Analysis of Tidal Straining as Driver for Estuarine Circulation in Well-Mixed Estuaries

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ABSTRACT

Tidal straining, which can mathematically be described as the covariance between eddy viscosity and vertical shear of the along-channel velocity component, has been acknowledged as one of the major drivers for estuarine circulation in channelized tidally energetic estuaries. In this paper, the authors investigate the role of lateral circulation for generating this covariance. Five numerical experiments are carried out, starting with a reference scenario including the full physics and four scenarios in which specific key physical processes are neglected. These processes are longitudinal internal pressure gradient forcing, lateral internal pressure gradient forcing, lateral advection, and the neglect of temporal variation of eddy viscosity. The results for the viscosity–shear covariance are correlated across different experiments to quantify the change due to neglect of these key processes. It is found that the lateral advection of vertical shear of the along-channel velocity component and its interaction with the tidally asymmetric eddy viscosity (which is also modified by the lateral circulation) is the major driving force for estuarine circulation in well-mixed tidal estuaries.

1. Introduction

There is a long history of discussion about which processes drive residual overturning circulation in tidal estuaries. The major motivation for analyzing dynamic processes leading to residual currents is to gain a quantitative understanding of estuarine salt fluxes (MacCready 2011) and sediment fluxes (Scully and Friedrichs 2007). Initially, there was consensus that estuarine circulation is mainly driven by gravitational circulation: that is, the subtidal longitudinal pressure gradient forcing residual near-bottom flow in the landward direction and a residual seaward flow near the surface (Pritchard 1952, 1954, 1956; Hansen and Rattray 1965; Chatwin 1976). This theory could be supported by a simple balance between vertical stress divergence and horizontal pressure gradient, leading to the classical analytical exchange profiles for constant eddy viscosity (Hansen and Rattray 1965), which was later modified for wind straining (Wong 1994; Ralston et al. 2008) and parabolic eddy viscosity (McGregor 1972; Ianniello 1977; Burchard and Hetland 2010). The role of tidal asymmetries of eddy viscosity and longitudinal vertical shear in shaping estuarine circulation has first been acknowledged by Jay and Musiak (1994). They found that the tidal straining mechanism suggested by Simpson et al. (1990) would lead not only to periodically oscillating stratification but also to landward-biased near-bottom velocity during ebb and flood. Note that ebb is here defined as the part of the tidal cycle where the flow is directed toward the higher density (lower buoyancy), and flood is vice versa. More recently, Lerczak and Geyer (2004) explored by means of an idealized numerical model study the fundamental role of lateral circulation for driving estuarine circulation in channels with a parabolic cross section. This lateral circulation process, which was first described by Nunes and Simpson (1985), leads to convergent near-bottom lateral flow during flood (including downwelling in the channel center) and divergent near-bottom lateral flow during ebb (including upwelling in the channel center). It has been found that ebb lateral circulation is...
generally weaker than flood lateral circulation (Lerczak and Geyer 2004; Scully et al. 2009; Burchard et al. 2011). During flood, excess landward momentum is vertically advected to the near-bottom region in the center of the channel and from there spreads along the bottom toward the sides by the laterally divergent bottom current. During ebb, the near-bottom, laterally convergent currents transport relatively weak seaward momentum from the sides toward the center of the channel, also contributing to increased near-bottom landward momentum. The role of this lateral circulation mechanism for generating estuarine circulation has recently been confirmed in a number of numerical model studies (Cheng and Valle-Levinson 2009; Scully et al. 2009; Cheng et al. 2009; Burchard et al. 2011) and has been explained in detail by Huijts et al. (2011).

However, in most model studies of estuarine circulation the effect of eddy viscosity tidal asymmetries (e.g., the flood values being significantly larger than the ebb values) has been neglected by applying a temporally constant eddy viscosity. To investigate this role of tidal straining, Burchard and Baumert (1998) set up a two-dimensional (2D) vertical flat-bottom tidal channel model. This model included a tidally varying eddy viscosity and gravitational circulation such that the contributions of both processes to the generation of estuarine circulation could be investigated. By alternatively suppressing eddy viscosity tidal asymmetries (by neglecting the buoyancy production term in the turbulence closure scheme) or gravitational circulation (by neglecting the internal pressure gradient in the longitudinal momentum balance), they could show that the contribution of tidal straining to estuarine circulation was significantly larger than the contribution from gravitational circulation. In a highly idealized analytical model study, Cheng et al. (2010) found for well-mixed estuaries that contributions from tidal asymmetries in eddy viscosity and contributions from gravitational circulation were of about the same order of magnitude.

To analyze the quantitative contributions from tidal straining and gravitational circulation to estuarine circulation in a nonintrusive way (i.e., without changing the physical balance), Burchard and Hetland (2010) set up a one-dimensional water column model subject to an oscillating tide and a constant longitudinal density gradient and decomposed the residual flow profile into contributions from both processes. The tidal straining flow profile was calculated as function of the covariance between eddy viscosity and vertical shear and of the tidally averaged eddy viscosity profile. As a result, the so-defined tidal straining amounted to about $\frac{2}{3}$ of the estuarine circulation over a large parameter space, although covering only well-mixed estuaries. This has recently been confirmed by Cheng et al. (2011) by means of a three-dimensional study of a narrow estuary without lateral circulation.

Recently, Burchard et al. (2011) extended the residual flow profile decomposition method to tidal flow in a channel with a parabolic cross section subject to a constant longitudinal density gradient (such that only well-mixed and weakly stratified estuaries could be studied). By doing so, the lateral advective straining and longitudinal gravitational circulation contributed to estuarine circulation in a similar way [which is in agreement with the findings by Lerczak and Geyer (2004) and Scully et al. (2009)]. For weakly stratified estuaries, the largest contribution came from the covariance between eddy viscosity and longitudinal vertical shear.

For estuaries with stronger buoyancy gradient forcing, Burchard et al. (2011) found that tidal straining is partially blocking the classical estuarine circulation, because of a three-layer residual current structure with up-estuary flows near the surface and near the bed and down-estuary flows in between. Similar results were found in the modeling study by Cheng et al. (2011).

In the strongly simplified one-dimensional case studied by Burchard and Hetland (2010), the tidal straining was the result of longitudinal processes only. In the case of a cross-sectional model, lateral processes modify the tidal straining by changing both the vertical shear of the longitudinal velocity field and the vertical viscosity. The residual flow profile decomposition method, however, does not allow for further decomposition of the covariance profile. Thus, covariance of eddy viscosity and shear due to direct coupling (longitudinal processes only) and covariance due to indirect coupling (e.g., eddy viscosity and shear both depending on lateral circulation) cannot be distinguished.

It is therefore the major aim of the present study to get insight into the mechanisms leading to the dominance of the covariance between eddy viscosity and vertical shear for generating estuarine circulation in well-mixed estuaries. Examples for such tidally energetic estuaries can be found in many coastal areas, such as in the tidal gullies of the southeastern Wadden Sea, for which Flöser et al. (2011) and Becherer et al. (2011) found Simpson numbers around 0.1, values that are characteristic for the transition between well-mixed and periodically stratified conditions (Simpson et al. 1990). Other examples for such estuaries are the Conwy Estuary in North Wales (Nunes and Simpson 1985; Simpson et al. 2001) and the Chesapeake Bay tributary York River during spring tides (Simpson et al. 2005). There are many other macrotidal estuaries that show well-mixed to periodically stratified
conditions, such as the Western Scheldt between the Netherlands and Belgium or the Wash in England, but many of those have not yet been analyzed for the effects of estuarine circulation. As a model tool, the weak buoyancy forcing setup as described in Burchard et al. (2011) will be used: that is, an oscillating flow through a channel with parabolic cross section under the impact of a constant longitudinal density gradient.

2. Methods

a. Basic equations

Assuming an infinitely long straight estuary of uniform cross-sectional bathymetry, driven by a constant longitudinal buoyancy gradient \([\bar{\rho}, b]\), and a periodically varying longitudinal barotropic pressure gradient function \([P_y](t)\), constant in space, the hydrostatic dynamic equations along and across the estuary can be written as (note that diagnostic variables are set in square brackets)

\[
\partial_t u + \partial_y (uv) + \partial_z (uw) - \partial_z (A_v \partial_z u) = \int_0^z [\bar{\rho}, b] \, dz - [P_y](t) \quad \text{and} \quad (1a)
\]

\[
\partial_t v + \partial_y (uv) + \partial_z (vw) - \partial_z (A_v \partial_z v) = \int_0^z b \, dz - P_y(y, t), \quad (1b)
\]

with the longitudinal velocity component \(u(y, z, t)\), the lateral velocity component \(v(y, z, t)\), and the vertical velocity component \(w(y, z, t)\). Here, \(A_v(y, z, t)\) is the vertical eddy viscosity. The latter is calculated by means of a two-equation \((k-\epsilon)\) turbulence closure model with an algebraic second moment closure (see Umlauf and Burchard 2005 for details). Advection of turbulent quantities is neglected in these simulations because the turbulence time scale is fairly short compared to the advective time scale. The buoyancy \(b(y, z, t) = -g(\rho(y, z, t) - \rho_0) / \rho_0\) [with density \(\rho(y, z, t)\) and reference density \(\rho_0 = 1000 \text{ kg m}^{-3}\)] is calculated here by means of a linear equation of state,

\[
b = -g \beta S, \quad (2)
\]

with the haline contractivity \(\beta\) for which we choose the constant value \(\beta = 7.8 \times 10^{-4} \text{ psu}^{-1}\). The flow is assumed to be nondivergent in the \(y-z\) plane,

\[
\partial_y v + \partial_z w = 0. \quad (3)
\]

The longitudinal barotropic pressure gradient \([P_y](t)\) is calculated in such a way that the cross-sectionally averaged longitudinal velocity is harmonic at the \(M_2\) tidal frequency with the velocity amplitude \(U_r\), which will be used for the nondimensional graphical representation of velocities. The lateral barotropic pressure gradient is chosen such that a rigid-lid condition is obtained [for details, see Huijts et al. (2009) and Burchard et al. (2011)]. The salinity is calculated by means of the following budget equation:

\[
\partial_t S + u[\partial_z S] + \nu[\partial_z S] + w[\partial_z S] - \partial_z (K_e \partial_z S) = 0, \quad (4)
\]

where \(K_e\) is the eddy diffusivity. The constant in time and space longitudinal salinity gradient, \([\partial_z S]\), is prescribed in such a way that it is consistent with the prescribed buoyancy gradient \(\partial_z b\) in the sense that the equation of state (2) is fulfilled. Additionally, a small contribution from lateral diffusion of momentum and salinity is added (terms not shown for simplicity), which do not significantly contribute to modification of estuarine circulation as shown by Burchard et al. (2011).

b. Residual flow profile decomposition

After tidally averaging, vertically integrating, decomposing eddy viscosity and shear into tidal mean, and fluctuating contributions and vertically integrating again [for details, see Burchard and Hetland (2010) and Burchard et al. (2011)], (1a) is transformed into an equation for the residual velocity profile,

\[
\langle u \rangle = -\int_{-H}^{z} \frac{\langle A_v^2 \partial_z u \rangle}{\langle A_v \rangle} \, d\xi - \int_{-H}^{z} \frac{\partial_y \langle uw \rangle \, d\zeta - \langle uw \rangle}{\langle A_v \rangle} \, d\xi + \frac{\langle [P_y] \rangle}{2} \int_{-H}^{z} \frac{\xi}{\langle A_v \rangle} \, d\xi - \partial_z \langle \rho \rangle, \quad (5)
\]

where the triangular brackets denote tidal averages, resulting from a decomposition of all variables \(X\) into tidal means \(\langle X \rangle\) and tidal fluctuations \(X'\) with \(X = \langle X \rangle + X'\) and \(\langle X' \rangle = 0\). Elimination of the external pressure gradient term finally leads to the following decomposition of the tidal residual velocity profile with zero vertical mean and a no-slip condition at the bed and local runoff profiles \(u_r\) with a no-slip bottom condition,

\[
\langle u \rangle = \langle u_x' \rangle + \langle u_y' \rangle + \langle u_z' \rangle + u_r. \quad (6)
\]

In (6), all variables are functions of \((y, z)\) and the subscripts denote the processes leading to the residual flow profile components (in brackets, the major variable is given on which the process depends):
entire cross section with a given density stratification,
cubic meter) required to instantaneously homogenize the
Simpson (1981) as the amount of mechanical energy (per
here as the potential energy anomaly defined according to
The cross-sectionally averaged stratification is computed

\[ U_{t} \] the tidal frequency
\[ \bar{\rho} \] and the geometry of the tidal channel

\[ \text{Un} = \frac{\omega \bar{H}}{U_{*}}. \] (8)
Furthermore, the dynamics depends on the relative bed
roughness \( z_{0} / \bar{H} \) and the geometry of the tidal channel
(shape and aspect ratio) [for details, see Burchard
(2009), Burchard and Hetland (2010), and Burchard
et al. (2011)].
The exchange flow intensity is here quantified following
Burchard et al. (2011) as a bulk shear,

\[ \mathcal{M}(\langle \hat{u}_{i} \rangle) = -\frac{1}{W} \int_{0}^{W} \frac{4}{[H(y)]^{2}} 0 \int_{-H(y)}^{H(y)} g z \hat{\rho}(y) - \rho(y, z) \, dz \, dy. \] (10)
with the depth mean density \( \bar{\rho} \) and the cross-sectional
area \( A \).
To allow for an analysis of tidal asymmetries, we define
for the channel center the ebb, flood, and absolute
velocity profiles,
\[ u_{\text{flood}}(z) = \frac{2}{T} \int_{0}^{T/2} u(z) \, dt; \quad u_{\text{ebb}}(z) = \frac{2}{T} \int_{T/2}^{T} u(z) \, dt; \]
\[ u_{\text{abs}}(z) = \pm \frac{1}{T} \int_{0}^{T} |u(z)| \, dt, \] (11)
respectively.

3. Numerical experiments
To investigate the importance of different mechanisms
contributing to the covariance between eddy viscosity and
vertical shear, a series of numerical experiments is carried
out for a Simpson number of \( Si = 0.0911 \) and an un-
steadiness number of \( Un = 0.0343 \). This experiment has
the same specifications as the reference scenario pre-
presented by Burchard et al. (2011) (see there for details
of the model parameters). A number of simulations
with varying physical complexity is then carried out and
analyzed:

Experiment A: full physics as in (1) such that the
scenario 1 by Burchard et al. (2011) is repeated (see
their Table 1). The tidal residual longitudinal flow
and the contribution from the covariance between
vertical longitudinal shear and eddy viscosity are
shown in Fig. 1.
Experiment B: neglect of the internal pressure
gradient in the longitudinal momentum budget,
leading to a shutoff of the gravitational circulation
\( \langle \hat{u}_{i} \rangle = 0 \). It should be noted that for this experiment
the longitudinal salinity gradient is nonzero in the
salinity equation, such that tidal straining and lateral circulation are still present.

Experiment C: neglect of lateral and vertical momentum advection terms to suppress momentum transfer from secondary circulation to longitudinal circulation, leading to \( \langle \mu'_y \rangle = 0 \). Note that, with the neglect of the lateral momentum advection, lateral salinity advection is still present such that processes as restratification due to lateral straining during slack tides are still included.

Experiment D: neglect of the internal pressure gradient in the lateral momentum budget, leading to a shutoff of the secondary circulation and to \( \langle \mu'_y \rangle = 0 \). Here, lateral salinity advection is shut off such that the results for each water column are equivalent to the one-dimensional simulations presented by Burchard and Hetland (2010).

Experiment E: full momentum budget as in (1), but with temporally constant and spatially varying eddy viscosity \( \langle A_\psi \rangle \), as computed using the scenario A. This will be leading to vanishing covariance between eddy viscosity and vertical longitudinal shear, \( \langle \mu'_y \rangle = 0 \).

To quantitatively assess the importance of different mechanisms leading to changes in the straining contribution to estuarine circulation between various numerical experiments, the straining term for the reference experiment is decomposed into four terms, Eq. (12)

\[
\langle [A'_i]_A \partial_z [u']_A \rangle = \langle [A'_i]_Y \partial_z [u']_Y \rangle + \langle [A'_i]_Y \partial_z [u']_A-Y \rangle + \langle [A'_i]_{A-Y} \partial_z [u']_Y \rangle + \langle [A'_i]_{A-Y} \partial_z [u']_{A-Y} \rangle
\]

with \( [A'_i]_A = [A'_i]_Y + [A'_i]_{A-Y} \) and \( [u'_i]_A = [u'_i]_Y + [u'_i]_{A-Y} \), where \([ ]_A\) denotes deviations from tidal means for the experiment A and \([ ]_{A-Y}\) denotes differences in the deviations from the mean between the experiments A and Y. Here, A denotes the reference experiment A and Y denotes an experiment without lateral advection (either C or D). Hence, the first term on the right-hand side of (12) approximates the contribution from the
covariance between shear and eddy viscosity in the reduced physics experiment Y. The second term on the right-hand side of (12) calculates the effects of shear deviation between experiments A and Y, the third term calculates the effects of the changed eddy viscosity deviation, and the fourth term gives the effects of correlated changes due to both deviations. When applying the decomposition (12) to (5) and carrying out the double integration of the tidally averaged momentum equations as explained in detail (Burchard et al. 2011), the residual circulation caused by the eddy viscosity–shear covariance circulation is decomposed into four contributions,

$$\langle u_c^e \rangle_{A,A} = \langle u_c^e \rangle_{Y,Y} + \langle u_c^e \rangle_{Y,A-Y} + \langle u_c^e \rangle_{A-Y,Y} + \langle u_c^e \rangle_{A-Y,A-Y},$$

(13)

where the sequence of terms corresponds to (12). The terms in (13) can easily be quantified by means of (9),

$$M(\langle u_c^e \rangle_{A,A}) = M(\langle u_c^e \rangle_{Y,Y}) + M(\langle u_c^e \rangle_{Y,A-Y}) + M(\langle u_c^e \rangle_{A-Y,Y}) + M(\langle u_c^e \rangle_{A-Y,A-Y}).$$

(14)

Note that the resulting velocity field $\langle u_c^e \rangle_{Y,Y}$ will differ from the original $\langle u_c^e \rangle$ calculated for experiment Y, because now the tidal mean eddy viscosity $\langle A_v \rangle$ from experiment A is used for the calculation shown in (5). This, however, does not pose a problem, because here it is the contribution of the reduced physical dynamics to the complete physical dynamics (including the relevant tidal mean eddy viscosity) that is quantified rather than that the reduced physical dynamics is quantified itself. Table 1 shows the intensity of exchange profiles resulting from experiments A–E.

### 4. Results

For the full physics experiment A, the results for the estuarine circulation intensity of Burchard et al. (2011) are closely reproduced [see Table 1; note that we here apply a higher vertical resolution as in Burchard et al. (2011), such that minor deviations occur]. For this experiment, the viscosity–shear covariance is the dominant process, explaining 67% of the residual exchange flow.

When the longitudinal internal pressure gradient (i.e., the gravitational circulation) is neglected (see experiment B), the flow situation is only slightly changed: estuarine circulation is reduced by 15%, mainly because of the vanishing contribution from gravitational circulation, whereas the viscosity-shear covariance and the lateral advection contribution are almost unchanged. The relative changes in cross-sectionally averaged eddy viscosity (Fig. 2a) and stratification (Fig. 2b) are small. This is in agreement with the findings of Burchard and Baumert (1998), who obtained weakened but still significant estuarine circulation in their estuarine model experiment also after neglecting gravitational circulation.

When, however, neglecting lateral momentum advection (experiment C), the residual exchange flow is reduced by 60% compared to experiment A. This reduction is largely due to a decrease of the covariance contribution by more than 50%, although it still explains 72% of the exchange flow with gravitational circulation, explaining most of the remaining residual (note that the advection contribution has vanished). The stratification is significantly decreased during late flood and early ebb (0.4 ≤ $t/T$ ≤ 0.8), which is consistent with an increase of the cross-sectionally averaged eddy viscosity (see Fig. 2a). Consequently, the tidally averaged eddy viscosity $\langle A_v \rangle$ is increased as well. This latter change is remarkable, because lateral advection of stratification is still enabled and the changes in stratification are only due to neglect of lateral advection of longitudinal momentum.

The lateral circulation is completely shut off by neglecting the lateral internal pressure gradient (experiment D). The consequence of this is a further reduction of the residual exchange flow (now being reduced by 66%, compared to experiment A), with the covariance now contributing 74% to the residual circulation. In this case, the remaining residual is still explained by the gravitational circulation, but its absolute contribution is reduced because of the increased mean tidal eddy viscosity $\langle A_v \rangle$ (Fig. 2a) caused by strongly reduced mean stratification (Fig. 2b) due to neglect of lateral circulation.
The situation completely changes in experiment E, when applying the tidally averaged (spatially varying) eddy viscosity calculated from experiment A [see Fig. 5b of Burchard et al. (2011) for a spatially resolved field of \( \langle A_y \rangle \) for this experiment]. The covariance becomes zero, but the gravitational circulation is unchanged with respect to experiment A (because it only depends on \( \partial_y \) and \( \langle A_y \rangle \)). Compared to experiment A, the advective contribution is significantly reduced because during slack tide no collapse of turbulence with subsequently enhanced lateral exchange flow occurs. As a consequence, in experiment E, the residual circulation is reduced by almost 80% with respect to experiment A.

What can be derived from the above results for the mechanisms leading to the covariance between eddy viscosity and shear for driving estuarine circulation? For experiment B, where the gravitational circulation had been switched off, Table 2 shows that longitudinal straining is dominant and that contributions from shear change and viscosity change cancel out each other.

The difference between experiments A and C shows that lateral advection of longitudinal momentum plays an important role in building the viscosity–shear covariance, because its shutoff leads to a substantial reduction of the covariance. The lateral advection of momentum could influence the covariance of eddy viscosity and shear in two ways: by increasing the ebb–flood shear asymmetry or by increasing the ebb–flood eddy viscosity asymmetry.

Analyzing the covariance between viscosity and shear across two different experiments according to (13) helps to understand the underlying dynamics. Figure 3 shows for experiment A the four contributions to the residual circulation resulting from the viscosity–shear covariance, when using the results of experiment C to decompose this covariance. These contributions are (i) the viscosity–shear covariance from experiment C (i.e., without lateral momentum advection) \( \langle u'_x \rangle_{C,C} \), (ii) the effect of the shear change due to lateral momentum advection and the eddy viscosity obtained from experiment C \( \langle u'_x \rangle_{C,A-C} \), (iii) the effect of the viscosity change due to lateral momentum advection and the shear from experiment C \( \langle u'_x \rangle_{A-C,C} \), and (iv) the covariance of both changes \( \langle u'_x \rangle_{A-C,A-C} \). In experiment C, the lateral advection of momentum is neglected, whereas the lateral advection of salinity is accounted for. Earlier analytical model results showed that, under appropriate conditions (see Huijts et al. 2011), the contributions of advective terms could be well captured using an asymptotic series expansion. Hence, it is expected that the difference in salinity field between experiments A and C is much smaller than the differences in the vertical shear. This is supported by the results in Table 2. The residual circulation contribution resulting from the shear change \( \langle u'_x \rangle_{C,A-C} \) is of similar magnitude as the viscosity–shear covariance from longitudinal straining \( \langle u'_x \rangle_{C,C} \), whereas the contribution from the viscosity change is small and negative and the contribution

**TABLE 2.** Decomposition of the residual flow contribution from the covariance between eddy viscosity and vertical shear as calculated from the difference between the reference experiment and experiments B, C, and D. The numbers show the exchange flow intensity for each of the components. For the calculation, the tidal mean eddy viscosity field \( \langle A_y \rangle \) from experiment A has been used.

<table>
<thead>
<tr>
<th>Expt</th>
<th>( M((u'<em>x)</em>{A-A}) )</th>
<th>( M((u'<em>x)</em>{Y-Y}) )</th>
<th>( M((u'<em>x)</em>{Y-A,A}) )</th>
<th>( M((u'<em>x)</em>{A-Y,Y}) )</th>
<th>( M((u'<em>x)</em>{A-Y,A}) )</th>
<th>( M((u'<em>x)</em>{A-Y,A}) )</th>
</tr>
</thead>
<tbody>
<tr>
<td>( Y = B )</td>
<td>+0.0272</td>
<td>+0.0264</td>
<td>+0.0051</td>
<td>-0.0043</td>
<td>-0.0009</td>
<td>+0.0001</td>
</tr>
<tr>
<td>( Y = C )</td>
<td>+0.0272</td>
<td>+0.0132</td>
<td>+0.0174</td>
<td>-0.0036</td>
<td>+0.0001</td>
<td>+0.0001</td>
</tr>
<tr>
<td>( Y = D )</td>
<td>+0.0272</td>
<td>+0.0154</td>
<td>+0.0086</td>
<td>-0.0056</td>
<td>+0.0088</td>
<td>+0.0088</td>
</tr>
</tbody>
</table>
from the combined change is negligible. The fact that
the latter two contributions are small can be understood
by the fact that the difference in vertically averaged
eddy viscosity is relatively small between experiments A
and C (see Fig. 2a).

In Figs. 4a,d, ebb- and flood-averaged velocity profiles
are given, in comparison to (symmetric) tidally averaged
profiles of the absolute value of the longitudinal velocity
profiles for experiments A and C, respectively. The same
is shown in Figs. 4b,e for the vertical shear of the velocity
profiles. The difference between the flood and ebb tide-
averaged shear profiles gives the tidally averaged shear amplitude profiles. In Figs. 4c,f, the flood- and ebb-
averaged eddy viscosity profiles are shown. The difference
in shear profiles between experiments A and C was far
greater than the difference in eddy viscosity. Therefore,
for weak stratification indeed the lateral advection of
vertical shear asymmetry seems to be the major process
in generating the covariance between eddy viscosity and
vertical shear.

When analyzing the differences between experiments
A and D, the additional effect of advection of stratifi-
cation is excluded, an effect that could lead to additional
changes in tidal asymmetry of eddy viscosity. Table 1
suggests that the differences between the experiments C
and D are relatively small. However, Table 2 shows the
influence of salinity advection on the composition of the
exchange flow. The contribution from the shear change
is still large (but now smaller than the longitudinal
straining \( \langle u_c^e \rangle_{D,D} \)), but now also the combined shear and
viscosity change matters significantly. The salinity field
between A and D has changed significantly because of
the exclusion of the lateral circulation in experiment D.
This change in salinity field has changed the vertical
viscosity in such a way that the difference between ex-
periments A and D now results in a large contribution to
the covariance. The results of experiment C and D are
consistent: between experiments A and C, the difference
in viscosity is relatively small; hence, the \( \langle u_c^e \rangle_{A-D} + \langle u_c^v \rangle_{A-D} \) contribution in Table 2 of experiment D
should be (almost) equal to the \( \langle u_c^e \rangle_{C,A-C} \) contribution in
experiment C. That this is the case can be seen in Fig. 5,
where the spatial distribution of the contributions to
the straining circulation is shown for experiment D.
Figures 4g–i confirms for the central water column that
between experiments A and D indeed both shear and
eddy viscosity asymmetry are reduced, meaning that
the lateral advection of stratification is also significantly
affecting the covariance between eddy viscosity and lon-
gitudinal shear.
5. Conclusions

This present idealized modeling study of channelized estuarine flow analyzes the role the tidal straining (defined as covariance between eddy viscosity and vertical shear) as a major process for driving estuarine circulation in tidal estuaries. The new result is that, for scenarios with relatively weak buoyancy gradient forcing, the classical longitudinal tidal straining as suggested by Simpson et al. (1990) and Jay and Musiak (1994) is only a minor part of total tidal straining. The major part is provided by the tidal asymmetry in lateral advection of along-channel vertical shear and its interaction with the tidal asymmetry in eddy viscosity.

We have investigated the composition of the total tidal straining by means of successively switching off relevant physical processes and quantifying (i) the resulting tidal straining, (ii) the changes in ebb and flood profiles of shear and eddy viscosity in the channel center, and (iii) the covariance of shear and eddy viscosity across numerical experiments with full and with reduced physics. The essential role of lateral advection for generating tidal straining is demonstrated by switching off lateral momentum advection (experiment C). Further switching off...
also lateral salinity advection has a minor effect on the strength of the tidal straining. Analysis of the profiles in the channel center suggests that lateral momentum advection mainly enhances the tidal shear asymmetry and has little influence on the tidal eddy viscosity asymmetry caused by direct longitudinal straining and modified by salinity advection.

A quantification of the processes contributing to tidal straining could only be achieved by calculating covariances of shear and eddy viscosity across different experiments. The covariance analysis across the reference experiment A and the reduced physics experiment C shows that lateral momentum advection enhances tidal straining substantially by increasing the tidal shear asymmetry. Analyzing covariances across experiments A and D (no lateral advection of momentum and salt) shows that, in addition to purely longitudinal straining (experiment D), the total tidal straining is predominantly composed of additional straining from increased tidal shear asymmetry (due to lateral momentum advection) and from the interplay of increased tidal shear and tidal eddy viscosity asymmetry (from lateral salinity advection).

The present study has been carried out for one particular combination of Simpson number, unsteadiness number, and aspect ratio $H/W$. As shown by Burchard et al. (2011), the composition of the processes driving estuarine circulation in such channelized well-mixed to periodically stratified estuarine flows is depending on these parameters. The present reference case, however, does well represent the basic dynamics of this class of flow. With varying aspect ratio, it would be expected that the effects of lateral circulation on tidal straining would play a decreasing role for the limits of very high and very small aspects ratios, because lateral circulation would decrease for these situations.

The contribution of lateral advection to the covariance of eddy viscosity and longitudinal shear and thus its indirect contribution to residual circulation must not be confused with the direct contribution of lateral advection to residual estuarine circulation. The latter process directly deposits longitudinal shear in support of estuarine circulation during flood and during ebb, as already discussed by Lerczak and Geyer (2004) and Burchard et al. (2011).

Acknowledgments. This paper is an answer to the question by Malcolm Scully (Norfolk, Virginia) posed at the 15th Biennial Conference on Physics of Estuaries and Coastal Seas in Sri Lanka (14–17 September 2010), to what degree lateral circulation would contribute to the covariance between eddy viscosity and vertical shear. Parts of the study have been carried out in the framework of the Role of Estuarine Circulation for Transport of Suspended Particulate Matter in the Wadden Sea.
REFERENCES


