



Numerical Simulation of the Boundary Layer Flow Generated in Monterey Bay, California, by the 2010 Chilean Tsunami: Case Study

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1 **Case study:**

2 **Numerical simulation of the boundary layer flow generated in**
3 **Monterey Bay, California by the 2010 Chilean tsunami**

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12 **ABSTRACT**

13 This work presents a case study involving the numerical simulation of the unsteady
14 boundary layer generated by the 2010 Chilean tsunami, as measured by field equipment in
15 Monterey Bay, California, USA. A one-dimensional vertical (1DV) boundary layer model is
16 utilized, solving Reynolds-averaged Navier-Stokes equations, coupled with two-equation $k-\omega$
17 turbulence closure. Local effects of convective acceleration (converging-diverging effects) on
18 the boundary layer due to the sloping bed are likewise approximated. Four cases are consid-
19 ered involving simulation of: (1) the long tsunami-induced boundary layer flow in isolation,
20 in combination with either (2) convective acceleration effects or (3) energetic short wind
21 waves, and finally (4) all effects combined. Reasonable agreement with field measurements
22 is achieved, with model results similarly showing that the tsunami-induced boundary layer
23 in this case only spans a fraction of the local water depth. Systematic comparison of the var-
24 ious cases likewise elucidates the likely significance of both local converging-diverging effects,

25 as well as interaction with the much shorter period wind waves, on the tsunami-generated
26 boundary layer. In the latter case, analogy is drawn to well-known wave-current boundary
27 layer interaction, with the boundary layer turbulence associated with the short wind waves
28 inducing an effective wave roughness felt by the tsunami-induced flow, which effectively plays
29 the role of the current.

INTRODUCTION

Tsunamis most commonly occur in the aftermath of seismic events and are thus inherently unpredictable in nature. While many studies of tsunamis emphasize their run-up and inundation, the nature of their-induced boundary layer flows (essential e.g. for understanding their induced sediment transport and scour around coastal structures) has also been studied realistically in controlled experimental (Larsen et al. 2018; Tanaka et al. 1999) or numerical (Williams and Fuhrman 2016; Larsen et al. 2017; Larsen and Fuhrman 2019a; Larsen and Fuhrman 2019b; Tinh and Tanaka 2019; Tanaka et al. 2020) environments. Due to their infrequent and unpredictable occurrence, actual field data involving the detailed boundary layer flow structure induced by tsunamis is elusive, seemingly being limited to the measurements of Lacy et al. (2012), who were the first to measure detailed velocities within the boundary layer of a tsunami-induced flow.

Specifically, the event in question corresponds to the 2010 Chilean tsunami, which was caused by the $M_w = 8.8$ (moment magnitude scale) 2010 Chilean earthquake on February 27 off the coast of Maule. The resulting tsunami devastated several coastal towns in south-central Chile and damaged the port at Talcahuano. Wave heights of about 3 m were reported from Chilean Islands, while the height of the tsunami in the deeper water was measured to be approximately 25 cm (see e.g. DART station 32412), which according an expert from the Pacific Tsunami Warning Centre (Fryer 2015), is large enough to issue a warning in the areas around the Pacific ocean.

Approximately 14 hr later, corresponding to propagation around nearly half the globe as shown in Fig. 1, on February 28, 2010 the tsunami arrived at Monterey Bay, California, USA. Fortunately, the United States Geological Survey (USGS) had at the time an upward-pointing acoustic Doppler current profiler (ADCP), and a downward-pointing pulse-coherent acoustic Doppler profiler (PCADP) deployed at a test site in the bay, and the latter was utilized to take the measurements within the local tsunami-induced boundary layer flow (Lacy et al. 2012). The sampling frequency was increased in anticipation of the tsunami's

57 arrival, and the data obtained from the PCADP are likewise the basis for the forthcoming
 58 case study. Directed toward the bed, the equipment took measurements from approximately
 59 0.05 mab (meters above the bed) to 0.62 mab with a spacing of 0.095 m vertically and at a
 60 frequency of 1 Hz for 25 minutes every half hour.

61 In the numerical case study which follows the one-dimensional vertical (1DV) model of
 62 Fuhrman et al. (2013) will be utilized to simulate and study the tsunami-induced boundary
 63 layer flow measured by Lacy et al. (2012). This model is based on the horizontal component
 64 of the incompressible Reynolds-averaged Navier-Stokes (RANS) equations in combination
 65 with the Wilcox (2006) $k-\omega$ turbulence closure model. The present case study attempts
 66 to model this event, motivated to better understand both the tsunami-induced boundary
 67 layer flow itself, as well as the influence of various external or local factors. Specifically, the
 68 study aims to elucidate the importance of both convective acceleration (converging-diverging
 69 effects due to the local bed slope) as well as interaction with energetic short (wind) waves
 70 that were present during the captured event. An overview of the basic parameters used in
 71 this model, as presented by Lacy et al. (2012), are presented in Table 1.

72 **MODEL DESCRIPTION**

73 **Governing Equations**

74 As mentioned above, the model employed in this investigation is that of Fuhrman et al.
 75 (2013), and it is likewise detailed in the recent book of Sumer and Fuhrman (2020) (see their
 76 Section 5.12). It solves a simplified version of the horizontal component of the incompressible
 77 Reynolds-averaged Navier-stokes (RANS) equations, and for the turbulence closure the two-
 78 equation $k-\omega$ of Wilcox (2006) and Wilcox (2008) is used. The RANS equation is as follows:

$$79 \quad \frac{\partial \bar{u}}{\partial t} = -\frac{1}{\rho} \frac{\partial \bar{p}}{\partial x} + \frac{\partial}{\partial y} \left(\frac{\tau}{\rho} \right) - \underbrace{\left(\bar{u} \frac{\partial \bar{u}}{\partial x} + \bar{v} \frac{\partial \bar{u}}{\partial y} \right)}_{(1)}$$

80 In the above t is time, (\bar{u}, \bar{v}) are the (Reynolds averaged, as indicated by the overbar)
 81 velocities in the (x, y) (horizontal and vertical) directions, \bar{p} is pressure, and τ is the total

82 (viscous plus turbulent) shear stress expressed as

$$83 \quad \frac{\tau}{\rho} = (\nu + \nu_T) \frac{\partial \bar{u}}{\partial y} \quad (2)$$

84 where ρ is the fluid density, with ν and ν_T respectively being the kinematic fluid and eddy
85 viscosity. In what follows the x axis can be more specifically interpreted as pointing in the
86 cross shore direction (positive onshore).

87 Note that the underlined terms in (1) (and also in forthcoming equations (7), (8) and
88 (14)) correspond to convective acceleration terms, which are included in the present study
89 to potentially account for converging-diverging effects associated with a sloping sea bottom.
90 These are approximated as follows. As described in detail by Fuhrman et al. (2013), the x
91 derivative in the convective terms can be locally approximated as:

$$92 \quad \frac{\partial}{\partial x} = \frac{S}{h} \quad (3)$$

93 where S is the local slope of the seabed and h is the local water depth. Subsequently, the
94 vertical velocity can be obtained from the local continuity equation:

$$95 \quad \frac{\partial \bar{u}}{\partial x} + \frac{\partial \bar{v}}{\partial y} = 0 \quad (4)$$

96 After invoking (3), and integrating from the local seabed at $y = 0$, the vertical velocity can
97 be obtained through cumulative integration:

$$98 \quad \bar{v} = -\frac{S}{h} \int_0^y \bar{u} dy \quad (5)$$

99 This method conveniently allows for the effects from x -variations (from the sloping bed) to
100 be incorporated in the 1DV model, while still only requiring discretization in the vertical
101 y -direction.

102 To achieve turbulence closure, we will utilize the two-equation k - ω model of Wilcox
 103 (2006). This model consists of a transport equation for the turbulent kinetic energy (per
 104 unit mass)

$$105 \quad k = \frac{1}{2}(\overline{u'^2} + \overline{v'^2} + \overline{w'^2}) \quad (6)$$

106 where the prime superscript denotes turbulent fluctuations (w' being the fluctuation in the
 107 unresolved horizontal along-shore z direction) and the overbar averaging, corresponding to:

$$108 \quad \frac{\partial k}{\partial t} = \nu_T \left(\frac{\partial \bar{u}}{\partial y} \frac{\partial \bar{u}}{\partial y} \right) - \beta^* k \omega + \frac{\partial}{\partial y} \left[\left(\nu + \sigma^* \frac{k}{\omega} \right) \frac{\partial k}{\partial y} \right] - \underbrace{\left(\bar{u} \frac{\partial k}{\partial x} + \bar{v} \frac{\partial k}{\partial y} \right)} \quad (7)$$

109 as well as a similar transport equation for the specific dissipation rate ω :

$$110 \quad \frac{\partial \omega}{\partial t} = \alpha \frac{\omega}{k} \nu_T \left(\frac{\partial \bar{u}}{\partial y} \frac{\partial \bar{u}}{\partial y} \right) - \beta \omega^2 + \frac{\partial}{\partial y} \left[\left(\nu + \sigma \frac{k}{\omega} \right) \frac{\partial \omega}{\partial y} \right] + \frac{\sigma_d}{\omega} \frac{\partial k}{\partial y} \frac{\partial \omega}{\partial y} - \underbrace{\left(\bar{u} \frac{\partial \omega}{\partial x} + \bar{v} \frac{\partial \omega}{\partial y} \right)} \quad (8)$$

111 The eddy viscosity ν_T is defined by

$$112 \quad \nu_T = \frac{k}{\tilde{\omega}}, \quad \tilde{\omega} = \max \left\{ \omega, C_{lim} \frac{|\partial \bar{u} / \partial y|}{\sqrt{\beta^*}} \right\}, \quad (9)$$

113 where $C_{lim} = 7/8$. In (8)

$$114 \quad \sigma_d = \mathcal{H} \left\{ \frac{\partial k}{\partial y} \frac{\partial \omega}{\partial y} \right\} \sigma_{do}, \quad (10)$$

115 where $\mathcal{H} \{ \cdot \}$ is the Heaviside step function, taking a value of zero when the argument is
 116 negative, and a value of unity otherwise.

117 In the right hand side of (7) the first term represents the **production** of turbulent
 118 kinetic energy (the rate at which kinetic energy is transferred from the mean flow to the
 119 turbulence), the second term represents **dissipation** (the rate at which turbulent kinetic
 120 energy is converted into thermal internal energy) and the third term includes both molecular
 121 and turbulent **diffusion**. The default model closure coefficients suggested by Wilcox (2006)
 122 are utilized: $\alpha = 13/25$, $\beta = \beta_0 f_\beta$, $\beta_0 = 0.0708$, $\beta^* = 9/100$, $\sigma = 1/2$, $\sigma^* = 3/5$, $\sigma_{do} = 1/8$.

123 Note that for two-dimensional flows, as considered herein, $f_\beta = 1$.

124 **Boundary conditions**

125 The 1DV model described above is subject to the following boundary conditions. The
126 bottom boundary is considered a friction wall, with a no-slip boundary condition imposed:

$$127 \quad \bar{u} = \bar{v} = 0, \quad y = 0 \quad (11)$$

128 Furthermore, at the bottom boundary, a zero-gradient condition is imposed for k , the tur-
129 bulent kinetic energy (per unit mass), $dk/dy = 0$, which imposes a zero flux of turbulent
130 kinetic energy through the sea bed. This condition is justified based on experimental mea-
131 surements for steady flows on rough beds by Fuhrman et al. (2010) and Sumer et al. (2003).
132 Fuhrman et al. (2010) also demonstrated that the zero-gradient condition allows for a nat-
133 ural development of the viscous sublayer near smooth walls while avoiding the creation of a
134 fictitious viscous sublayer near rough walls. This is in contrast to simply forcing $k = 0$ at the
135 wall, which creates (and hence requires resolution of) a viscous sublayer in all circumstances.
136 This zero-gradient condition has also previously been successfully employed in simulations of
137 oscillatory wave boundary flows by Fuhrman et al. (2011), on both smooth and rough beds.

138 The bottom boundary condition for the specific dissipation rate, ω , is taken from Wilcox
139 (2006):

$$140 \quad \omega = \frac{U_f^2}{\nu} S_R, \quad y = 0, \quad (12)$$

141 where

$$142 \quad S_R = \begin{cases} \left(\frac{200}{k_N^+}\right)^2, & k_N^+ \leq 5, \\ \frac{K_r}{k_N^+} + \left[\left(\frac{200}{k_N^+}\right)^2 - \frac{K_r}{k_N^+}\right] e^{5-k_N^+}, & k_N^+ > 5. \end{cases} \quad (13)$$

143 Here $k_N^+ = k_N U_f / \nu$ is the roughness Reynolds number, $k_N = 2.5d$ is Nikuradse's equivalent
144 sand grain roughness (d being the sediment grain diameter), and $U_f = \sqrt{|\tau_b|/\rho}$ is the

145 instantaneous friction velocity, where $\tau_b = \tau(y = 0)$ is the bed shear stress. The value
 146 $K_r = 180$ (a calibration constant) is utilized in this model, which has been demonstrated by
 147 Fuhrman et al. (2010) to produce standard logarithmic velocity profile solutions for steady
 148 boundary layers when used in combination with the $\partial k/\partial y = 0$ bottom boundary condition.

149 **Pressure gradient**

150 A specified horizontal pressure gradient is used to drive the flow within the model, based
 151 on a prescribed velocity u_0 at the top of the model domain, which will be based directly on
 152 measurements from Lacy et al. (2012). The pressure gradient has thus been implemented as
 153 follows:

$$154 \quad \frac{1}{\rho} \frac{\partial \bar{p}}{\partial x} = -\frac{\partial u_0}{\partial t} - \frac{u_0}{\partial x} + \frac{1}{\rho} \frac{\partial \tau}{\partial y} \Big|_{y=y_{top}} \quad (14)$$

155 This is seen to account for both unsteadiness and convective acceleration, in the standard way
 156 for free stream regions. Additionally, the third term in the right-hand side of (14) accounts
 157 for potential shear stress at the top-most point of the model domain ($y = y_{top}$), which in the
 158 present case study may be non-zero. This addition is necessary, since the top of the model
 159 domain has been here chosen to coincide with the largest-elevation field measurements, where
 160 the shear stress may not be exactly zero. (The velocimeter was mounted at a fixed elevation
 161 above the bed, and it is therefore not, at any given moment, certain whether the top-most
 162 velocity measurement is inside or outside the boundary layer; We have found that without
 163 including this term there will be a slow drift in the top-most velocity away from what was
 164 measured.)

165 Finally it is worth mentioning that the model as described in Fuhrman et al. (2013) is also
 166 capable of accounting for so-called “conventional” boundary layer streaming, which involves
 167 additional approximations for the convective terms to account for spatial variations inherent
 168 within progressive (regular) waves. For progressive (regular) waves, the importance of these
 169 effects is well known to scale as $U_{0m}/c \sim ak_w$, where U_{0m} is the characteristic free-stream

170 velocity magnitude, c is the wave celerity, $k_w = 2\pi/L$ is the wave number, L is the wave
171 length, and $a = U_{0m}/\omega_w$, where $\omega_w = 2\pi/T$ is the angular frequency and T the wave period.
172 Such effects have been found to be negligible for primary tsunami wave scales considered here,
173 however, due to the extremely long wave lengths (hence very small wave number k_w and very
174 large celerity c) associated with tsunamis. This is easily confirmed by substituting typical
175 tsunami-scale values into the scaling parameter mentioned just above. These conventional
176 streaming effects are thus not included in any of the simulations which follow, for the sake of
177 simplicity. (This is also convenient, since neither the transient tsunami nor the irregular short
178 wave field to be considered could be very well approximated by a regular wave assumption,
179 hence inclusion of such additional effects would not be straight forward.)

180 RESULTS

181 Description of cases

182 All model results will be based on simulated flows driven by velocity signals measured
183 at the top-most elevation by Lacy et al. (2012). The full data set is, as described by Lacy
184 et al. (2012), organized in bursts of 25 min each, with 5 min between each burst. However,
185 the main tsunami event transpired during the third burst, and this will therefore be the
186 focus of the current analysis, and these measurements, corrected for ambiguity errors as
187 described in Lacy and Sherwood (2004), are depicted in Fig. 2. In this figure, the grey
188 line depicts the full measured signal i.e. including both the long tsunami-induced flow, as
189 well as that induced by energetic short wind waves, which were also present. Additionally,
190 the dark black line in Fig. 2 depicts the low-pass filtered signal from Lacy et al. (2012),
191 where the flow associated with the short waves has been removed, leaving the tsunami-
192 induced velocity signal in isolation. Note that we have divided the tsunami-induced flow
193 into regions of acceleration (depicted by the diamonds) and deceleration (represented by 'x'
194 markers), where each time instant (representing a 10 s average, each separated by approx. 2
195 min). is marked by a different color. The coloring at these instants will be maintained in the
196 presentation of several forthcoming velocity profile comparisons. Note that the peak velocity

197 (represented by both a diamond and an x on Fig. 2) is taken to represent both the (end of
198 the) accelerating region and the (beginning of the) decelerating region, to serve as a clear
199 and shared state of reference.

200 We will consider four separate cases, as summarized in Table 2. Each will be driven by
201 one of the velocity signals depicted in Fig. 2, as stated previously. (The only modification
202 is that the initial flow is ramped up from zero in order to match the prescribed initial
203 conditions; This takes place long before the arrival of the main tsunami wave, and hence
204 does not affect the results of present interest.) As seen in Table 2, Case 1 is the most
205 basic, considering simulation of the transient boundary layer which develops beneath the
206 long tsunami-induced flow in isolation (black line in Fig. 2). Cases 2 and 3, respectively,
207 consider the addition of either convective terms (again, accounting for converging flow due
208 to a sloping bed) or the short wave field (i.e. now using the grey line in Fig. 2). Finally,
209 Case 4 combines all three effects. In all cases the local water depth is taken as $h = 9$ m,
210 and the bed slope as $S = 0.012$, which has been estimated from the contour plot provided
211 as Figure 1 from Lacy et al. (2012). 1. Nikuradse’s equivalent sand grain roughness is taken
212 as $k_s = 2.5d$. Note finally that Lacy et al. (2012) found that the tsunami and short wave
213 field were approximately co-linear. Hence no attempt to account for effects associated with
214 varying directionality between the tsunami and short waves has been made, though it is
215 recognized that this is only an approximation due to likely directional spreading within the
216 irregular short wave field.

217 It should finally be mentioned that switching on the convective terms (as in Cases 2
218 and 4) creates a slight drift in the velocities (an additional form of wave boundary layer
219 streaming, see e.g. Chapter 6 of Sumer and Fuhrman (2020)). To ensure proper comparison
220 with the field measurements the exact times when the model results are chosen for these
221 cases is adjusted slightly, such that the velocity at (or near) the top matches the field data,
222 enabling the most direct and clear comparison with the other measurement locations further
223 below.

224 **Case 1: Tsunami alone**

225 The most natural place to start the case-study of this tsunami event is by modeling
226 the tsunami-induced flow in isolation, Case 1. Comparison between the measured (circles)
227 and simulated (full lines) results, at the acceleration/deceleration instants marked in Fig. 2,
228 are depicted in Fig. 3. It is seen that even in this case, the simulated evolution of the
229 tsunami-induced wave boundary layer resembles that measured, especially considering that
230 the present model is uni-dimensional. Results during the acceleration stages are more ac-
231 curate than in the deceleration stage. This is also shown quantitatively in Table 3, which
232 summarizes the root-mean-square error in the simulated velocity profiles (relative to the
233 measurements) during periods of acceleration and deceleration, at the peak, as well as in
234 totality (i.e. the average of all depicted times). Consistent with the findings of Lacy et al.
235 (2012), it is seen that the boundary layer induced by the tsunami is largely contained within
236 the modeled near-bed domain i.e. it spans only a fraction of the 9 m water depth. This find-
237 ing is likewise consistent with the later general findings of Williams and Fuhrman (2016),
238 and is important as it implies that quasi-steady friction formulations (e.g. widely used Man-
239 ning or Chezy formulations which effectively assume the boundary layer spans the full flow
240 depth) should be used with great care in tsunami simulations. It is emphasized that this
241 simulated case accounts for neither the effects of the short waves nor the beach slope on the
242 resultant boundary layer. As it is likely of the most practical interest, as well as for the sake
243 of simplicity and to ease comparison, much of the discussion for the various cases that follow
244 will focus on the effects of the profile at the flow peak. In the present case it is clear that the
245 modeled boundary layer velocities near the bed at peak flow are significantly underestimated.

246 **Case 2: Tsunami and convective terms (converging-diverging flow effects)**

247 As presented in the model description, the model is able to include second-order terms
248 in the simulation, accounting for convective acceleration (converging-diverging flow) effects
249 associated with a sloping bed. Fuhrman et al. (2009a) and Fuhrman et al. (2009b) have
250 previously demonstrated that, for the idealized case involving sinusoidal free stream flows,

251 the importance of these effects will scale according to the dimensionless parameter:

$$252 \quad \frac{Sa}{h} \quad (15)$$

253 where $a = U_{0m}T/(2\pi)$ is a length scale, corresponding to the amplitude of the free stream
254 fluid particle motion. Invoking the characteristic values from Table 1, along with the local
255 bottom slope $S = 0.012$ (estimated from the contour plot provided as Figure 1 in Lacy et al.
256 (2012)) yields: $Sa/h \approx 0.07$. This is of the order 10%, thus non-negligible and indicating
257 the likely importance of convective acceleration due to beach slope in this context. For
258 comparison, doing similarly but using the wind wave scales from Table 1 instead yields:
259 $Sa/h \approx 0.0008$ i.e. two orders of magnitude smaller (hence likely insignificant). As the
260 wave- and tsunami-induced velocities are similar in magnitude, and the bottom slope itself
261 is relatively mild, it is clearly the much larger period of the tsunami (estimated as 16 min in
262 this case, compared to a characteristic 10 s for the wind waves) which significantly increases
263 the likely importance of convective acceleration effects due to beach slope in the present
264 application.

265 Based on the assessment above, it was therefore expected that the converging flow effects
266 will have a noticeable effect on the tsunami-induced wave boundary layer. This is largely
267 confirmed from the simulation of Case 2, which is otherwise identical to Case 1, but now
268 with the convective-acceleration terms active. The resulting horizontal velocity profiles are
269 presented in Fig. 4, including comparison with those measured in the same fashion as in
270 Fig. 3. It is seen that, with these effects added, the model fits the measured velocity profile
271 during acceleration and near peak flow better. This is in line with physical expectations, since
272 it is well-known that near bed velocities are enhanced due to favorable pressure gradients in
273 a converging flow. If we focus e.g. on the two measurement elevations nearest the bed at the
274 peak flow, it is now seen that the computed near-bed boundary layer velocities at the peak
275 are, in fact, slightly over-predicted in this case. The results during the deceleration stage

276 are again much less accurate, however, and these differences are even more exaggerated than
277 in Case 1. These differences in accuracy beneath the accelerating and decelerating flow can
278 also be seen quantitatively in Table 3.

279 **Case 3: Tsunami and wind waves**

280 In contrast to Cases 1 and 2, Case 3 will now consider the boundary layer flow driven
281 by the full signal (grey line) in Fig. 2, containing measured velocities from both the tsunami
282 as well as the short wind wave field. Convective acceleration effects are switched off for
283 this case. Lacy et al. (2012) hypothesized that the presence of the short wave field would
284 result in an increased “apparent roughness” on the much longer tsunami, analogous to that
285 in classical wave-current interactions (Grant and Madsen 1979). The computed velocity
286 profiles for this case are shown at selected instants against the measured profiles in Fig. 5,
287 similar to before. Comparing this case with Case 1, the increased “apparent roughness”
288 effects are indeed directly confirmed, which retard the flow near the bed considerably. This
289 is particularly apparent at and near the peak flow. Addition of the short waves likewise
290 seems to improve the computed boundary layer flow throughout the event, especially during
291 the deceleration stage. Indeed, unlike Cases 1 and 2, the simulated flow is now more accurate
292 during deceleration than during acceleration, as is also quantified in Table 3.

293 **Case 4: Tsunami and wind waves with convective terms**

294 Finally, Case 4 will combine the full velocity signal (both tsunami and short waves)
295 with convective acceleration terms also active, thus representing the most complete physical
296 simulation to be considered in the present case study. In Fig. 6 the velocity profiles are
297 once again compared to those measured in the field, and it can be concluded that the
298 most complete model does describe the evolution of the tsunami wave boundary layer in a
299 satisfactory and balanced fashion. Indeed, investigation of Fig. 6 reveals that this simulation
300 gives the best prediction of the boundary layer structure at peak flow, while also maintaining
301 reasonably similar error during both periods of acceleration and deceleration (Table 3). The
302 predictions at other times are not perfect, but are still reasonable given the difficulty of

303 simulating complicated field conditions, and likely importance of other factors not captured
304 within a simple 1DV approach. Such factors are briefly discussed below.

305 **DISCUSSION**

306 Perhaps the most compelling result from the present case study is the direct demon-
307 stration and confirmation via computational fluid dynamics simulation that the wind waves
308 impose an additional and clear apparent roughness on the transient boundary layer flow
309 induced by the tsunami. This effect is well known from the theory of wave-current in-
310 teractions, see Grant and Madsen (1979) as well as e.g. Fredsøe and Deigaard (1992). It
311 was also hypothesized to have occurred in this case by Lacy et al. (2012). A simple wave-
312 plus-current analogy is also considered in a wave friction coefficient approach by Sumer and
313 Fuhrman (2020). The present case study is believed to be the first to clearly and quanti-
314 tatively demonstrate such an effect utilizing a computational fluid dynamics model of the
315 turbulent boundary layer, driven directly by field measured tsunami-induced velocities. To
316 illustrate the effects of the apparent roughness more clearly, selected velocity profiles from
317 Case 4 (all effects included) and Case 2 (without short waves) are compared in Fig. 7. The
318 measurements at the same instants are also included for completeness. The retarding effect
319 of the short-wave-induced apparent roughness on the tsunami-induced flow is very clear,
320 especially near the bed. Without accounting for this effect, if energetic wind waves are also
321 locally present it is clear that tsunami models would likely over-predict near bed velocities.
322 It can also be seen that neither model result predicts a velocity overshoot near the time of
323 maximum free stream flow, though a slight overshoot is evident in the field measurements.
324 Velocity overshoot commonly occurs e.g. in oscillatory wave boundary layers, as a conse-
325 quence of phase leads in the near bed flow from a previous cycle. It does not, therefore,
326 occur in the numerical simulations since the driving velocity signal induces an essentially a
327 transient (rather than periodic) event.

328 Additionally, this case study also confirms the likely importance of convective-acceleration
329 (converging flow) effects in the boundary layer flows induced by transient tsunamis due to

330 a sloping bed, even when the local bed slope is seemingly rather mild, being $O(0.01)$ in the
331 present case. This was hypothesized in this context by Fuhrman et al. (2009a), who argued
332 that the experimental conditions considered by Sumer et al. (1993) were most likely relevant
333 at tsunami scales, based on a/h being $O(10)$, which feeds directly into the scaling parameter
334 identified in (15) above. This effect is best elucidated by comparing the simulated flow
335 from Case 4 (all effects included) with Case 3 (without convective acceleration) at selected
336 instants, as done in Fig. 8. Here it is seen that the inclusion of convective acceleration
337 increases the flow near the bed significantly. This again makes physical sense, since this will
338 induce a favorable pressure gradient when the flow is positive i.e. in the converging direction.

339 We finally compare computed results for the friction velocity (hence bed shear stress) with
340 those estimated by Lacy et al. (2012) in Fig. 9. Lacy et al. (2012) estimated the tsunami-
341 induced friction velocities via a fit to a logarithmic velocity profile based on velocities (low-
342 pass filtered and averaged over 10 s to eliminate effects from wind waves) from the lowest
343 three measurement locations, at times when the tsunami-induced boundary layer extended
344 at least 30 cm from the bed. Conversely, the time variation of the computed bed shear
345 stresses stem directly from the model i.e. equation (2) applied at $y = 0$. Results from Cases
346 1, 2 and 4 are specifically compared in Fig. 9, such that effects associated with the various
347 contributing factors may be clearly elucidated. As Cases 1 and 2 exclude the effects of
348 the shorter wind waves, the model result may be compared directly with the estimates of
349 Lacy et al. (2012). It is seen that while the inclusion of convective-acceleration effects (as
350 in Case 2) increases the friction velocity slightly relative to Case 1, both model results are
351 well beneath the field estimates. Since Case 4 has resolved the short wind waves directly,
352 the model results for the friction velocity are not directly comparable to the estimates of
353 Lacy et al. (2012), which again correspond to the tsunami-induced contribution. To enable
354 comparison for Case 4 we have therefore low pass (Butterworth) filtered the computed bed
355 shear stress (τ_b) results, to eliminate the higher-frequency short-wave contributions. These
356 filtered results have then been converted to the friction velocities depicted as the solid black

357 line in Fig. 9. Note that this follows a similar methodology as used by Lacy et al. (2012)
358 to isolate the tsunami-induced velocity signal from the raw signal, leading to the results
359 depicted in Fig. 2. While the peak in the friction velocity from the model occurs somewhat
360 earlier than estimated from the field data, the magnitude and overall variation for this case
361 are much closer to the field estimates than either Cases 1 or 2. This finding is generally in line
362 with our previous discussions, and quantitatively illustrates the additional flow resistance
363 felt by the long tsunami due to the presence of the short wind wave field.

364 Obviously, none of the simulations considered in the present case study utilizing a 1DV
365 approach have resulted in a perfect match with the field measurements. Such differences are
366 likely due, at least in part, to the treatment of the tsunami- and wind-wave-induced flow
367 as simply co-linear. This simplification neglects entirely any effects of directional spreading
368 within the short wave field, which will inevitably give rise to a three-dimensional flow and
369 turbulence field. Additional effects associated with any bed forms (e.g. increased roughness)
370 that may have been present would also serve to complicate matters further. Despite such
371 differences, the match achieved in the present case study can be considered as quite reason-
372 able, especially given the well-known difficulty of reproducing complicated field conditions
373 in numerical models. Comparisons with desired effects switched on and off, as done herein,
374 have likewise proved useful in quantitatively elucidating their likely importance in such field
375 conditions.

376 **CONCLUSIONS**

377 A numerical case study simulating the 2010 Chilean tsunami-induced boundary layer
378 flow, as captured in field measurements at Monterey Bay, California, USA by Lacy et al.
379 (2012), has been conducted. The study utilizes a one-dimensional vertical model, based on
380 the horizontal component of the Reynolds-averaged Navier-Stokes (RANS) equation, coupled
381 with $k-\omega$ turbulence closure, with flow driven by pressure gradients derived directly from the
382 measured velocities. Four cases have been considered involving the boundary layer induced
383 by: (1) the tsunami in isolation, (2) additional inclusion of convective acceleration effects

384 due to a sloping bed, (3) the tsunami in the presence of the short wind wave field, and (4) all
385 effects combined simultaneously. While none of the results can be considered as perfect, likely
386 due at least in part due to limitations with a one-dimensional vertical approach compared
387 with full three-dimensional field conditions, the boundary layer flow is reasonably captured
388 by the simulations. In particular, the simulation including all effects combined captures
389 the peak flow most accurately, while also maintaining similar errors during regions of both
390 flow acceleration and deceleration. Comparison of the various cases clearly demonstrates the
391 likely importance of both convective acceleration (due to a local bed slope) as well as the
392 presence of a “wave-induced roughness” effect similar to those in wave-plus-current flows,
393 where the developing boundary layer under tsunami effectively plays the role of the current.
394 The latter finding largely confirms that speculated by Lacy et al. (2012). Hence, both of
395 these effects should be accounted for in practical tsunami simulations, especially those where
396 the detailed boundary layer flow near the bed is of interest, e.g. studies of scour or sediment
397 movement.

398 **DATA AVAILABILITY**

399 The near-bed velocity measurements in Monterey Bay during the arrival of the 2010
400 Chilean tsunami are available at the U.S. Geological Survey data release: [https://doi.](https://doi.org/10.5066/P9T90P01)
401 [org/10.5066/P9T90P01](https://doi.org/10.5066/P9T90P01) (Ferreira et al. 2020).

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TABLE 1. Characteristic parameters for the observed tsunami event, as reported in Lacy et al. (2012).

Parameter	Symbol	Value	
Tsunami	Maximum depth averaged velocity	$U_{m,T}$	0.36 m/s
	Period	T_T	16 min = 960 s
	Maximum wave height	H_T	1.12 m
	Reynolds number	$Re_T = T_T U_{m,T}^2 / (2\pi\nu)$	2.0×10^7
Wind waves	Representative orbital velocity	$U_{m,w}$	0.4 m/s
	Period	T_w	10 s
	Significant wave height	H_s	1.3 m
	Reynolds number	$Re_w = T_w U_{m,w}^2 / (2\pi\nu)$	2.5×10^5
Water depth	h	9 m	
Mean sediment grain size	d	0.315 mm	

TABLE 2. The four distinct cases that are the basis for all the results

Case	Tsunami	Convective terms	Short waves
1	×		
2	×	×	
3	×		×
4	×	×	×

TABLE 3. Tabulation of the model root-mean-square error (RMSE) for each of the four tsunami wave boundary layer simulation cases during acceleration, at the peak, during deceleration. The total RMSE is also shown, corresponding to the mean of all depicted time levels.

Case	Figure	RMSE (m/s)			
		Acceleration	Peak	Deceleration	Total
1	3	0.0099	0.0119	0.0183	0.0139
2	4	0.0068	0.0087	0.0248	0.0151
3	5	0.0213	0.0211	0.0121	0.0171
4	6	0.0133	0.0056	0.0155	0.0136

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FIG. 1. A view of the more than 9000 km the tsunami covered from the earthquake epicenter to the site in Monterey Bay just south of the wharf. Figure created with Google maps.

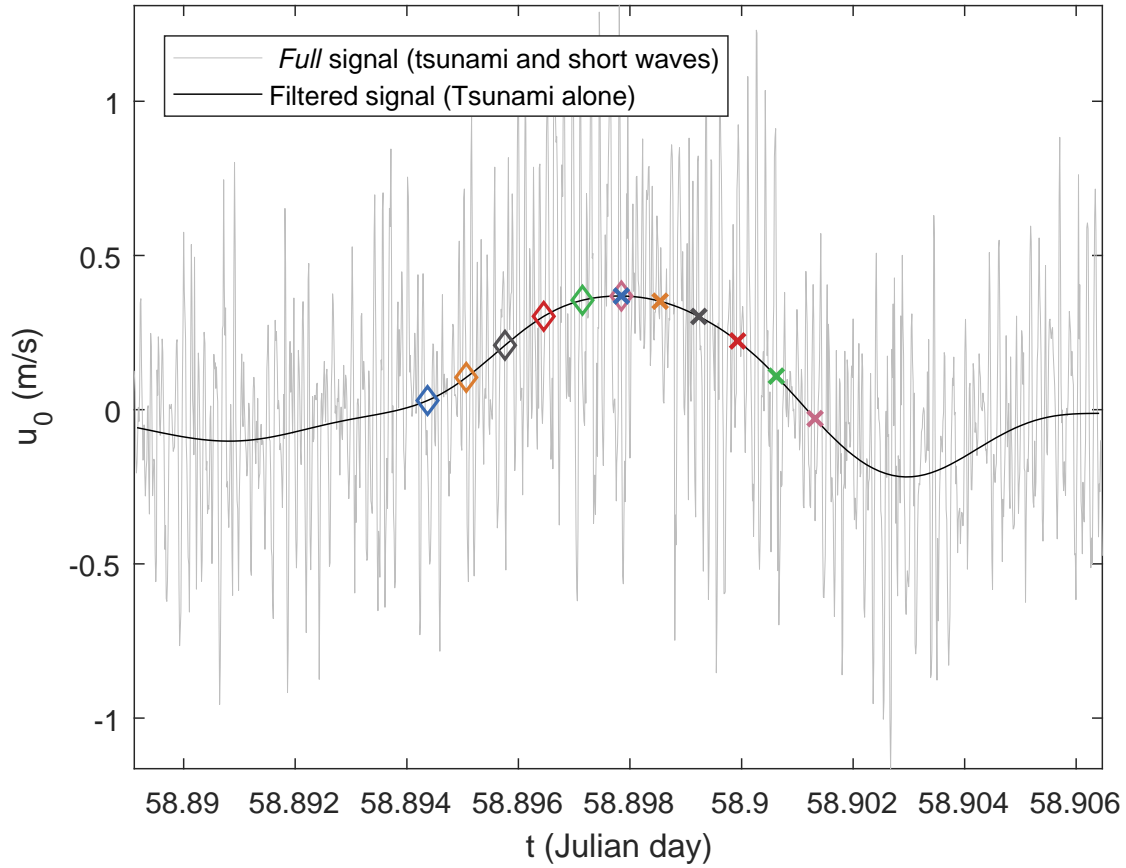


FIG. 2. A plot showing the field data of the full flow, the tsunami filtered from it, and the times at which comparisons will be made.

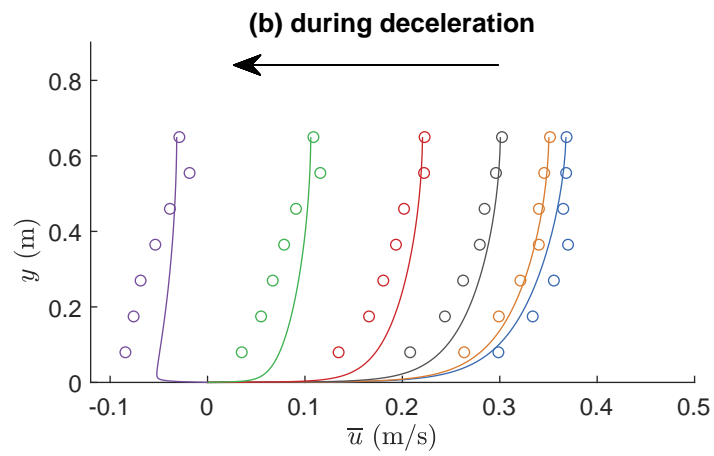
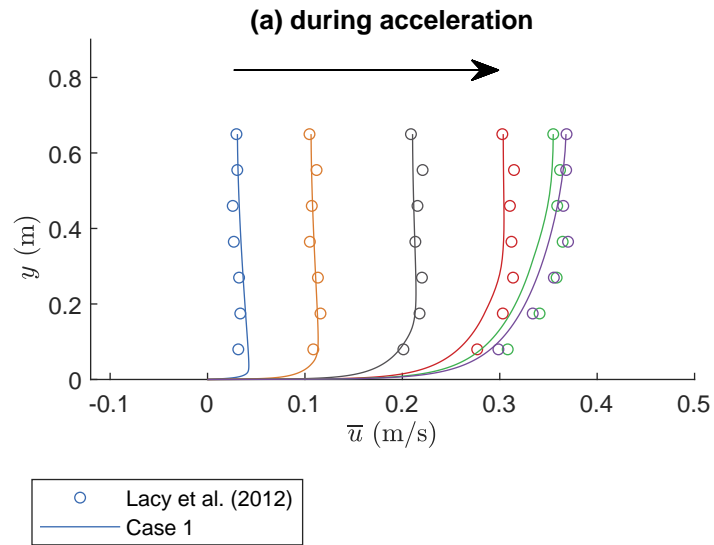


FIG. 3. Results of the tsunami simulation (Case 1) compared to the field data.

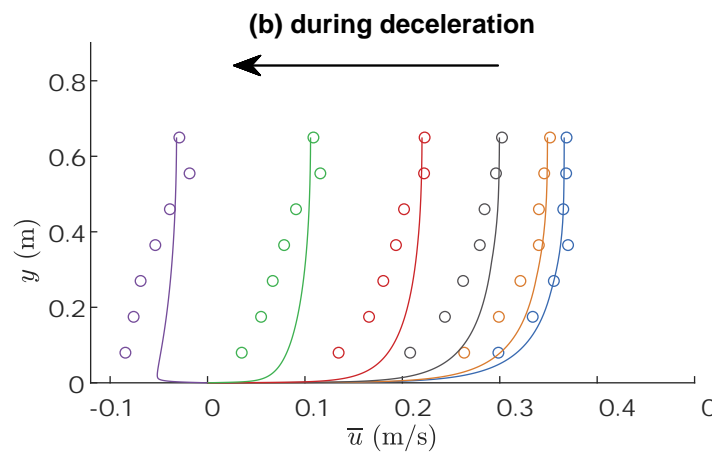
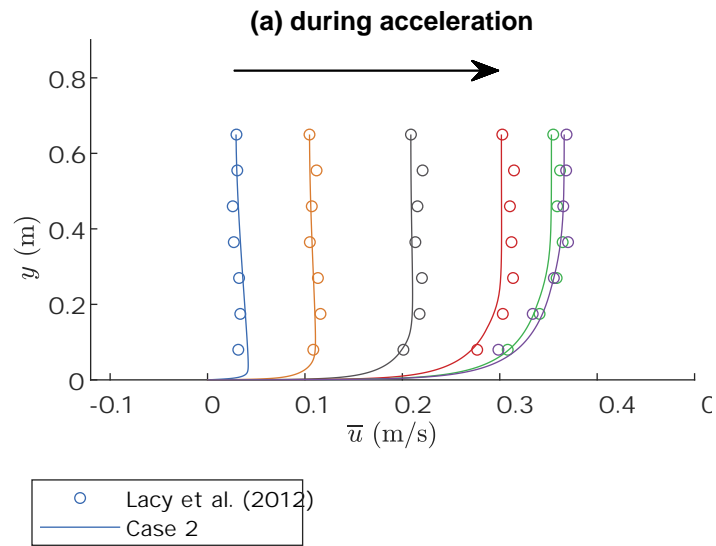


FIG. 4. Results of the tsunami simulation including converging-diverging effects (Case 2) compared to the field data.

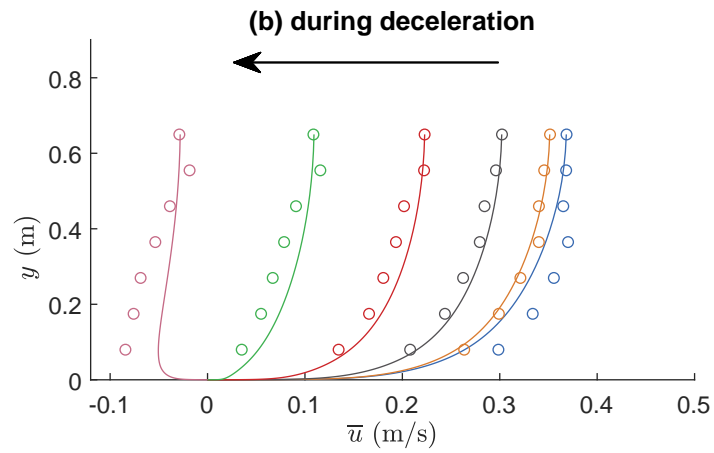
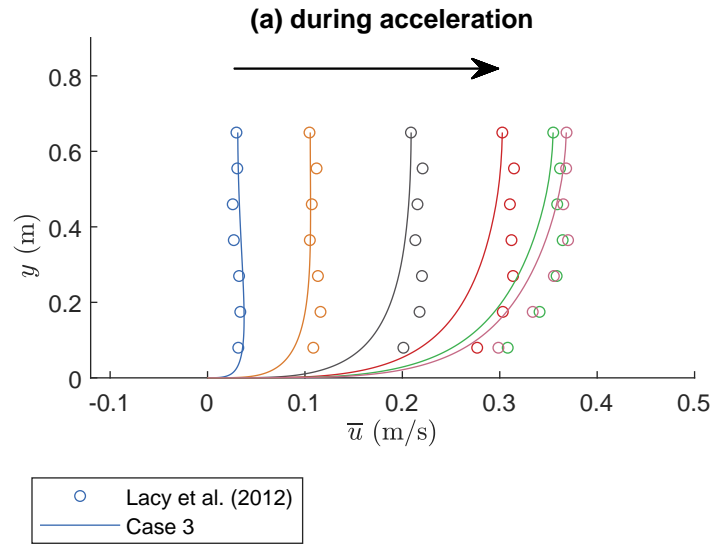


FIG. 5. Results of the tsunami and wind waves simulation (Case 3) compared to the field data.

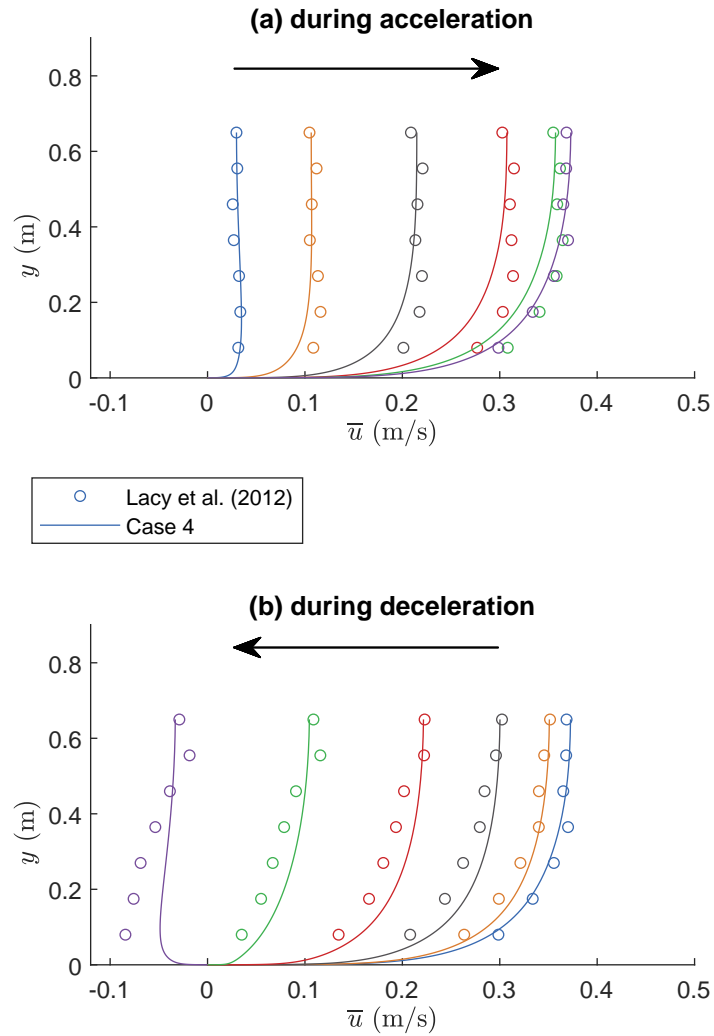


FIG. 6. Results of the tsunami and wind waves simulation including converging-diverging effects (Case 4) compared to the field data.

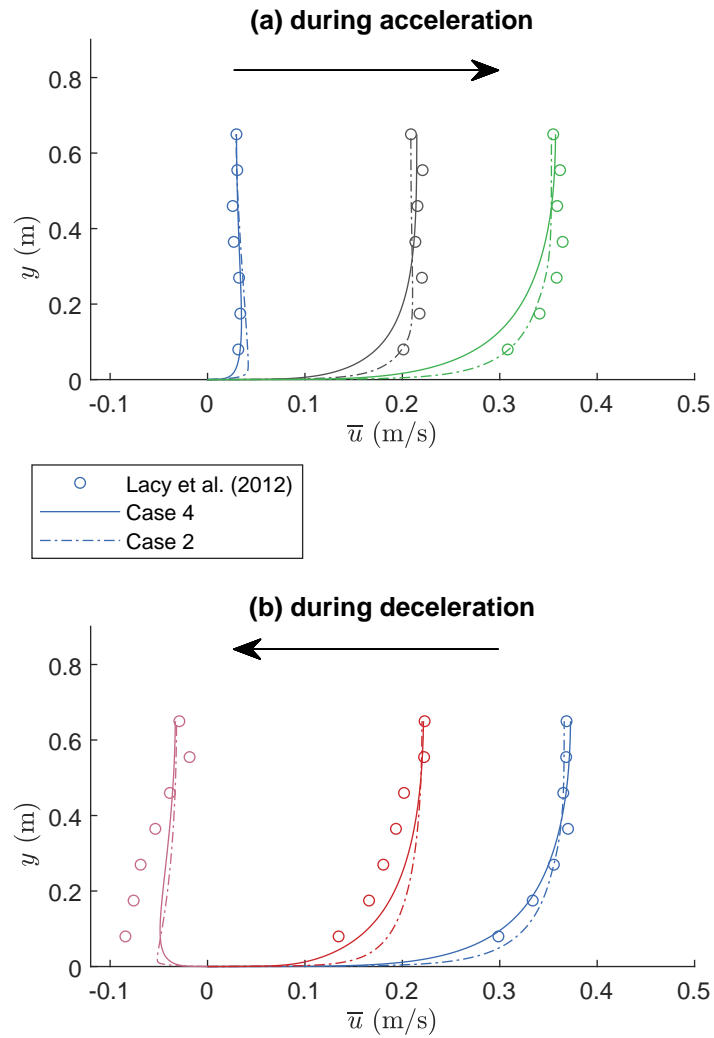


FIG. 7. The effect of the wind waves on the evolution of the boundary layer compared to only modeling the tsunami. The converging diverging effects are included in the simulation.

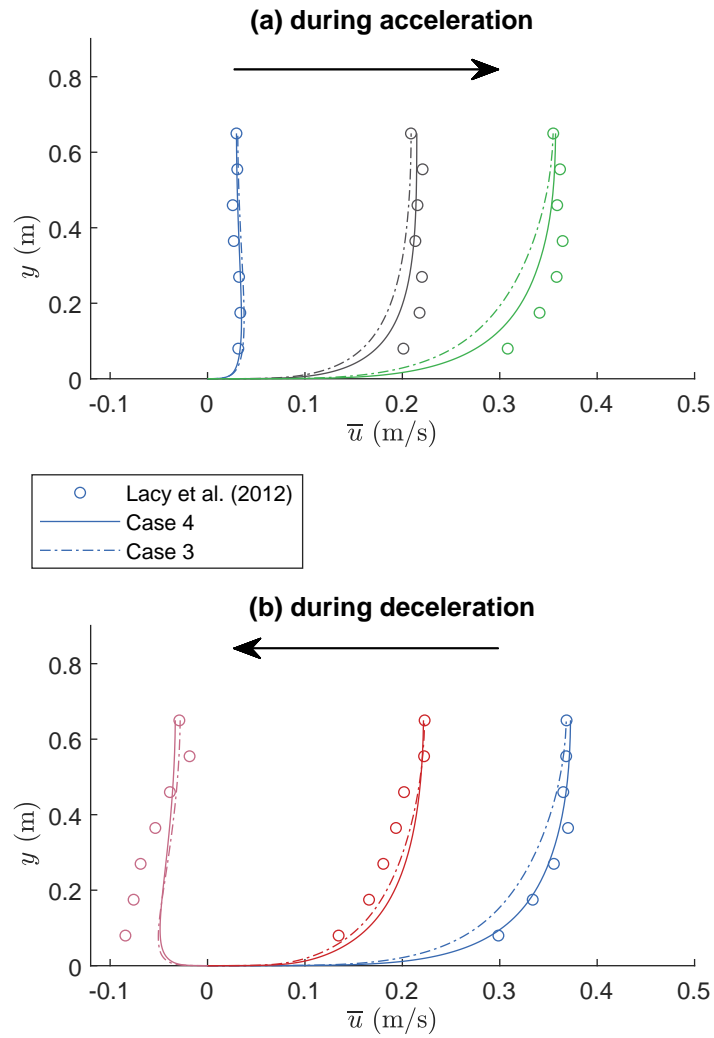


FIG. 8. The effect of the bed slope and therefore the converging-diverging effect on the evolution of the boundary layer. Both simulations include tsunami and wind waves.

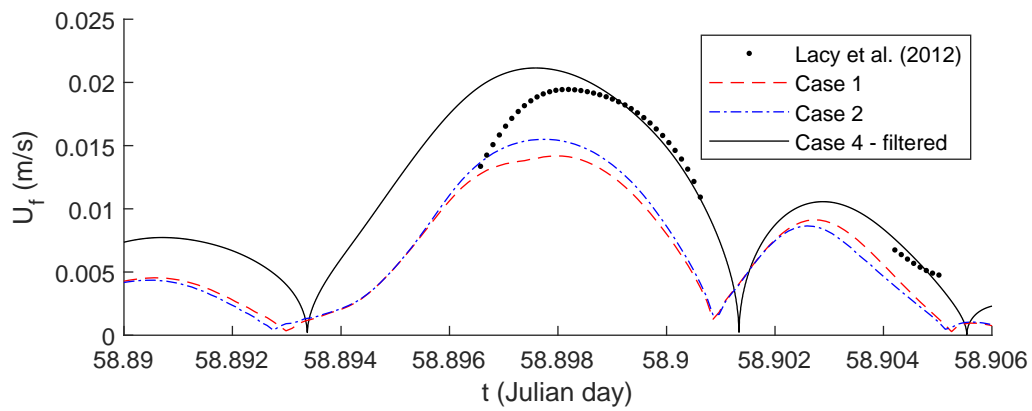


FIG. 9. Comparison of the computed tsunami-induced friction velocities (Cases 1, 2 and 4) with the estimated field values of Lacy et al. (2012). To isolate the tsunami-induced contribution the Case 4 results have been obtained from low-pass Butterworth filtering of the computed bed shear stress.