

1                    **A Subtidal Box Model based on the Longitudinal**  
2                    **Anomaly of Potential Energy for Narrow Estuaries. An**  
3                    **Application to the Guadalquivir River Estuary (SW**  
4                    **Spain).**

5                    **M. Cobos<sup>1</sup>, A. Baquerizo<sup>1</sup>, M. Díez-Minguito<sup>1</sup>, and M.A. Losada<sup>1</sup>**

6                    <sup>1</sup>Andalusian Institute for Earth System Research, University of Granada, Avda. del Mediterráneo s/n,  
7                    18006, Granada, Spain

8                    **Key Points:**

- 9                    • A balance equation of the longitudinal anomaly of the potential energy (LAPE)  
10                    was derived.  
11                    • The balance was applied in boxes to the GRE and explains the salt intrusion and  
12                    turbidity maxima.  
13                    • The dominant LAPE transports in the GRE vary spatially and also depend on  
14                    the hydraulic regime.

---

Corresponding author: Manuel Cobos, [mcobosb@ugr.es](mailto:mcobosb@ugr.es)

16  
17  
18  
19  
20  
21  
22  
23  
24  
25  
26  
27  
28  
29  
30  
31  
32  
33  
34  
35  
36

The objective of the present study is to demonstrate the informative capacity of the longitudinal anomaly of potential energy (LAPE) in the analysis of the magnitude and spatiotemporal variability of estuarine processes. For this purpose, a LAPE balance equation is formulated. The LAPE integrates and varies with the vertical and longitudinal density distribution. The formulation is applied on a subtidal scale to each box or stretch of the Guadalquivir River estuary, a narrow, highly turbid, weakly stratified, and strongly anthropized estuary. Data recorded by a large network of monitoring stations in 2008 and 2009 are used to quantify advective transports as well as the transports associated with longitudinal dispersion and vertical turbulent mixing in different hydraulic regimes. In low-river flow conditions, (river flows  $Q < 40 \text{ m}^3\text{s}^{-1}$ ), the magnitude of LAPE transports decreases upstream and varies locally, depending on neap-spring tidal cycles. The direction of the net LAPE transport creates convergence zones that are particularly consistent with maximum levels of estuarine turbidity. During high-river flows ( $Q > 400 \text{ m}^3\text{s}^{-1}$ ), this convergence disappears and the maximum longitudinal density gradient moves towards the mouth. More specifically, tidal pumping -induced LAPE governs during these conditions and manages to compensate the sum of the mean non-tidal and dispersive and differential advective LAPE transports. However, during the post-riverflood period, the mechanisms controlling recovery downstream from the mouth are the longitudinal dispersive and differential advective LAPE transports. Furthermore, the convergence zone reappears with a longitudinal gradient of the net LAPE transport that is even greater than in low-river flow conditions.

37

## 1 Introduction

38

### 1.1 Knowledge integration and simplification for estuary management

39  
40  
41  
42  
43  
44  
45  
46  
47

Human action is the main reason why many estuaries have experienced significant changes in recent decades. Such changes include: (i) modification of their hydrographic basins because of water diverted for irrigation, and the damming of the main water course and its tributaries; (ii) industrial and farming activities on their banks; and (iii) changes in estuary geometry because of depth and maintenance dredging (e.g.) (Hobbie, 2000; Ruiz et al., 2015; Reyes-Merlo et al., 2017). The anthropization of an estuary alters its biochemical and physical dynamics by modifying nutrient mixing as well as the sediment and oxygen transport in water bodies adjacent to rivers and seas (Simpson & Sharples, 2012; Officer, 2013).

48  
49  
50  
51  
52  
53  
54  
55  
56  
57  
58  
59

In certain estuaries, such as the Guadalquivir River Estuary (SW Spain), these interventions are reflected in changes in hydro-morphodynamics all along the estuary as well as in different hydraulic regimes of the fluvial agents (low-river flow, high-river flow, and post-riverflood recovery) and marine agents (neap and spring tides) (Ruiz et al., 2015). From an estuary management perspective, it is a considerable challenge to integrate this scientific evidence in simple, manageable models such as box models that are capable of quantifying the effects of such changes on estuarine dynamics. Box models are generally used to describe processes and to establish balances based on volumetric averages of the main oceanographic and estuarine variables with a suitable spatiotemporal scale (Officer, 1980; Hamilton et al., 1985; Austin, 2002; Rice et al., 2008; Díez-Minguito et al., 2014; Sun et al., 2017). This type of model can facilitate decision-making based on knowledge of the sustainable use of the estuary (Leschine et al., 2003).

60  
61  
62  
63  
64

Such decisions should be regulated by a legal framework such as the Water Framework Directive 2000/60/EC (WFD), which establishes the need to adopt measures to protect and improve transitional water bodies, such as estuaries, whose quality depends on their ecological status (Ganju et al., 2016). The WFD describes the ecological status of an estuary in terms of the values of hydromorphological parameters, such as the

65 current velocity, salinity, temperature, turbidity and mixing characteristics, among oth-  
 66 ers.

## 67 1.2 Longitudinal Anomaly of Potential Energy (LAPE)

68 The potential energy anomaly is a magnitude that integrates those hydromorpho-  
 69 logical parameters (Simpson & Hunter, 1974; Simpson, 1981; Simpson et al., 1990). In  
 70 energy terms, the potential energy anomaly,  $\phi_S$  ( $\text{Jm}^{-3}$ ), quantifies the water column strat-  
 71 ification and is defined as the amount of mechanical energy per  $\text{m}^3$  needed to instantane-  
 72 ously homogenize the water column (Simpson, 1981). This is mathematically expressed  
 73 as follows:

$$\phi_S(\mathbf{x}, t) = \frac{g}{d} \int_{-h}^{\eta} (\rho - \underline{\rho}) z dz, \quad \text{where} \quad \underline{\rho}(\mathbf{x}, t) = \frac{1}{d} \int_{-h}^{\eta} \rho dz, \quad (1)$$

74 where  $g$  is the gravity constant;  $t$  is the time;  $\mathbf{x}$  defines the horizontal position;  $z$ , rep-  
 75 represents the vertical coordinate;  $\rho(\mathbf{x}, z, t)$  is the water density, and  $\underline{\rho}(\mathbf{x}, t)$  is the reference  
 76 density in complete mixing conditions, both of which include the suspended sediment  
 77 density;  $d$  is the total depth that is the sum of  $\eta(\mathbf{x}, t)$ , the elevation over mean sea level  
 78 and  $h(\mathbf{x})$ , the mean depth. From the time when it was first proposed in the mid 1970s  
 79 (Simpson & Hunter, 1974), the potential energy anomaly has been widely used to iden-  
 80 tify physical processes that produce water exchanges in shelf seas (Simpson et al., 1990;  
 81 Hofmeister et al., 2009; Yang et al., 2017), regions influenced by freshwater (Simpson,  
 82 1997; De Boer et al., 2008), estuaries (Garvine & Whitney, 2006; Rice et al., 2008; Sun  
 83 et al., 2017), and lakes (Zhao et al., 2018).

84 Crucial to the present research study is the work of Garvine and Whitney (2006)  
 85 (GW2006), who used the potential energy anomaly to calculate the flow exchanges be-  
 86 tween the Delaware estuary and the adjacent continental shelf. The work of GW2006  
 87 establishes the potential energy anomaly budget in which the reference density is at a  
 88 point,  $\mathbf{x}_0$  located downstream on the continental shelf instead of the mean density in the  
 89 water column (eq. 1). The definition of  $\phi$  in GW2006 is henceforth referred to as the Lon-  
 90 gitudinal Anomaly of Potential Energy (LAPE) and is formulated as follows:

$$\phi(\mathbf{x}, t) = \frac{g}{d} \int_{-h}^{\eta} (\rho - \underline{\rho}_0) z dz, \quad \text{where} \quad \underline{\rho}_0 = \frac{1}{d} \int_{-h}^{\eta} \rho(\mathbf{x}_0, t) dz, \quad (2)$$

91 where  $\rho_0$  is the mean density of slope water beyond the shelf at the location  $\mathbf{x}_0$ . This  
 92 density is the reference density in complete mixing conditions, which are in this case, lon-  
 93 gitudinal. GW2006 represents the characteristics of the system with a single box to rep-  
 94 resent the estuary, which is directly connected to the continental platform at the sub-  
 95 tidal scale. They considered the change rate in the LAPE as induced by freshwater in-  
 96 flow, water outflow to the shelf, tidal and wind mixing, and surface heat flux. However  
 97 in relatively shallow estuaries that are weakly stratified or well-mixed, in addition to grav-  
 98 itational circulations, it is also necessary to consider other contributions to the LAPE,  
 99 such as those associated with longitudinal dispersion, tidal pumping, and differential ad-  
 100 vection (van Aken, 1986; Burchard & Hetland, 2010; Becherer et al., 2016). Differential  
 101 advection includes two terms. The first term is associated with depth-mean straining due  
 102 to the vertical mean horizontal density gradient strained by the deviation from the depth-  
 103 mean velocity vector (depth-mean straining). The second term is associated with the de-  
 104 viations of the longitudinal velocities and the gradient of density deviations (non-depth-  
 105 mean straining) (Burchard & Hofmeister, 2008).

## 106 1.3 Objectives and Organization

107 The objective of this research was to evaluate the joint spatiotemporal variability  
 108 of state variables in the WFD, such as density. This includes the contribution of the sus-  
 109 pended solids concentration and the flow velocity throughout the estuary and in differ-  
 110 ent fluvial regimes (i.e. low-river flow, high-river flow, and post-riverflood recovery) and

111 tidal regimes (neap and spring tides). This study led to the formulation of a subtidal-  
 112 scale dynamic equation that is able to evaluate the Longitudinal Anomaly of Potential  
 113 Energy (LAPE), and which integrates the previously mentioned state variables and the  
 114 transports that control them. This equation includes the impact of intratidal processes  
 115 on the subtidal scale by means of the tidal decomposition method (Díez-Minguito et al.,  
 116 2013; Scully & Friedrichs, 2007; Becherer et al., 2016; Burchard et al., 2018). It also ex-  
 117 tends the work of Garvine and Whitney (2006) by including the influence of the follow-  
 118 ing: (i) longitudinal dispersion; (ii) tidal pumping associated with cross-correlations be-  
 119 tween density, elevations, and currents; and (iii) differential advection due to depth-mean  
 120 straining, and non-depth-mean straining. These terms are particularly relevant to nar-  
 121 row estuaries (Jay & Musiak, 1994; Burchard & Hetland, 2010).

122 The LAPE general balance equation was applied to the dynamics of the Guadalquivir  
 123 River Estuary (GRE). Since the GRE is a narrow, weakly stratified and highly anthropized  
 124 estuary (Díez-Minguito et al., 2012, 2013, 2014; Ruiz et al., 2015), these same charac-  
 125 teristics made it possible to simplify the balance equation and extend Garvine and Whit-  
 126 ney (2006) by segmenting the GRE in  $N$  boxes. This box model was applied to the GRE  
 127 over a one-year period with all the previously mentioned hydraulic regimes. To better  
 128 explain the spatiotemporal variability of the LAPE, we analyzed a high-resolution data  
 129 set regarding salinity, suspended solids, tidal currents and elevations, and fluvial discharges,  
 130 which had been recorded in 2008-2009. These data were used to evaluate the LAPE and  
 131 the different LAPE transport terms in each box.

132 The rest of this paper is organized as follows. §2 presents the theoretical deriva-  
 133 tion of LAPE balance equation. In §3, the study zone is described, along with the mon-  
 134 itoring network and the configuration of the box model. §4 analyzes the processes ob-  
 135 served and relates them to the LAPE and the resulting transports of the model. After  
 136 Section 5 discusses the results and the validity of the model, Section 6 presents the con-  
 137 clusions derived from this research.

## 138 2 Formulation of the LAPE

139 This study focused on a relatively narrow estuary in which the Coriolis effect is neg-  
 140 ligible and the river flow is transversally uniform (Hansen & Rattray, 1966; Valle-Levinson,  
 141 2008, 2010). The system of coordinates  $(x, z)$  has the origin at the estuary mouth, ly-  
 142 ing over the mean sea level. The positive  $x$ -axis follows the thalweg and the  $z$ -axis points  
 143 upwards;  $d = \eta + h$  is the total depth;  $h(x)$ , the mean depth of the bed; and  $\eta(x, t)$  is  
 144 the water surface that varies over time,  $t$ .

### 145 2.1 Interpretation of the LAPE

146 According to GW2006, the LAPE,  $\phi$ , is the mechanical energy per  $\text{m}^3$  needed to  
 147 instantaneously homogenize the water mass between sections  $x_0$  and  $x$  (eq. (2)). In weak  
 148 stratification conditions, the LAPE can be simplified as follows:

$$\phi(x, t) \simeq \frac{[\rho_0 - \rho(x, t)]gd}{2}, \quad (3)$$

149 with  $x_0 < x$ , where  $x_0$  is located beyond the shelfbreak (Garvine & Whitney, 2006).  
 150 As can be observed in eq. (3), calculating  $\phi(x, t)$  is the same as calculating the poten-  
 151 tial energy of the semi-difference of densities between  $x$  and  $x_0$ . For a single box, eq. (3)  
 152 establishes that if the longitudinal density gradient is zero [ $\rho_0 = \rho(x, t)$ ], then  $\phi$  is zero.  
 153 This is the same as finding conditions of complete longitudinal mixing, as can be observed,  
 154 for example, in tidal river reaches. In normal conditions, density decreases upstream (Officer,  
 155 2013) and  $\phi > 0$ . However, in inverse estuaries or in regions where the profile of  $\rho(x)$   
 156 is locally inverted, density increases upstream and  $\phi < 0$ .



## 2.2 Development of the LAPE balance equation

The density,  $\rho(x, z, t)$ , is governed by the balance equation or the dynamic density equation (van Aken, 1986; Burchard & Hofmeister, 2008) defined as:

$$\mathbf{D}(x, z, t) = 0, \quad (4)$$

where  $\mathbf{D}$  is the density equation defined as:

$$\mathbf{D}(x, z, t) = \partial_t \rho + \partial_x(u\rho) + \partial_z(w\rho) - \partial_x(k_x \partial_x \rho) - \partial_z(k_z \partial_z \rho) - \mathbf{S}. \quad (5)$$

In eq. (5),  $\partial_\chi \zeta$  denotes the derivative of the generic variable,  $\zeta$ , in regard to the independent variable  $\chi$ . The terms  $k_x$  and  $k_z$  denote the longitudinal and vertical eddy diffusivities that result from the parameterizations of turbulent fluxes (Burchard & Hofmeister, 2008). The term  $\mathbf{S}(x, z, t)$ , encompasses the local source and sink terms, which includes the non-linear effects that appear from the equation of state for sea water (Burchard & Hofmeister, 2008), the incident solar radiation, and the suspended sediment:

$$\begin{aligned} \mathbf{S}(x, z, t) = & k_x [\partial_{\theta, x}^2 \rho (\partial_x \theta)^2 + \partial_{s, x}^2 \rho (\partial_x s)^2] + k_z [\partial_{\theta, z}^2 \rho (\partial_z \theta)^2 + \partial_{s, z}^2 \rho (\partial_z s)^2] \\ & + \frac{g\rho\alpha_w}{\rho_0 c_p} I e^{-\varepsilon(\eta-z)} + \left(1 - \frac{\rho}{\rho_s}\right) C \partial_z(w_s C), \end{aligned} \quad (6)$$

where  $\partial_{\chi, \varphi}^2 \zeta = \frac{\partial^2 \zeta}{\partial \chi \partial \varphi}$ ; the light penetration in the water column follows the Lambert-Beer Law, and where  $\alpha_w$ , is the water expansion coefficient;  $c_p$  is the heat capacity of the water;  $\varepsilon$  is the light absorption coefficient;  $\rho_s$  is the sediment density;  $w_s$  is the settling velocity; and  $C$  is the suspended sediment concentration. Since the values for precipitation, evaporation, and water diverted for irrigation are relative small compared to the fluvial discharges, they are not taken into account. The first term of the RHS in eq. (6) can be usually neglected because it is several orders of magnitude smaller than its counterpart in eq. (5). Assuming weakly-stratified to well-mixed conditions, the second term of the RHS of eq. (6) can also be neglected. This simplification is similar to considering a linear equation of state for seawater.

The dynamic equation of  $\phi$  is based on the expressions of  $\mathbf{D}$  (eq. (5)), evaluated in  $x$  and  $x_0$ , and the equation of the longitudinal anomaly of the potential energy (eq. (2)). When  $\mathbf{D}(x_0, z, t)$ , is depth-averaged, taking into account the continuity equation,  $\partial_x u + \partial_z w = 0$ , and the dynamic and kinematic boundary conditions (see Annex), the result is the following:

$$\begin{aligned} \underline{\mathbf{D}}(x_0, t) = & \partial_t(\rho_0) + \partial_x(\underline{u\rho_0}) - \partial_x(\underline{k_x \partial_x \rho_0}) - d^{-1}[k_x \partial_x \rho_0]_\eta \partial_x \eta - d^{-1}[k_z \partial_z \rho_0]_\eta \\ & - d^{-1}[k_x \partial_x \rho_0]_{-h} \partial_x h - d^{-1}[k_z \partial_z \rho_0]_{-h} - \frac{g\rho\alpha_w I}{\rho_0 c_p d} \left(1 - e^{-\varepsilon(\eta-z)}\right) - \left(1 - \frac{\rho}{\rho_s}\right) \frac{w_s}{2d} ([C^2]_\eta - [C^2]_{-h}), \end{aligned} \quad (7)$$

where subindexes  $_\eta$  and  $_{-h}$  indicate the respective surfaces on which the terms ( $z = \eta$  and  $z = -h$ ) are evaluated. The chain rule is applied to the derivatives and Leibniz's rule to integrals with variable limits. To calculate the average of  $\underline{\mathbf{D}}(x_0, t)$ , each dependent variable  $\zeta$  is decomposed in the superposition of the depth-averaged value  $\bar{\zeta}$ , and a deviation,  $\zeta_v$ . The same operation is performed with the term  $\underline{\mathbf{D}}(x, t)z$ . By substituting the results obtained in eqs. (7) and (A5), the following dynamic equation is obtained for the LAPE:

$$\begin{aligned} g(\underline{\mathbf{D}}(x_0, t)z - \underline{\mathbf{D}}(x, t)z) \equiv & \underbrace{\partial_t \phi}_{t_0} + \underbrace{g \partial_x([\rho_0 - \rho]u z)}_{t_1} + \underbrace{\partial_x(u \phi_S)}_{t_A} \\ & - \underbrace{g \partial_x(\rho u_v z)}_{t_B} - \underbrace{g \partial_x(u_v \rho_0 z - u_v \rho_v z)}_{t_C} + \underbrace{g(w_v \rho_v)}_{t_D} \end{aligned}$$

$$\begin{aligned}
& - \underbrace{\frac{g\rho\alpha_w I}{\rho_0 c_p d} \left[ \left(1 - e^{-\varepsilon(\eta-z)}\right) \underline{z} + \left(\varepsilon^{-1} - e^{-\varepsilon(\eta-z)} (d + \varepsilon^{-1})\right) \right]}_{t_G} + \underbrace{\frac{w_s}{2d} \left( [C^2]_{\eta}(1 + \eta) - [C^2]_{-h}(1 - h) - d\underline{C} \right)}_{t_E} \\
& + \underbrace{gd^{-1}k_z \left( [z\partial_z \rho_{0,v}]_{\eta} + [z\partial_z \rho_{0,v}]_{-h} \right)}_{t_F} + \underbrace{\partial_x \left( g[k_x \partial_x (\underline{\rho} + \rho_v)(z - \underline{z})] \right)}_{t_H} \\
& + \underbrace{gd^{-1}(z - \eta)[k_x \partial_x (\underline{\rho} + \rho_v)]_{\eta} \partial_x \eta - gd^{-1}(z - h)[k_x \partial_x (\underline{\rho} + \rho_v)]_h \partial_x h}_{t_I} = 0. \quad (8)
\end{aligned}$$

189 The first term  $t_0$  is the time variation of  $\phi$  (De Boer et al., 2008, eq. 6). The term  
190  $t_1$  represents the horizontal advection of the longitudinal density difference in relation  
191 to the average density. The term  $t_A$  is the longitudinal advection of the potential energy  
192 anomaly as defined by Simpson (1981),  $\phi_S$  (eq. (1)), due to the vertical average of the  
193 longitudinal velocity;  $t_B$  is the depth-mean straining, based on the vertical mean hor-  
194 izontal density gradient strained by the deviation from the depth-mean velocity vector  
195 (Bowden, 1981), also called the differential advection term by van Aken (1986) (see also  
196 Burchard and Hofmeister (2008));  $t_C$  is the non-averaged differential advection based on  
197 the deviations of the longitudinal velocities and the gradient of the density deviations.  
198 As for the potential energy anomaly,  $t_B$  and  $t_C$  include velocity deviations  $u_v$  and their  
199 longitudinal gradients. The term  $t_D$  is the vertical advection due to the deviations in re-  
200 gard to the vertical average of the vertical velocity. The term  $t_E$  is an estimation of the  
201 turbulent flux of density that depends on the eddy diffusivity coefficient (van Aken, 1986;  
202 Simpson & Bowers, 1981; Simpson & Sharples, 2012). The surface and bottom buoyancy  
203 transports are included in the term  $t_F$ . The source and sink terms related to heat trans-  
204 port on the free surface ( $R = (1 - e^{-kd}) \underline{z} + [\varepsilon^{-1} - e^{-kd} (d + \varepsilon^{-1})]$ ), (first term) and  
205 the bed sediment resuspension (second term) are indicated in  $t_G$ . Finally, the mean tur-  
206 bulent transport divergence throughout the estuary is defined by  $t_H$  whereas  $t_I$  is the  
207 divergence of the mean turbulent transport through the boundaries  $z = \eta$  and  $z = -h$ .  
208 To facilitate their tracking, these terms were renamed as follows:  $\Psi_{\zeta} = gd^{-1}(z - \zeta)[k_x \partial_x (\underline{\rho} +$   
209  $\rho_v)]_{\zeta} \partial_x \zeta$  for  $\zeta = \eta, -h$ .

### 210 **2.2.1 Application to weakly stratified estuaries**

211 In weakly stratified or well-mixed estuaries, the vertical density profile is practi-  
212 cally constant, i.e.  $\rho_v \simeq \rho_{0,v} \simeq 0$ . The term  $t_E$  cannot be neglected since it is the mech-  
213 anism that sustains mixing in the water column and determines the vertical mass ex-  
214 change in weakly stratified estuaries (Jay & Smith, 1990). Assuming this hypothesis, eq. (8)  
215 is simplified as follows:

$$\begin{aligned}
& \partial_t \phi + g \partial_x ((\underline{\rho}_0 - \underline{\rho}) \underline{u} \underline{z}) \\
& - g \partial_x (\underline{\rho} \underline{u}_v \underline{z}) - g \underline{k}_z \partial_z \underline{\rho} - \frac{g \rho \alpha_w}{\rho_0 c_p} R + \Psi_{\eta} - \Psi_{-h} = 0. \quad (9)
\end{aligned}$$

216 In the case of other estuaries, different hypotheses can be adopted that simplify the  
217 general equation (eq. (8)).

### 218 **2.3 Box model of the LAPE**

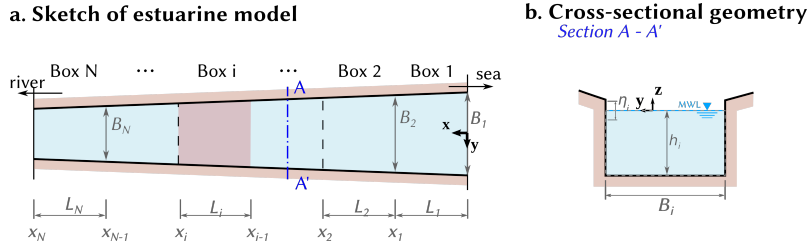
219 The following section describes how the estuary was divided into a series of inter-  
220 connected boxes that are representative of the subtidal-scale flow characteristics. For a  
221 box  $i$  (figure 1), eq. (9) is integrated with respect to a  $x$  between two sections,  $x_i$  and

222  $x_i+L_i$ . When the tidal decomposition of the resulting variables is substituted, this gives  
 223 the LAPE equation for the box at the subtidal scale:

$$\begin{aligned}
 & \underbrace{\frac{\Delta_t \bar{\Phi}}{\tau_0}}_{\tau_0} + \underbrace{\frac{gB}{2} \Delta_x F_2(d, \rho_0 - \rho, u)}_{\tau_1} - \underbrace{gB \Delta_x F_1(\rho, u, z)}_{\tau_2} \\
 & + \underbrace{\frac{4\epsilon C_d \bar{u}^3 A_s}{3\pi h}}_{\tau_3} + \underbrace{\frac{\delta k_s \rho_a \bar{W}^3 A_s}{h}}_{\tau_3} - \underbrace{\frac{g\rho\alpha_w A_s \bar{R}}{\rho_0 c_p}}_{\tau_4} \\
 & + \underbrace{\langle \Psi_\eta \rangle - \langle \Psi_{-h} \rangle}_{\tau_5} = 0, \tag{10}
 \end{aligned}$$

224 where for a generic variable  $\zeta$ ,  $\langle \zeta \rangle = \int_{x_i}^{x_i+L_i} \zeta dx$  is the integral in the length of a box,  
 225  $L_i$ ; and,  $\bar{\zeta} = T_{M_2}^{-1} \int_t^{t+T_{M_2}} \zeta dt$  is the time average in a tidal cycle  $T_{M_2}$ , which permits  
 226 the decomposition of  $\zeta = \bar{\zeta} + \zeta'$ , where  $\zeta'$  is the deviation in regard to the average. At  
 227 the same time,  $\Delta_t \zeta$  represents the time difference of  $\zeta$  evaluated in  $t$  and  $t+T_{M_2}$ ; and  
 228  $\Delta_x \zeta$  represents the spatial difference of  $\zeta$  evaluated in  $x$  and  $x+L$ . The functions  $F_1$   
 229 and  $F_2$  are given by:

$$\begin{aligned}
 F_1(\zeta_1, \zeta_2) &= \bar{\zeta}_1 \bar{\zeta}_2 + \overline{\zeta_1' \zeta_2'} \\
 F_2(\zeta_1, \zeta_2, \zeta_3) &= \bar{\zeta}_1 \bar{\zeta}_2 \bar{\zeta}_3 + \overline{\zeta_1' \zeta_2' \zeta_3'} + \overline{\zeta_2' \zeta_1' \zeta_3'} + \overline{\zeta_3' \zeta_1' \zeta_2'} + \overline{\zeta_1' \zeta_2' \zeta_3'}. \tag{11}
 \end{aligned}$$



**Figure 1.** Diagram of the estuarine box model and reference frame. *a)* Global diagram of the estuarine model. The red-shaded region stands for the surface area of box- $i$ ,  $A_{s,i}$ ; *b)* geometry of the cross-section. The gray dashed rectangle is the geometry of the box cross-section,  $A_i$ .

230 In eq. (10),  $\Phi = \int_{A_s} \phi dA$  ( $\text{J}\cdot\text{m}^{-1}$ ) represents the LAPE of the box. The term  $\tau_0$   
 231 is the variation experienced by  $\Phi$  from the beginning to the end of the tidal cycle. The  
 232 term  $\tau_1$  includes the superposition of the mean transport [ $\tau_{1,1} = h\bar{u}(\rho_0 - \rho)$ ], the Stokes  
 233 transport [ $\tau_{1,2} = \overline{(\rho_0 - \rho)\eta u'}$ ], and three tidal pumpings related to correlations between  
 234  $\eta + h$ ,  $u'$  and  $(\rho_0 - \rho)'$  [ $\tau_{1,3} = \overline{u\eta(\rho_0 - \rho)'}$ ;  $\tau_{1,4} = \overline{hu'(\rho_0 - \rho)'}$ ;  $\tau_{1,5} = \overline{\eta u'(\rho_0 - \rho)'}$ ].  
 235 The term  $\tau_2$  considers the depth-averaged advection density, spatiotemporal depth changes,  
 236 and mean vertical mass transports. The term  $\tau_3$ , which is obtained from ( $t_E$ ) in eq. (8),  
 237 is empirically related to tidal and wind mixing (van Aken, 1986; Simpson, 1981). Ac-  
 238 cording to Simpson (1981); Simpson and Bowers (1981); Simpson et al. (1991), the mix-  
 239 ing term is described by the cube of the tidal velocity  $\bar{u}$  and the wind velocity  $\bar{W}$ . The  
 240 constant parameters  $\epsilon$ ,  $\delta$ ,  $C_d$  and  $k_s$  represent the effective surface drag force, bottom drag  
 241 force, tidal current mixing, and wind mixing, respectively. The term  $\rho_a$  denotes the air  
 242 density at sea level. The surface solar radiation is  $\tau_4$ . Finally,  $\tau_5$  describes the turbulent

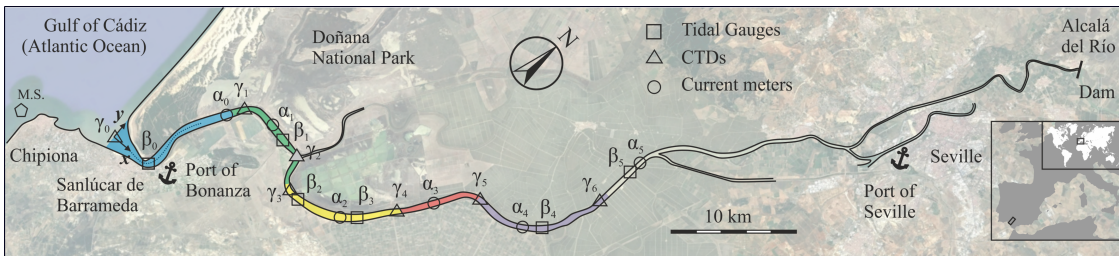
243 horizontal transport difference between  $z = \eta$  and  $z = -h$ . The LAPE transports due  
 244 to tidal pumping ( $\tau_{1,4}$ ), differential advection ( $\tau_2$ ) and longitudinal dispersion ( $\tau_5$ ) were  
 245 not contemplated by (Garvine & Whitney, 2006) in their box model.

246 In what follows, all variables are averaged in a tide cycle, although for the sake of  
 247 simplicity, the lower bar in the equations indicating this has been eliminated.

### 248 3 Application to the Guadalquivir River Estuary

#### 249 3.1 Study zone

250 The Guadalquivir River Estuary (GRE) is located in the southeast of Spain ( $36^\circ 43'N$   
 251 -  $37^\circ 32'N$ ,  $5^\circ 56'W$  -  $6^\circ 30'W$ ). It is a transition zone between the river basin of the same  
 252 name and the Gulf of Cadiz (figure 2). The estuary comprises the last 110 km, 85 of which  
 253 are navigable as far as Seville. The Alcalá del Río dam is the upstream tidal limit. The  
 254 estuary is weakly convergent: channel widths vary from 800 m near the mouth to 150 m  
 255 at the head. The channel mean depth is approximately equal to 7 meters, artificially main-  
 256 tained in many stretches by dredging to favor navigation.



**Figure 2.** Map of study area with the location of the monitoring network stations (Navarro et al., 2011). The monitoring network is composed of current-meter profilers (ADCPs), tidal gauges, a meteorological station (M.S.) and environmental quality probes or CTDs. The origin of the along-channel coordinate (km 0 at  $\gamma_0$ ) was established at the mouth and follows the thalweg. Colored stretches designate the extension of the boxes in the model (see Table 1). This color code is the same throughout the manuscript.

257 The GRE is a positive estuary, dominated by the flood tide and mesotidal range  
 258 (tidal range of  $<4$  m at the mouth), where the most significant tidal constituents are  $M_2$ ,  
 259  $S_2$ , and  $N_2$ . The GRE is generally dominated by tidal processes and is weakly strat-  
 260 ified when fluvial discharges are not greater than  $40 \text{ m}^3\text{s}^{-1}$ , which occurs over 75% of  
 261 the year (Díez-Minguito et al., 2012). Mean density deviations from the vertical density  
 262 average are less than  $1.42 \text{ kg}\cdot\text{m}^{-3}$ . The estuary number is less than 0.05; the densimet-  
 263 ric Froude number is smaller than 0.014; and the Richardson number of the estuary is  
 264 less than 0.3 (Díez-Minguito et al., 2013).

#### 265 3.2 Data collection

266 The data for the study were collected in real time from 2008 until 2010 with a re-  
 267 mote monitoring system installed by the Institute of Marine Sciences of Andalusia (ICMAN-  
 268 CSIC) (Navarro et al., 2011). Figure 2 shows the instrumentation at the estuary, namely,  
 269 six environmental-quality monitoring stations, six water level sensors, six acoustic Doppler  
 270 velocimeters, and a meteorological station, represented by  $\alpha$ ,  $\beta$ ,  $\gamma$ , and M.S., respectively.  
 271 The environmental quality station recorded temperature ( $T$ ), salinity ( $S$ ), and suspended  
 272 solids concentration every thirty minutes at four different depths ( $z=-1, -2, -3$  and  $-4$  m).

273 These data were used to calculate the fluid density  $\rho(S, T, C, p)$  (sea water and suspended  
 274 sediment mixture) as  $\rho = \rho_w + (1 - \rho_w/\rho_s)C$ , in which the water density,  $\rho_w(S, T, p)$   
 275 was obtained from the equation of state for seawater. The sediment density,  $\rho_s$ , was con-  
 276 sidered to be equal to  $2650 \text{ kg}\cdot\text{m}^{-3}$  (Díez-Minguito et al., 2014). Water level sensors recorded  
 277 the position of the free water surface every 10 minutes, and these data were used to cal-  
 278 culate the tidal range. Current data were measured every 15 minutes by means of acous-  
 279 tic current profilers. At the head of the estuary, daily data pertaining to the discharges  
 280 from the Alcalá del Río dam were obtained from the Regional Water Management Agency  
 281 (Agencia Andaluza del Agua, Junta de Andalucía). Wind velocity and direction and so-  
 282 lar radiation were measured every 10 minutes in the meteorological station at the mouth  
 283 of the GRE.

### 284 3.3 Configuration of the model

285 The previously described LAPE formulation was applied to the GRE from 07/24/2008  
 286 to 07/24/2009. Figure 3.f shows the following hydraulic regimes that were observed dur-  
 287 ing this one-year period (Díez-Minguito et al., 2012): (1) low-river flow; (2) high-river  
 288 flow; and (3) an intermediate riverflood recovery period. These regimes are discussed at  
 289 length in the sections that follow.

290 The spatial domain extended from the mouth,  $x = 0 \text{ km}$ , to  $x = 57.6 \text{ km}$  upstream,  
 291 the average position reached by the 2- $\text{psu}$  isohaline (Monismith et al., 2002), which char-  
 292 acterizes salt intrusion. The upper stretch of the estuary, which normally has the char-  
 293 acteristics of a tidal river, was not considered. The GRE was divided into five boxes, whose  
 294 length, width, and depth are defined in Table 1. These boxes are representative of es-  
 295 tuarine dynamics regarding tidal wave propagation, salt transport, and suspended par-  
 296 ticulate material transport along the GRE.

297 Equation (3) was used to calculate  $\phi$  in the five boxes, using the data observed (fig-  
 298 ure 3). In each box, the reference density,  $\rho_{0,i}$ , is the density at the point farthest down-  
 299 stream;  $\rho_i$  is the density at the point farthest upstream; and  $h_i$  is the average depth for  
 300  $i = 1, 2, \dots, 5$ . Typical average density values at these points are also shown in Table 1.  
 301 The equation (10) evaluates the transports that control the LAPE, which are derived  
 302 in Section 2.2.1, also based on the data observed.

303 In order to compare the results of our study with those of the seminal work of Garvine  
 304 and Whitney (2006) in the Delaware estuary, the average LAPE was also calculated through-  
 305 out the study zone,  $\phi_e$ , as though it were a single box. This was obtained with the fol-  
 306 lowing expression:

$$\phi_e = A_{s,t}^{-1} \sum_{i=1}^N \phi_i A_{s,i} \quad i = 1, 2, \dots, 5, \quad (12)$$

307 where  $A_{s,t} = \sum_{i=1}^5 A_{s,i}$ , the total surface area of the study zone. In the same way, the  
 308 net transport of the estuary was calculated for a term  $j$ , as follows:

$$\tau_{e,j} = A_{s,t}^{-1} \sum_{i=1}^N \tau_{j,i} A_{s,i} \quad i = 1, 2, \dots, 5. \quad (13)$$

## 309 4 Results

### 310 4.1 Observational background

311 This section analyzes the observations in order to ascertain the influence of the sus-  
 312 pended solids concentration, water temperature, and salinity on water density for the  
 313 different fluvial regimes during the study period. This analysis clarifies the interpreta-  
 314 tion of the spatiotemporal variability of the LAPE. In figure 3, panels (a), (b), (c) and

**Table 1.** Morphological parameters of the boxes used in the subtidal box model. Each box is defined by its initial downstream position  $x_i$ ; its final upstream position  $x_{i+1}$ , length  $L_i$ ; and the limiting cross-sections of areas  $A_{i+1}$  and  $A_i$  (figure 1). The surface area of each box is defined by  $A_{s,i}$ . Also shown are typical vertical averaged density values and their deviations in regard to the vertical average value,  $\rho_v$ , in each box. Besides the density values, the percentiles 5 and 95 of the time-series of density deviations are shown in brackets.

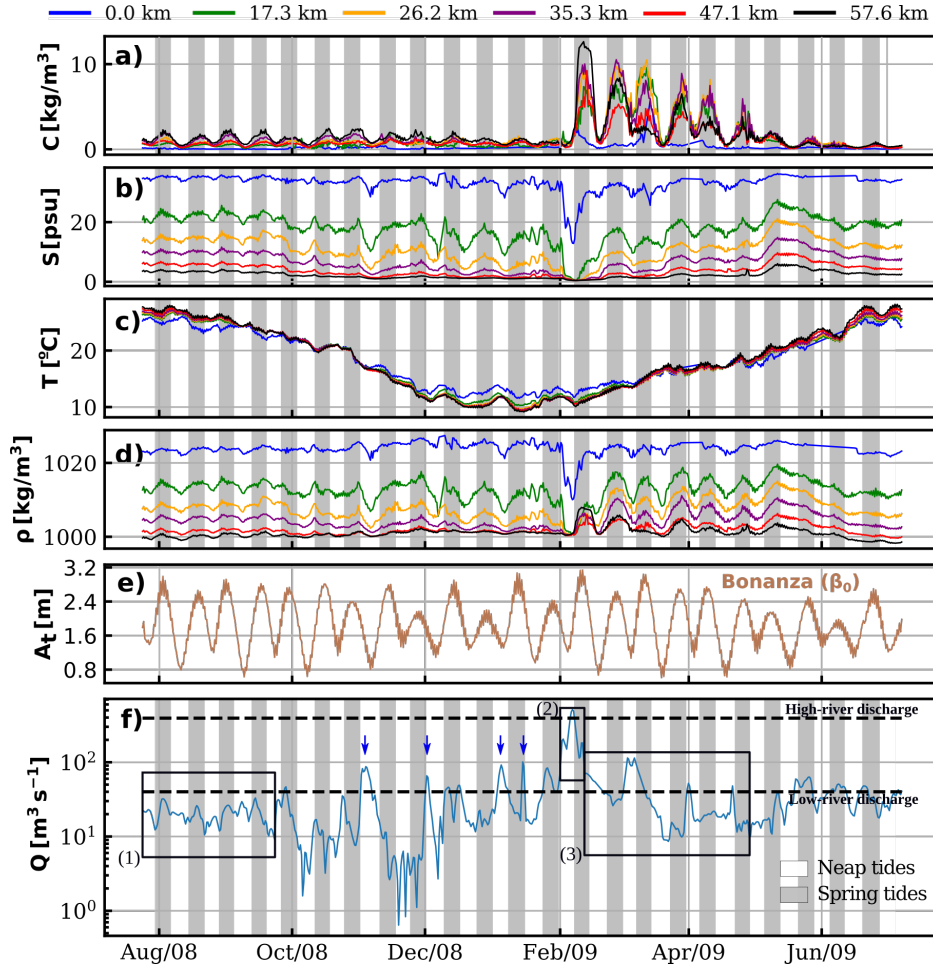
no.	Box		Geometry				Density variability	
	Color	$x_i$ (km)	$L_i$ (km)	$h_i$ (m)	$A_i$ ( $10^2 \text{ m}^2$ )	$A_{s,i}$ ( $10^6 \text{ m}^2$ )	$\bar{\rho}$ (5%, 95%) $\text{kg} \cdot \text{m}^{-3}$	$\rho_v$ (5%, 95%) $\text{kg} \cdot \text{m}^{-3}$
1	cyan	0	17.3	7.26	58.39	12.91	1023.75 (-4.30, 3.12)	0.41 (-0.31, 0.60)
2	green	17.3	8.9	7.13	43.82	5.22	1012.64 (-7.01, 6.82)	0.21 (-0.04, 0.24)
3	orange	26.2	9.1	7.05	37.80	4.54	1007.01 (-4.48, 4.70)	0.11 (-0.11, 0.44)
4	red	35.3	11.8	6.95	32.51	4.90	1004.22 (-3.12, 5.31)	1.42 (-1.20, 0.50)
5	purple	47.1	10.5	6.85	26.72	3.56	1001.67 (-1.99, 3.31)	0.07 (-0.05, 0.24)
-	-	57.6	-	6.80	22.45	-	1000.97 (-2.54, 3.37)	0.25 (-0.14, 0.20)
<b>TOTAL</b>		-	57.6	-	-	31.13	-	-

315 (d) show, respectively, the tidal-averaged time series of the vertically-averaged suspended  
316 solids concentration, salinity, water temperature, and density for different locations in  
317 the estuary (colored curves). Panel (e) shows the tidal range at the estuary mouth, and  
318 panel (f), the fluvial discharge during the study period.

319 During the summer of 2008 (1), the estuary remained in a low-river flow regime  
320 with fluvial discharges of less than  $40 \text{ m}^3\text{s}^{-1}$ . In these conditions, tidal currents deter-  
321 mine estuarine dynamics. The suspended particulate material throughout the estuary  
322 scarcely reached  $2 \text{ kg}\cdot\text{m}^{-3}$  during the summer. In autumn and the beginning of the win-  
323 ter in that same year, pulsed discharges reached  $100 \text{ m}^3\text{s}^{-1}$  and lasted for various days  
324 (blue arrows in figure 3.f). This produced a marked decrease in salinity and density ( $\Delta\rho \simeq$   
325  $5 \text{ kg}/\text{m}^3$ ).

326 In the middle of winter (2), intense rains made it necessary to open the floodgates  
327 of the dam upstream. The discharge reached the threshold of extreme fluvial conditions  
328 in the GRE ( $Q > 400 \text{ m}^3\text{s}^{-1}$ ). Density dropped sharply to freshwater levels in the en-  
329 tire estuary, except at the mouth, where the density reduction was  $15 \text{ kg}/\text{m}^3$ . The flu-  
330 vial discharge caused the salt to be transported seaward from the estuary. It also increased  
331 the quantity of sediments suspended from the bed, which led to a turbidity peak approach-  
332 ing  $10 \text{ kg}\cdot\text{m}^{-3}$  (fig. 3.a). During this period, the mouth showed the maximum horizon-  
333 tal density gradient of the estuary.

334 This episode was followed by a recovery period (3) in which the fluvial discharge  
335 of the dam again decreased with values below the threshold of  $40 \text{ m}^3\text{s}^{-1}$ . The salt that  
336 had been transported towards the mouth by the discharge began to move upstream again  
337 (figure 3.b). These conditions persisted until the beginning of the summer of 2009, when  
338 salinity levels and the sediment concentration were once more typical of low river flows.  
339 The summer was characterized by a slight linear reduction in density detected by all sen-  
340 sors. This was caused by the increase in water temperatures because of short-wave ra-  
341 diation (Padilla et al., 2016).



**Figure 3.** Time series of the tidal-averaged *a)* vertically-averaged concentration of suspended particulate material; *b)* vertically-averaged salinity; *c)* vertically-averaged temperature; *d)* vertically-averaged density; *e)* tidal range at Port of Bonanza; and, *f)* daily river discharges from the Alcalá del Río dam during the study period. Several boxes are displayed in panel *f)*: (1) refers to the low-river flow regime; (2) refers to the high-river flow regime; and (3) stands for the recovery period. Spring and neap periods are also displayed. Blue arrows indicate pulsed discharges lasting various days, reaching  $100 \text{ m}^3 \text{ s}^{-1}$ , which were not normal conditions for the estuary.

342

## 4.2 Evaluation of the LAPE

343

### 4.2.1 Longitudinal variation of $\phi(x)$

344

345

346

347

348

349

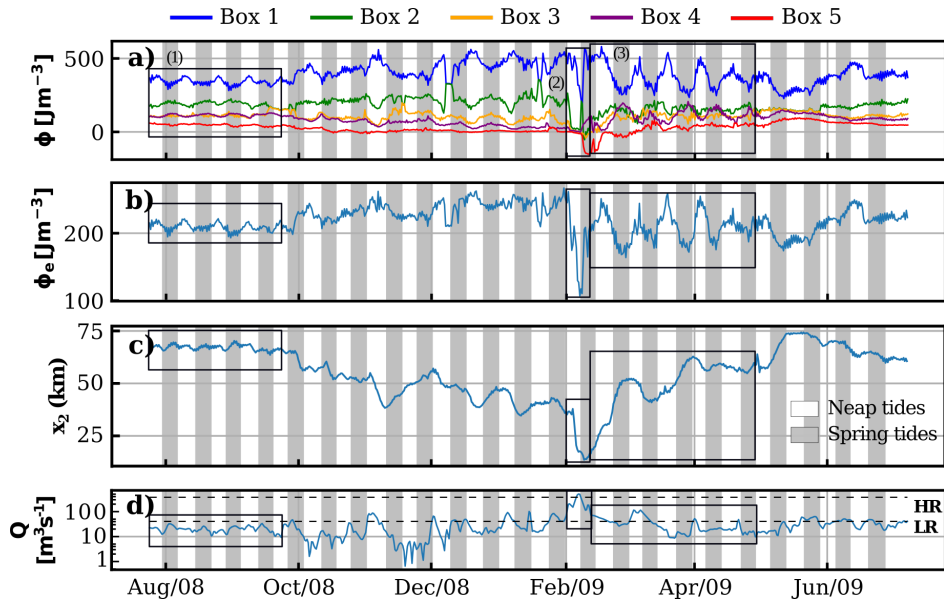
350

351

352

Figure 4.a shows the results of eq. (3) at several sections representative of the conditions at the boxes defined in Table 1, where  $\phi_i$  indicates the longitudinal anomaly of the potential energy (LAPE or  $\phi$ ) at  $x_i$ . The first third of the estuary (boxes 1 and 2) shows the highest values of  $\phi$ , where the maximum average longitudinal density gradients occur as well as the greatest time variability. The LAPE for box 1 during the study period is  $\phi_1 = 395.04 \pm 122.68 \text{ Jm}^{-3}$ , where the first term is the mean value and  $\pm X$  is the standard deviation. The amplitude of the fortnightly modulation of  $\phi_i$  and its mean value decrease upstream ( $\phi_5 \simeq 90 \pm 45 \text{ Jm}^{-3}$ ). In box 1,  $\phi$  oscillates out of phase with the fortnightly constituent of the tidal range with minimums of  $\phi$  during spring tides and

353 maximums during neap tides. Upstream, just the opposite occurs. The reason of the fort-  
 354 nightly change of  $\phi$  will be discussed in the next section. The dependence of  $\phi$  on the  
 355 neap-spring-tide cycles was no longer visible from October 2008 until February 2009 be-  
 356 cause the intermediate fluvial pulsed discharges became more numerous ( $\simeq 100 \text{ m}^3/\text{s}$ ) (fig-  
 357 ure 4.d). The 2-psu isohaline (figure 4.c) reaches box 5, where  $\phi_i$  remains almost const-  
 358 ant during low riverflow conditions. However, it retreats to  $x = 40 \text{ km}$  with the pulsed  
 359 discharges observed at the end of autumn which is manifested in a reduction of  $\phi$  in boxes  
 360 4 and 5. Then, it moves to box 1 during the flood (2), causing a marked decrease in  $\phi$   
 361 in all boxes. On a yearly scale, at the mouth, positive temperature increases generated  
 362 negative increases in  $\phi$ . Upstream, the temperature oscillation is in phase with  $\phi$ .



**Figure 4.** Time series of the longitudinal potential energy anomaly. *a)*  $\phi_i$  at every box; *b)*  $\phi_e$  of the estuary (eq. (12)); *c)* salt intrusion during the study period (2-psu isohaline from figure 8 of Díez-Minguito et al. (2013)); and, *d)* daily river discharges from the Alcalá del Río dam during the study period. HR and LR indicate high-river flow and low-river flow regime. Boxes represent the periods discussed in the main text.

363 Generally, the value of  $\phi_e$  remains more or less constant in low-river flow conditions  
 364 (1), which is to be expected because of the small time variations in the density observed  
 365 at the estuary mouth and head (figure 3.d). During the high-river flow regime (2),  $\phi_e$   
 366 decreases by half because of the seaward movement of the salt intrusion. The dependence  
 367 of the salt intrusion increases with the freshwater flow (Reyes-Merlo et al., 2013). The  
 368 salt intrusion during (2) moves upstream to kilometer 15. During the subsequent recov-  
 369 ery period (3),  $\phi_e$  oscillates out of phase  $\pi$  with the tides. The mean value and variabil-  
 370 ity of  $\phi_e \simeq 218.82 \pm 36.89 \text{ Jm}^{-3}$  are in accordance with the values obtained by Garvine  
 371 and Whitney (2006) in the Delaware estuary where the mean estuary depth is 8 m. The  
 372 amplitude of the oscillation triples during the recovery period ranging from a little more  
 373 than  $15 \text{ Jm}^{-3}$ , corresponding to what is observed in normal conditions, to almost  $50 \text{ Jm}^{-3}$ ,  
 374 mostly conditioned by the box of the estuary mouth.



375

#### 4.2.2 Variation of $\phi(x)$ depending on fluvial regime

376

377

378

Figure 5 shows the profiles of  $\rho(x)$  and  $\phi(x)$  for the different regimes (low-river flow, high-river flow, and post-riverflood recovery). In all cases, it was observed that the variability in regard to that mean density profile (orange line, figure 5) decreases upstream.

379

380

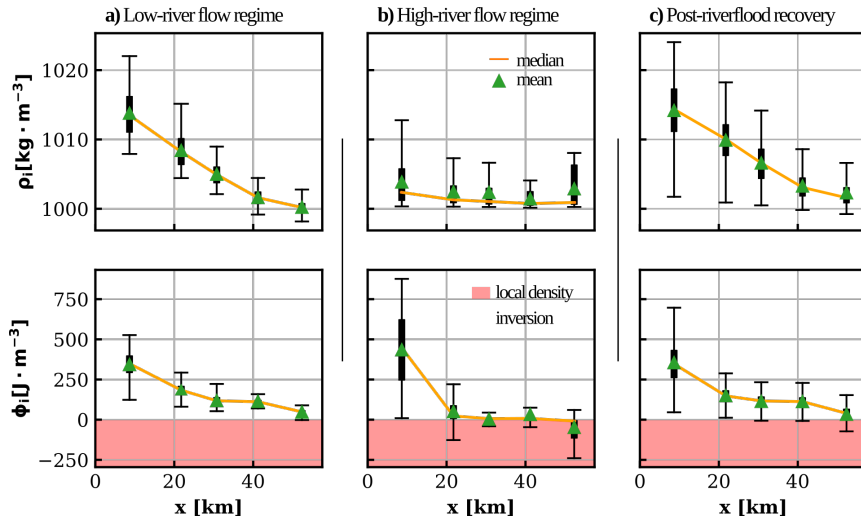
381

382

383

384

In normal conditions (1) (fig. 5.a), the mean density in each box decreases from the mouth upstream. This decrease is linear to km 40. The LAPE decreases more gradually upstream. During the high-river flow regime (2) (figure 5.b), almost the entire estuary is composed of freshwater, which gives values of  $\phi \simeq 0$  except for the mouth. In boxes 3 and 5,  $\phi$  is negative, which means that the density is locally inverted ( $\rho_3 > \rho_2$  and  $\rho_5 > \rho_4$ ).



**Figure 5.** Along-channel density and longitudinal anomaly of potential energy for different regimes: *a*) low-river flow; *b*) high-river flow; and, *c*) post-riverflood recovery period. The rectangle in error bars indicate the corresponding  $\pm 1$  quartile from the median whereas line edges indicate 5 and 95 percentiles. Shaded areas indicate local density inversion  $\rho_i > \rho_{i-1}$ .

385

386

387

388

389

In the recovery period (3), the along-channel density (fig. 5.c, upper panel) shows a similar profile to that observed in normal conditions, where the variability in  $\rho(x)$  as well as  $\phi(x)$  is maximum during recovery. The negative values of  $\phi$  persist until the density observed in normal conditions is recovered because the values approach the estuarine turbidity maximum (ETM) (Díez-Minguito et al., 2014).

390

### 4.3 Evaluation of LAPE transports

391

#### 4.3.1 Longitudinal variation of advective transports of $\phi$

392

393

394

395

396

397

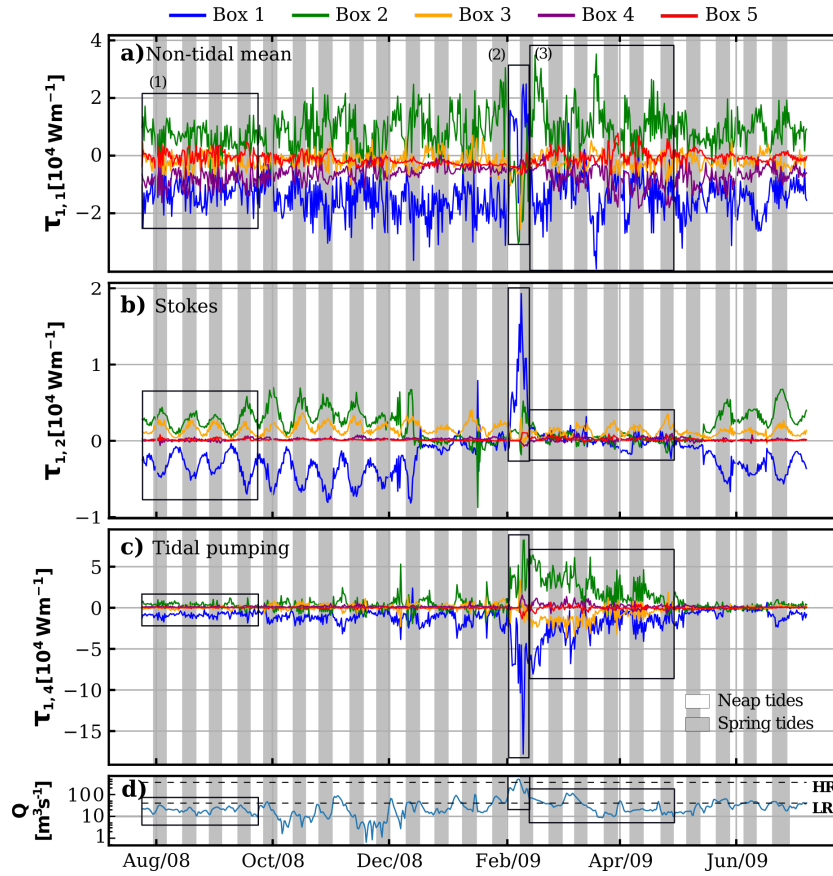
398

399

400

Figure 6 shows the main advective transport terms in each box, calculated in accordance with equation (10). Terms  $\tau_{1,1}$ ,  $\tau_{1,2}$  and  $\tau_{1,4}$  represent the average non-tidal advective  $\phi$ -transport, the LAPE transport due to Stokes and the tidal pumping  $\phi$ -transport induced by the correlation between currents and densities. These terms represent more than 99% of the advective transport of  $\phi$ . The terms  $\tau_{1,3}$  and  $\tau_{1,5}$  are not shown since they are various orders of magnitude lower than the latter ones ( $\simeq 10^2 \text{ Wm}^{-1}$ ). At the estuary mouth, the mean non-tidal advective LAPE transport,  $\tau_{1,1}$ , is negative  $\simeq -1.5 \cdot 10^4 \text{ Wm}^{-1}$  (blue curve in fig. 6.a), which indicates that the non-tidal residual currents flow seaward. Upstream, this mean value becomes positive (green curve), then negative

401 (purple curve), and then positive (red curve) again, which indicates the convergence and  
 402 divergence zones of  $\tau_{1,1}$  all along the estuary. This longitudinal  $\phi$ -transport variation is  
 403 also observed in Stokes-induced and tidal pumping transport processes.



**Figure 6.** Time series of the most significant advective terms of the LAPE along the estuary. *a)* Non-tidal mean; *b)* Stokes; *c)* tidal pumping due to the correlation between currents, elevations and densities; and, *d)* daily river discharges from the Alcalá del Río dam over the study period. HR and LR indicate the high-river and low-river flow regime, respectively. Positive values are upstream. Boxes represent the periods discussed in the main text.

404 During the low-river flow regime (1),  $\phi$ -transports due to Stokes ( $\tau_{1,2}$ ) and tidal  
 405 pumping ( $\tau_{1,4}$ ) are in phase with the tidal range. At the estuary mouth, the dependence  
 406 of  $\tau_{1,1}$  on the tidal range is not as clear. The term,  $\tau_{1,1}$  has a greater magnitude than  
 407  $\tau_{1,4}$  and  $\tau_{1,2}$ . More specifically,  $\tau_{1,2}$  is approximately one-fourth of  $\tau_{1,1}$ .

408 The non-tidal mean advective  $\phi$ -transport and tidal pumping are primarily respon-  
 409 sible for the fortnightly change in density during the low-river flow ( $\simeq 2 \cdot 10^4 \text{ Wm}^{-1}$ , in  
 410 absolute value). Transport magnitude decreases upstream except in box 3 where it in-  
 411 creases. Upstream transports (box 4 and 5) become negligible.

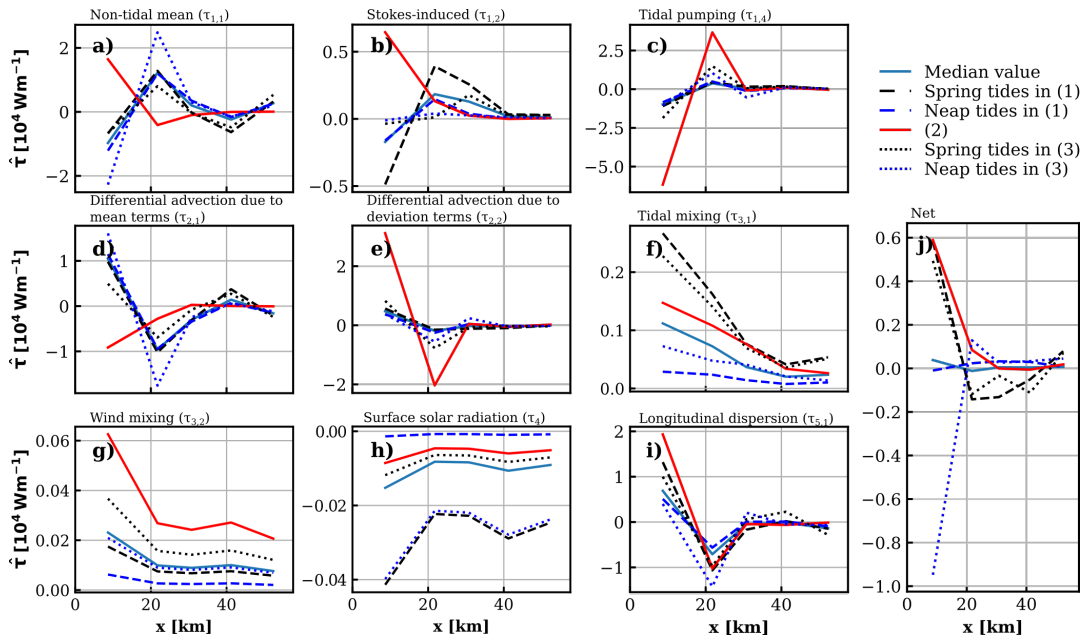
412 In the high-river flow regime (2),  $\tau_{1,1}$  is inverted and the transport in boxes 2 and  
 413 3 drops to  $\simeq -2.5 \cdot 10^4 \text{ Wm}^{-1}$ , whose absolute value exceeds that of the first box. Non-  
 414 tidal mean advective and Stokes  $\phi$ -transports at the estuary mouth are in the upstream  
 415 direction in contrast to what occurs in the following box. The  $\tau_{1,2}$  and  $\tau_{1,4}$  transports  
 416 intensify with the discharge. This means that tidal pumping ( $\tau_{1,4}$ ) is now the dominant

417 process in the first third of the estuary, moving seaward at the mouth ( $\simeq -15 \cdot 10^4 \text{ Wm}^{-1}$ )  
 418 and upstream in the following box ( $\simeq 5 \cdot 10^4 \text{ Wm}^{-1}$ ).

419 During the recovery period (3), the mean transport regains its values in low-river  
 420 flow conditions and when it is in phase with the tide. The Stokes-induced LAPE trans-  
 421 port ( $\tau_{1,2}$ ) recovers the values observed in (1). In the case of tidal pumping, maximums  
 422 observed during the discharge progressively attenuated until the total recovery of the es-  
 423 tuary in May 2009.

#### 4.3.2 Longitudinal variability of $\phi$ -transports for different fluvial regimes

425 This section jointly analyzes, in terms of LAPE transport, the longitudinal vari-  
 426 ability of the dominant processes observed in the GRE during periods of neap and spring  
 427 tides in normal conditions, recovery conditions, and during the high-river flow regime.  
 428 Figure 7 shows the median values ( $\hat{\tau}$ ) of the LAPE transport results.



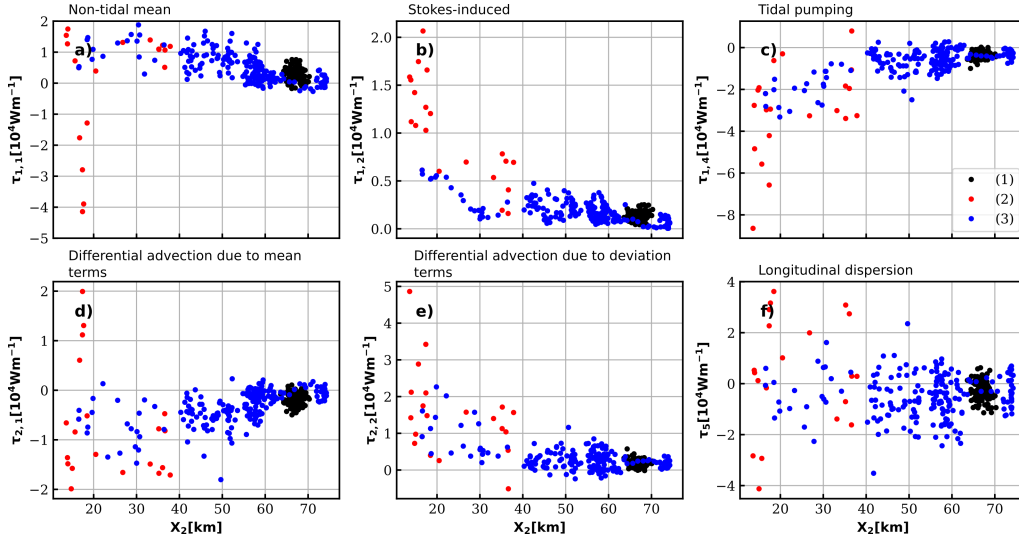
**Figure 7.** Median values of several LAPE transports along the estuary under different flow conditions. Panel *j*) shows the sum of terms *a*) to *i*). Low-river flow, high-river flow and recovery periods are indicated as (1), (2), and (3), respectively. Positive values are upstream.

429 During (1), non-tidal mean, Stokes-induced, and tidal pumping  $\phi$ -transports at the  
 430 mouth are seaward. This behavior is inverted upstream ( $x=26.2 \text{ km}$ ), where the value  
 431 is of the same order, but positive. From here upstream, these  $\phi$ -transports decrease and  
 432 show a negative value in  $x=47.5 \text{ km}$ . This variability causes various points of energy trans-  
 433 port convergence and divergence to appear throughout the estuary. More specifically,  
 434 near kilometer 15, there is a divergence point of  $\tau_{1,1}$ ,  $\tau_{1,2}$  and  $\tau_{1,4}$ . Near  $x=35 \text{ km}$ , an-  
 435 other convergence point for the same transports is also identified. In spring tides, the  
 436 mean LAPE transport value is greater near the mouth and decreases after the divergence  
 437 point. The term  $\tau_{1,1}$  varies considerably along the estuary, due to phase changes between  
 438 the longitudinal velocity and density. Although the longitudinal transport profile intensifies  
 439 during spring tides in the estuary recovery period, in neap tides, this same value  
 440 visibly decreases in  $x=17.3 \text{ km}$ . It is important to highlight how the longitudinal pro-  
 441 file of  $\tau_{1,1}$  is inverted during the fluvial discharge (2). Moreover, as can be observed the

442 net LAPE transport in spring tides (fig. 7.j) is positive at the mouth and negative, up-  
 443 stream, whereas in neap tides, it is practically zero in the entire estuary.

444 The Stokes-induced  $\phi$ -transport (fig. 7.b) shows a similar pattern during flow condi-  
 445 tions (1) and (2) except for the fact that it becomes practically negligible upstream  
 446 from the divergence point and during the recovery phase. In any case, the Stokes-induced  
 447  $\phi$ -transport has a magnitude that is slightly less than that obtained for the non-tidal mean  
 448 advection. The upstream decrease in the Stokes-associated  $\phi$ -transport is also controlled  
 449 by the phase lag between elevations and tidal currents. This phase lag approached  $\pi/2$   
 450 upstream because of the partial standing nature of tidal wave propagation due to the  
 451 tide reflection in the dam at the head of the estuary (Díez-Minguito et al., 2012). Tidal  
 452 pumping  $\tau_{1,4}$  (figure 7.c) also has the same profile with the absolute maximum value of  
 453 its magnitude at the estuary mouth. During the high-river flow regime (2),  $\tau_{1,4}$  main-  
 454 tains the same longitudinal profile and significantly increases so as to become the domi-  
 455 nant process in these conditions.

456 As can be observed in panels (d), (e), and (i) of figure 7, the LAPE transports due  
 457 to differential advection and dispersion also have an important role in the balance. Trans-  
 458 port values due to longitudinal dispersion are greater near the mouth and decrease up-  
 459 stream. In low-river flow conditions, the transport median value approaches  $1.3 \cdot 10^4 \text{ Wm}^{-1}$   
 460 in spring tides and is about half that value in neap tides. This shows that dispersive trans-  
 461 port is very important in the bottom third of the estuary with a contribution that in cer-  
 462 tain conditions reaches 30% of the total transport. Transport induced by differential advec-  
 463 tion of the average terms is the second dominant process in magnitude during (2),  
 464 and is comparable to the dispersive transport,  $\tau_5$ . After kilometer 35.3, transport induced  
 465 by the differential advection of the mean terms and dispersion practically disappears.



**Figure 8.** Net-estuarine transports of LAPE versus salt intrusion (2-psu isohaline). *a)* Non-tidal mean advection; *b)* Stokes advection; *c)* tidal pumping advection due to the correlation between currents, elevations and densities; *d)* differential advection due to mean terms; *e)* differential advection due to deviation terms; and, *f)* longitudinal dispersion. Colors refer to normal conditions (black), high-river flow (red), and post-riverflood (blue) conditions. Positive values are upstream.

466 The contribution of the surface solar radiation and mixing terms to  $\phi$ -transport (pan-  
 467 els (f), (g), and (h) of figure 7) is not significant. In the Delaware estuary studied in Garvine

468 and Whitney (2006) and Rice et al. (2008), the solar radiation and tidal and wind mix-  
469 ing  $\phi$ -transports are significant buoyancy terms. On the other hand, in the GRE case,  
470 the contribution of these terms are not significant because they depend on the surface  
471 area, and the boxes considered in the model (see Table 1) are two order of magnitude  
472 smaller than the surface area of the Delaware estuary ( $2.1 \cdot 10^9 \text{ m}^2$ , Table 2 in GW2006).

473 Figure 8 shows the most significant net LAPE transports and their relation to the  
474 2-psu isohaline in different hydraulic regimes. The magnitude of  $\tau_i$  decreased upstream  
475 because the longitudinal density distribution depends to a great extent on the salinity  
476 of the estuary. This behavior is similar to that of the salt transport observed in Díez-  
477 Minguito et al. (2013). As can be observed, salt intrusion due to the river flow (Reyes-  
478 Merlo et al., 2013) is linked to LAPE transports. Although the relation between both  
479 variables is different for each transport, certain similarities can be observed. Firstly, in  
480 normal conditions (1), the absolute transport value does not reach  $10^4 \text{ Wm}^{-1}$  and the  
481 salt intrusion remains between kilometers 63 and 70. In high-river flow conditions (2),  
482 the salt intrusion travels approximately to kilometer 12. Insofar as their absolute value,  
483 transports are maximum although the non-tidal mean advective moves seaward the same  
484 as the tidal pumping, Stokes transport, and differential advectives move upstream. LAPE  
485 transport due to longitudinal dispersion is a special case because during (2), its magni-  
486 tude increases in absolute value. Nevertheless, its dependence on the spring and neap  
487 tide cycles is important and dispersion increases its variability and maintains its mean  
488 value close to zero. During the recovery stage (3), the salt intrusion moves from the es-  
489 tuary mouth upstream and transports regain their values in (1).

490 Averaged  $\phi$ -transports,  $\tau_e$  in eq. (13) were compared with values obtained by GW2006  
491 in the Delaware estuary. In normal conditions, the mean LAPE transport in the GRE,  
492 ranged from -1 to  $1 \cdot 10^4 \text{ Wm}^{-1}$ . In fact, all individual transport modes were found to  
493 have similar magnitude  $\sim 10^4 \text{ Wm}^{-1}$ . During the high-river flow regime, it is observed  
494 that the values of individual transports do depend significantly on the flow conditions  
495 even though the order of magnitude of  $\tau_e$  hardly differs from its value in normal condi-  
496 tions. Furthermore, the order of magnitude of the mean transport coincides with the es-  
497 timates made by Garvine and Whitney (2006) for high-river flows in the Delaware. This  
498 occurs because in our model the transport associated with tidal pumping compensates  
499 the transport associated with longitudinal dispersion and differential advection. This first  
500 term was not considered in Garvine and Whitney (2006).

## 501 5 Discussion

502 This study adopted the definition of the longitudinal anomaly of the potential en-  
503 ergy in GW2006 and extended it to the zone of tidal river reaches by segmenting the es-  
504 tuary in stretches. The basic physics premises used led to the derivation of a general bal-  
505 ance equation (eq. (8)). In realistic scenarios such as this one, balances are determined  
506 by only a few terms that can be different in time and space (Díez-Minguito et al., 2013;  
507 Burchard & Hofmeister, 2008). The initial hypotheses of a weakly stratified or well-mixed  
508 estuary of limited width allowed us to simplify the LAPE balance equation in the GRE  
509 for a box and average it in a tidal cycle (eq. (10)). The model includes the major intrati-  
510 dal processes in the GRE. The non-linear terms that appeared from the equation of state  
511 for sea water were neglected because their influence under weakly stratified to well-mixed  
512 conditions is limited. In addition, adiabatically corrected temperature and density (i.e.,  
513 potential temperature and density, respectively), do not play a role in the GRE, where  
514 mean depths are about 7 m. However, these terms could be important in stratified es-  
515 tuaries. It is worth to note that the vertical mixing of density is expressed in terms of  
516 eddy diffusivity, and it is simplified with empirical mixing terms at a subtidal scale (van  
517 Aken, 1986; Simpson, 1981; Simpson & Bowers, 1981; Simpson et al., 1990). A more rig-  
518 orous definition of the mixing terms would require the analysis of high resolution spa-  
519 tial and temporal measurements that is out of the scope of this study.

520 The longitudinal variability of LAPE and of  $\phi$ -transports in other fluvial regimes  
521 is evident as reflected in the previous analyses. In the low-river flow regime (1), the time  
522 variability of the density in neap and spring tides modulated the LAPE field and its trans-  
523 ports all along the estuary. Although in our study,  $\tau$  includes mass anomaly fluxes, the  
524 results show alterations similar to those observed in mass fluxes in other estuaries (Vaz  
525 et al., 1989), where the alterations were caused by the turbulence modulation induced  
526 by tidal currents, basically in fortnightly cycles. The longitudinal variability of  $\tau$  pro-  
527 duced convergence regions that corresponded to at least one of the ETMs, more specif-  
528 ically, to the one located near  $\sim 17$  km (Díez-Minguito et al., 2014). The net estuary  
529 transport was positive in box 1 and negative in 2, 3, and 4 during spring tides, whereas  
530 the opposite occurred during neap tides. Nevertheless, the magnitude of the net trans-  
531 port decreased considerably in neap tides in the box of the estuary mouth compared to  
532 its value in spring tides. It is striking that the same did not occur in high riverflows (2),  
533 where the net transport was directed upstream. Nor did it occur in the post-riverflood  
534 recovery (3), where the transport at the estuary mouth had the same magnitude in spring  
535 tides and neap tides, which mutually compensate each other.

536 After intermediate fluvial discharges (i.e. flows that are neither low nor high) of  
537 flow  $Q \simeq 100 \text{ m}^3\text{s}^{-1}$ , there was a slight increase in the contribution of the  $\phi$ -transport  
538 associated with tidal pumping. If those pulsed discharges continued during the fortnightly  
539 semi-cycle, the magnitude of the Stokes-induced transport significantly decreased in the  
540 lower stretch of the estuary (boxes 1 to 3). This reduction in magnitude was apparently  
541 caused by the movement of the salt intrusion towards the estuary mouth (Reyes-Merlo  
542 et al., 2013). The phase lag between elevations and tidal currents, which to a great ex-  
543 tent controls the Stokes transport, was not altered by these intermediate discharges.

544 During the high-river flow episodes (2), the longitudinal profile of  $\rho$  was constant  
545 inside the estuary and approximately equal to the freshwater density. As a result, the  
546 LAPE was minimum in these conditions except at the estuary mouth where it was at  
547 its maximum. This situation mainly intensified the magnitudes of the non-tidal mean  
548 advective transport and dispersive transport, which was expected. However, it also sig-  
549 nificantly increased the LAPE transports associated with tidal pumping and differen-  
550 tial advection. For this reason, these contributions cannot be ignored in the GRE in high-  
551 river flow conditions. These results widen the vision of the GRE and extend it to other  
552 hydraulic regimes since in previous studies (Díez-Minguito et al., 2013, 2014), only trans-  
553 ports in low-river flow conditions were considered.

554 During the initial stages of the post-riverflood recovery phase (3), differential ad-  
555 vection terms and longitudinal dispersion terms were positive (upstream) near the es-  
556 tuary mouth. This may suggest that there are mechanisms that control the recovery of  
557 the LAPE in the estuary, such as the differential advection or the strain-induced peri-  
558 odic stratification (Simpson et al., 1990). However, more research is needed to confirm  
559 this possibility.

560 The influence of sediment suspension was also relevant in the post-riverflood re-  
561 covery phase. The sediment that was suspended with the discharge remained so because  
562 of the turbulence generated by the tidal currents (Losada et al., 2017). The high con-  
563 centration of suspended solids during the recovery phase was reflected in greater time  
564 fluctuations of the LAPE and its transports. The larger difference between neap and spring  
565 tides also increased the landward transport of the tidal pumping (Scully & Friedrichs,  
566 2007) in boxes 2 and 3.

567 In addition, the increase in the non-tidal mean  $\phi$ -transport in the same boxes could  
568 be promoting salt transport towards the estuary. Erosion and suspension of the bed sed-  
569 iment favored tidal suspension (Van Leussen & Dronkers, 1988). Along with this,  $\phi$ -transport  
570 due to longitudinal dispersion became significant during the recovery phase. At the in-  
571 tratidal scale, these conditions created periods during which stratification-induced tur-

bulence decreased, which indicated an increase in the capacity to retain suspended sediment (Geyer, 1993; Jay & Musiak, 1996).

As for the applicability of the model, it should be mentioned that in the theoretical development of the equation (9), the estuary is assumed to be weakly stratified. The observations show that the vertical density variability, whether in low-river or high-river flow conditions, ranges from  $-1.20$  to  $0.60 \text{ kg} \cdot \text{m}^{-3}$  (Table 1), which justifies this hypothesis. The second hypothesis is that transversal processes are negligible. This hypothesis is quantified in the Kelvin ( $K_e$ ) and Ekman ( $E_k$ ) numbers parameter space (Valle-Levinson, 2008). The channel width is related to the internal Rossby radius of deformation through the Kelvin number whereas the Ekman number establishes the relation between the viscosity and Coriolis forces. For the mean value of the vertical eddy viscosity coefficient, typical of the GRE  $\sim 3.3 \cdot 10^{-4} \text{ m}^2\text{s}^{-1}$  (Reyes-Merlo et al., 2013), values,  $K_e \sim -2$  and  $E_k \sim -2.5$ , clearly signal a vertically-sheared exchange. Near the estuary mouth, where channel widths are greater and where the values of the vertical eddy viscosity coefficient are  $\sim 1.2 \cdot 10^{-3} \text{ m}^2\text{s}^{-1}$  ( $K_e \sim -2$  and  $E_k \sim -1.25$ ), a certain degree of lateral variability is expected.

Finally, this study highlights the dependence of the LAPE on the following: (i) origin of the forcing agent; (ii) along-estuary position; (iii) neap and spring tide cycles; and (iv) hydraulic regime. A snapshot of the LAPE along the estuary provides information about the hydrodynamic and mixing conditions of the estuary and could be useful for the management and regulation of the freshwater flow in the head dam. Its use would allow estuary management at the subtidal and local scale (e.g. freshwater discharges in certain stretches of the estuary by taking advantage of the irrigation network).

## 6 Conclusions

The objective of this study was to analyze the spatiotemporal evolution of estuarine processes within the framework of a LAPE box-averaged approach. For this purpose, we developed a dynamic equation of the longitudinal anomaly of the potential energy at the subtidal scale. This equation includes the non-tidal mean advective and Stokes-induced LAPE transport, as well as empirical approximations of the tidal vertical turbulent mixing and the mixing due to wind and solar radiation. In addition, this research extends previous work by including the influence of (i) longitudinal dispersion; (ii) tidal pumping associated with cross-correlations between density, elevations, and currents; and (iii) differential advection due to depth-mean straining, and non depth-mean straining. The general balance equation was applied to the GRE, whose characteristics allowed for simplification and segmentation in five boxes. The analyzed data pertains to the salinity distribution, temperature, suspended solids, tidal elevation and currents, and fluvial discharges recorded along the estuary over a one-year period (2008-2009) in which low-river flow, high-river flow, and post-riverflood recovery conditions were found. These observations were used to evaluate the LAPE and transports throughout the GRE. The results can help to better understand the processes in a narrow and shallow water estuary under different hydraulic regimes.

The time variability of LAPE transport in low-river flows was modulated with neap and spring tides. At the estuary mouth, they were in phase with each other, and further upstream, they were out of phase. Time variability in regard to the mean value was similar in spring tides and neap tides ( $\pm 10.65\%$ ). The only exception was the Stokes-induced LAPE transport, whose magnitude significantly increased in spring tides. The LAPE transports of all processes in the GRE changed sign on various occasions at approximately kilometers 17 and 35. This gave rise to transport convergence and divergence zones, respectively. The transport convergence region was consistent with the location of the ETM in previous studies. Transports due to dispersive and differential advection moved upstream from the estuary mouth, whereas the others went seaward.

623 In high-river flows, the longitudinal profiles of the LAPE transport were inverted.  
 624 The opposite was true for tidal pumping, longitudinal dispersion and differential advec-  
 625 tion of the deviation in regard to the mean terms, which became considerably greater.  
 626 The convergence and divergence zones of the non-tidal mean advective LAPE transport  
 627 disappeared as well as those of the Stokes-induced, and differential advection transport  
 628 because of the mean terms. Tidal pumping exported the LAPE through the estuary mouth,  
 629 and was thus balanced with the transport associated with the longitudinal dispersion  
 630 and differential advection because of the deviation terms that import it.

631 During the post-riverflood recovery, all transport processes intensified except for  
 632 the Stokes-induced transport. In the vicinity of the tidal river reach, the magnitude of  
 633 all LAPE transports was no longer significant. The differential advection and longitu-  
 634 dinal dispersion terms moved upstream in the estuary mouth, which suggests that they  
 635 are the terms that recover the LAPE of the estuary. The time variability in these con-  
 636 ditions was considerable in the LAPE transport because of the longitudinal dispersion  
 637 ( $\pm 19.49\%$  of the mean value). The greater difference between neap and spring tides in-  
 638 creased the landward transport of tidal pumping in the middle stretch of the estuary.  
 639 In addition to this, there was a significant increase in the non-tidal mean LAPE trans-  
 640 port. In these conditions, the longitudinal gradient of LAPE transport was amplified,  
 641 which caused the convergence and divergence zones to reappear.

## 642 Appendix A Derivation of the LAPE Equation

643 The balance equation of the potential energy anomaly in a box (eq. (8)) is obtained  
 644 by evaluating the dynamic density equation  $\mathbf{D}(x, t)$  in one reference position  $x$  and an-  
 645 other  $x_0$  and substituting the values in the following expression:

$$g[\underline{\mathbf{D}(x_0, t)}_{\bar{z}} - \underline{\mathbf{D}(x, t)}_z] = 0, \quad (\text{A1})$$

646 where  $\bar{z}$  represents the depth average; the term,  $\mathbf{D}(x, t)$ , is defined by equation 5, (re-  
 647 peated below to clarify the derivation):

$$\mathbf{D}(x, z, t) = \partial_t \rho + \partial_x (u\rho) + \partial_z (w\rho) - \partial_x (k_x \partial_x \rho) - \partial_z (k_z \partial_z \rho) - \mathbf{S}, \quad (\text{A2})$$

648 where  $\mathbf{S}$  includes the density variation due to heat transfer by the surface layer and the  
 649 resuspension of sediment from the estuary bottom as:

$$\mathbf{S} = \frac{g\rho\alpha_w}{\rho_0 c_p} I e^{-\varepsilon(\eta-z)} + \left(1 - \frac{\rho}{\rho_s}\right) C \partial_z (w_s C), \quad (\text{A3})$$

650 where non-linear terms that appear from equation of state for sea water are neglected  
 651 (see eq. (6));  $\alpha_w$  is the water expansion coefficient;  $c_p$  is the heat capacity of the water;  
 652  $\varepsilon$  is the light absorption along the vertical profile;  $w_s$  is the sedimentation velocity of the  
 653 particulated material, assumed to be constant for a given grain size; and,  $C$  is the sus-  
 654 pended solids concentration. The parameterization of heat transfer follows the Lambert-  
 655 Beer Law where  $\varepsilon$  is the attenuation coefficient (Ruiz et al., 2013). In shallow basins, a  
 656 significant portion of solar radiation can reach the bottom where it is transformed into  
 657 heat (Wiles et al., 2006). By calculating

$$g(\underline{\mathbf{S}(x_0, t)}_{\bar{z}} - \underline{\mathbf{S}(x, t)}_z) = -\frac{\alpha_w \rho g I}{\rho_0 c_p} \left[ \left(1 - e^{-\varepsilon(\eta-z)}\right)_{\bar{z}} + \left(\varepsilon^{-1} - e^{-\varepsilon(\eta-z)} (d + \varepsilon^{-1})\right) \right] \\ + \frac{w_s}{2d} ([C^2]_{\eta} - [C^2]_{-h} + \eta[C^2]_{\eta} + h[C^2]_{-h} - d\underline{C}), \quad (\text{A4})$$

658 term  $t_G$  of eq. (8) is obtained.

659 The processes related to transversal circulation are negligible in narrow low-friction  
 660 channels (Valle-Levinson, 2008) and thus dependent on  $x$ ,  $z$ , and time  $t$ .



661 For the sake of simplicity, the methodology is applied to the terms on the left of  
 662 eq. (A2), related to time variation, advection, dispersion, and density mixing. The term  
 663 most to the left in eq. (8) is further developed in eq. (A5) by applying the chain rule and  
 664 Leibniz's rule to derivatives with variable limits and kinematic boundary conditions on  
 665 the water surface  $w_\eta = \partial_t \eta + u_\eta \cdot \partial_x \eta$ , for a fixed bed,  $w_{-h} = 0$ :

$$\begin{aligned}
 & \int_{-h}^{\eta} z [\partial_t \rho + \partial_x (u\rho) + \partial_z (w\rho) - \partial_x (k_x \partial_x \rho) - \partial_z (k_z \partial_z \rho)] dz = \\
 & \partial_t \int_{-h}^{\eta} z \rho dz + \partial_x \int_{-h}^{\eta} (zu\rho) dz - \int_{-h}^{\eta} w\rho dz - \partial_x \int_{-h}^{\eta} zk_x \partial_x \rho dz + \int_{-h}^{\eta} k_z \partial_z \rho dz \\
 & + [(z\rho)_\eta \partial_t \eta + (zu\rho)_\eta \partial_x \eta + (zw\rho)_\eta] + [(z\rho)_{-h} \partial_t h + (zu\rho)_{-h} \partial_x h + (zw\rho)_{-h}] - \\
 & (zk_x \partial_x \rho)_\eta \partial_x \eta - (zk_x \partial_x \rho)_{-h} \partial_x h + (zk_z \partial_z \rho)_\eta + (zk_z \partial_z \rho)_{-h} = \\
 & \partial_t (dz\rho) + \partial_x (dzu\rho) - (dw\rho) + \partial_x (dzk_x \partial_x u\rho) + (dk_x \partial_z \rho) \\
 & (zk_x \partial_x \rho)_\eta \partial_x \eta + (zk_x \partial_x \rho)_{-h} \partial_x h - (zk_z \partial_z \rho)_{-h} - (zk_z \partial_z \rho)_{-h}, \quad (\text{A5})
 \end{aligned}$$

666 where subindexes  $\eta$  and  $-h$  indicate that the terms are evaluated in  $z = \eta$  and  $z = -h$ .  
 667 Inserting eqs. (7) and (A5) in the eq. (A1),

$$\begin{aligned}
 & \partial_t (g(\rho_0 \underline{z} - \rho z)) + \partial_x (g(u\rho_0 \underline{z} - u\rho z)) - g(w\rho) \\
 & - g\underline{k_z \partial_z \rho} + g d^{-1} (zk_z \partial_z \rho)^\eta - g d^{-1} (zk_z \partial_z \rho)^{-h} + \partial_x (g[k_x \partial_x \rho](z - \underline{z})) \\
 & + g d^{-1} (\underline{z} - \eta) (k_x \partial_x \rho)^\eta \partial_x \eta - g d^{-1} (\underline{z} + h) (k_x \partial_x \rho)^{-h} \partial_x h. \quad (\text{A6})
 \end{aligned}$$

668 A general variable  $\zeta = \underline{\zeta} + \zeta_v$  is composed of the sum of a depth-averaged value  
 669 and its deviation. When this variable is applied in the equation, equation (8) is obtained.  
 670 As in Burchard and Hofmeister (2008) and De Boer et al. (2008), no hypothesis has been  
 671 formulated regarding the form of the longitudinal velocity profile.

## 672 Acknowledgments

673 This research was partially funded by the *Campus de Excelencia Internacional del*  
 674 *Mar (CeI-MAR)* and the Spanish Ministry of Economy and Competitiveness, Project  
 675 *PIRATES (CTM2017-89531-R)*. It was also supported by *AQUACLEW*. Project *AQUA-*  
 676 *CLEW* is part of ERA4CS, an ERA- NET initiated by JPI Climate, and funded by FOR-  
 677 *MAS (SE)*, *DLR (DE)*, *BMFW (AT)*, *IFD (DK)*, *MINECO (ES)*, *ANR (FR)* with co-  
 678 *funding by the European Commission*. We would like to thank two anonymous review-  
 679 *ers for the thoughtful contribution that has significantly improved the quality of the pa-*  
 680 *per*. Datasets for this research are available in <http://doi.org/10.5281/zenodo.3459610>.  
 681 This study is a tribute to the memory of Richard W. Garvine, whose research was the  
 682 inspiration for our work.

## 683 References

- 684 Austin, J. A. (2002). Estimating the mean ocean-bay exchange rate of the chesa-  
 685 peake bay. *Journal of Geophysical Research: Oceans*, 107(C11).  
 686 Becherer, J., Flöser, G., Umlauf, L., & Burchard, H. (2016). Estuarine circulation  
 687 versus tidal pumping: Sediment transport in a well-mixed tidal inlet. *Journal*  
 688 *of Geophysical Research: Oceans*, 121(8), 6251–6270.  
 689 Bowden, K. (1981). Turbulent mixing in estuaries. *Ocean Management*, 6(2-3), 117–  
 690 135.  
 691 Burchard, H., & Hetland, R. D. (2010). Quantifying the contributions of tidal  
 692 straining and gravitational circulation to residual circulation in periodically  
 693 stratified tidal estuaries. *Journal of Physical Oceanography*, 40(6), 1243–1262.  
 694 Burchard, H., & Hofmeister, R. (2008). A dynamic equation for the potential energy  
 695 anomaly for analysing mixing and stratification in estuaries and coastal seas.  
 696 *Estuarine, Coastal and Shelf Science*, 77(4), 679–687.

- 697 Burchard, H., Schuttelaars, H. M., & Ralston, D. K. (2018). Sediment trapping in  
698 estuaries. *Annual review of marine science*, *10*, 371–395.
- 699 De Boer, G. J., Pietrzak, J. D., & Winterwerp, J. C. (2008). Using the potential  
700 energy anomaly equation to investigate tidal straining and advection of stratifi-  
701 cation in a region of freshwater influence. *Ocean Modelling*, *22*(1-2), 1–11.
- 702 Díez-Minguito, M., Baquerizo, A., de Swart, H., & Losada, M. (2014). Structure  
703 of the turbidity field in the Guadalquivir estuary: Analysis of observations and  
704 a box model approach. *Journal of Geophysical Research: Oceans*, *119*(10),  
705 7190–7204.
- 706 Díez-Minguito, M., Baquerizo, A., Ortega-Sánchez, M., Navarro, G., & Losada,  
707 M. (2012). Tide transformation in the Guadalquivir estuary (SW Spain) and  
708 process-based zonation. *Journal of Geophysical Research: Oceans*, *117*(C3).
- 709 Díez-Minguito, M., Contreras, E., Polo, M., & Losada, M. (2013). Spatio-temporal  
710 distribution, along-channel transport, and post-riverflood recovery of salinity in  
711 the Guadalquivir estuary (SW Spain). *Journal of Geophysical Research: Oceans*,  
712 *118*(5), 2267–2278.
- 713 Ganju, N. K., Brush, M. J., Rashleigh, B., Aretxabaleta, A. L., Del Barrio,  
714 P., Grear, J. S., ... others (2016). Progress and challenges in coupled  
715 hydrodynamic-ecological estuarine modeling. *Estuaries and Coasts*, *39*(2),  
716 311–332.
- 717 Garvine, R. W., & Whitney, M. M. (2006). An estuarine box model of freshwater  
718 delivery to the coastal ocean for use in climate models. *Journal of Marine Re-  
719 search*, *64*(2), 173–194.
- 720 Geyer, W. R. (1993). The importance of suppression of turbulence by stratification  
721 on the estuarine turbidity maximum. *Estuaries*, *16*(1), 113–125.
- 722 Hamilton, P., Gunn, J. T., & Cannon, G. A. (1985). A box model of Puget Sound.  
723 *Estuarine, Coastal and Shelf Science*, *20*(6), 673–692.
- 724 Hansen, D. V., & Rattray, M. (1966). New dimensions in estuary classification. *Lim-  
725 nology and Oceanography*, *11*(3), 319–326.
- 726 Hobbie, J. E. (2000). Estuarine science: the key to progress in coastal ecological  
727 research. *Estuarine science: A synthetic approach to research and practice*, 1–  
728 11.
- 729 Hofmeister, R., Burchard, H., & Bolding, K. (2009). A three-dimensional model  
730 study on processes of stratification and de-stratification in the Limfjord. *Conti-  
731 nental Shelf Research*, *29*(11-12), 1515–1524.
- 732 Jay, D. A., & Musiak, J. D. (1994). Particle trapping in estuarine tidal flows. *Jour-  
733 nal of Geophysical Research: Oceans*, *99*(C10), 20445–20461.
- 734 Jay, D. A., & Musiak, J. D. (1996). Internal tidal asymmetry in channel flows: Ori-  
735 gins and consequences. *Coastal and Estuarine Studies*, 211–249.
- 736 Jay, D. A., & Smith, J. D. (1990). Residual circulation in shallow estuaries: 2.  
737 weakly stratified and partially mixed, narrow estuaries. *Journal of Geophysical  
738 Research: Oceans*, *95*(C1), 733–748.
- 739 Leschine, T. M., Ferriss, B. E., Bell, K. P., Bartz, K. K., MacWilliams, S., Pico, M.,  
740 & Bennett, A. K. (2003). Challenges and strategies for better use of scien-  
741 tific information in the management of coastal estuaries. *Estuaries*, *26*(4),  
742 1189–1204.
- 743 Losada, M., Díez-Minguito, M., & Reyes-Merlo, M. (2017). Tidal-fluvial interaction  
744 in the Guadalquivir river estuary: Spatial and frequency-dependent response of  
745 currents and water levels. *Journal of Geophysical Research: Oceans*, *122*(2),  
746 847–865.
- 747 Monismith, S. G., Kimmerer, W., Burau, J. R., & Stacey, M. T. (2002). Struc-  
748 ture and flow-induced variability of the subtidal salinity field in northern San  
749 Francisco Bay. *Journal of Physical Oceanography*, *32*(11), 3003–3019.
- 750 Navarro, G., Gutiérrez, F. J., Díez-Minguito, M., Losada, M. A., & Ruiz, J. (2011).  
751 Temporal and spatial variability in the Guadalquivir estuary: a challenge for

- 752 real-time telemetry. *Ocean Dynamics*, 61(6), 753–765.
- 753 Navarro, G., Ruiz, J., Cobos, M., Baquerizo, A., Díez-Minguito, M., Ortega-  
754 Sanchez, M., & Losada, M. n. (2019, September). *Atmospheric, hydrody-  
755 namic and water quality observations from environmental-quality stations,  
756 water level sensors, acoustic Doppler velocimeters, and meteorological sta-  
757 tions located at the Guadalquivir river estuary (2008 - 2010)*. Retrieved from  
758 <https://doi.org/10.5281/zenodo.3459610> doi: 10.5281/zenodo.3459610
- 759 Officer, C. B. (1980). Box models revisited. *Estuarine and wetland processes with  
760 emphasis on modeling*, 65-114.
- 761 Officer, C. B. (2013). *Physical oceanography of estuaries*. John Wiley.
- 762 Padilla, E. M., Díez-Minguito, M., Ortega-Sánchez, M., & Losada, M. A. (2016).  
763 A subtidal model of temperature for a well-mixed narrow estuary: the  
764 Guadalquivir river estuary (SW Spain). *Estuaries and Coasts*, 39(3), 605–620.
- 765 Reyes-Merlo, M. Á., Díez-Minguito, M., Ortega-Sánchez, M., Baquerizo, A., &  
766 Losada, M. Á. (2013). On the relative influence of climate forcing agents  
767 on the saline intrusion in a well-mixed estuary: Medium-term Monte Carlo  
768 predictions. *Journal of Coastal Research*, 65(sp2), 1200–1205.
- 769 Reyes-Merlo, M. Á., Ortega-Sánchez, M., Díez-Minguito, M., & Losada, M.  
770 (2017). Efficient dredging strategy in a tidal inlet based on an ener-  
771 getic approach. *Ocean and Coastal Management*, 25(3–4), 37 - 52. doi:  
772 10.1016/0378-3839(95)00008-Y
- 773 Rice, A. E., Whitney, M. M., Garvine, R. W., & Huq, P. (2008). Energetics in  
774 Delaware Bay: Comparison of two box models with observations. *Journal of  
775 Marine Research*, 66(6), 873–898.
- 776 Ruiz, J., Macías, D., Losada, M., Díez-Minguito, M., & Prieto, L. (2013). A simple  
777 biogeochemical model for estuaries with high sediment loads: Application to  
778 the Guadalquivir river (SW Iberia). *Ecological Modelling*, 265, 194–206.
- 779 Ruiz, J., Polo, M. J., Díez-Minguito, M., Navarro, G., Morris, E. P., Huertas, E., . . .  
780 Losada, M. A. (2015). The Guadalquivir estuary: a hot spot for environmen-  
781 tal and human conflicts. In *Environmental management and governance* (pp.  
782 199–232). Springer.
- 783 Scully, M. E., & Friedrichs, C. T. (2007). Sediment pumping by tidal asymmetry in  
784 a partially mixed estuary. *Journal of Geophysical Research: Oceans*, 112(C7).
- 785 Simpson, J. (1981). The shelf-sea fronts: implications of their existence and be-  
786 haviour. *Phil. Trans. R. Soc. Lond. A*, 302(1472), 531–546.
- 787 Simpson, J. (1997). Physical processes in the rofi regime. *Journal of marine sys-  
788 tems*, 12(1-4), 3–15.
- 789 Simpson, J., & Bowers, D. (1981). Models of stratification and frontal movement in  
790 shelf seas. *Deep Sea Research Part A. Oceanographic Research Papers*, 28(7),  
791 727–738.
- 792 Simpson, J., Brown, J., Matthews, J., & Allen, G. (1990). Tidal straining, density  
793 currents, and stirring in the control of estuarine stratification. *Estuaries and  
794 Coasts*, 13(2), 125–132.
- 795 Simpson, J., & Hunter, J. (1974). Fronts in the Irish Sea. *Nature*, 250(5465), 404.
- 796 Simpson, J., & Sharples, J. (2012). *Introduction to the physical and biological  
797 oceanography of shelf seas*. Cambridge University Press.
- 798 Simpson, J., Sharples, J., & Rippeth, T. (1991). A prescriptive model of stratifica-  
799 tion induced by freshwater runoff. *Estuarine, Coastal and Shelf Science*, 33(1),  
800 23–35.
- 801 Sun, Q., Whitney, M. M., Bryan, F. O., & Tseng, Y.-h. (2017). A box model for  
802 representing estuarine physical processes in earth system models. *Ocean Mod-  
803 elling*, 112, 139–153.
- 804 Valle-Levinson, A. (2008). Density-driven exchange flow in terms of the Kelvin and  
805 Ekman numbers. *Journal of Geophysical Research: Oceans*, 113(C4).

- 806 Valle-Levinson, A. (2010). *Contemporary issues in estuarine physics*. Cambridge  
807 University Press.
- 808 van Aken, H. M. (1986). The onset of seasonal stratification in shelf seas due to dif-  
809 ferential advection in the presence of a salinity gradient. *Continental Shelf Re-*  
810 *search*, 5(4), 475–485.
- 811 Van Leussen, W., & Dronkers, J. (1988). Physical processes in estuaries: An intro-  
812 duction. In *Physical processes in estuaries* (pp. 1–18). Springer.
- 813 Vaz, R. A. N., Lennon, G. W., & de Silva Samarasinghe, J. R. (1989). The negative  
814 role of turbulence in estuarine mass transport. *Estuarine, Coastal and Shelf*  
815 *Science*, 28(4), 361–377.
- 816 Wiles, P. J., van Duren, L. A., Häse, C., Larsen, J., & Simpson, J. H. (2006). Strat-  
817 ification and mixing in the limfjorden in relation to mussel culture. *Journal of*  
818 *marine systems*, 60(1-2), 129–143.
- 819 Yang, W., Wei, H., & Zhao, L. (2017). Observations of tidal straining within  
820 two different ocean environments in the east china sea: Stratification and  
821 near-bottom turbulence. *Journal of Geophysical Research: Oceans*, 122(11),  
822 8876–8893.
- 823 Zhao, Q., Ren, Y., & Wang, J. X. (2018). Temporal and spatial characteristics of  
824 potential energy anomaly in lake taihu. *Environmental Science and Pollution*  
825 *Research*, 1–10.