparent dashed lines: boundary layer height given by the ERA5 reanalysis, calculated from the bulk Richardson number (Seidel et al. 2012). Plain lines: vertical (boundary layer) and linear (free atmosphere) fits to estimate the quantities in Fig. S10. **b:** Examples of ill-processed vertical profiles at 3 different time steps (blue - 2/12/2013 - 2300 UTC, orange - 20/03/2017 - 0000 UTC, green - 14/07/2017 - 1400 UTC) at the South Namib station. **c:** Hourly distribution of ill-processed vertical profiles. **d:** Monthly distribution of ill-processed vertical profiles. These profiles are ill-processed because the temperature found at the boundary layer from the linear fit in the free-atm is smaller than the average one inside the boundary layer. This is an unstable situation, which does not allow to calculate the surface Froude number.



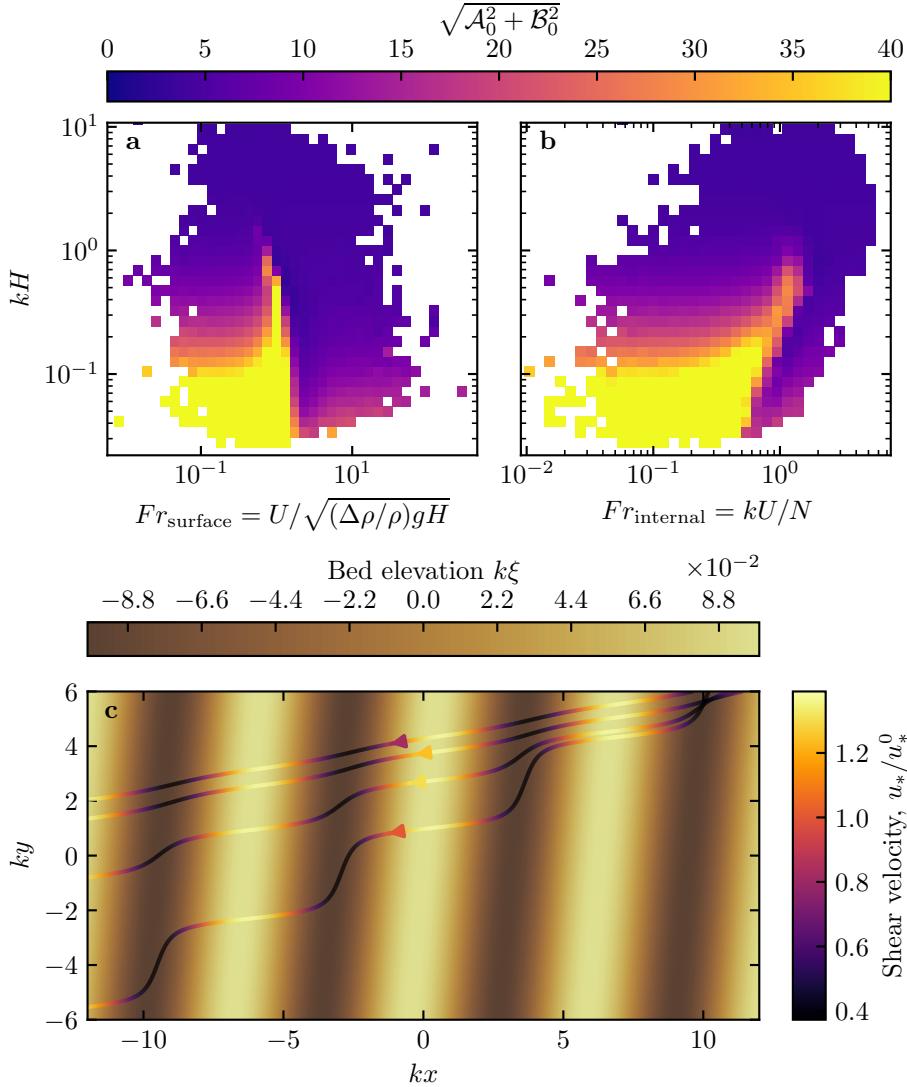
**Fig. S10** Distributions of the meteorological parameters resulting from the processing of the Era5-Land data for the South Namib (blue) and the Deep Sea (orange) stations.



**Fig. S11** Non-dimensional parameters distributions. For the marginal distributions, the orange correspond to the South Namib station, and the blue to the Deep Sea station.



**Fig. S12** Regime diagrams of the wind deviation  $\delta_\theta$  and relative attenuation/amplification  $\delta_u$  in the spaces  $(\mathcal{Fr}_I, kH)$  and  $(\mathcal{Fr}_I, \mathcal{Fr}_S)$ , containing the data from both the Deep Sea and South Namib stations. Blue dashed lines empirically delimit the different regimes. The point density in each bin of the diagrams is shown in Fig. S11. The regime diagrams in the space  $(\mathcal{Fr}_S, kH)$  are shown in Fig. 5 of the main article.



**Fig. S13** Physical interpretation of the flow disturbance. (a) and (b) Magnitude of the disturbance induced by a sinusoidal topography calculated from the time series of the non-dimensional numbers presented in Figures 4 and 5 using the linear model of Andreotti et al. (2009). (c) Shear velocity streamlines represented in the case of the Deep Sea station, for increasing values of  $\sqrt{\mathcal{A}_0^2 + \mathcal{B}_0^2}$ . From the upper to the lower streamline, values of  $(kH, Fr_{\text{surface}}, Fr_{\text{internal}}, \mathcal{A}_0, \mathcal{B}_0, \sqrt{\mathcal{A}_0^2 + \mathcal{B}_0^2})$  are  $(1.9, 0.6, 1.5, 3.4, 1.0, 3.5)$ ,  $(1.5, 0.3, 0.4, 4.8, 1.4, 5.0)$ ,  $(0.1, 3.5, 1.0, 8.6, 0.1, 8.6)$ ,  $(0.5, 0.05, 0.04, 9.6, 2.5, 9.9)$ .