The Relative Importance of Wind Straining and Gravitational Forcing in Driving Exchange Flows in Tidally Energetic Estuaries

XAVIER LANGE AND HANS BURCHARD

Leibniz Institute for Baltic Sea Research Warnemünde, Rostock, Germany

(Manuscript received 5 March 2018, in final form 3 November 2018)

ABSTRACT

In straight tidal estuaries, residual overturning circulation results mainly from a competition between gravitational forcing, wind forcing, and friction. To systematically investigate this for tidally energetic estuaries, the dynamics of estuarine cross sections is analyzed in terms of the relation between gravitational forcing, wind stress, and the strength of estuarine circulation. A system-dependent basic Wedderburn number We_b is defined as the ratio between wind forcing and opposing gravitational forcing at which the estuarine circulation changes sign. An analytical steady-state solution for gravitationally and wind-driven exchange flow is constructed, where tidal mixing is parameterized by parabolic eddy viscosity. For this simple but fundamental situation, $We_b \approx 0.15$ is calculated, meaning that the up-estuary wind forcing needs to be 15% of the gravitational forcing to invert estuarine circulation. In three steps, relevant physical processes are added to this basic state: (i) tidal dynamics are resolved by a prescribed semidiurnal tide, leading to $We_b \approx 0.45$ caused by tidal straining; (ii) lateral circulation is added by introducing cross-channel bathymetry, smoothly increasing We_b from 0.47 (flat bed) to 1.3 (parabolic bed) due to an increasing effect of lateral circulation on estuarine circulation; and (iii) full dynamics of a real tidally energetic inlet with highly variable forcing, where $We_b = 0.59 \pm 0.15$ results from a two-dimensional linear regression.

1. Introduction

In tidally energetic estuaries characterized by relatively strong tidal mixing and low runoff (Geyer and MacCready 2014), the gravitational forcing is typically augmented by tidal processes such as *eddy viscosity shear covariance* (ESCO; Dijkstra et al. 2017) mainly driven by tidal straining [high turbulence during flood when stratification is weakened and vice versa during ebb, see Jay and Musiak (1994)] and *lateral advection circulation* when, during a flood, buoyancy in the center of the tidal channel is weaker than at the edges and vice versa during ebb (Lerczak and Geyer 2004). In such situations, which occur for example in macrotidal estuaries during spring tide or in the Wadden Sea of the southeastern North Sea, ESCO-driven circulation may dominate over gravitational circulation (Burchard et al. 2011; Cheng et al. 2011; Geyer and MacCready 2014).

ESCO and lateral circulation have in common that they scale with the nondimensional estuarine buoyancy gradient represented by the Simpson number (Stacey et al. 2010; Burchard et al. 2011). The buoyancy gradient varies in time and space with freshwater runoff, tides, and weather conditions, but is generally supporting classical estuarine circulation (landward near the bed), with the exception of buoyancy gradient inversions occurring at times for example in the Wadden Sea (Burchard et al. 2008; Flöser et al. 2011; Burchard and Badewien 2015) or in arid regions (Wolanski 1986; Largier et al. 1997; Burling et al. 1999; Valle-Levinson et al. 2001). Apart from buoyancy-gradient-related forcing, estuarine circulation may also be modulated by convergence [along-estuary gradient of cross-sectional area; Ianniello (1979); Burchard et al. (2014); Geyer and Ralston (2015)], which is limited to topographic features.

Furthermore, wind straining caused by the along-estuary component of the wind stress (Weisberg and Sturges 1976; Geyer 1997; Scully et al. 2005; Chen and Sanford 2009; Purkiani et al. 2016) is an important driver of estuarine circulation. In contrast to buoyancy gradient and estuarine convergence forcing, the wind straining forcing is highly variable in time. Down-estuary wind forcing

Denotes content that is immediately available upon publication as open access.

Corresponding author: Xavier Lange, xaver.lange@io-warnemuende.de

DOI: 10.1175/JPO-D-18-0014.1

© 2019 American Meteorological Society. For information regarding reuse of this content and general copyright information, consult the AMS Copyright Policy (www.ametsoc.org/PUBSReuseLicenses).

enhances buoyancy-gradient-driven estuarine circulation, and up-estuary wind forcing will weaken or invert estuarine circulation. Geyer (1997) found that the simple Hansen and Rattray (1965) model based on a constant eddy viscosity could represent observations of estuarine circulation and its wind-driven inversions in a microtidal estuary when well-tuned. Chen and Sanford (2009) used the dimensionless Wedderburn number W to quantify the balance between gravitational and wind forcing in an idealized tidal estuary. The Wedderburn number is the ratio of wind stress to gravitational acceleration caused by the estuarine buoyancy gradient, where negative values represent down-estuarine wind. In their idealized simulations, Chen and Sanford (2009) confirmed that values of $W \approx 1$ indicate cancellation of estuarine circulation because of up-estuary wind forcing. In addition to momentum, wind stress also increases turbulence in the water column, such that for down-estuary wind the stratifying straining and the destratifying entrainment compete.

By replacing the estuarine density gradient by the local density gradient, Purkiani et al. (2016) defined the local Wedderburn number (We) as the ratio of the Simpson number and the nondimensional along-estuary wind stress component averaged over an estuarine cross section. They showed that it is highly correlated with the strength of estuarine circulation in a Wadden Sea tidal channel when wind forcing is strong.

The present study has the goal to systematically investigate the role of wind straining on estuarine circulation in tidally energetic inlets. The major question is how wind stress is related to the inversion of buoyancy-driven estuarine circulation using the local Wedderburn number as the descriptive parameter. The existence and significance of a critical Wedderburn number is first motivated by a new analytical stationary solution for wind and buoyancy-driven exchange flow, based on a balance of pressure-gradient and friction including parabolic eddy viscosity and diffusivity as tidal mixing parameterization. In three regimes of increasing complexity various driving mechanisms of estuarine circulation are added to the gravitational circulation as only driver included in the analytical model: (i) adding tidal oscillations and stratification effects, (ii) adding effects of lateral circulation, and (iii) adding the full complexity of realistic transient forcing of a real tidally energetic inlet.

The paper is structured as follows: first, the basic equations and nondimensional parameters are introduced in section 2, followed by an idealized parameter space study in section 3, including the derivation of an idealized analytical solution for the residual velocity profile in estuaries (section 3a) and a comparative 1D numerical study in the same nondimensional parameter space to

obtain a basic relation for the inversion of estuarine circulation (section 3b). Afterward the effect of lateral circulation on this relation is investigated by means of a 2D cross-sectional model in section 3c. In section 4, the applicability of the theoretical findings for real tidally energetic inlets is tested for a realistic 3D model of a tidal inlet in the Wadden Sea. The results are discussed in section 5, and conclusions are drawn in section 6.

2. Theory

a. Basic equations

Along-estuary one-dimensional models are useful tools to understand fundamental processes of small-scale turbulence and mixing on estuarine circulation (Hansen and Rattray 1965; Simpson et al. 1990; Burchard and Hetland 2010; Burchard et al. 2013; Dijkstra et al. 2017). The water column equation of motion ignoring Coriolis acceleration and nonlinear terms can be formulated as

$$\partial_t u - \partial_z (A_v \partial_z u) = -z \partial_x b - \frac{1}{\rho_0} \partial_x p^{\text{ext}}(t), \quad (1)$$

where $-H \leq z \leq \eta$ denotes the upward Cartesian coordinates with constant water depth H and sea surface height η , x is the eastward coordinate pointing up-estuary, and the corresponding eastward velocity is u . In Eq. (1), $\partial_x b$ denotes a prescribed constant horizontal buoyancy gradient with the buoyancy $b = -g(\rho - \rho_0)/\rho_0$, the potential density ρ , the gravitational acceleration $g = 9.81 \text{ m s}^{-2}$, and the reference density $\rho_0 = 1000 \text{ kg m}^{-3}$. Note that a constant buoyancy gradient in a tidal regime is observed by Rippeth et al. (2001) in the Liverpool Bay area of the Irish Sea and successfully reproduced by Simpson et al. (2002) by applying a 1D model with constant gradients of salinity and temperature. The eddy viscosity A_v in the diffusion term on the left-hand side of Eq. (1) is calculated by means of a turbulence closure model to include effects of stratification and shear on vertical stresses [here we use the k - ϵ models, see e.g., Umlauf and Burchard (2005) for details]. The external pressure gradient function $\partial_x p^{\text{ext}}(t)$ represents the barotropic pressure gradient from tidal constituents and any sea surface slope in general. By calculating the depth-averaged tidal-mean velocity and substituting the pressure gradient term with the residual flow velocity u_r , it is ensured that the tidal mean transport is equal to the residual transport (see Burchard and Hetland 2010):

$$U(t) = \frac{1}{D} \int_{-H}^{\eta} u(z, t) dz = u_r + U_t \sin(\omega t), \quad (2)$$

where $D = H + \eta$ is the total water depth, $\omega = 2\pi/T$ is the tidal frequency, T is the tidal period, and U_t is the

depth-averaged tidal velocity. Boundary conditions for Eq. (1) are given by the no-slip condition at the bottom $u(-H, t) = 0$ and the prescribed momentum flux at the surface:

$$\frac{\tau_x^s}{\rho_0} = A_v \partial_z u|_{z=\eta} = u_*^s |u_*^s|, \quad (3)$$

with the wind stress τ_x^s and the surface friction velocity u_*^s . The water column approach chosen here is neglecting three-dimensional effects such as lateral circulation, Earth rotation, and longitudinal variability, and is used here only to derive some fundamental relations and to study the basic effects of tidal motion and stratification.

b. Nondimensional parameters

The dynamic equations are scaled by means of the scaling parameters ω (tidal frequency, generally the M_2 tide unless stated differently), H (reference water depth), and U_* (reference bottom friction velocity). Nondimensional variables are defined as follows:

$$\begin{aligned} \tilde{t} &= t\omega, & \tilde{z} &= \frac{z}{H}, & \tilde{u} &= \frac{u}{U_*}, \\ \tilde{u}_r &= \frac{u_r}{U_*}, & \tilde{z}_0^b &= \frac{z_0^b}{H}, & \tilde{A}_v &= \frac{A_v}{U_* H}, \end{aligned} \quad (4)$$

where z_0^b denotes the bottom roughness length. As shown by Burchard (2009), the whole set of dynamic equations can be made nondimensional with these parameters [note that Burchard (2009) used a different velocity scale], retaining the complete physics included in the turbulence closure model. For the presentation of model results, the reference bottom friction velocity $U_* = \langle (u_*^s)^2 \rangle^{1/2}$ is defined by the bottom drag coefficient C_D for the vertically integrated water column as derived from the law of the wall (see Burchard et al. 2011):

$$C_D = \left[\frac{u_*^b}{U(t)} \right]^2 = \left[\frac{\kappa}{(1 + \tilde{z}_0^b) \ln \left(\frac{1}{\tilde{z}_0^b} + 1 \right)} - 1 \right]^2, \quad (5)$$

where angular brackets denote the tidal average:

$$\langle X \rangle(t_0) = \frac{1}{T} \int_{t_0 - T/2}^{t_0 + T/2} X(t) dt. \quad (6)$$

Essential nondimensional state parameters are the following:

- Simpson number: $\text{Si} = \partial_x b H^2 / U_*^2$, giving the ratio of stratifying forces (density gradient forcing) to the destratifying forces (vertical mixing) (Stacey et al. 2010; Burchard et al. 2011),

- Unsteadiness number: $\text{Un} = \omega H / U_*$, scaled tidal frequency (Burchard and Hetland 2010; Burchard et al. 2011),
- Nondimensional wind stress: $\text{Ts} = |u_*^s| u_*^s / U_*^2$, relative strength of the wind stress to tidal stress.

The local Wedderburn number (see Purkiani et al. 2016) is then defined as

$$\text{We} = \frac{|u_*^s| u_*^s}{\partial_x b H^2} = \frac{\text{Ts}}{\text{Si}}, \quad (7)$$

which is a measure for the competition of density-driven estuarine circulation and wind straining. Note that this *local* Wedderburn number We is motivated by an estuary-wide Wedderburn number originally introduced by Thompson and Imberger (1980).

3. Idealized parameter space studies

a. Analytical 1D scenario with parabolic eddy viscosity

Decomposition of the momentum balance in Eq. (1) into a tidal mean and a tidal fluctuation part $X = \langle X \rangle + X'$ and subsequently application of Eq. (6) leads to

$$\partial_z (\langle A_v \rangle \partial_z \langle u \rangle) = z \partial_x b + \frac{1}{\rho_0} \partial_x \langle p^{\text{ext}} \rangle - \partial_z (\langle A_v' \partial_z u' \rangle), \quad (8)$$

where $\langle X' \rangle = 0$ for periodic flow and therefore $\langle A_v \partial_z u \rangle = \langle A_v \rangle \partial_z \langle u \rangle + \langle A_v' \partial_z u' \rangle$ has been used. The last term is the ESCO (Dijkstra et al. 2017) and describes the effect of tidal straining, which is known to amplify the density-driven estuarine exchange flow (see Jay and Musiak 1994). Assuming a tidally constant eddy viscosity for the moment (e.g., $A_v' = 0$), Eq. (8) can be integrated from an arbitrary position in the water column to the zero mean sea surface to

$$\partial_z \langle u \rangle = \frac{\partial_x \langle p^{\text{ext}} \rangle}{\rho_0 \langle A_v \rangle} z + \frac{\partial_x b}{2 \langle A_v \rangle} z^2 + \frac{\langle u_*^s | u_*^s | \rangle}{\langle A_v \rangle}. \quad (9)$$

The eddy viscosity A_v can be parameterized analytically in terms of the bottom friction velocity U_* and the bottom roughness length z_0^b as a depth-dependent parabolic profile (see Burchard et al. 2011):

$$A_v = -\kappa U_* (z + z_0^b + H) \left(\frac{z}{H} \right), \quad (10)$$

with the Karman-constant $\kappa = 0.41$. This formulation does not consider feedback from stratification. Note that the eddy viscosity vanishes at the surface. Inserting the eddy viscosity profile into Eq. (9) and subsequently integrating from the bottom $-H$ to position z in the water

column leads with the no-slip boundary condition to an expression that can be rewritten by eliminating the external pressure gradient term caused by calculation of the residual flow velocity defined in Eq. (2):

$$u(\tilde{z}) = \frac{\partial_x b H^2}{2\kappa U_*} \left[\frac{\ln B(\tilde{z})}{2C} - (\tilde{z} + 1) \right] + u_r \frac{\ln B(\tilde{z})}{C} - \frac{u_*^s |u_*^s|}{\kappa U_*} \frac{1}{1 + \tilde{z}_0^b} \left[\ln(-\tilde{z}) + \frac{\ln B(\tilde{z})}{C} \right], \quad (11)$$

with the integration constant

$$C = \int_{-1}^0 \ln B(\tilde{z}) d\tilde{z} = (1 + \tilde{z}_0^b) \ln \left(\frac{1}{\tilde{z}_0^b} + 1 \right) - 1 \quad (12)$$

and the dimensionless depth parameter $B(\tilde{z}) = (\tilde{z} + \tilde{z}_0^b + 1)/\tilde{z}_0^b$. A nondimensional form is obtained by scaling the solution in Eq. (11) with the bottom friction velocity scale U_* defined in Eq. (4):

$$\begin{aligned} \tilde{u}(\tilde{z}) &= \frac{\text{Si}}{2\kappa} \left[\frac{\ln B(\tilde{z})}{2C} - (\tilde{z} + 1) \right] + \tilde{u}_r \frac{\ln B(\tilde{z})}{C} \\ &\quad - \text{Ts} \frac{1}{\kappa(1 + \tilde{z}_0^b)} \left[\ln(-\tilde{z}) + \frac{\ln B(\tilde{z})}{C} \right] \\ &= \tilde{U}_g + \tilde{U}_r + \tilde{U}_w. \end{aligned} \quad (13)$$

With this, the tidal mean velocity profile comprises a density-driven component $\tilde{U}_g$, a residual flow component $\tilde{U}_r$, and a wind-driven component $\tilde{U}_w$. An equivalent formulation for constant eddy viscosity has already been derived by Hansen and Rattray (1965). For $\text{Ts} = 0$, Eq. (13) conforms to the analytical solution suggested by Burchard and Hetland (2010), where $\tilde{U}_r$ is the usual flow under the law of the wall and $\tilde{U}_g$ is the gravitational circulation caused by a horizontal density gradient. In the following the focus will be on the newly derived wind straining component $\tilde{U}_w$. The wind component shows a logarithmic behavior with strong increase near the surface and the law of the wall at the bottom. At the surface the new velocity profile goes toward infinity because of the zero eddy viscosity at $z = 0$. An eddy viscosity profile solving this problem should have a surface value of $A_v|_{z=0} = |u_*^s| \kappa z_0^s$, where z_0^s is the surface roughness length, but with this, the problem could not be solved analytically in closed form. Deviating from this behavior, the idealized Hansen and Rattray (1965) solution with vertically constant eddy viscosity has a finite value at the surface. However, because z_0^s is a very small value and the vertical integral is convergent and equals zero the solution is reasonable for further analytical studies.

Figure 1 shows an overview of the vertical structure of the nondimensional velocity components with constant values of the bottom roughness $\tilde{z}_0^b = 1.5 \times 10^{-4}$ and the residual flow velocity $\tilde{u}_r = -0.1$ for a varying Simpson number and nondimensional wind stress. It can be seen that down-estuary wind stress augments the density-driven flow while up-estuary wind will reduce the estuarine circulation or can even invert the circulation direction if Ts becomes large enough with respect to Si . Additional residual flow leads to stronger surface velocities that are harder to cancel out for counteracting wind. To quantify the strength of the estuarine circulation Burchard et al. (2011) proposed the integral of the depth-weighted scaled velocity as a dimensionless exchange flow intensity $\tilde{M}$. It is defined in a way that a steplike exchange flow around middepth $\tilde{u}(-H, \dots, -H/2) = +k$ and $\tilde{u}(-H/2, \dots, 0) = -k$ results in $\tilde{M}(\tilde{u}) = k$. We will use a modified definition to take the lateral changes in the bathymetry into account:

$$\tilde{M} = -\frac{4}{\langle A \rangle} \int_0^L \langle D(y) \rangle \int_{-1}^0 \langle \tilde{u}(y, \tilde{z}) \rangle \left(\tilde{z} + \frac{1}{2} \right) d\tilde{z} dy, \quad (14)$$

where $D(y)$ is the total water depth, the transect area $\langle A \rangle = \int_0^L \langle D(y) \rangle dy$, and the width of the transect is L . The one-dimensional case reduces to

$$\tilde{M} = -4 \int_{-1}^0 \langle \tilde{u}(\tilde{z}) \rangle \left(\tilde{z} + \frac{1}{2} \right) d\tilde{z}, \quad (15)$$

where $\tilde{M} > 0$ corresponds to classical estuarine circulation and $\tilde{M} < 0$ is an inverted circulation direction. Note that the total $\tilde{M}$ is equal to the sum of all contributions $\tilde{M}_i$ (linearity). Insertion of Eq. (13) into Eq. (15) leads to an analytical expression for $\tilde{M}$:

$$\tilde{M} = \frac{\text{Si}}{2\kappa} \left(0.5\gamma + \frac{1}{3} \right) + \tilde{u}_r \gamma - \text{Ts} \frac{1 + \gamma}{\kappa(1 + \tilde{z}_0^b)}, \quad (16)$$

with $\gamma = 2\tilde{z}_0^b - 1/C$, which is only dependent on the non-dimensional bottom roughness length. The critical condition $\tilde{M} = 0$ is then given by two expressions describing the transition of the circulation direction:

$$\text{Ts} = \text{Si} \times \text{We}_b + \alpha \tilde{u}_r \Leftrightarrow \text{We} = \text{We}_b + \frac{\alpha \tilde{u}_r}{\text{Si}}, \quad (17)$$

with $\alpha = \gamma \kappa (1 + \tilde{z}_0^b) / (1 + \gamma)$ and the basic Wedderburn number $\text{We}_b = (0.5\gamma + 1/3)(1 + \tilde{z}_0^b) / (2 + 2\gamma)$. In Eq. (17), Ts and We and represent the dimensionless wind stress and Wedderburn number, respectively, needed for the cancellation of estuarine circulation because of wind stress opposing gravitational and residual flow forcing.

FIG. 1. Composition of the nondimensional velocity $\tilde{u}$ (red) in terms of the residual flow $\tilde{u}_r = -0.1$ (black), density-driven component (blue), and wind-driven component (green) for varying Simpson number Si , nondimensional wind stress T_s , and constant nondimensional bottom roughness $\tilde{z}_0^b = 1.5 \times 10^{-4}$. The density-driven estuarine circulation is enhanced for down-estuary wind ($T_s < 0$) while up-estuary wind ($T_s > 0$) counteracts and can even invert the circulation direction.

Figure 2 shows the exchange flow intensity $\tilde{M}$ calculated from Eq. (15) with help of the stationary analytical velocity profile from Eq. (13) for varying Si , We , and T_s . In the expression within the Si – T_s parameter space (Figs. 2a,c) positive values of T_s correspond to up-estuary wind and vice versa. The isolines of $\tilde{M}$ are straight parallel lines with a positive slope and describe different scenarios of wind stress and horizontal density gradient with the same exchange flow intensity. Here $\tilde{M} > 0$ is the classical estuarine circulation that is for $T_s = 0$ purely density driven and increases with larger Simpson numbers. Down-estuary wind increases the value of $\tilde{M}$ and supports the density-driven circulation, while up-estuary wind decreases the exchange flow intensity. The bold line shows the critical condition where the up-estuary wind stress is strong enough to cancel out the density-driven circulation. For larger positive (up-estuary) values of T_s the circulation direction is inverted and the exchange flow is dominated by wind. Additional residual flow (Fig. 2c) does not change the slope of the $\tilde{M} = 0$ isoline but leads to a horizontal shift in a way that more wind is needed to invert the estuarine circulation. In the expression within the Si – We parameter space (Figs. 2b,d) the critical value for inversion of the circulation is the link between the actual density gradient and the theoretical wind stress needed for the inversion of

the estuarine circulation. This value comprises a basic Wedderburn number We_b determining the smallest wind stress needed for the inversion for a given density gradient and a variable depending on the nondimensional residual flow velocity and the Simpson number. The analytical value of the basic Wedderburn number is about $We_b \approx 0.15$ but is slightly dependent on the nondimensional bottom roughness length $\tilde{z}_0^b$. Because of the residual flow the critical Wedderburn number for inversion of estuarine circulation is dependent on the actual Simpson number and goes asymptotically against the We_b value for large Si . The slope of the $\tilde{M} = 0$ isoline in an Si – T_s diagram gives information about this value, which is equal to $1/We_b$.

b. 1D tidally periodic scenarios

The influence of a tidally varying eddy viscosity leading to an enhanced estuarine circulation is well known as tidal straining (see Simpson et al. 1990; Jay and Musiak 1994; Burchard et al. 2011). This effect as well as feedback from stratification on vertical mixing was neglected in the derivation of the analytical solution in Eq. (8) by assuming $A'_v = 0$. In the following the one-dimensional water column model GOTM [k – ϵ model with algebraic second-moment closure; see Umlauf and Burchard (2005) for details] will be used to calculate realistic velocity

FIG. 2. Strength of estuarine circulation $\tilde{M}$ calculated from an analytical solution for the velocity profile for varying Simpson number Si , nondimensional wind stress Ts , and Wedderburn number We (c), (d) with a residual flow velocity of $\tilde{u}_r = -1.0$ and (a), (b) without residual flow. Here $\tilde{M} > 0$ corresponds to classical estuarine circulation while $\tilde{M} < 0$ shows an inverted circulation direction. The bold line indicates the critical condition for the inversion of the circulation, which is without residual flow described by the basic Wedderburn number $We = We_b \approx 0.15$ in (b). Additional residual flow leads to a critical We , which is dependent on Si in (d), but does not change the slope of the critical isoline in the Si - Ts parameter, which is $1/We_b$. The dashed line marks the zero runoff condition for reference.

profiles forced by prescribed winds, tides, and horizontal buoyancy gradients. In contrast to the analytical solution an evolving stratification is now taken into account by solving a buoyancy equation leading to tidally varying eddy viscosity. The nondimensional parameters were varied over a range of $Ts = [-0.5, \dots, 0.5]$ and $Si = [-0.2, \dots, 0.5]$ in steps of 0.01, the depth-averaged tidal velocity amplitude U_t is prescribed and the friction velocity scale U_* is calculated with help of Eq. (5). With this, the surface stress and the horizontal salinity gradient are defined by the prescribed Simpson number Si^0 and the nondimensional wind stress Ts^0 :

$$\frac{\tau_x^s}{\rho_0} = Ts^0 (U_*^0)^2; \quad \partial_x S = -\frac{Si^0 (U_*^0)^2}{H^2 g \beta}, \quad (18)$$

with the salinity S and the haline contraction coefficient $\beta = 7.4 \times 10^{-4} \text{ kg g}^{-1}$. The actual tidally averaged bed stress is dependent on wind stress, horizontal density gradient, and residual flow. Because the definition of

U_*^0 in Eq. (5) assumes zero wind stress and horizontal density gradient, the prescribed parameters (superscript 0) have to be rescaled afterward with the actual tidal mean friction velocity:

$$Ts = Ts^0 \left(\frac{U_*^0}{U_*} \right)^2; \quad Si = Si^0 \left(\frac{U_*^0}{U_*} \right)^2; \quad Un = Un^0 \frac{U_*^0}{U_*}. \quad (19)$$

Afterward the dimensionless exchange flow intensity $\tilde{M}$ is calculated with help of Eq. (15). Figure 3 compares the numerical results of $\tilde{M}$ with residual flow $\tilde{u}_r = -1$ in Figs. 3c,d and without residual flow in Figs. 3a,b with a common Unsteadiness number of $Un^0 = 0.04$, which is equivalent to an M_2 tide with an velocity amplitude of $U_t = 1.0 \text{ m s}^{-1}$ in a 10-m water column.

In the expression within the Si - Ts parameter space (Figs. 3a,c) the isolines of $\tilde{M}$ are straight lines for positive values and are more nonlinear for negative values of Ts . Deviating from the analytical case the isolines are not

FIG. 3. As in Fig. 2, but the strength of the estuarine circulation $\tilde{M}$ is calculated with tidally averaged velocity profiles from a 1D numerical model with varying salinity gradient and wind stress and an Unsteadiness number of $Un = 0.04$. The slope of the critical condition (bold line) in (a) is smaller than in the analytical solution in (b) because of a larger value of We_b . Residual flow shifts the isoline of $\tilde{M} = 0$ to larger values of Ts in (c) but does not change the slope. The gray line indicates the analytical condition for $\tilde{M} = 0$ for reference.

equidistant, instead the distance increases with higher wind velocities. The slope of the zero line is smaller than in the analytical case indicating a larger value of We_b . Similar to the analytical study the residual flow leads to a horizontal shift of the isolines. In the expression within the Si - We parameter space, the zero residual flow isoline in Fig. 3b is approximately a straight vertical line like in the analytical case and has a limiting value of $We_b = 0.46$. Note that changing the underlying turbulence model by varying the steady-state Richardson number leads only to minor changes from $We_b = 0.46$ ($Ri_{st} = 0.25$) to $We_b = 0.47$ ($Ri_{st} = 0.23$) and $We_b = 0.45$ ($Ri_{st} = 0.27$), respectively. To study the dependency of the critical value for inversion of the circulation direction with respect to tidal straining, the simulations were repeated with varying Un . As can be seen in Fig. 4, the zero residual flow We_b is about 0.45 in tidally dominated cases $Un \gg 1$ and decreases with larger Unsteadiness numbers up to the analytical value of about 0.15 in the limiting case $Un \ll 1$. With this, 3 times more wind forcing is needed, in the presence of tides, to compensate the gravitational forcing. This is consistent

with the result that tidal straining amplifies the gravitational circulation and contributes to about 2/3 of the estuarine circulation (see Burchard and Hetland 2010), with 1/3 ($We_b = 0.15$) being the nontidal part and 2/3 ($We_b = 0.30$) due to the effect of tides leading to an overall value of $We_b = 0.45$.

c. Cross-sectionally resolving model

Effects of lateral circulation like enhanced flood–ebb asymmetry, as originally proposed by Lerczak and Geyer (2004), as well as the additional effect of this asymmetry on increasing the ESCO (Burchard and Schuttelaars 2012) are known for increasing density-driven estuarine circulation. To study the influence of this process on We_b , a 2D cross-estuarine simulation is carried out using the General Estuarine Transport Model (GETM; Burchard and Bolding 2002), including the same turbulence closure model as in section 3b. Comparable model setups have been applied by Burchard et al. (2011), Burchard and Schuttelaars (2012), and Schulz et al. (2015) to study dynamic effects of lateral circulation on the exchange flow. A bathymetry of the following form:

FIG. 4. Dependency of the basic Wedderburn number We_b for the inversion of the estuarine circulation on the Unsteadiness number Un . Larger values of Un correspond with smaller tidal amplitudes. For large Unsteadiness numbers the numerical value of We_b matches approximately the analytical solution without tidal straining with $We_b = 0.15$. For small values of Un the critical condition is reached at values of about $We_b \approx 0.45$. A Un of 3 corresponds to a tidal amplitude of 0.01 m s^{-1} in a 10-m water column.

$$H(y) = \frac{3\bar{H}}{3-r} \left[1 - \left(\frac{y - 0.5L}{0.5L} \right)^2 r \right], \quad (20)$$

is used, where $r = 1 - H_{\min}/H_{\max}$ is the varying ratio of minimum water depth at the banks and maximum water depth at the center of the transect with the special case of $r = 0$ denoting a flat bottom. The bathymetry in Eq. (20) is defined in a way that guarantees an equal transect mean water depth $\bar{H}$ and therefore a constant cross-section area through all simulations using $\bar{H} = 10 \text{ m}$ and a width of $L = 1000 \text{ m}$. Similar to the simulations in section 3b a constant longitudinal salinity gradient and wind stress is prescribed and feedback resulting from vertical stratification is considered by the turbulence model GOTM as described above. Additionally, the lateral buoyancy distribution is now calculated by solving the buoyancy equation in GETM. An Unsteadiness number of $Un = 0.04$ is used for all simulations. Figure 5 shows the nondimensional tidal mean velocity at the left for $Si = 0$ and $Ts = 0.25$ for three ratios $r = 0, 0.5, 1.0$.

For a flat bottom ($r = 0$), the exchange flow is laterally homogeneous and identical to the result from the dynamic 1D model (Fig. 5a). The black line in Figs. 5b, 5d, and 5f shows the condition for the transition from classical to inverted estuarine circulation ($\bar{M} = 0$) obtained by a planar fit. With this, the corresponding basic

Wedderburn number We_b is then calculated from the slope of this line. For a flat bottom, this results in $We_b = 0.47$ (Figs. 5a,b), which is consistent with findings of the dynamic 1D model. For increasing lateral bathymetry variation, the return flow is increased and more confined to the center of the channel, resulting in depth-mean down-wind flow on the shoals and up-wind flow in the channel center [Figs. 5c,e, see also Csanady (1973)]. For moderate bathymetry variation ($r = 0.5$), We_b increases to 0.74 (Fig. 5d), whereas for a fully parabolic profile, the value of 1.29 is reached (Fig. 5f). With this, for identical tidal and density gradient forcing, lateral variation increases estuarine exchange flow in such a strong way that 3 times the up-estuarine wind stress is needed to neutralize it. Figure 6 shows that We_b varies almost linearly with the bathymetry steepness parameter r .

4. Application of the theory to a tidal channel in the Wadden Sea

a. Scenario description

The theory developed in the former sections is here applied to results of a three-dimensional model of a tidal inlet in the Wadden Sea recently presented by Purkiani et al. (2015, 2016). The goal here is to determine the basic Wedderburn number We_b of that inlet and compare it to the results of the idealized simulations with reduced physics. The buoyancy is now calculated using the full 3D buoyancy equation. The tidal inlet connects the semienclosed Sylt-Rømø Bight to the southeastern North Sea. The Sylt-Rømø Bight (see Fig. 7) is part of the Wadden Sea and has an area of 410 km^2 including 190 km^2 intertidal flats. Two small rivers discharge about $4\text{--}10 \text{ m}^3 \text{ s}^{-1}$ into the bight, establishing together with a net precipitation of about 0.8 m yr^{-1} a negative salinity gradient and thus in general a positive buoyancy gradient into the Sylt-Rømø Bight (Burchard et al. 2008). Tidal currents are dominantly semidiurnal with a tidal range of about 1.8 m and tidal currents reaching up to 1 m s^{-1} in the inlet of about 30-m maximum depth and a width of 2 km .

Observations in the tidal inlet show significant tidal straining (Becherer et al. 2011), which is strongly modified by lateral circulation mainly triggered by curvature (Purkiani et al. 2015). Analysis of numerical model results for the Sylt-Rømø Bight shows a strong dependence of the estuarine circulation on wind straining (Purkiani et al. 2016).

b. Model description

The model simulations by Purkiani et al. (2016), which focus on February–March 2008, have been carefully

FIG. 5. (a),(c),(e) Nondimensional velocity and (b),(d),(f) strength of estuarine exchange flow for varying ratios of $H_{\min}$ and $H_{\max}$. The black line indicates the $\langle \tilde{u} \rangle = 0$ isoline in (a), (c), and (e) and the $\tilde{M} = 0$ isoline in (b), (d), and (f).

calibrated to observations. As turbulence closure model, the same $k-\epsilon$ model as described above is used. The horizontal resolution of the model is 100 m, and the vertical discretization consists of 25 equidistant σ layers. Meteorological forcing is calculated from output of the German Weather Service Local Model (DWD-LM) with a horizontal resolution of 7 km. Sea levels and temperature and salinity at the open boundary are generated by larger-scale outer models for the North Sea and the German Bight [see Purkiani et al. (2015) for details].

To compare model results to observations, Purkiani et al. (2015) constructed a transect across the tidal inlet through positions of two moorings. For consistency reasons, the same transect was used by Purkiani et al. (2016), although its orientation turned out to be not optimal, since the monthly mean cross-sectionally averaged cross-channel circulation amounted to about 0.05 m s^{-1} . To avoid this contamination of along-channel diagnostics by cross-channel dynamics, the orientation was changed to obtain a zero-mean cross-channel circulation to allow for a clean separation in cross-channel and along-channel velocity components and buoyancy gradients.

Moreover we evaluate the dimensionless value of $\tilde{M}$ instead of the dimensional one. The velocities, stresses,

and density gradients were projected on the transect and tidally averaged with help of a depth-weighted version of Eq. (6). Subsequently the strength of the estuarine exchange flow is calculated with help of Eq. (14) and the other nondimensional parameters according to the definitions in section 2b.

c. Results

The nondimensional forcing parameter calculated for the transect in Fig. 7 is shown in Fig. 8 for February 2008. The analytical $\tilde{M}$ (shown in dashed), calculated according to Eq. (16), underestimates the one calculated using the velocity profiles because of the missing effects as shown above, but follows the trend. For stronger up-estuary wind ($Ts > 0$) the estuarine circulation can be inverted. As shown in the analytical solution in Eq. (16) the value of $\tilde{M}$ is dependent on the nondimensional residual flow $\tilde{u}_r$, the Simpson number Si , and the nondimensional wind stress Ts and Wedderburn number We , respectively. The exchange flow parameter $\tilde{M}$ for February from the study in Purkiani et al. (2016) is shown in Fig. 9 in the Si - Ts and Si - We parameter space distinguished between positive (red) and negative (blue) values representing the classical and the

FIG. 6. Basic Wedderburn number for varying ratios $r = 1 - H_{\min}/H_{\max}$ for a parabolic bathymetry in a 2D tidally periodic slice model with an Unsteadiness number of $Un = 0.04$. Here $r = 0$ denotes a flat bottom and agrees with results of the 1D numerical study in 3b. Decreasing ratios enhances ESCO, which leads to increasing values of We_b .

inverted estuarine circulation, respectively. The separation of positive and negative values of $\tilde{M}$ in Fig. 9, the amplification of the estuarine circulation for down-estuary wind and the reversion for up-estuary wind corresponds well with the analytical and numerical results in Eq. (16). In the Si–We parameter space the isolines of $\tilde{M}$ describe hyperbolas that go asymptotically against the critical value for higher values of Si where the residual flow term in the analytical solution in Eq. (17) becomes small compared to We_b in accordance with the theoretical results in section 3. To obtain the basic Wedderburn number for the realistic scenario a least squares surface regression of the following form:

$$\tilde{M} = aTs + bSi + c, \quad (21)$$

is conducted, where (a, b, c) are fitting parameters. This is valid since the analytical equation in Eq. (16) describes a plane. With this, the basic Wedderburn number results in $We_b = -b/a$. To eliminate immoderate events, a threshold of $\tilde{M} \leq 0.4$ is used for the regression. With this, we find a value of $We_b = 0.59 \pm 0.15$, which is higher than in the 1D stationary case ($We_b = 0.15$) and the 1D tidally oscillating case ($We_b = 0.45$), but is located in the range found in the 2D tidally oscillating cross-sectional model including lateral effects ($We_b = 0.45, \dots, 1.3$).

FIG. 7. Map of the study area with the North Sea (bottom-right inset) and details of the Sylt-Rømø bight (top-left inset). The transect under investigation is shown as a red line. The local coordinate system, with the origin at the intersection of the transect with the coast of Sylt, the y axis northward along the transect, and the x axis eastward orthogonal to the transect is indicated.

5. Summary

Considering the parameter space spanned by the non-dimensional wind stress and the Simpson number, a variety of exchange profiles for tidally energetic estuaries and their decomposition is qualitatively represented by the analytical solution of steady-state equations in Eq. (13), of which the wind straining term has been newly derived here (see Fig. 1). Note that a similar decomposition could also be presented by the classical solution by Hansen and Rattray (1965) based on a vertically constant eddy viscosity.

To characterize the transition between classical and inverse estuarine circulation, Purkiani et al. (2016) have defined the critical local Wedderburn number We_c . The analytical solution for stationary conditions of estuarine circulation suggests that We_c should still depend on the residual flow velocity [see Eq. (17)], which might in reality be highly variable. We propose therefore the asymptotic value for vanishing residual flow, the basic Wedderburn number We_b , as a robust measure for critical conditions leading to cessation of estuarine circulation under up-estuarine wind forcing. Under idealized analytical conditions, this number only weakly depends on the bed roughness, and not on wind forcing and pressure gradient forcing, such that it should be a measure characterizing an estuarine system rather than a state of an estuarine

FIG. 8. Time series of (a) nondimensional wind stress, (b) nondimensional exchange flow intensity (solid line), and (c) nondimensional density gradient for the transect marked in Fig. 7. The dashed line in (b) is calculated using the analytical solution in Eq. (16).

system. For the stationary case, we obtain a basic Wedderburn number of $We_b \approx 0.15$, which means that the wind straining needs to be 15% of the gravitational forcing to neutralize estuarine circulation. When adding tidally dynamic effects such as tidal straining circulation by using a tidally resolving dynamic model, the basic Wedderburn number increases to a value of $We_b \approx 0.45$. This can be explained by the fact that now the stratifying

gravitational forces are increased because of additional processes of estuarine circulation, such that stronger wind straining (now 45%) is needed to neutralize estuarine circulation. Interestingly, this factor of 3 in increased efficiency of gravitational forcing is consistent with findings by Burchard and Hetland (2010) who showed that for such simple one-dimensional scenarios of energetic tidal flow tidal straining circulation

FIG. 9. Nondimensional exchange flow intensity $\tilde{M}$ for the transect in Fig. 7 in February 2008. The distribution of $\tilde{M}$ in the (a) Si–Ts and (b) Si–We parameter space agrees with the results in sections 3a and 3b, where positive values (red) indicate classical estuarine circulation and negative ones (blue) indicate a reversed circulation direction. The basic Wedderburn number is found by 3D regression in (a) and is marked as a dashed line with a slope of $1/We_b$ with $We_b = 0.59 \pm 0.15$. This value is the asymptotic condition for high Si in (b) where residual flow becomes unimportant.

contributes 2/3 of the total estuarine circulation. As shown in Fig. 4, the basic Wedderburn number depends on the unsteadiness of the tidal flow, as quantified by the Unsteadiness number Un : for small Un (highly unsteady flow), We_b converges toward a maximum value of $We_b \approx 0.48$, and the limit for steady flow ($Un \rightarrow \infty$, stationary conditions) is similar to the stationary analytical solution, $We_b \approx 0.15$. Small differences between this limit and the analytical solution are due to

inconsistencies in the eddy viscosity formulation between the analytical solution and the turbulence closure model. When adding lateral bathymetry variation to the tidal flow, resulting in substantially increased exchange flow (Lerczak and Geyer 2004; Burchard et al. 2011, We_b is increased from 0.45 (flat bottom) to 1.3 (parabolic bottom), that is, an up to threefold up-estuarine surface stress is required to compensate density-driven circulation (see section 3c):

	No tides + flat bottom	→	Tides + flat bottom	→	Tides + parabolic bottom
$We_b =$	0.15	...	0.45	...	1.3

When moving to a realistic three-dimensional estuarine system such as a tidal channel in the Wadden Sea, the analysis of the basic Wedderburn number becomes difficult because of the variability in the residual flow. Careful analysis of the dynamics in the tidal channel of the Wadden Sea results in $We_b \approx 0.6$ (see Fig. 9b). This value can be explained by the fact that, in addition to the influence of tides, effects of lateral circulation are supporting estuarine circulation such that stronger up-estuarine wind (60% of the gravitational forcing) is needed to neutralize estuarine circulation.

With this, the response of an estuarine exchange flow to wind forcing is controlled by the geometry of the estuary and the predominating tidal properties. The latter suggests that the importance of wind straining compared to gravitational forcing increases in nontidal estuaries. This is observed by Arneborg and Liljebladh (2009) in the Gullmar fjord and by Geyer (1997) in a shallow estuary in Cape Cod, where landward wind forcing inhibited the estuarine circulation, highlighting the importance of the alignment of the estuary in comparison to the predominating main wind direction. The importance of lateral circulation on gravitational circulation has been extensively investigated by Becherer et al. (2011) and Purkiani et al. (2015) for the tidal channel exemplarily analyzed in section 4, or in estuaries like the Hudson River (Scully et al. 2009) and Chesapeake Bay (Li and Li 2012) with stronger lateral effects decreasing the sensitivity of the estuary to wind stress. The value for the critical local Wedderburn number of 2 analyzed by Purkiani et al. (2016) turned out to be an overestimation because of a slight misalignment of the transect with respect to the tidal flow.

6. Conclusions

Wind straining is an essential process of estuarine dynamics, since in most tidally energetic estuarine systems

up-estuarine wind straining has the potential to reverse estuarine circulation, with a potentially substantial impact on transport of suspended particulate matter. The geographical orientation of an estuarine channel relative to the main wind direction does therefore strongly influence the sensitivity of the subtidal flow to wind straining. For the case of the Wadden Sea (see large map in Fig. 7), it does matter if the coastline is facing northward or westward, since the typical wind direction for storms is westerly (Duran-Matute et al. 2014). It is thus expected that tidal channels in the southern Wadden Sea with northward-facing tidal channels are less susceptible to wind-driven reversals of estuarine circulation than the tidal channel of the Sylt-Rømø Bight, which is investigated in section 4 of the present study.

As the major result of this study, the local Wedderburn number (ratio of wind straining to gravitational straining averaged over an estuarine cross section) turns out to be a good indicator for the competition between destratification resulting from up-estuarine wind forcing and stratification caused by processes related to the longitudinal buoyancy gradient. Analytical results suggest that the basic local Wedderburn number We_b (defined as the inverse of the slope of the isoline for zero estuarine circulation in a wind stress–Simpson number diagram) should be largely independent of the Simpson number (the ratio between gravitational forcing and turbulent mixing). In practice, when analyzing model results for a real tidally energetic estuary in the wind stress–Simpson number parameter space, the latter relation using the basic local Wedderburn number turned out to be a promising approach allowing us to apply a two-dimensional linear regression to calculate We_b .

Acknowledgments. This manuscript benefited from the thoughtful comments of two anonymous reviewers. The study was conducted within the framework of the

Research Training Group “Baltic TRANSCOAST” funded by the Deutsche Forschungsgemeinschaft (DFG) under Grant GRK 2000/1. This is Baltic TRANSCOAST Publication Number GRK2000/0022.

REFERENCES

- Arneborg, L., and B. Liljebladh, 2009: Overturning and dissipation caused by baroclinic tidal flow near the sill of a fjord basin. *J. Phys. Oceanogr.*, **39**, 2156–2174, <https://doi.org/10.1175/2009JPO4037.1>.
- Becherer, J., H. Burchard, G. Flüser, V. Mohrholz, and L. Umlauf, 2011: Evidence of tidal straining in well-mixed channel flow from micro-structure observations. *Geophys. Res. Lett.*, **38**, L17611, <https://doi.org/10.1029/2011GL049005>.
- Burchard, H., 2009: Combined effects of wind, tide, and horizontal density gradients on stratification in estuaries and coastal seas. *J. Phys. Oceanogr.*, **39**, 2117–2136, <https://doi.org/10.1175/2009JPO4142.1>.
- , and K. Bolding, 2002: GETM—A general estuarine transport model. European Commission Tech. Rep. EUR 20253 EN, 157 pp.
- , and R. D. Hetland, 2010: Quantifying the contributions of tidal straining and gravitational circulation to residual circulation in periodically stratified tidal estuaries. *J. Phys. Oceanogr.*, **40**, 1243–1262, <https://doi.org/10.1175/2010JPO4270.1>.
- , and H. M. Schuttelaars, 2012: Analysis of tidal straining as driver for estuarine circulation in well-mixed estuaries. *J. Phys. Oceanogr.*, **42**, 261–271, <https://doi.org/10.1175/JPO-D-11-0110.1>.
- , and T. H. Badewien, 2015: Thermohaline residual circulation of the Wadden Sea. *Ocean Dyn.*, **65**, 1717–1730, <https://doi.org/10.1007/s10236-015-0895-x>.
- , G. Flüser, J. V. Staneva, T. H. Badewien, and R. Riethmüller, 2008: Impact of density gradients on net sediment transport into the Wadden Sea. *J. Phys. Oceanogr.*, **38**, 566–589, <https://doi.org/10.1175/2007JPO3796.1>.
- , R. D. Hetland, E. Schulz, and H. M. Schuttelaars, 2011: Drivers of residual estuarine circulation in tidally energetic estuaries: Straight and irrotational channels with parabolic cross section. *J. Phys. Oceanogr.*, **41**, 548–570, <https://doi.org/10.1175/2010JPO4453.1>.
- , H. M. Schuttelaars, and W. R. Geyer, 2013: Residual sediment fluxes in weakly-to-periodically stratified estuaries and tidal inlets. *J. Phys. Oceanogr.*, **43**, 1841–1861, <https://doi.org/10.1175/JPO-D-12-0231.1>.
- , E. Schulz, and H. M. Schuttelaars, 2014: Impact of estuarine convergence on residual circulation in tidally energetic estuaries and inlets. *Geophys. Res. Lett.*, **41**, 913–919, <https://doi.org/10.1002/2013GL058494>.
- Burling, M. C., G. N. Ivey, and C. B. Pattiaratchi, 1999: Convectively driven exchange in a shallow coastal embayment. *Cont. Shelf Res.*, **19**, 1599–1616, [https://doi.org/10.1016/S0278-4343\(99\)00034-5](https://doi.org/10.1016/S0278-4343(99)00034-5).
- Chen, S.-N., and L. P. Sanford, 2009: Axial wind effects on stratification and longitudinal salt transport in an idealized, partially mixed estuary. *J. Phys. Oceanogr.*, **39**, 1905–1920, <https://doi.org/10.1175/2009JPO4016.1>.
- Cheng, P., A. Valle-Levinson, and H. E. de Swart, 2011: A numerical study of residual circulation induced by asymmetric tidal mixing in tidally dominated estuaries. *J. Geophys. Res.*, **116**, C01017, <https://doi.org/10.1029/2010JC006137>.
- Csanady, G., 1973: Wind-induced barotropic motions in long lakes. *J. Phys. Oceanogr.*, **3**, 429–438, [https://doi.org/10.1175/1520-0485\(1973\)003<0429:WIBMIL>2.0.CO;2](https://doi.org/10.1175/1520-0485(1973)003<0429:WIBMIL>2.0.CO;2).
- Dijkstra, Y. M., H. M. Schuttelaars, and H. Burchard, 2017: Generation of exchange flows in estuaries by tidal and gravitational eddy viscosity-shear covariance (ESCO). *J. Geophys. Res. Oceans*, **122**, 4217–4237, <https://doi.org/10.1002/2016JC012379>.
- Duran-Matute, M., T. Gerkema, G. J. De Boer, J. J. Nauw, and U. Gräwe, 2014: Residual circulation and freshwater transport in the Dutch Wadden Sea: A numerical modelling study. *Ocean Sci.*, **10**, 611–632, <https://doi.org/10.5194/os-10-611-2014>.
- Flöser, G., H. Burchard, and R. Riethmüller, 2011: Observational evidence for estuarine circulation in the German Wadden Sea. *Cont. Shelf Res.*, **31**, 1633–1639, <https://doi.org/10.1016/j.csr.2011.03.014>.
- Geyer, W. R., 1997: Influence of wind on dynamics and flushing of shallow estuaries. *Estuarine Coastal Shelf Sci.*, **44**, 713–722, <https://doi.org/10.1006/ecss.1996.0140>.
- , and P. MacCready, 2014: The estuarine circulation. *Annu. Rev. Fluid Mech.*, **46**, 175–197, <https://doi.org/10.1146/annurev-fluid-010313-141302>.
- , and D. K. Ralston, 2015: Estuarine frontogenesis. *J. Phys. Oceanogr.*, **45**, 546–561, <https://doi.org/10.1175/JPO-D-14-0082.1>.
- Hansen, D. V., and M. Rattray, 1965: Gravitational circulation in straits and estuaries. *J. Mar. Res.*, **23**, 104–122.
- Ianniello, J. P., 1979: Tidally induced residual currents in estuaries of variable breadth and depth. *J. Phys. Oceanogr.*, **9**, 962–974, [https://doi.org/10.1175/1520-0485\(1979\)009<0962:TIRCIE>2.0.CO;2](https://doi.org/10.1175/1520-0485(1979)009<0962:TIRCIE>2.0.CO;2).
- Jay, D. A., and J. D. Musiak, 1994: Particle trapping in estuarine tidal flows. *J. Geophys. Res.*, **99**, 20 445–20 461, <https://doi.org/10.1029/94JC00971>.
- Largier, J. L., J. T. Hollibaugh, and S. V. Smith, 1997: Seasonally hypersaline estuaries in Mediterranean-climate regions. *Estuarine Coastal Shelf Sci.*, **45**, 789–797, <https://doi.org/10.1006/ecss.1997.0279>.
- Lerczak, J. A., and W. R. Geyer, 2004: Modeling the lateral circulation in straight, stratified estuaries. *J. Phys. Oceanogr.*, **34**, 1410–1428, [https://doi.org/10.1175/1520-0485\(2004\)034<1410:MTLCIS>2.0.CO;2](https://doi.org/10.1175/1520-0485(2004)034<1410:MTLCIS>2.0.CO;2).
- Li, Y., and M. Li, 2012: Wind-driven lateral circulation in a stratified estuary and its effects on the along-channel flow. *J. Geophys. Res.*, **117**, C09005, <https://doi.org/10.1029/2011JC007829>.
- Purkiani, K., J. Becherer, G. Flüser, U. Gräwe, V. Mohrholz, H. M. Schuttelaars, and H. Burchard, 2015: Numerical analysis of stratification and de-stratification processes in a tidally energetic inlet with an ebb tidal delta. *J. Geophys. Res. Oceans*, **120**, 225–243, <https://doi.org/10.1002/2014JC010325>.
- , —, K. Klingbeil, and H. Burchard, 2016: Wind-induced variability of estuarine circulation in a tidally energetic inlet with curvature. *J. Geophys. Res. Oceans*, **121**, 3261–3277, <https://doi.org/10.1002/2015JC010945>.
- Rippeth, T. P., N. R. Fisher, and J. H. Simpson, 2001: The cycle of turbulent dissipation in the presence of tidal straining. *J. Phys. Oceanogr.*, **31**, 2458–2471, [https://doi.org/10.1175/1520-0485\(2001\)031<2458:TCOTDI>2.0.CO;2](https://doi.org/10.1175/1520-0485(2001)031<2458:TCOTDI>2.0.CO;2).
- Schulz, E., H. M. Schuttelaars, U. Gräwe, and H. Burchard, 2015: Impact of the depth-to-width ratio of periodically stratified tidal channels on the estuarine circulation. *J. Phys. Oceanogr.*, **45**, 2048–2069, <https://doi.org/10.1175/JPO-D-14-0084.1>.
- Scully, M. E., C. Friedrichs, and J. Brubaker, 2005: Control of estuarine stratification and mixing by wind-induced straining of the estuarine density field. *Estuaries*, **28**, 321–326, <https://doi.org/10.1007/BF02693915>.

- , W. R. Geyer, and J. A. Lerczak, 2009: The influence of lateral advection on the residual estuarine circulation: A numerical modeling study of the Hudson River estuary. *J. Phys. Oceanogr.*, **39**, 107–124, <https://doi.org/10.1175/2008JPO3952.1>.
- Simpson, J. H., J. Brown, J. Matthews, and G. Allen, 1990: Tidal straining, density currents, and stirring in the control of estuarine stratification. *Estuaries*, **13**, 125–132, <https://doi.org/10.2307/1351581>.
- , H. Burchard, N. R. Fisher, and T. P. Rippeth, 2002: The semi-diurnal cycle of dissipation in a ROFI: Model-measurement comparisons. *Cont. Shelf Res.*, **22**, 1615–1628, [https://doi.org/10.1016/S0278-4343\(02\)00025-0](https://doi.org/10.1016/S0278-4343(02)00025-0).
- Stacey, M. T., M. L. Brennan, J. R. Burau, and S. G. Monismith, 2010: The tidally averaged momentum balance in a partially and periodically stratified estuary. *J. Phys. Oceanogr.*, **40**, 2418–2434, <https://doi.org/10.1175/2010JPO4389.1>.
- Thompson, R., and J. Imberger, 1980: Response of a numerical model of a stratified lake to wind stress. *Second Int. Symp. on Stratified Flows*, Trondheim, Norway, Norwegian Institute of Technology, 562–570.
- Umlauf, L., and H. Burchard, 2005: Second-order turbulence models for geophysical boundary layers: A review of recent work. *Cont. Shelf Res.*, **25**, 795–827, <https://doi.org/10.1016/j.csr.2004.08.004>.
- Valle-Levinson, A., J. A. Delgado, and L. Atkinson, 2001: Reversing water exchange patterns at the entrance to a semiarid coastal lagoon. *Estuarine Coastal Shelf Sci.*, **53**, 825–838, <https://doi.org/10.1006/ecss.2000.0813>.
- Weisberg, R. H., and W. Sturges, 1976: Velocity observations in the West Passage of Narragansett Bay: A partially mixed estuary. *J. Phys. Oceanogr.*, **6**, 345–354, [https://doi.org/10.1175/1520-0485\(1976\)006<0345:VOITWP>2.0.CO;2](https://doi.org/10.1175/1520-0485(1976)006<0345:VOITWP>2.0.CO;2).
- Wolanski, E., 1986: An evaporation-driven salinity maximum zone in Australian tropical estuaries. *Estuarine Coastal Shelf Sci.*, **22**, 415–424, [https://doi.org/10.1016/0272-7714\(86\)90065-X](https://doi.org/10.1016/0272-7714(86)90065-X).