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1 **Exploring wind-driven subsurface water flow with an acoustic Doppler**
2 **velocity profiler**

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ABSTRACT

9 In this study, we report on experiments carried out in a large wind-wave tank
10 to investigate the potential of the acoustic Doppler velocity profiler for deter-
11 mining the structure of the subsurface water boundary layer. This flow located
12 just beneath the air-water interface forms whenever wind blows. The profiler
13 is first tested for a steady flow generated by pumps beneath a flat water surface.
14 Measurements of the velocity field at different stages of development of the
15 wind-induced shear flow, from laminar to fully-turbulent, are then analyzed.
16 The best way to obtain reliable data under these flow conditions is thoroughly
17 examined. Despite the inherent difficulty of seeding acoustic tracers homo-
18 geneously in such a boundary layer, the profiler has the major advantage of
19 providing records of the instantaneous profiles of the subsurface velocity field
20 referenced to surface elevation. This feature makes it possible to estimate sta-
21 tistical properties of the water motions at various scales in a wave-following
22 coordinate system, and thus greatly increases the physical significance of the
23 measured quantities. The variation with fetch of the main characteristics of
24 the mean drift current, orbital wave motions and turbulent flow disturbances
25 estimated in this coordinate system is then presented and discussed in detail.

26 **1. Introduction**

27 Turbulent motions generated in the thin water boundary layer driven by wind and waves at the
28 free surface play a key role in heat and mass exchanges across the air-sea interface. They control
29 processes of critical importance for predicting air-sea interaction and behavior of many natural
30 environment systems, such as the transfer of greenhouse gases and the dispersion of pollutants
31 and micro-organisms. However, the experimental investigation of water surface flow dynamics
32 remains very challenging because of the inherent difficulty of performing non-intrusive measure-
33 ments just beneath a moving air-water interface of complex and highly variable geometry. Another
34 difficulty lies in the structure of the subsurface water boundary layer in which several instability
35 processes coexist and interplay, generating motions of comparable magnitude and which over-
36 lap in the space-time domain. Therefore, the properties of the mean shear current, orbital wave
37 motions, and other organized and turbulent motions of various scales occurring in water are still
38 poorly known.

39 Over the last decades, better knowledge of the small-scale processes within the surface sub-
40 layer has been achieved by using sophisticated optical techniques, such as infrared imaging, laser
41 Doppler or particle image velocimetry. However, the new insights are confined to a limited number
42 of specific experimental conditions for which one of the processes involved is essentially domi-
43 nant such as wave microbreaking or initial generation of shear flow and turbulence (Thais and
44 Magnaudet 1996; Banner and Peirson 1998; Peirson and Banner 2003; Siddiqui et al. 2001; Veron
45 and Melville 2001; Caulliez et al. 2007). Furthermore, there is still a significant gap in our under-
46 standing between the detailed description of individual phenomena as provided by spatial imaging
47 techniques and the statistical flow properties as estimated from time measurements by single-point
48 instruments.

49 Meanwhile, a technique for measuring water velocity fields remotely by means of acoustic
50 pulses has been developed. Based on the Doppler effect, it was designed and further improved
51 primarily for conducting ocean observations. The spatial resolution and accuracy of these mea-
52 surements have been gradually increased by the implementation of advanced techniques or new
53 probing concepts, enabling the assessment of the water turbulence behavior. More recently, a fur-
54 ther step forward has been achieved with the commercial launch of the acoustic Doppler velocity
55 Vectrino profiler by Nortek. This instrument offers the means to measure the three water veloc-
56 ity components along a short flow segment with relatively high space and time resolution (1 mm
57 and 0.01 s). This ready-to-use device thus enables the investigation of the small-scale dynamics of
58 turbulent flows as currently observed within boundary layers over various types of bed-water inter-
59 face in laboratory, lakes, rivers or coastal zones (Rusello and Allard 2012; Thomas and McLelland
60 2015; Brand et al. 2016; Koca et al. 2017; Leng and Chanson 2017). However, to our knowledge,
61 no detailed study of the capabilities of this new probe for describing the complex structure of the
62 natural free surface boundary layers has been undertaken yet. The aim of the present work is to
63 address this question.

64 After a brief description of the instrument and the experimental procedure adopted in this study,
65 we first examine the performance of the acoustic Doppler velocity profiler for measuring the ve-
66 locity field just below the air-water interface when this instrument is deployed in a steady flow
67 generated by water pumps. Then, the observations made by the Vectrino profiler within the wind-
68 driven subsurface water boundary layer are presented in detail. They enable us to characterize
69 with a good accuracy the variations of the mean and fluctuating flow features when the surface
70 boundary layer changes spatially from a laminar to a fully-developed turbulent structure.

71 **2. Experimental arrangement**

72 *a. Experimental configuration*

73 The tests of the Vectrino profiler and the observations performed with this instrument for in-
74 vestigating the water surface flow were carried out in the large *UMS Pytheas* wind-wave facility
75 in Marseille-Luminy (Fig.1a). The facility is composed of a 40 m long, 2.6 m wide, and 0.9 m
76 deep water tank and an air channel 1.5 m in height at the test section. The air flow is generated by
77 an axial fan located in the recirculation flume and then passes through divergent and convergent
78 sections including a settling chamber equipped with turbulence grids. This arrangement facilitates
79 the generation of a homogeneous and very low-turbulence wind at the entrance to the test section.
80 The computer-controlled wind speed can vary between 1 and 15 m s⁻¹. In addition, two water
81 pumps can generate a steady current of a few centimeters per second throughout the tank. At the
82 end of the flume, a long permeable beach damps the wave reflection. In order to investigate the
83 coupled generation of wind waves and surface drift current at short fetches without any influence
84 of preexisting turbulence in the air, particular attention was focused on the air-water junction at the
85 entrance to the water tank, to make it as smooth as possible. To this end, the junction between the
86 bottom floor of the air channel and the water surface was leveled by a 1.5 m long weakly-inclined
87 floating device fixed at the end of the air duct. This prevents the development of large-scale air
88 flow disturbances within the water surface boundary layer for all wind speed conditions.

89 The water flow driven by wind inside the subsurface boundary layer was investigated by means
90 of a Nortek Vectrino Acoustic Doppler Velocimeter Profiler (referenced as profiler or ADVP).
91 Its mode of operation and its set up will be detailed in Sect.2b. The water surface flow velocity
92 was monitored using Particle Tracking Velocimetry (PTV). For this purpose, small paper drifters
93 were injected from the air flow just above the water surface two meters upstream of the measuring

94 section and their motions were recorded by a video camera looking down vertically from the top
95 of the tunnel. The field of view was 0.2 m in the wind direction with the head of the profiler visible
96 in the downwind part of the image to provide a precise fetch reference. On average, for one flow
97 condition, twenty drifters were monitored by the camera at a frame rate of 5 Hz. The images were
98 then processed by using the PTV algorithm developed by Brevis et al. (2011).

99 To observe wind wave growth simultaneously with drift current development, water surface dis-
100 placements were measured by two high resolution capacitance wave gauges 8 mm apart stream-
101 wise and located 45 mm downstream of the center of the ADVP transceiver. To avoid ripple
102 disturbances generated at the water surface by any rigid mount, these probes are made of two thin
103 sensitive wires (0.3 mm in diameter) hung vertically in the water with a weight. The phase veloc-
104 ity of dominant waves is determined from both wave signals using a cross-spectral method. Both
105 components of the water surface slope were also measured by a single-point laser slope gauge. As
106 first described by Lange et al. (1982), this system is based on the detection of the refraction angle
107 of a He-Ne laser beam at the water surface by an optical receiver. The latter includes a Fresnel
108 lens, a diffusing screen and a dual-axis position sensing diode. The He-Ne laser was mounted
109 vertically at the top of the tunnel and the receiver was immersed at a depth of 0.4 m below the
110 water surface and centered at the same fetch 0.35 m spanwise from the profiler head. To make
111 visualizations easier, all instruments were set up at the test section of the air tunnel equipped with
112 large glass windows located 28 m from the entrance to the tank. To adjust the fetch X , the water
113 surface was covered by a long floating plastic sheet of appropriate length. The fetch refers here to
114 the distance between the downwind edge of the plastic sheet and the profiler head. The reference
115 wind velocity was measured with a Pitot tube located at the center of the tunnel, 0.7 m above the
116 water level at rest and 8 m upwind of the profiler.

117 *b. The Acoustic Doppler Velocimeter Profiler*

118 The Nortek Vectrino profiler is a multi-static acoustic Doppler velocimeter that can measure
119 simultaneously the three water velocity components over a 30 mm long flow section with a spatial
120 resolution as fine as 1 mm and a sampling rate of up to 100 Hz. The sensor head of the device
121 consists of a central active transducer, 6 mm in diameter and emitting at a 10 MHz frequency,
122 surrounded symmetrically by an array of four receivers positioned in two perpendicular vertical
123 planes with a slanting angle of 30 deg towards the center (see Nortek (2013)). This probe was
124 mounted on a fixed, 0.36 m long rigid stem connected to the main waterproof electronics housing.
125 It was deployed vertically in the upward-looking position at a variable distance from the water
126 surface but with the latter generally embedded within the measuring profile. The lowest sample
127 cell of the profile is located 40 mm above the central emitter, i.e. in a region where the flow
128 disturbances caused by the probe are mostly negligible (Rusello et al. 2006). The instrument
129 used in these experiments was owned by Nortek-Med and was operating with the 2013 MIDAS
130 Software to control and collect data.

131 This high-resolution profiler uses a pulse-to-pulse coherent Doppler procedure for measuring
132 instantaneous scattering particle velocity at the intersection of the acoustic transmitter and receiver
133 beams, as first described by Lhermitte and Serafin (1984). Its mode of operation as well as its
134 measurement capabilities and accuracy were reviewed in detail recently in a comprehensive work
135 by Thomas et al. (2017). Therefore, we will just summarize briefly the underlying physics that
136 control the measurements. To perform velocity measurements, successive pairs of acoustic wave
137 pulses separated by a time interval Δt (ping interval) are emitted by the central transducer at a
138 'pulse-repetition-frequency' and scattered back by the acoustic tracers suspended in water to be
139 detected by the receivers. The phase shift between the emitted and the received signals is then

140 analyzed to provide estimates of the velocity of the scattering elements. Due to its geometrical
 141 configuration, the profiler measures the velocity component along the angle bisector delimited
 142 by the transmitter and each receiver beam. For a perfectly-manufactured instrument in which
 143 receivers 1 and 3 are aligned with the streamwise plane of the flow, the four-measured velocities
 144 are transformed into a Cartesian coordinate system, the longitudinal u and the first vertical w_1
 145 velocity components being obtained from beam velocities 1 and 3 and the transverse v and the
 146 second vertical w_2 components from beam velocities 2 and 4 according to Eqs. (1):

$$\begin{aligned}
 u &= \frac{b_1 - b_3}{2 \sin \alpha} & \text{and } w_1 &= \frac{b_1 + b_3}{2 \cos \alpha}, \\
 v &= \frac{b_2 - b_4}{2 \sin \alpha} & \text{and } w_2 &= \frac{b_2 + b_4}{2 \cos \alpha}.
 \end{aligned}
 \tag{1}$$

147 where b_i , with $i = 1$ to 4, are the beam velocities and α which varies around 15° along the measur-
 148 ing profile is the bisector angle between the transmitted and any received beams. In practice, the
 149 instrument configuration is not perfect and the transformation of beam velocities into Cartesian
 150 velocities involves a 4×4 matrix obtained by a manufacturer calibration and set up directly into
 151 the instrument software. However, from Eqs. (1), it is easy to see that the noise affecting velocity
 152 signals will be significantly larger on the longitudinal u and crosswise v components than the ver-
 153 tical ones, w_1 and w_2 , the ratio of the respective variances varying typically as $\tan^{-2} \alpha$, i.e. in the
 154 magnitude range from c.a. 7 to 27.

155 As acoustic Doppler velocimeters are known to work poorly in clear water due to the low acous-
 156 tic energy level backscattered to receivers, the quality of velocity data collected by these instru-
 157 ments is largely dependent on the quality and the density of the acoustic tracers present or in-
 158 troduced into the water flow (Blanckaert and Lemmin 2006; Thomas et al. 2017). Owing to the
 159 difficulties of seeding in bulk for a long period of time a volume of water as large as 100 m^3 with-
 160 out severe pollution due to particle sedimentation, the best method we found for seeding this flow

161 efficiently was to generate hydrogen microbubbles by electrolysis, as suggested by Blanckaert and
162 Lemmin (2006). For generating a regular and reproducible population of tiny hydrogen bubbles
163 at a well-controlled rate in water, we chose to use 0.1 mm thick copper wires for the cathode and
164 a 4 mm diameter aluminum rod for the anode. The bubble flux was then adjusted by varying the
165 DC voltage between the two electrodes (around 20 V). For these experiments, the seeding system
166 was composed of two copper wires and one aluminum rod, 0.56 m long in the transverse direction
167 and 5 and 30 mm apart in the vertical direction (Figs.1b-c). The electrodes were fixed horizontally
168 at the front of a frame by means of two vertical insulating stems, 0.1 m long, and the frame itself
169 was set up on a vertical displacement system to adjust the wire depth for each flow condition.
170 Note that during these experiments, this frame did not cause significant disturbances of the sur-
171 face flow because it was generally immersed in the return flow which develops below the surface
172 drift current but in the opposite direction and the 12 cm long vertical stems were located 0.28 m
173 crosswise on either side of the sensor head (Fig.1c). In addition, to avoid as far as possible bubble
174 buoyancy-induced disturbances inside the water surface boundary layer at the measuring section,
175 but keeping mostly a homogeneous distribution of bubbles within this layer, the seeding system
176 was placed at a distance of about 0.1 m upstream of the sample velocity profile. We will discuss
177 more extensively the behavior of these tiny bubbles as passive flow tracers hereafter because it was
178 found to be strongly dependent on wave and turbulence conditions. Note that the localized bubble
179 seeding of the uppermost water layer enables visualization of the flow and thus provides a first
180 overview of its main turbulent features. However, this seeding set-up does not allow simultaneous
181 measurement of the velocity field in the return current.

182 3. Test Measurements

183 As pointed by many authors (see for instance Rusello and Allard (2012); Koca et al. (2017)),
184 to perform measurements in the immediate vicinity of a solid wall or an interface by means of an
185 acoustic Doppler velocimeter raises specific difficulties caused by the acoustic wave reflections
186 at these boundaries. Therefore, before investigating more complex water flow dynamics as the
187 wind-induced surface flow, the ability of the profiler to measure velocity fields in the thin layer
188 just beneath the water surface has been examined in detail. To better assess the quality and the
189 accuracy of the velocity data collected in this layer, preliminary tests were performed for a steady
190 flow generated by two recirculating water pumps. For these tests, the profiler sensor was placed
191 on the centerline of the large wind-wave tank at 28 m fetch. The transceiver was fixed vertically
192 at a distance of 6.8 cm from the water surface at rest, as measured by the bottom check facility of
193 the profiler just before the experiments, and the seeding system was immersed 0.10 m upstream
194 with the upper wire at a distance of 2 cm from the surface. Velocity profiles were sampled over 30
195 levels with 1 mm high measuring cells. As recommended by the manufacturer, the *adaptive once*
196 *ping interval mode* was used, the velocity range and the power level being adjusted at 0.1 m s^{-1}
197 and 'High-' setting values, respectively. For each measuring condition, time sequences of 240 s
198 were recorded at a sampling frequency of 30 Hz. The statistical analysis of flow properties was
199 performed on raw velocity data, the Goring and Nikora (2002) phase-space thresholding method
200 for detecting spikes revealing that those represent less than 1% of the collected data in test experi-
201 ments. Furthermore, this method is not applicable for flows oscillating randomly as observed just
202 beneath a wavy air-water interface.

203 To illustrate these measurements, the instantaneous profiles of the four measured velocity com-
204 ponents are displayed in a space-time representation during 8 s in Figs. 2b-e. The velocity values

205 are given by a color code and referenced to a right-handed Cartesian coordinate system with the
206 longitudinal x axis oriented downstream and the vertical z axis toward the water surface. In ad-
207 dition, the beam-averaged acoustic power backscattered to the receivers is plotted in Fig. 2a in
208 a similar representation. Acoustic power will be used in this work for evaluating the data qual-
209 ity rather than signal-to-noise ratio (SNR), both quantities being linked by a linear relationship
210 (Thomas et al. 2017). For this flow, in the absence of well-developed small-scale turbulence,
211 Fig. 2a shows that the seeding bubbles are carried by the fluid at the location of the profiler over
212 about a 1.5 cm thick layer in which the acoustic power is very high, of the order of -20 to -10 dB.
213 On both sides of this layer, the acoustic power drops drastically owing to the combined effects in
214 the decrease of the scattering particle density in water and the decrease in the scattering volume as
215 analyzed in detail in Brand et al. (2016) and Thomas et al. (2017). Very close to the water surface,
216 the high power values observed should be ascribed to the strong acoustic wave reflection occurring
217 at this boundary. Otherwise, the instantaneous streamwise velocity profiles observed in the sam-
218 ple layer (Fig. 2b) are rather flat, with values of the order of 8 cm s^{-1} , except in the region very
219 close to the water surface where velocity vanishes. The spanwise velocity profiles present most of
220 the time values around zero. These profiles also exhibit large-scale velocity fluctuations, typically
221 of 1 to 2 cm s^{-1} , due very likely to persistent vortices generated by the pumps and advecting
222 downstream without significant damping. The w_1 and w_2 vertical velocity profiles are estimated
223 independently from bubble echoes backscattered in the longitudinal and transverse vertical planes
224 respectively. These profiles are very similar, displaying values close to zero. In the upper levels
225 however, positive values up to 1 cm s^{-1} can be observed. This small vertical velocity gradient may
226 come from a slight buoyancy force acting on bubbles, in particular on the largest ones.

227 The properties of the acoustic return signals and the velocity field averaged over the whole time
228 sequence are presented in Fig. 3. Fig. 3a shows the mean profiles of the acoustic power (also called

229 amplitude) scattered to the four receivers versus the distance h_t to the central transducer. These
230 profiles vary very similarly, except that the beam 1 amplitude is 10 to 15 dB lower than the other
231 three. The various results obtained here, in particular the fact that the beam 1 variation with depth
232 matches the other variations, suggest this distinct behavior is very likely due to a lower response of
233 the receiver 1 electronics rather than a receiver misalignment. Accordingly, phase measurements
234 made by this device will be more sensitive to noise. Otherwise, the beam amplitude profiles exhibit
235 the same typical shape, with a maximum at a level of c.a. 5 cm. This region is generally associated
236 with the *sweet spot* for which the overlap between the transducer and receiver beams is optimal
237 (Brand et al. 2016; Thomas et al. 2017). On both sides of the maximum, the beam amplitude
238 decreases regularly but more rapidly in the layer above owing to the decrease of the bubble density
239 at these flow levels. This decrease in bubble density also shifts the location of the amplitude
240 maximum to a distance of 4.9 cm from the emitter rather than 5.0 to 5.2 cm as normally expected
241 owing to the sensor geometry (i.e. around cells 10 to 12). Except for receiver 1, the average
242 strength of the return signals scattered by bubbles remains very high, above -30 dB, thus indicating
243 the satisfactory seeding quality (note that for all beams, a -30 dB amplitude value corresponds
244 roughly to a 30 dB SNR). In the subsurface layer above 6.5 cm, the four beam amplitudes increase
245 drastically due to contamination of the receiver signals by acoustic wave reflections from the water
246 surface. The power maxima observed at the 6.87 cm level coincides very well, i.e. within the 1 mm
247 profiler cell resolution, with the height of the water surface at rest as indicated by a dashed line in
248 the graph. For such water flows with a flat air-water interface, the depth of the region contaminated
249 by the surface echo is thus estimated at c.a. 4 mm (i.e. the depth including the first three cells just
250 below the water surface).

251 The high quality of the seeding obtained when using electrolysis hydrogen bubbles is also illus-
252 trated by the high values of the correlation coefficients observed in the surface layer, as seen in

253 Fig. 3b. Except for beam 1, these values are higher than 97%, barely varying with height. The
254 variations of the correlation coefficient on both sides of the sweet spot are much more pronounced
255 for beam 1, owing to the weak return signal amplitude, but the observed values remain within
256 the generally acceptable magnitude range for accurate velocity measurements, i.e. above 90%
257 (Rusello and Allard 2012).

258 The vertical variation of the four mean flow velocity components measured by the Vectrino
259 profiler is given in Fig. 3c. The mean streamwise velocity observed within the surface layer is
260 constant except in the region contaminated by the surface echo. There, the measured velocity
261 drops nearly to zero. The velocity values observed below oscillate slightly with depth around a
262 value of 7.9 cm s^{-1} but are in a very good agreement with the average velocity estimated from the
263 small float displacements at the water surface recorded by the camera. As the air-water interface is
264 approached, pulse-to-pulse phase measurements and hence velocity values are affected by surface
265 echo. This is corroborated by the observed trend of the beam 1 amplitude (or correlation) with
266 transducer distance, its decrease comes to a halt at a depth of 3.9 mm and then reverses. Otherwise,
267 as expected for a sensor well aligned with the longitudinal flume axis, the mean spanwise velocity
268 wavers slightly with depth around zero, reaching at most $\pm 0.5 \text{ mm s}^{-1}$. Both estimates of the
269 mean vertical velocity vary with depth in the same way, from small negative values (less than
270 1 mm s^{-1}) at the bottom of the sample layer to more significant positive values, up to 3.5 mm s^{-1} ,
271 close to the water surface. This variation may result from buoyancy effects dependent on bubble
272 size distribution, the largest bubbles being preferentially conveyed at the highest levels. Note that
273 spurious vertical and spanwise velocity measurements are also obtained in the subsurface layer
274 when the acoustic backscattered power from bubbles is contaminated by surface echo. Hereafter,
275 we will discard all velocity measurements within this c.a. 4.0 mm deep layer.

276 To evaluate the performance of the profiler for investigating the turbulent flow structure, the
277 vertical variations of velocity variances are presented in Fig. 3d. Despite the fact that the instru-
278 ment does not satisfy all the assumptions required for this evaluation, in particular the condition
279 of isotropy owing to the lower value of the beam 1 amplitude, the method proposed by Hurther
280 and Lemmin (2001) has been used to provide a rough estimate of the noise variance affecting each
281 velocity component, as shown in Fig. 3e. Such estimates can be made because for evaluating the
282 velocity field in a given plane, the contribution of the beam velocities from the other transverse
283 plane remains small. So, as a first approximation, we can assume that the respective noises asso-
284 ciated with both measured vertical velocities are not correlated. This method provides here only
285 an order of magnitude estimate of noise variances. We should point out that the more elaborate
286 method proposed recently by Thomas et al. (2017) modifies only marginally estimates of these
287 quantities, very likely because the sensor does not fulfil the condition of noise isotropy between
288 beam 1 and beam 3 required for applying it.

289 As seen in Fig. 3d, the longitudinal velocity variance $\langle u'^2 \rangle$ exhibits a variation with height
290 typical of highly noisy ADVP measurements (Blanckaert and Lemmin 2006). In fact, this quan-
291 tity remains almost constant, of the order $0.5 \text{ cm}^2 \text{ s}^{-2}$, in the region centered at the sweet spot
292 around 5 cm, but increases considerably on both sides of the sampled layer, reaching at such
293 levels values higher than $1 \text{ cm}^2 \text{ s}^{-2}$. Thus, this profile appears largely controlled by the noise
294 variance dependency on sensor distance (Fig. 3e). Compared to the longitudinal component, the
295 transverse velocity variance profile presents a more expected shape for this type of turbulent flow,
296 the observed velocity fluctuations keeping a constant intensity throughout the surface layer free
297 from contamination by surface echo. Nevertheless, one can note that $\langle v'^2 \rangle$ increases a little,
298 up to 30%, for heights corresponding to the deepest measuring cells of the profiler, without any
299 significant increase in the noise variance there (Fig. 3e). The noise variance contributes only 10

300 to 20 % to $\langle v'^2 \rangle$ at all depths. Both estimates of the vertical velocity variance $\langle w_1'^2 \rangle$ and
 301 $\langle w_2'^2 \rangle$ as well as their normally noise-free covariance $\langle w_1'w_2' \rangle$ are very low as expected for
 302 a flow developing below a horizontal surface. Their respective values do not exceed $0.1 \text{ cm}^2 \cdot \text{s}^{-2}$
 303 and vary with height very similarly. The profiles exhibit a parabolic-like shape, $\langle w_2'^2 \rangle$ and
 304 $\langle w_1'w_2' \rangle$ following practically the same curve but $\langle w_1'^2 \rangle$ varying with depth with a much more
 305 pronounced curvature. Consistently, the estimated $\langle w_2'^2 \rangle$ noise variance is almost constant over
 306 the whole boundary layer, with values of less than 15% of $\langle w_1'w_2' \rangle$ while that of w_1' fits a highly
 307 symmetrical parabolic-like curve, reaching values from 30 to 90% of $\langle w_1'w_2' \rangle$.

308 To get a better idea of the origin of the noise, time series and frequency spectra of the velocity
 309 signals are displayed in Fig. 4 and Fig. 5 for three heights, one close to the sweet spot at $h_t = 5.2 \text{ cm}$
 310 and the two others at the edges of the sampled layer, i.e. at $h_t = 4.1 \text{ cm}$ and 6.35 cm . In Fig. 5, we
 311 also report the normally noise-free cospectrum of both vertical velocity records w_1 and w_2 . The
 312 spectra are computed using the classical Welch method over successive time-series segments of
 313 2048 data points weighted by a Hanning window of same length with an overlap of 1024 points.
 314 At the sweet spot, the vertical velocity signals vary with time in the same way, exhibiting only
 315 tiny fluctuations at high frequencies, and sporadically more significant spikes (Fig. 4). The w_2
 316 spectrum and the w_1w_2 cospectrum almost collapse into the same curve, except at frequencies
 317 above 3 Hz where w_2 spikes may contribute to higher w_2 spectral levels (Fig. 5). The w_1 spectrum
 318 deviates from that of w_2 but to a lesser extent, being slightly higher at all frequencies. The u and
 319 v time records first exhibit large-scale fluctuations clearly visible in both signals and most likely
 320 associated with turbulent flow vortices. At high frequencies however, they display much more
 321 random fluctuations than vertical velocity signals. In fact, the ratio of the horizontal and vertical
 322 velocity spectra above 10 Hz agrees well with the ratio of the related beam geometrical coefficients

323 at the sweet spot (i.e. 14, see Eqs.(1)). This indicates that noise fluctuations rather than turbulent
324 motions dominates the spectral energy densities in this frequency range.

325 In Figs. 4b-c, the velocity signals observed at the edges of the sampling layer are undoubtedly
326 much noisier, displaying much wider fluctuations at high frequencies. Accordingly, below 1 Hz,
327 longitudinal and transverse velocity spectra match quite well those observed at the sweet spot
328 (i.e. within the confidence interval) but beyond this frequency, they exhibit much higher levels,
329 in particular the longitudinal component at the nearest height above the transducer (up to a factor
330 10). Conversely, the vertical velocity spectra differ from those observed at the sweet spot for all
331 frequencies, and this is more noticeable at low frequencies at the deeper level ($h_t = 4.1$ cm) but
332 above 1 Hz at the upper level ($h_t = 6.35$ cm).

333 Fig. 3f presents the variations with height of velocity variances corrected from noise by two
334 different methods: i) by using the Hurther and Lemmin (2001) method as discussed above (con-
335 tinuous lines), ii) by summing velocity spectra at low frequencies, up to a cut-off of 1 Hz (dotted
336 lines), following Brand et al. (2016). This figure reveals that for this low-turbulence flow, the
337 second method provides much better results. One can see that $\langle u'^2 \rangle$ and $\langle v'^2 \rangle$ are constant
338 and of similar magnitude, typically $0.35 \text{ cm}^2 \text{ s}^{-2}$, over the whole measuring layer. The filtering
339 method appears less efficient for correcting vertical velocity variances from noise, $\langle w_{1,2}'^2 \rangle$ still
340 exhibiting a minimum at the sweet spot and increasing values on both sides of this region. This
341 increase is however stronger at the bottom levels of the surface layer than at the upper levels, sug-
342 gesting a different origin from only a beam decorrelation which affects Reynolds stress estimates
343 (Brand et al. 2016). As expected, the Hurther and Lemmin (2001) method provides rather poor
344 results for the profiler used in this experiment. It appears that this noise correction on $\langle u'^2 \rangle$ is
345 overestimated at the largest distances from the transducer but underestimated at the lowest ones,
346 giving accurate results only in the sweet spot region around 5.2 cm. In addition, this method does

347 not improve estimates of the spanwise velocity variance at the lower heights below 4.5 cm. In
348 essence, according to the underlying assumptions of the method, both vertical velocity variance
349 estimates should collapse into the supposed noise-free $\langle w'_1 w'_2 \rangle$ covariance. If the values ob-
350 served for this quantity at the sweet spot seem to be in good agreement with those obtained by
351 filtering, the $\langle w'^2_{1,2} \rangle$ estimates outside this region look strongly affected by noise, or perhaps by
352 other instrument defects. In fact, the analysis of velocity field properties performed above clearly
353 shows that velocity measurements made by the Vectrino profiler used here are much more prone to
354 noise contamination owing to the low level of the beam 1 amplitude but also it indicates that these
355 measurements may suffer from minor calibration errors, in particular at low distances from the
356 transducer. Below the sweet spot, it appears indeed that velocity measurements in the transverse
357 plane are contaminated by the flow velocity field in the longitudinal plane. Similar anomalies have
358 been detected previously by several authors, such as Zedel and Hay (2011), and a new calibration
359 procedure was proposed by Nortek in 2016. However, the Vectrino profiler software used in the
360 experiments carried out in 2014 does not take into account this new algorithm.

361 **4. Observations of the wind-driven subsurface water flow**

362 The potential of the Vectrino profiler for investigating the dynamical properties of the wind-
363 driven near-surface water boundary layer have been examined in a series of experiments where
364 the velocity field inside the subsurface flow was measured at different stages of wind wave devel-
365 opment. The observations were conducted at two wind speeds U (4.5 and 5.5 m s⁻¹) and eleven
366 fetches X ranging from 1 to 15 m, i.e. for wind wave fields of wave height not exceeding 30 mm.

367 The overall arrangement of the instrumentation adopted for these experiments was the same as
368 described in Sect. 2. The depth of the electrolysis wires and the profiler head was adjusted for each
369 wind and wave condition to get the best bubble seeding within the sample layer and the longest

370 velocity profile below the highest wave crests observed. Thus, the depth of the upper electrolysis
371 wire was kept between 1 and 3 cm and the sensor head was always immersed at a depth greater
372 than 4.0 cm below wave troughs. The velocity range and the acoustic power level of the Vectrino
373 profiler were set at 0.1 m s^{-1} and 'High-'. For each forcing condition, the velocity profiles were
374 recorded simultaneously with wave height and wave slope signals during three time sequences of
375 5 min separated by a time interval of 2 to 5 min. Wave signals were digitized at 256 Hz frequency
376 while the sampling rate for velocity profile records was set at 30 Hz. These records were made
377 after a minimum time period of 30 to 45 min necessary for stabilizing the water flow within the
378 tank after a change in wind or fetch conditions.

379 *a. Wind waves observed at the water surface*

380 Before analyzing the velocity field generated by wind and waves in water, we describe wind
381 wave properties observed at fetches and wind speeds selected for these experiments. Typical time
382 series and spectra of wave height and wave slope observed at 4.5 m s^{-1} wind speed are shown in
383 Figs. 6a-d for several fetches. In addition, the fetch dependence of a few related statistical wave
384 parameters is given in Figs. 6e-h. At the first fetch upstream, the water surface looks practically flat
385 with only tiny oscillations observed in two distinct frequency ranges (6a-b). Wave motions around
386 3 Hz correspond to the small transverse oscillations of the water surface detected at the entrance
387 of the tank, while those around 10 Hz, well-discernible only in the slope spectrum, correspond to
388 the first wind-generated waves. Immediately downstream, Fig. 6a shows that wind waves grow
389 approximately at the same frequency up to 3 m fetch, and a wide dominant peak is emerging from
390 noise at around 10 Hz. In parallel, waves of higher frequencies start to develop, very likely asso-
391 ciated with dominant wave harmonics. The dominant waves clearly display nonlinear shape with
392 round crests and sharp troughs and are arranged in three-dimensional surface patterns (see Fig.2 in

393 Caulliez and Collard (1999)). Beyond 3 m fetch, the dominant peak keeps growing significantly
394 in energy but with a regular shift towards lower frequencies while a saturation range develops at
395 frequencies higher than 20 Hz. The formation of this saturation range results from the rapid devel-
396 opment of parasitic capillary ripples propagating at the front of the highest dominant wave crests,
397 as seen in Fig. 6c-d. The wave slope spectra displayed in Fig. 6b exhibit the same variation with
398 fetch, except that the spectra contain proportionally more energy at high frequencies. At this wind
399 speed, the dominant wave height increases from a few tenths of a millimeter for fetches below 3
400 m to about one centimeter at 15 m fetch, as seen in Fig. 6c or better quantified in Fig. 6e using the
401 variation with fetch of the RMS wave amplitude. In addition, Fig. 6g shows that dominant waves
402 keep roughly the same wavelength up to 4 m fetch, typically 4 cm at 4.5 m s^{-1} , but downstream,
403 this quantity increases gradually up to 15 cm. The nonlinear shape and the related dynamic prop-
404 erties of dominant waves thus shift from capillary-gravity to short gravity wave types. The wave
405 slope spectra enable us to evaluate separately the total mean square slope (*mss*) associated with
406 dominant waves and small-scale wave roughness, respectively. The former has been estimated by
407 spectrum integration from 0.5 to 1.5 times the dominant peak frequency f_d while the latter has
408 been estimated by integration over a scale domain ranging above the smallest of both frequencies,
409 namely 20 Hz and 3.1 times f_d . The variation with fetch of the average dominant wave steepness
410 (i.e. the square root of the dominant wave *mss* denoted ak_d for simplicity) and the small-scale
411 wave *mss* thus estimated are displayed in Figs. 6f-h. We see that ak_d first increases regularly but
412 slowly at fetches lower than 3 m whereas downstream, the growth accelerates significantly up to
413 6 m. Beyond this fetch, the growth slows down drastically, ak_d becoming almost constant, then
414 rating 0.14 for 4.5 m s^{-1} wind speed. In parallel, the short wave *mss*, mss_{eq} , being negligible at
415 the first fetches, starts to grow rapidly beyond 3 m to reach a maximum at 6 m. Downstream,
416 this quantity decreases slowly to a fetch of 8 m beyond which a plateau is observed, amounting

417 to c.a. 5×10^{-3} . The changes with fetch in wave height and slope spectra and the related wave
418 parameters observed at 5.5 m s^{-1} wind speed are very similar, except that they are much faster, the
419 rapid growth of ak_d and mss_{eq} starting at 1.5 m fetch, both quantities then reaching equilibrium
420 values around 8 m but with a mss_{eq} overshoot located at c.a. 3 m.

421 *b. Detection of surface motions by the profiler*

422 A major difficulty when performing single point velocity measurements in the immediate prox-
423 imity of a wavy surface is the lack of precise estimate of the instantaneous location of the velocity
424 sample collected with respect to the surface. This distance to the surface is required for recon-
425 structing in an appropriate way the vertical variation of the averaged properties of the various
426 water flow motions, namely the mean wind-induced drift current, the wave-induced orbital mo-
427 tions and the turbulent motions. In fact, as pointed out in Caulliez (1987), averaging over time
428 measurements made at a fixed Eulerian location in such a shear flow introduces significant bias
429 in estimates of the flow characteristics because the variation in time of the actual depth and the
430 orbital motions are linked in phase. As first noted in Section 3, the Vectrino profiler offers the
431 possibility of retrieving the instantaneous measuring depth when the water surface is embedded
432 in the sampled layer. Accordingly, this knowledge of the colocalized motion of the water surface
433 enables us to reconstruct the velocity field below waves in a curvilinear reference system in which
434 orbital motions do not contribute to the mean flow and vice versa, at least at the first order.

435 Fig.7a presents a time sequence of the vertical profiles of the beam-averaged acoustic power
436 backscattered to the receivers when wind waves are well-developed at the water surface. The z
437 axis origin refers to the water level at rest as measured by the profiler bottom check, i.e a distance
438 of 62 mm from the transceiver. As for the flat-surface channel flow investigated previously in
439 Section 3, the very high acoustic power observed at the water surface ($> -5 \text{ dB}$) compared to

440 the acoustic power backscattered by hydrogen bubbles advected by the flow makes this boundary
441 clearly distinguishable. Here again, this phenomenon is due to the high acoustic wave reflection
442 occurring at the air-water interface. Consistently, taking advantage of this observation, we have de-
443 veloped an algorithm to determine with a resolution of 1 mm the water surface elevation associated
444 with each sampled velocity profile. The procedure is based on the detection of the backscattered
445 acoustic power field maxima. First, each instantaneous acoustic power profile is filtered from noise
446 by applying a smoothing window 3 pixels in height. The maximum of the profile at a given time is
447 then detected from the location of the previous one by searching it within a limited height interval,
448 the latter being estimated as the largest surface wave displacement observed by the capacitance
449 gauge located just downstream from the sensor. To illustrate this procedure, the time sequence
450 of the surface elevation associated with acoustic power maxima observed in Fig.7a is shown in
451 Fig.7b.

452 This method for retrieving the instantaneous water surface elevation at the location of the profiler
453 was validated by comparing the corresponding wave spectrum with that derived from wave probe
454 records. To facilitate the comparison, the wave probe signal was resampled in time at the nearest
455 frequency of the velocity data acquisition rate, i.e. 32 Hz, and in height with a 1 mm resolution.
456 Fig. 7c shows that both wave height spectra overlap very well over the dominant peak centered
457 around 5.8 Hz, the estimates of associated RMS dominant wave amplitude differ by less than
458 1 %. However, the wave spectrum derived from acoustic detection contains higher noise at all
459 frequencies which precludes the detection of higher wave harmonics. This background noise
460 results undoubtedly from the low resolution in height of the wave signal and the lack of high
461 frequency filtering.

462 Using these colocalized wave height measurements and assuming as a first approximation that
463 the orbital motions of dominant waves satisfy the linear wave theory, the velocity field in the upper

464 near-surface layer and its statistical flow properties will be described hereafter along the dominant
465 wave streamlines centered at a distance z from the mean surface level, i.e. along the vertical
466 coordinate \tilde{z} given by the equation:

$$\tilde{z}(z, t) = z + \eta(t) \cdot e^{kz}, \quad (2)$$

467 in which $\eta(t)$ is the instantaneous vertical displacement of the water surface estimated at the
468 location of the profiler and k , the dominant wavenumber derived from the measured dominant
469 wave frequency f_d and phase speed c_d .

470 *c. Mean subsurface flow observed in water*

471 Figs. 8a-b present the vertical distribution of the mean longitudinal velocity as observed in the
472 wind-generated boundary layer just beneath the water surface for all fetches and both wind speeds.
473 When waves ruffle the water surface, these time-averaged velocity profiles are estimated by using
474 the curvilinear coordinate system introduced above. The velocity is then averaged along the dom-
475 inant wave streamlines \tilde{z} centered at a distance z from the mean surface level. To determine the
476 average flow properties from high quality velocity data, sequences of velocity signals for which
477 the acoustic power is higher than -30 dB during a time interval of at least 512 successive points
478 have been selected and then, time-averaged quantities have been computed for the whole selec-
479 tion. In addition, in Figs. 8a-b, velocity data obtained in the uppermost 3.5 mm thick sublayer have
480 been discarded from the profiles (except for the first one as illustration) because there, the acous-
481 tic power backscattered by bubbles to the receivers is contaminated by the water surface echo, as
482 analyzed in detail in Sect. 3.

483 The mean values of the surface drift current are also reported in Figs. 8a-b. They are estimated
484 using PTV with an interpolation at the appropriate fetch. To better appraise the spatial development

485 of the water boundary layer, the variation with fetch of the mean flow velocity observed at the
486 surface and at 10 mm depth is shown in Figs. 8c-d.

487 As marked by three different symbols, the vertical distribution of the mean velocity exhibits three
488 distinct behaviors when the boundary layer develops spatially under wind forcing. At the entrance
489 to the water tank, when the water surface is still flat, the flow is confined in a thin layer of about
490 15 mm depth. There, in absence of waves, the formation of the boundary layer results only from the
491 direct action of the air viscous stress exerted by wind at the interface and the viscous diffusion of
492 the longitudinal momentum downward throughout the fluid. Although the profiler does not enable
493 investigation of the upper part of the flow, we can assume, given the high velocity values observed
494 at the surface, that the rapid decrease of the mean velocity is likely to be linear in the uppermost
495 millimeters of the boundary layer, as shown previously in Caulliez et al. (2007) from a detailed
496 laser Doppler velocimeter (LDV) investigation performed in a smaller tank. Below this highly-
497 sheared sublayer, the mean flow decreases much more slowly at a progressively falling rate towards
498 the lower edge of the boundary layer. Note that the depth where the mean velocity vanishes or even
499 reverses due to the development of a return current in the flume cannot be determined precisely
500 when local particle seeding is used. For this experimental arrangement, ADVP measurements can
501 be made only when tracers move towards the sensor, i.e. for a flow in the wind direction but not
502 for a vanishing-speed or reverse flow. During this laminar stage of flow development, the mean
503 velocity gradually increases with fetch over the entire boundary layer. As shown in Figs. 8c-d at
504 the shortest fetches, these measurements confirm rather well that the surface velocity varies with
505 fetch according to a power law $X^{1/3}$ as expected for a viscous boundary layer accelerated at the
506 surface by a constant wind stress (Caulliez et al. 2007).

507 When waves start to grow more rapidly at the water surface, i.e. at 4 and 5 m fetches for
508 4.5 m s^{-1} wind speed (Fig. 6e-f) and at 2.0 and 2.5 m for 5.5 m s^{-1} , the shape of the mean ve-

509 locity profiles changes dramatically compared to those observed previously at the laminar stage
510 of the boundary layer formation. The mean surface drift velocity drops by about 30 % whereas
511 the boundary layer deepens sharply, significant velocity values being observed at all depths of the
512 measured profiles. This behavior suggests that the boundary layer undergoes an abrupt laminar-
513 turbulent transition. The onset of this phenomenon can also be detected by the high variability
514 of the surface drifter motion observed in this region, as indicated by the large error bars plotted
515 in Figs. 8c-d or checked qualitatively by viewing the recorded microbubble images. This phe-
516 nomenon will be analyzed on a more quantitative basis hereafter, by investigating the turbulent
517 flow properties. Note that the inflectional profile observed at 2 m fetch for 5.5 m s^{-1} wind speed
518 reveals that the boundary layer at this stage of transition is highly unstable.

519 Further downstream, at fetches up to $X = 8 \text{ m}$, the velocities of the surface drift current and the
520 bulk flow increase again by $1 \text{ to } 3 \text{ cm s}^{-1}$ at 4.5 m s^{-1} or a few millimeters per second at 5.5 m s^{-1} .
521 The most striking feature of both flows at this stage of development is that the profiles are very
522 flat, which indicates strong vertical mixing within the boundary layer. Hence, it seems difficult
523 to identify a region where the profiles follow a logarithmic decrease with depth, as expected for
524 turbulent boundary layers along a moving rigid wall. For both wind speeds, the best agreement
525 with such a law would occur between 5 and 12 mm depth but the corresponding values of the
526 friction velocity u^*_w , practically invariant with fetch, prove to be very low, roughly 30 to 40 %
527 of the values estimated from the measured friction velocities in air u^*_a when Reynolds stress
528 continuity is assumed across the air-water interface. At the bottom of the sample boundary layer,
529 i.e. below 20 mm depth, most of the profiles exhibit a small increase in velocity. This behavior is
530 probably due to measurement errors introduced by the Vectrino profiler at the first levels above the
531 sensor owing to calibration defects, as previously detected in Sect. 3. Finally, beyond 8 m fetch
532 for both wind speeds, the variation with fetch of the mean flow velocity observed at all depths

533 including the water surface is characterized by a slow but well-identified decrease. However, the
534 profile shape does not change significantly.

535 Note that the vertical velocity field measured in the boundary layer does not vanish on aver-
536 age, the hydrogen bubbles used as tracers being driven by slight buoyancy forces. The generally
537 positive velocity values observed vary a little with depth, from 2 to 3.5 mm s⁻¹ when the flow
538 is laminar (i.e. values similar to those observed in Sect. 3) to less than 1 mm s⁻¹ when intense
539 turbulent motions develop at the longer fetches.

540 *d. Fluctuating velocity field in water*

541 To evaluate the performance of the profiler for investigating the spatial development of turbulent
542 and wave motions inside the water boundary layer, frequency spectra of the three velocity compo-
543 nent time series observed at 10 mm depth for all fetches are shown in Fig. 9 for 4.5 m s⁻¹. These
544 spectra are computed from each time sequence of 512 points selected for estimating the mean flow
545 profiles and then averaged, as explained above. Note that these individual time sequences are short
546 enough to allow a high rate of data selection, i.e. generally between 60 and 100 % except for the
547 two to three lower cells at the bottom of the sample layer where the bubble seeding is less regular.
548 However, they are long enough (17 s) to capture most of the fluctuating motions of the flow except
549 probably for the conditions for which the flow is highly nonhomogeneous, i.e. at fetches where the
550 laminar-turbulent transition occurs or large-scale longitudinal vortices develop downstream. This
551 will be described in detail later on. To minimize the noise contribution, the vertical velocity spec-
552 tra are computed from the cospectra of both vertical velocity signals measured in the longitudinal
553 and transverse planes respectively.

554 One can see in Fig. 9a that the vertical velocity spectra $S_{w_1 w_2}$ exhibit two different trends as-
555 sociated with two frequency ranges and two types of water motions. At low frequencies, below

556 c.a. 1 Hz, $S_{w_1w_2}$ decreases regularly with frequency, first quite slowly up to 0.5 Hz, then a lit-
557 tle more rapidly. The spectral level increase with fetch observed in this frequency range is due
558 to the development of vertical turbulent motions in water. At frequencies above 1 to 2 Hz, the
559 vertical velocity spectra show one or two well-marked peaks located at the same frequencies as
560 those observed in wave spectra displayed in Fig. 6. Within this range, $S_{w_1w_2}$ appear to be largely
561 dominated by the orbital motions of surface waves, their shape and their energy level changing
562 with fetch in the same way as wave spectra. Consistently, the wide maximum observed around
563 2.5 Hz at fetches shorter than 3 m refers to transverse surface oscillations detected at the entrance
564 of the water tank while the conspicuous peak that grows sharply with fetch with a shift towards
565 low frequencies from 10 to 3.5 Hz, refers to wind-amplified dominant waves. However, the orbital
566 motions associated with small-scale waves that contribute to the spectral energy saturation range
567 are not clearly distinguishable in velocity spectra primarily because of the low rate of velocity
568 sampling used here (30 Hz) but also due to the faster decrease of such motions with depth.

569 The streamwise and spanwise velocity spectra shown in Figs. 9b-c contain more energy than the
570 vertical ones at frequencies below 1 Hz, their level being multiplied by a factor of 10 at fetches
571 shorter than 3 m and a factor of 2 to 3 at longer fetches when turbulent motions develop. At the
572 shortest fetches too, the streamwise velocity spectra above 1 Hz change in slope and start to level
573 off at a spectral value of $1-3 \times 10^{-3} \text{ cm}^2 \text{ s}^{-1}$. This spectral floor may originate from small-scale
574 velocity fluctuations which develop simultaneously with larger disturbances of laminar boundary
575 layers or the first surface motions but also the profiler noise that is much higher for the u and v
576 components than w (see Sect.3). As a consequence, the longitudinal orbital motions emerge from
577 this background floor only at 3 m fetch. Downstream, at 4 m fetch, the orbital motions also barely
578 emerge from the continuous spectral decay of highly-energetic turbulent motions which develop
579 intensively at scales above 1 Hz following the laminar-turbulent boundary layer transition. Further

580 downstream, the spectral peaks associated with higher-amplitude dominant wave motions become
581 clearly distinguishable, especially since the spectral level of turbulent motions decreases a little
582 there. Nevertheless, the spectral dominant peaks observed for u contain noticeably less energy
583 than those observed for w . The behavior of the spanwise velocity spectra look very similar to
584 those of the streamwise component, except for two features. At low frequencies, the spanwise
585 velocity fluctuations, being of the same order of magnitude as the streamwise ones at 0.1 Hz, decay
586 much more slowly as frequency increases, the spectra remaining practically flat up to 1 Hz. At
587 frequencies above 3 Hz, the peak of dominant wave motions can be clearly identified only for 15 m
588 fetch. The ratio of the orbital motions in the spanwise direction to those in the streamwise direction
589 then looks unexpectedly small, of the order of one tenth. When referred to the linear wave theory,
590 this ratio should normally be comparable to the mss ratio observed for these dominant waves, i.e.
591 of the order of 0.3. At fetches downstream 3 m, the spectral decay characteristic of turbulent flow
592 motions at high frequencies thus can be detected almost to 10 Hz for the v component but the
593 relatively high spectral level observed in this frequency range cannot explain the orbital velocity
594 ratio discrepancy.

595 The variation with fetch of the three velocity component spectra below 1 Hz as observed in Fig. 9
596 reflects quite well the peculiar growth of turbulent motions in water at the subsurface flow laminar-
597 turbulent transition. At the three shortest fetches, the spectral level of velocity fluctuations is very
598 low, of the order of $10^{-2} \text{ cm}^2 \text{ s}^{-1}$ for u and v and $10^{-3} \text{ cm}^2 \text{ s}^{-1}$ for w . The significant increase
599 of the velocity spectra observed at 2.5 and 3 m fetches indicates that stronger flow disturbances
600 start to grow. The development of the first viscous shear flow instabilities is corroborated by
601 visualizing the bubble images in which the formation of narrow elongated streaks can be clearly
602 distinguished, as shown for instance in Fig. 10a. The surface flow laminar-turbulent transition
603 primarily revealed at 4 m fetch by the sudden change in shape of the mean velocity profile (Fig. 8)

604 can also be detected in Fig. 9 from the drastic change of spectra both in shape and in energy. At
605 10 mm depth, the breakdown of the boundary layer is characterized by a rapid increase of the
606 spectral energy density at low frequencies, the streamwise component reaching its highest level at
607 4 m fetch, typically $3 \text{ cm}^2 \text{ s}^{-1}$. Similarly, S_v and S_w reach high levels at low frequencies but more
608 strikingly, at intermediate scales too, i.e. within the range from 0.5 to 5 Hz. This feature depicts
609 undoubtedly the development within the flow of intense energy turbulent microstructures resulting
610 from the breakdown of larger-scale streamwise vortices. Note that the ratio of S_w to S_u increases
611 significantly during the transition, reflecting the enhancement of the vertical turbulent mixing. At
612 5 and 6 m fetches, the substantial decrease of velocity spectra at low to intermediate frequencies
613 indicates that the generation of turbulent motions falls. This suggests that momentum transfer
614 across the interface and throughout the subsurface shear layer is not high enough to maintain the
615 turbulence production within the boundary layer at its highest level. However, the development of
616 an inertial subrange at intermediate frequencies above 0.8 Hz as identified by the classical $-5/3$
617 power decay shows that a certain energy balance between turbulence production and dissipation
618 processes tends to be reached at fine scales. Note too that the spectral level of spanwise velocity
619 fluctuations remains the highest at all scales, a distinctive feature which highlights the specific
620 nature of the wind-driven near-surface turbulent boundary layer. Finally, at fetches larger than
621 8 m, Fig. 9 shows that S_v and S_w increase again at low frequencies. These spectral changes may
622 result from the growth of large-scale coherent streamwise vortices better known as Langmuir cells,
623 as strongly suggested by the bubble trajectories visualized in the recorded images (Fig. 10b).

624 The vertical distributions of the standard deviations of the three velocity components associated
625 with low frequency turbulent motions and dominant wave orbital motions observed for all fetches
626 at 4.5 m s^{-1} are shown in Figs. 11 and 12. These quantities are computed by spectral integration
627 over respectively the low frequency range, from 0.05 to 1 Hz, and the dominant peak, from 0.5 to

628 1.5 times f_d . Like the mean velocity profiles, the RMS velocity profiles associated with turbulent
629 fluctuations exhibit three distinctive shapes associated with the three stages of the boundary layer
630 development. At fetches shorter than 3 m, when the flow is laminar, the turbulence intensities are
631 quite small, of the order of or less than 2 mm s^{-1} , apart from the streamwise component at 2.5
632 and 3 m fetches. The increase occurring there at all depths reflects the growth of the longitudinal
633 shear flow disturbances associated with the formation of viscous streaks aligned in the streamwise
634 direction (Fig. 10a). The development of such coherent structures characterized by high-speed and
635 low-speed regions randomly distributed in the spanwise direction makes the near-surface velocity
636 field observed at a fixed location more and more variable due to the increasing nonhomogeneity
637 of the flow. The viscous flow instabilities originate and grow in the highly-sheared subsurface
638 layer, creating there the strongest velocity disturbances. This would explain the regular increase
639 of $\sqrt{\langle u^2 \rangle_{LF}}$ observed at the uppermost depths, the noise contribution below 1 Hz being small
640 (see Sect. 3).

641 The most drastic change in the profiles of turbulence intensities occurs at 4 m fetch, following
642 the laminar-turbulent transition of the near-surface flow. These quantities increase by a factor
643 of 3 to 6, the largest growth being observed for the spanwise and vertical components at depths
644 between -15 and -20 mm, i.e. just below the bottom of the laminar boundary layer observed
645 upstream. The strong but localized high-speed fluid ejections from the subsurface region towards
646 the bulk water associated with the breakdown of the boundary layer can give rise at these depths
647 to very wide space and time velocity fluctuations. When approaching the surface, the turbulent
648 fluctuations decrease slowly for the longitudinal component but more significantly for the vertical
649 and transverse components. These variations can be explained when it is assumed firstly, that
650 the less intense energy and small-scale flow disturbances remain confined within the uppermost
651 region of the flow, and secondly, that the presence of the water surface prevents in this region flow

652 instabilities from growing in the upward direction, and thus in the spanwise direction too. Note
653 again that at all depths, the turbulence intensity in the vertical direction is less than half that of the
654 horizontal ones.

655 Consistent with the decrease of the mean flow at fetches immediately downstream of the laminar-
656 turbulent transition, the turbulent velocity fluctuations noticeably decrease there too, as previously
657 pointed out from the velocity spectrum variations. The drop, of the order of $1/3$, is particularly
658 marked for the streamwise component at all depths. For the spanwise and vertical components,
659 the drop is confined to the lower part of the sample layer. Thus, the variations with depth of such
660 quantities greatly diminish, indicating that the structure of the turbulent flow tends to be more
661 uniform and steady throughout the surface boundary layer. Unlike the mean flow which grows
662 noticeably between 5 and 8 m fetches, the turbulence intensities at this stage of flow development
663 remain remarkably constant, their variations with fetch being limited to c.a. 1 mm s^{-1} . At
664 fetches larger than 8 m however, when the mean flow decreases again, the spanwise and vertical
665 turbulence intensities start to grow significantly with fetch within most of the surface layer, except
666 naturally the vertical intensity at the smallest depths. Instead, the streamwise turbulence intensity
667 remains capped at the same level in the central part of the boundary layer. As suggested previously,
668 this fluctuation behavior appears to be linked with the development of Langmuir cells within the
669 surface flow.

670 To complete this analysis, in particular to assess the degree of confidence of turbulence intensity
671 measurements close to the surface and at the bottom of the sample layer, complementary flow
672 properties have been examined to appraise noise contamination, since the Hurther and Lemmin
673 (2001) method cannot be used for the present anisotropic ADV sensor. As discussed previously,
674 the first criterion is based on the variation with depth of the backscattered acoustic power which
675 enables determination of its trend reversal depth when approaching the surface, from a slow de-

crease to a fast increase. Secondly, the Doppler noise from two sampling cells being not correlated, the level of the spectral coherence between velocity records observed at two successive depths appears to be a good indicator for estimating noise contribution to velocity fluctuations of scales larger than one cell size. For fetches larger than 3 m, when the flow is fully turbulent, we thus found that the spectral coherence at frequencies lower than 1 Hz is always higher than 0.9 for the vertical component and 0.8 for the spanwise one, indicating that the contribution of noise to these turbulence intensities is rather weak. At short fetches, when the flow is laminar, the velocity fluctuations at low frequencies being much smaller, the spectral coherence drops to 0.6 or less, in particular at the uppermost depths, indicating that noise contributes much more significantly to the raw turbulence intensity estimates. For the streamwise component, the beam 1 acoustic power being quite low (see Sect. 3), the spectral coherence is generally weaker than for the two other velocity components, reaching values higher than 0.8 at frequencies up to c.a. 0.3 Hz but diminishing rapidly beyond. The decrease is particularly marked at the two smallest depths for fetches longer than 6 m for which outlier turbulence intensity values are observed. In fact, at fetches where wave height becomes significant, i.e. 1 cm typically, the minimum of the beam 1 acoustic power moves progressively from 3.5 to 5.5 mm depth. The streamwise velocity spectra observed in this region appear much noisier at all frequencies too. Therefore, these data are considered to be highly contaminated by noise and have been discarded from the streamwise turbulence intensity profiles (plotted just as symbols in Fig. 11).

The turbulence intensity profiles also present a few outlier data at the bottom of the sample layer, in particular for the streamwise component at 11 m fetch. The velocity measurements made at this fetch differ from the others by a low rate of selection of time sequences for which the averaged acoustic power remains above -30 dB. Below 18 mm depth, this rate decreases to less than 50 % due to a lack of bubble seeding. This very low percentage is linked undoubtedly to

700 the onset of Langmuir circulations and the formation of divergence zones in which fluid from
701 the deep water rises upward without bubbles (see Fig. 10b). This phenomenon may introduce
702 bias in measurements of averaged quantities as mean velocities or turbulence intensities, since
703 the flow properties observed under such conditions are more representative of convergence zones
704 characterized by downward motions than divergence zones characterized by water upwellings.

705 Fig. 13 displays the dependence on depth and fetch of the vertical turbulent momentum flux
706 $\langle -u'w'_2 \rangle$ estimated by integration of the longitudinal and vertical velocity cospectra from 0.05
707 to 1 Hz frequencies. Although the effects of beam decorrelation are certainly more pronounced,
708 this covariance was chosen to estimate the vertical flux rather than $\langle -u'w'_1 \rangle$ because its contam-
709 ination by the Doppler noise proves to be lower for the instrument used here. Owing to the fact
710 that the beam 1 and beam 3 acoustic powers and then the related noise variances differ widely, the
711 noise variance should contribute significantly to $\langle -u'w'_1 \rangle$ estimates (Brand et al. 2016). Fig. 13
712 confirms quite clearly the changes in fetch of the near-surface boundary layer structure described
713 above in detail. During the laminar flow regime observed for fetches less than 3 m, $\sqrt{\langle u'w'_2 \rangle}$
714 has insignificant values, thus corroborating the absence of vertical momentum transport by the
715 weakly fluctuating velocity field observed at this stage of flow development. The sudden increase
716 of the vertical Reynolds stress observed at 4 m fetch, in particular at the greater depths of the sam-
717 ple layer, attests that the breakdown of the surface boundary layer occurs there or more likely just
718 upstream. The formation of large-scale turbulent eddies and the associated bursting events which
719 characterizes this sharp by-pass transition to turbulence as described in Caulliez et al. (2007), is
720 then responsible for strong downward advection of the streamwise momentum from the subsur-
721 face layer towards deeper flow regions. Consistent with the mean flow and turbulence intensity
722 decrease with fetch, the vertical momentum flux also decreases drastically downstream of the
723 laminar-turbulent transition but become practically invariant with depth and fetch. This suggests

724 that turbulent mixing throughout the surface boundary layer becomes homogeneous even if the
 725 respective contribution to momentum transport from the viscous instabilities generated close to
 726 the water surface and the large-scale Langmuir circulations may change, seeming specific to each
 727 wind and wave field condition.

728 The vertical distribution of the RMS amplitude of dominant wave orbital motions estimated
 729 from the three velocity spectra is given in Fig. 12 for 4.5 m s^{-1} and all fetches. The exponential
 730 decrease with depth of the vertical RMS orbital velocity as derived from the RMS dominant wave
 731 amplitude by using the linear wave theory is also plotted in Fig. 12c for fetches larger than 5 m
 732 (dashed lines). This decrease is expressed by the following equation:

$$\sqrt{\langle \tilde{u}_{\text{th}} \rangle^2} = \sqrt{\langle \tilde{w}_{\text{th}} \rangle^2} = \sqrt{\langle \eta^2 \rangle} kc \cdot e^{kz} \quad (3)$$

733 First, Fig. 12 shows that the vertical component of the orbital motions decreases with depth and
 734 increases regularly with fetch but the observed values deviate widely from those estimated by the
 735 linear theory. However, as highlighted in Fig. 12d, the ratio of these quantities does not vary with
 736 depth, keeping remarkably constant values in the layer where the vertical orbital velocity remains
 737 significant (compared to noise or turbulent fluctuations), i.e. down to a depth of about 12 mm at
 738 5 m fetch and 20 mm at 15 m fetch. This ratio also increases regularly with fetch, from roughly 0.6
 739 at 5 m to 0.8 at 15 m. These findings indicate that the RMS values of the vertical orbital velocity
 740 observed in this wind-driven sheared boundary layer decays exponentially with depth, following
 741 a $\exp -kz$ trend with k being estimated from f_d and the measured phase speed c_d . Fig. 12d also
 742 shows that the related amplitude can be approximated by $ak_d(c_d - U_s)$ rather than ak_dc_d as derived
 743 from Eq. (3), i.e. by removing the effect of the drift current on the measured phase speed as
 744 predicted by the surface continuity equation. To support this outcome, further experimental but

745 also theoretical works are required because, to our knowledge, this basic question has not been
746 solved for these specific shear flow conditions.

747 The variation with depth of the dominant wave orbital velocity in the streamwise direction, as
748 shown in Fig. 12a, appears much more complex. At short fetches up to 8 m, $\sqrt{\langle u^2 \rangle_d}$ decreases
749 rapidly in the upper layer of the flow above 10 mm but below this depth, this quantity remains
750 practically invariant, taking rather substantial values in particular at the first stages of the transi-
751 tion to turbulence at fetches just larger than 4 m. As previously detected in Fig. 9b, this trend
752 corroborates the fact that quite intense turbulent motions develop at small scales above 5 Hz, thus
753 contributing significantly to the fluctuating velocity field within the dominant wave spectral range.
754 At fetches longer than 8 m, as orbital motions become dominant in the wave frequency range,
755 $\sqrt{\langle u^2 \rangle_d}$ varies with depth in a similar way as that of $\sqrt{\langle w^2 \rangle_d}$ but only in the central part
756 of the sample layer. When approaching the surface, the $\sqrt{\langle u^2 \rangle_d}$ variation with depth differs
757 dramatically, a drastic slowdown of the growth even leading to a trend reversal being observed.
758 This unexpected behavior may result from the approximate reconstruction of the orbital velocity
759 streamline depth from the linear theory because small errors in \tilde{z} estimates could introduce large
760 deviations in streamwise orbital velocity measurements. In this surface layer, owing to its high
761 shear, these errors, even minute, may indeed affect directly average outputs obtained when the
762 fluctuating velocity field is dissociated from the mean flow. The variation with depth of $\sqrt{\langle v^2 \rangle_d}$
763 shown in Fig. 12b exhibits a similar behavior as the $\sqrt{\langle u^2 \rangle_d}$ one. Note however that the re-
764 spective weight of orbital to turbulent motions is noticeably lower in the spanwise direction, the
765 near exponential decay being distinguishable only at fetches larger than 8 m. At 4 to 6 m fetches,
766 the $\sqrt{\langle v^2 \rangle_d}$ profiles then look quasi-flat since the contribution of small-scale spanwise turbulent
767 fluctuations to $\sqrt{\langle v^2 \rangle_d}$ proves to be largely dominant.

768 **5. Concluding remarks**

769 A series of tests and laboratory observations has been carried out in a large wind-wave tank
770 to explore the potential of a multistatic Nortek Vectrino profiler for investigating the structure
771 of the water boundary layer induced by wind and waves just beneath the surface. Compared to
772 single-point velocimeters, such as LDV, this instrument has been found to have two major advan-
773 tages. First, it can provide measurements of the three components of the velocity field along a
774 short vertical segment within the surface sublayer. The available spatial and temporal resolutions,
775 respectively 1 mm and 0.01 s, appears to be sufficiently high for such measurements, but the min-
776 imum distance to the surface to be reached for the data to remain valid is at best 3.5 to 4 mm,
777 the data collected above are contaminated by surface echoes. Second, when embedded within the
778 measuring profile, the water surface level can be detected simultaneously with the velocity field by
779 tracking acoustic power profile maxima. Thus, it enables the determination of the instantaneous
780 surface elevation with 1 mm accuracy and the instantaneous depth of velocity measurements. In
781 future work, this feature may enable us to perform conditional analysis of turbulent structures with
782 surface motions or wave groups and thus, to investigate in more details coupling between waves
783 and water subsurface flow turbulence. However, the main difficulty in getting reliable velocity
784 data, i.e. data weakly contaminated by noise, lies in the seeding quality. In the laboratory, when
785 seeding of the bulk water is not easily feasible owing to the large size of the tank or the risk of pol-
786 lution, hydrogen bubbles generated by electrolysis wires scatter a sufficiently high acoustic power
787 to offer a good and cheap alternative to seeding particles. Simple methods have also been tested
788 for appraising the signal quality and discarding velocity estimates contaminated by noise when
789 seeding conditions were poor or intermittent, in particular, in immediate proximity to the surface

790 or the bottom of the boundary layer. These methods are particularly useful for investigations in
791 which the Hurther and Lemmin (2001) method for removing noise is not applicable.

792 The velocity field inside the 2 cm thick water surface boundary layer driven by wind was ex-
793 plored along the first meters of the Marseille-Luminy wind wave tank. The spatial variation of the
794 mean drift current profiles was described in a curvilinear coordinate system at the various stages of
795 surface flow development. Thus, when fetch increases, we could identify the initial growth of the
796 laminar water boundary layer driven by air surface wind stress, the development of viscous streaks,
797 the abrupt laminar-turbulent transition, the coupled growth of orbital wave motions and finally, the
798 formation of Langmuir cells. The characteristic scales and spatial growth rate of velocity fluctua-
799 tions associated with the different types of motions have been described in detail. In particular, the
800 RMS amplitude of the dominant wave orbital motions observed in such a highly-sheared subsur-
801 face boundary layer has been quantified. A few peculiarities of this turbulent boundary layer were
802 also highlighted by comparison with those observed over a rigid wall. Among them, we report the
803 noticeably high level of spanwise turbulent velocity disturbances relative to the streamwise ones.

804 This quite successful investigation of the wind-driven boundary layer by means of an ADV pro-
805 filer also shows that further substantial measurement improvements can readily be made. Firstly,
806 it is clear that measurement quality would be significantly improved by using an instrument with
807 isotropic characteristics, enabling the diminution of data noise contamination and correction of
808 the statistical flow properties from residual noise. Moreover, to improve water seeding, it would
809 be of interest to design an array of electrolysis wires arranged upstream of the measuring profile
810 and probably downstream too. Because the turbulent flow is characterized by very large upward
811 and downward motions, such arrangement would enable investigation of the water flow dynamics
812 throughout the surface layer in a more homogeneous way, and thus, the reduction of the potential
813 bias introduced in averaging flow properties. In addition, to estimate properly the average flow

814 properties, this study clearly indicates the necessity of reconstructing the velocity field in a curvi-
815 linear coordinate system following the surface wave motion streamlines. Therefore, to improve the
816 separation of the three foremost flow components contributing to the velocity field observed, i.e.
817 the mean drift current, the dominant wave orbital motions and the turbulent fluctuations, a more
818 robust and manageable model describing orbital motions within a highly-sheared flow should be
819 developed. This knowledge would help to noticeably improve accuracy and reliability of various
820 flow characteristics estimates, this enabling more extensive survey of the finescale dynamics of
821 this very specific turbulent boundary layer.

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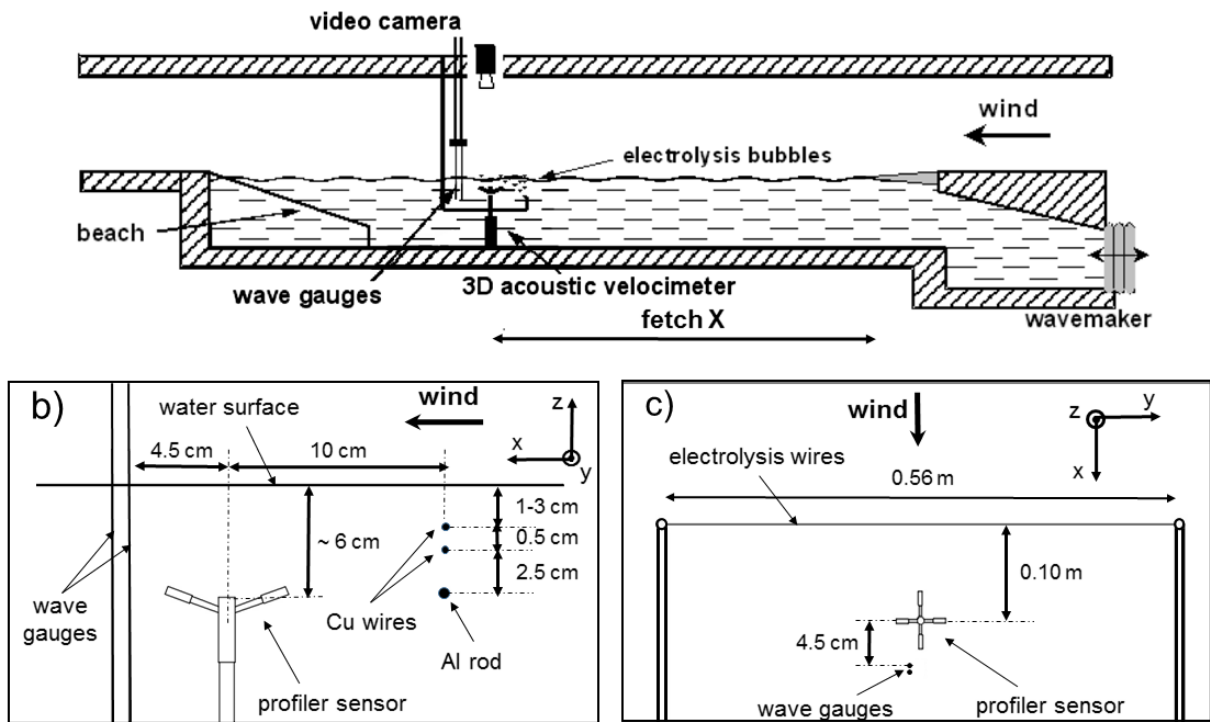
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939 electrolysis wire (visualized by the thick horizontal white line) is 3 and 2 cm respectively.
940 Wind blows from the bottom to the top of the images. 52

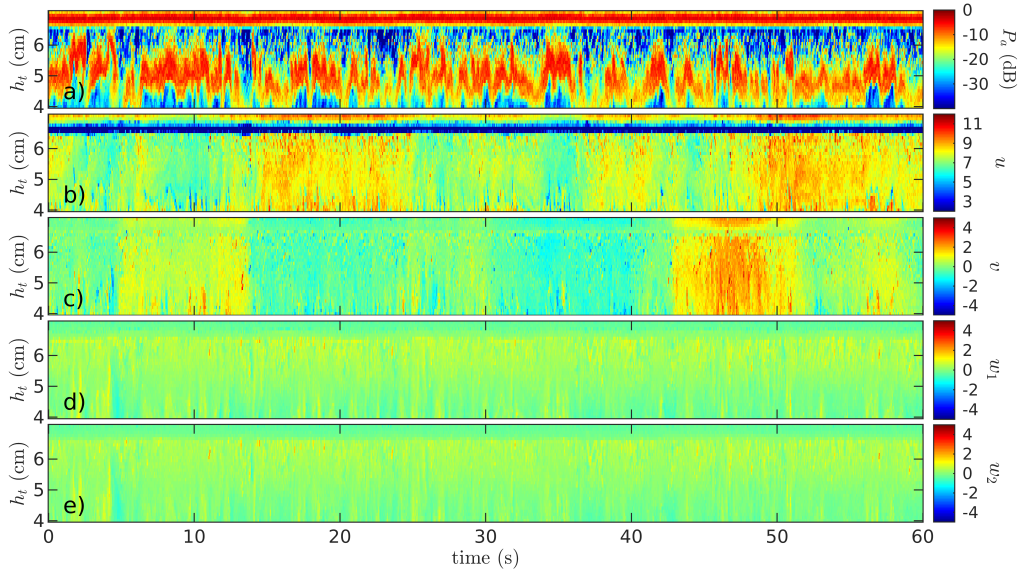
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950 dotted lines, to the respective ratios $ak_d(c_d - U_s)/ak_dc_d$ 54

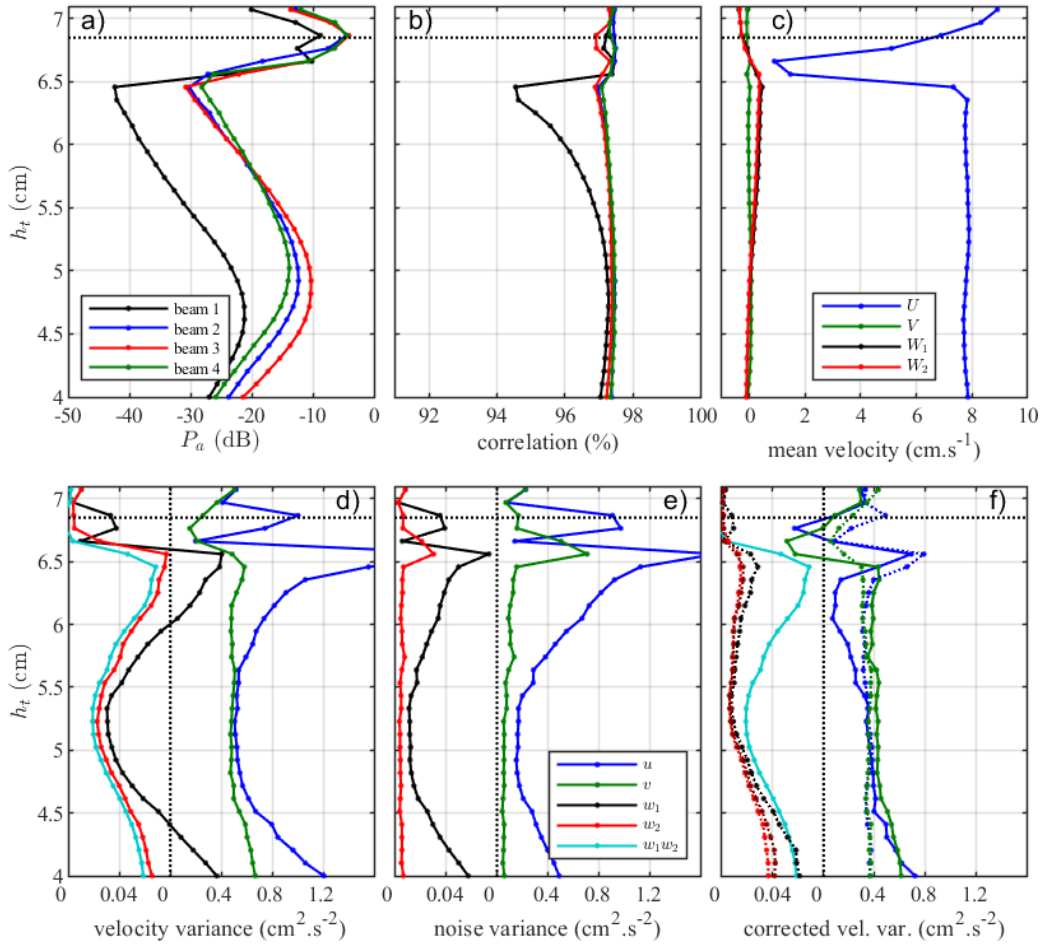
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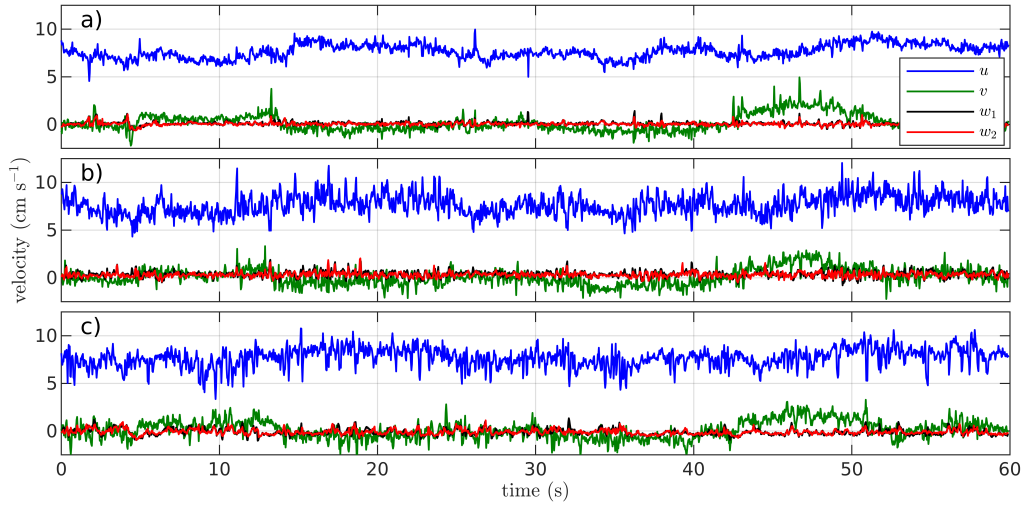
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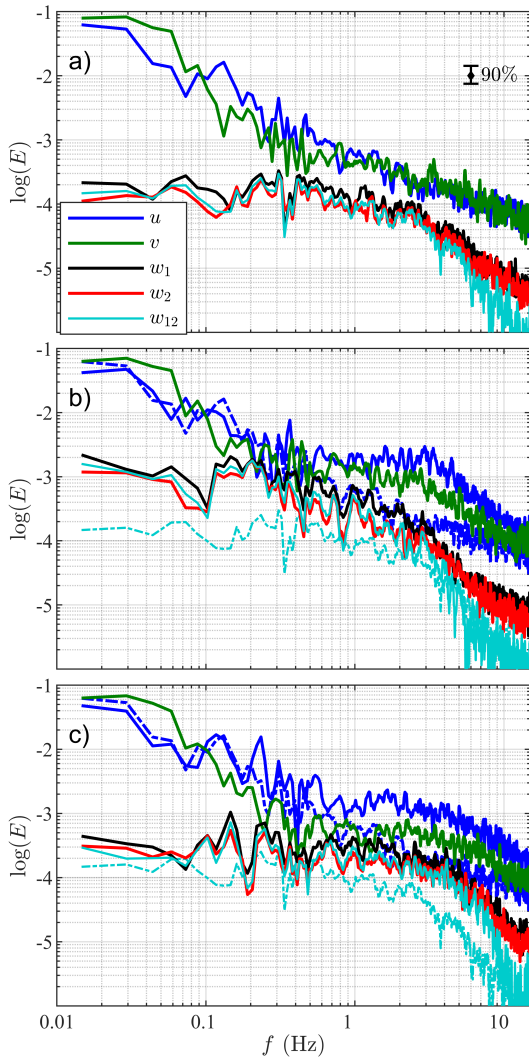
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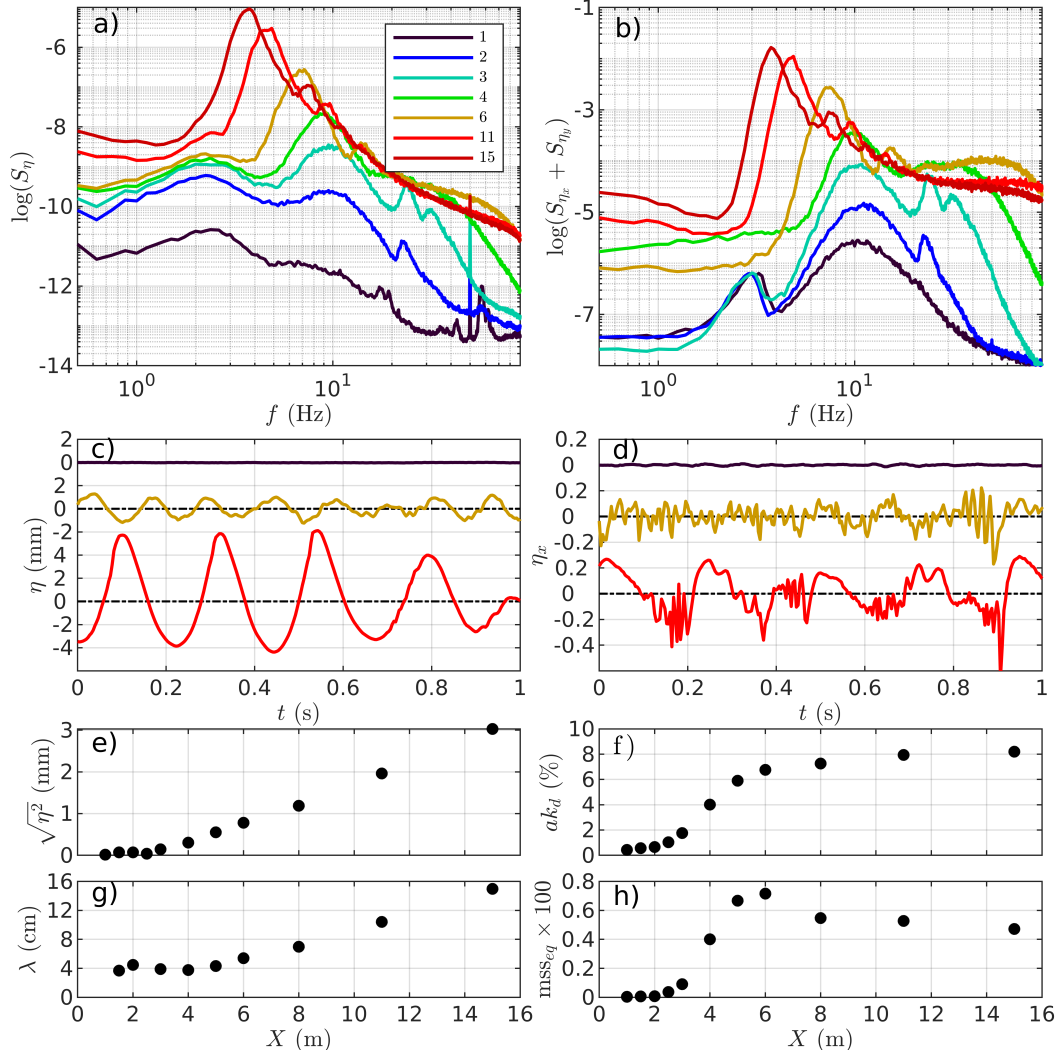
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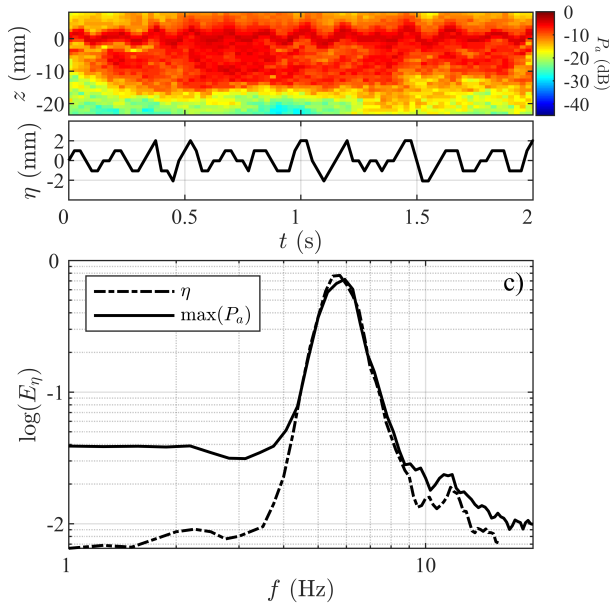
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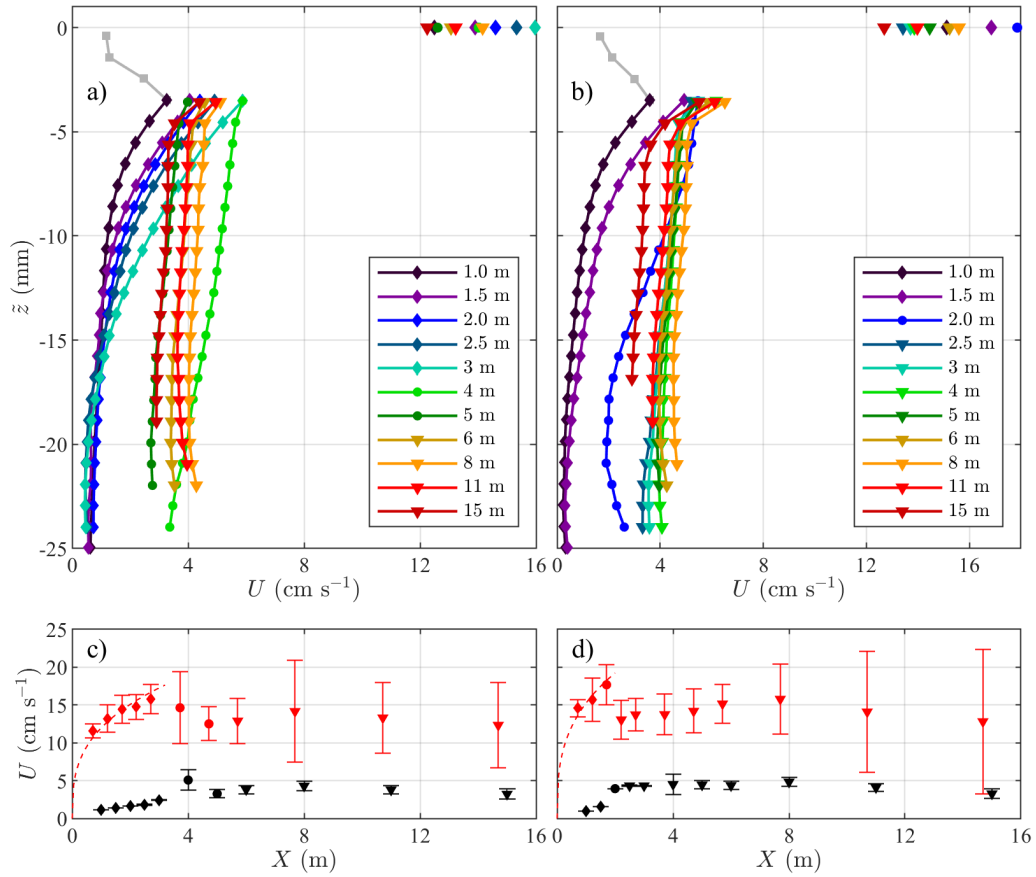
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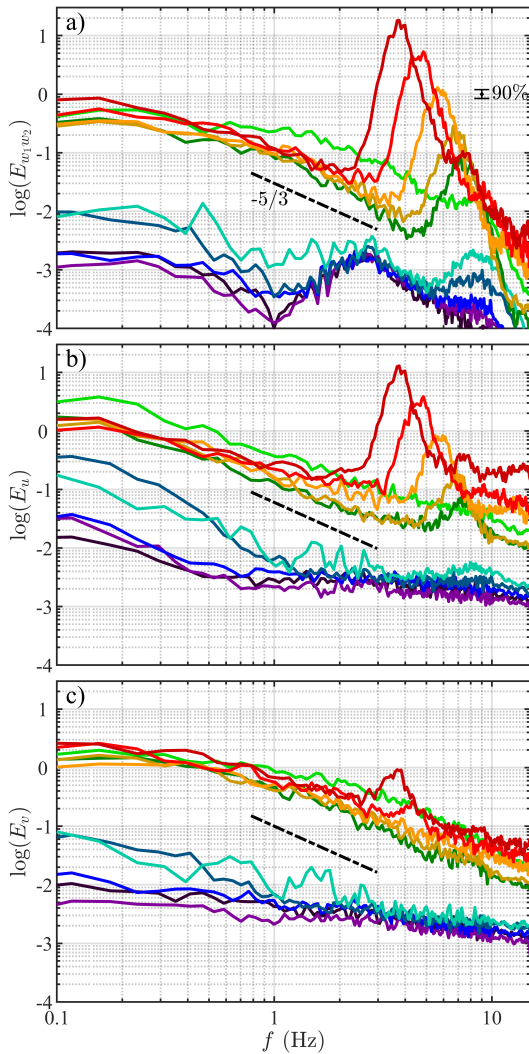
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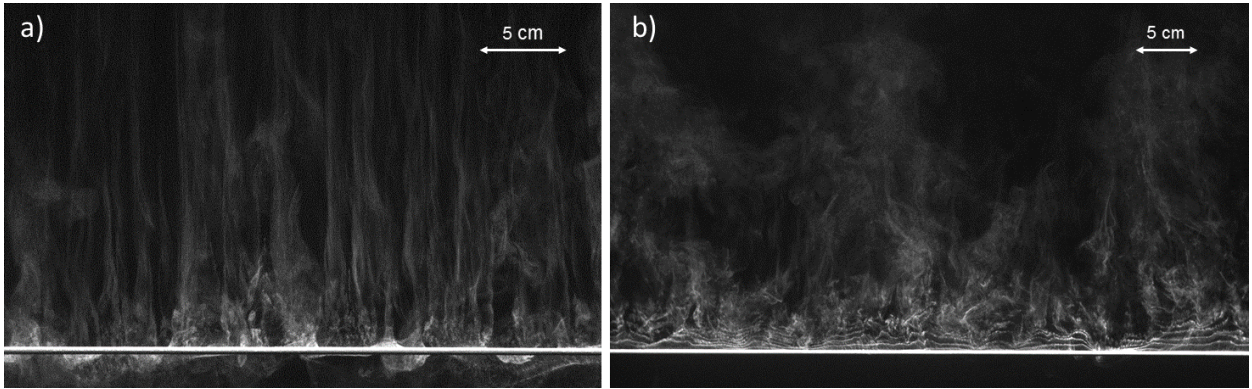
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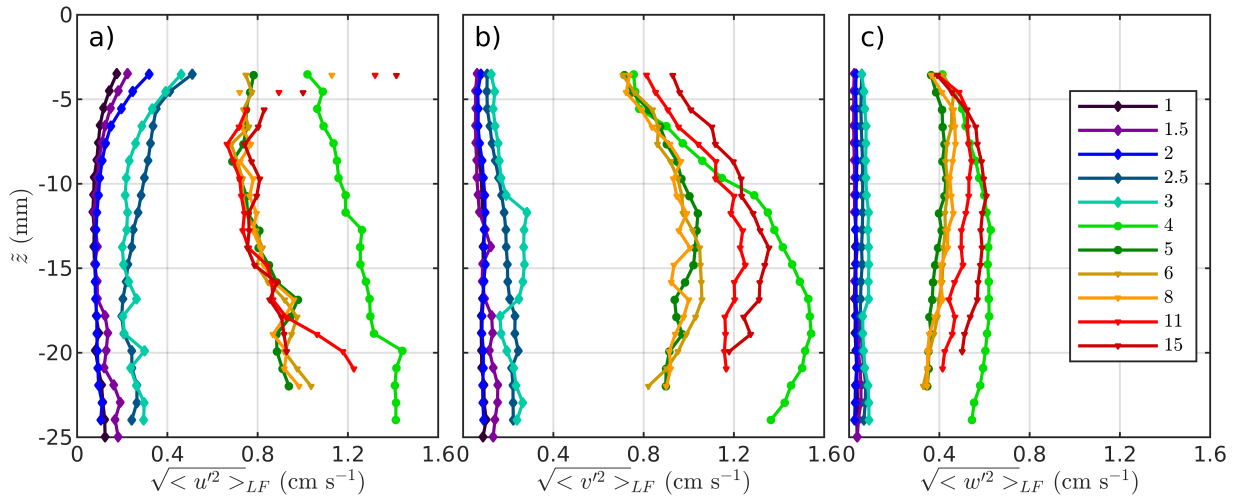
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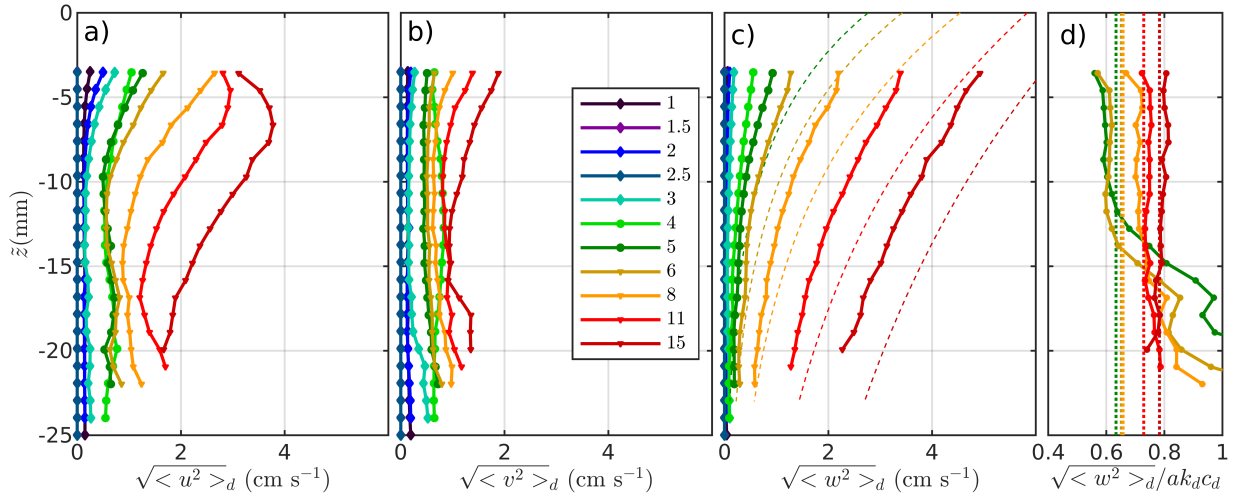
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 993 confidence interval of spectra is given at the top left-hand side of the panel a).



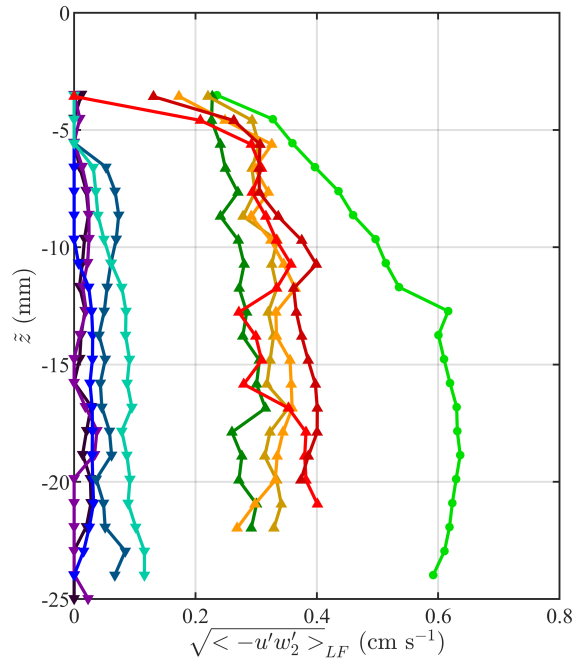
994 FIG. 10. Views from above of three-dimensional water subsurface flow structures made visible by electrolysis
995 bubbles. a): Laminar streaks observed just before the breakdown to turbulence of the wind-driven water bound-
996 ary layer at 2 m fetch and 5 m/s wind speed. b): Langmuir circulations observed at 13 m fetch and 4.0 m/s wind
997 speed. The depth of the upper copper electrolysis wire (visualized by the thick horizontal white line) is 3 and
998 2 cm respectively. Wind blows from the bottom to the top of the images.



999 FIG. 11. Variation with dominant wave streamline depth z of turbulence intensities estimated from integration
 1000 of velocity spectra at low frequencies (i.e. below 1 Hz) displayed for all fetches (given in the legend in
 1001 meters) and 4.5 m s^{-1} wind speed: a): u' , b): v' , c): w' .



1002 FIG. 12. (a)-(c): Variation with dominant wave streamline depth \tilde{z} of the RMS values of the three velocity
 1003 components of dominant wave orbital motions as estimated from integration of velocity spectra over the dom-
 1004 inant peak, displayed for all fetches and 4.5 m s^{-1} : a): u , b): v , c): w_2 . d): Ratio between the measured RMS
 1005 vertical orbital velocity value and this estimated from measured surface motions by using the linear wave theory
 1006 as plotted in (c) by dashed lines. Color lines refer to fetches as given in meters in the legend box and the dotted
 1007 lines, to the respective ratios $ak_d(c_d - U_s)/ak_dc_d$.



1008 FIG. 13. Variation with dominant wave streamline depth \tilde{z} of the square root of the vertical momentum flux
 1009 $\sqrt{\langle -u'w'_2 \rangle}$ estimated from integration of velocity cospectra at low frequencies (below 1 Hz), for all fetches and
 1010 4.5 m s^{-1} wind speed. Line colors refer to fetches as given in meters in Fig. 11 legend box.