



Asymmetric current reversals and 3D inter-channel connectivity in a deep multichannel and macrotidal estuary

M. Le Mouel^{a,b}, O. Lelandais^b, A. Hammouti^{a,b}, D. Pham Van Bang^{a,b,*} 

^a Department of Construction Engineering, Ecole de Technologie Supérieure (ÉTS), Université du Québec, 1100 rue Notre-Dame Ouest, Montréal, QC, H3C 1K3, Canada

^b Environmental Hydraulic Laboratory, Institut National de la Recherche Scientifique (INRS), 490 rue de la Couronne, Québec, QC, Canada

ARTICLE INFO

Keywords:

Tidal reversals
Dynamic connectivity
Inter-channel exchanges
3D hydrodynamics
Macrotidal estuary
Shoal-channel interactions
Lateral circulation
Tidal asymmetry

ABSTRACT

Multichannel estuaries exhibit complex tidal circulation, yet the mechanisms controlling tidal asymmetry and hydraulic connectivity in deep macrotidal systems remain insufficiently quantified. In particular, the transient dynamics governing tidal reversals and short-lived lateral exchanges between adjacent channels are poorly documented.

This study analyzes hydrodynamic connectivity in the multichannel St. Lawrence Estuary using a high-resolution three-dimensional barotropic numerical model resolving intra-tidal variability. The model captures longitudinal and lateral velocities, reversal-front propagation, and momentum inter-channel redistribution over the separating shoal.

Results show strong tidal asymmetry and pronounced inter-channel phase lags. Maximum longitudinal velocities reach 1.8–2.2 m s⁻¹ during flood and 1.5–1.9 m s⁻¹ during ebb. Lateral velocities over the shoal intensify during reversals, frequently exceeding 0.6 m s⁻¹, with peak lateral discharge between C1 and C2 reaching 6000–8900 m³ s⁻¹. Reversal fronts propagate at 25–39 km h⁻¹, with asynchronous transitions of 10–30 min between channels. These short phases generate transient shear zones, counter flows, and localized momentum redistribution over the shoal.

These findings demonstrate that tidal asymmetry and inter-channel connectivity are governed by transient three-dimensional processes driven by reversal asynchrony and intensified lateral momentum transfer. They provide a quantitative basis for future investigations of sediment pathways, salinity intrusion, and tidal-network evolution.

1. Introduction

Multichannel estuaries exhibit intrinsically three-dimensional flow structures arising from the interaction between tidal propagation, channelized bathymetry, and lateral connectivity. Rather than behaving as uniform longitudinal systems, these environments develop strong spatial heterogeneity, with sharp longitudinal gradients and laterally confined jets structured by deep channels, shoals, and archipelagos (Chu et al., 2022; Wang et al., 2021). As a result, the cross section is dynamically partitioned into lateral zones with different velocity, reflecting a transient hierarchy in which channels alternately behave as hydraulically connected or partially isolated pathways over the tidal cycle (Hlaing et al., 2024). In multichannel estuaries, this connectivity can occur through lateral channels (Hiatt et al., 2020), tidal junctions

(Le Mouel et al., 2025), secondary branches (Gundlach et al., 2021), or shoals (Zhou et al., 2025), each providing distinct pathways for momentum and mass exchange. This organization highlights that estuarine circulation emerges from the coupling between longitudinal, lateral, and vertical processes rather than from longitudinal forcing alone.

At the network scale, flow partitioning between channels is governed by continuity constraints at junctions and shoals, where water-level gradients and mass conservation control discharge distribution (Wang et al., 2021). Channel geometry (including width, depth, and curvature), influences this partitioning, producing time-dependent flow regimes (Hlaing et al., 2024). In some systems, the branch-to-main discharge ratio varies significantly over the tidal cycle, underscoring the role of internal hydrodynamic feedbacks. However, estuaries with similar morphologies may exhibit markedly different circulation

* Corresponding author. Department of Construction Engineering, Ecole de Technologie Supérieure (ÉTS), Université du Québec, 1100 rue Notre-Dame Ouest, Montréal, QC, H3C 1K3, Canada.

E-mail address: damien.pham-van-bang@etsmtl.ca (D. Pham Van Bang).

<https://doi.org/10.1016/j.ecss.2026.110261>

Received 6 May 2026; Received in revised form 14 September 2026; Accepted 15 September 2026

Available online 16 September 2026

0272-7714/© 2026 The Authors. Published by Elsevier Ltd. This is an open access article under the CC BY-NC-ND license (<http://creativecommons.org/licenses/by-nc-nd/4.0/>).

patterns, revealing a mismatch between planform organization and hydrodynamic behavior (Hiatt et al., 2020, 2022; Hlaing et al., 2024). This mismatch arises because momentum redistribution is governed by inherently three-dimensional and time-dependent processes that cannot be captured by depth-averaged approaches.

Vertical shear, bathymetric gradients, and lateral exchanges interact to produce complex flow structures. Sharp transitions between deep channels and adjacent shoals generate strong shear layers, vorticity production, and recirculation structures (He et al., 2023; Le Mouel et al., 2025). Bathymetric obstacles such as islands and headlands further enhance these processes through vortex shedding and secondary circulation.

In channel–shoal systems, exchange flow is typically confined to deep channels, as shallow shoals dissipate shear and cannot sustain residual circulation (Geyer et al., 2020). Nevertheless, shoals adjust exchange indirectly by imposing lateral density gradients, storing and releasing water masses, and altering tidal straining. In stratified systems, during neap tides, exchange is density-driven, whereas during spring tides tidal dispersion prevails. Shoals further regulate residual circulation through lateral storage and release (Geyer et al., 2020; Zhou et al., 2020). Shallow areas act as transient reservoirs that reorganize transport pathways (Chu et al., 2022).

A defining feature of multichannel estuaries is the strong temporal variability of inter-channel exchanges. Hydrodynamic connectivity evolves continuously over the tidal cycle and exhibits pronounced asymmetries between flood and ebb phases (Sandbach et al., 2018; Van Veen et al., 2005; Zhou et al., 2025). Individual channels may display distinct tidal dominance, and counter flows can develop in secondary branches while the main channel continues to flow seaward (Hlaing et al., 2024). Lateral circulation reinforces these asymmetries through velocity–stratification–Coriolis coupling (Lerczak and Geyer, 2004), while bathymetric transitions generate shear zones and convergence fronts that induce multiple lateral reversals (Collignon and Stacey, 2012). Even in well-mixed macrotidal systems, secondary circulation enhances vertical shear and turbulence on the surface (Ross et al., 2019). Shoal–channel system can produce opposite tidal dominance within the same cross section, with shoals often flood-dominant and deep channels ebb-dominant (Ralston et al., 2012; Sandbach et al., 2018; Zhou et al., 2025). However, these asymmetries do not capture the short-lived but dynamically dominant phases that structure inter-channel connectivity.

Despite this growing body of work, most studies focus on tidally averaged or persistent flood- or ebb-dominant behavior (Sandbach et al., 2018; Van Veen et al., 2005; Zhou et al., 2025), providing limited insight into the transient dynamics that occur during tidal transitions. Yet these short-lived phases are dynamically critical: near slack water, weakened longitudinal velocities intensify lateral pressure gradients, triggering asynchronous reversals and momentary flow opposition between adjacent branches. These conditions generate intense shear zones, recirculation cells, and localized convergence fronts that do not persist over the full tidal cycle. Although such behaviors have been reported in junctions (Bilal et al., 2020), they remain poorly quantified in shoals, and their three-dimensional structure is still largely unresolved. This gap is particularly significant because inter-channel exchanges and momentum redistribution peak during these transitional phases, when the system departs most strongly from its tidally averaged state.

The St. Lawrence Estuary provides an ideal setting to investigate these processes. Its hydrodynamics are structured by multiple deep tidal channels separated by extensive shoals and interconnected through complex junctions, including the Île d'Orléans and the Île-aux-Grues archipelago. This configuration generates strong bathymetric gradients and multiple interaction zones, producing complex coupling between longitudinal gradients, vertical shear, and lateral exchanges under strong tidal forcing. These characteristics make it well suited to examining how tidal asymmetry and hydrodynamic connectivity emerge in a deep macrotidal environment.

This study provides the first process-based characterization of

hydrodynamic connectivity in the multichannel St. Lawrence Estuary, resolving how tidal reversals, lateral shear, and cross-channel exchanges emerge and propagate within a deep macrotidal system. Using a high-resolution 3D hydrodynamic model (validated in Le Mouel et al., 2025), the intra-tidal variability of inter-channel exchanges is quantified during both fully developed flood and ebb conditions and transient flow-reversal periods. The objectives are to analyze (i) the propagation of tidal reversals and phase lags across channels, (ii) the emergence of lateral shear, recirculation, and cross-channel pressure gradients during transitional phases, and (iii) the role of shoals and channel junctions in controlling momentum redistribution. The overall hydrodynamics of the study area are first described, followed by a focused analysis of ebb-to-flood and flood-to-ebb transitions. The three-dimensional inter-channel connectivity is then examined, and the mechanisms responsible for this connectivity are subsequently identified.

2. Methodology

2.1. Site configuration

The study area is located at the fluvio-intermediate St. Lawrence estuary (Quebec, Canada). The boundary between the two zones is at the seaward tip of Île d'Orléans (IO), the main asymmetric tidal junction of the estuary (Fig. 1a). Immediately upstream of the island, a local depression (D1) reaches 55 m, with the navigable channel (C1) following the IO southern branch before rejoining the main upstream channel.

Downstream, flow splits into three channels: the deep, narrow C1 (20–80 m); intermediate C2, fragmented by the 21-island Île-aux-Grues (IG) archipelago; and the broader, shallower southern C3 (~20 m). This section shows highly variable bathymetry, from intertidal shallows to depths > 60 m (near IC). Lateral connectivity is controlled by junctions (J1, J2), shoals, and lateral channels; sandy shoals are particularly notable between C1 and C2 (Fig. 1b).

River discharges are regulated by the International St. Lawrence River Control Board, with mean flow at Québec ~12,200 m³ s⁻¹ (doubling during spring flood) (Matte et al., 2017b). The study site includes the salt intrusion limit and maximum turbidity zone, both located downstream of IO. Saline water intrudes preferentially into C1 under well-mixed conditions (Dionne, 2002; Simons et al., 2010).

The estuary is tide-dominated with a mixed diurnal-semi-diurnal regime. Maximum amplification occurs at IC (>7 m, spring tides) remaining 3–6 m at Québec (Matte et al., 2019). A tidal range of 7.17 m was even recorded at Québec during the Elliott storm on 23 December 2022 (Abair et al., 2025).

Tide propagation induces rapid water-level changes (>1 m h⁻¹), strong currents (up to 5 times of the mean flow), and frequent reversals under low discharge (Matte et al., 2014b). River–tide interactions and spring–neap variability cause temporal tidal changes, while channel convergence and bed friction distort the tidal wave, producing overtides and flow asymmetry. Tidal damping between IC and Québec remains moderate, with flow largely driven by the incident tide (Matte et al., 2014b). Maximum velocities (>2 m s⁻¹) occur in constrictions such as Québec Bridge, upstream IO, and shallow island channels. (Le Mouel et al., 2025; Matte et al., 2019).

2.2. Numerical model

Hydrodynamic simulations were performed with the three-dimensional TELEMAC-MASCARET module. The model solves the Navier-Stokes equations with a finite-element, hydrostatic formulation using the Boussinesq approximation for density variations (Hervouet, 2007). Turbulence used eddy-viscosity closure: $k-\epsilon$ horizontally and mixing-length vertically. No external forces are considered in the model.

An unstructured triangular mesh spans from Neuville upstream to Saint-Joseph-de-la-Rive downstream (Fig. 1a). The computational mesh

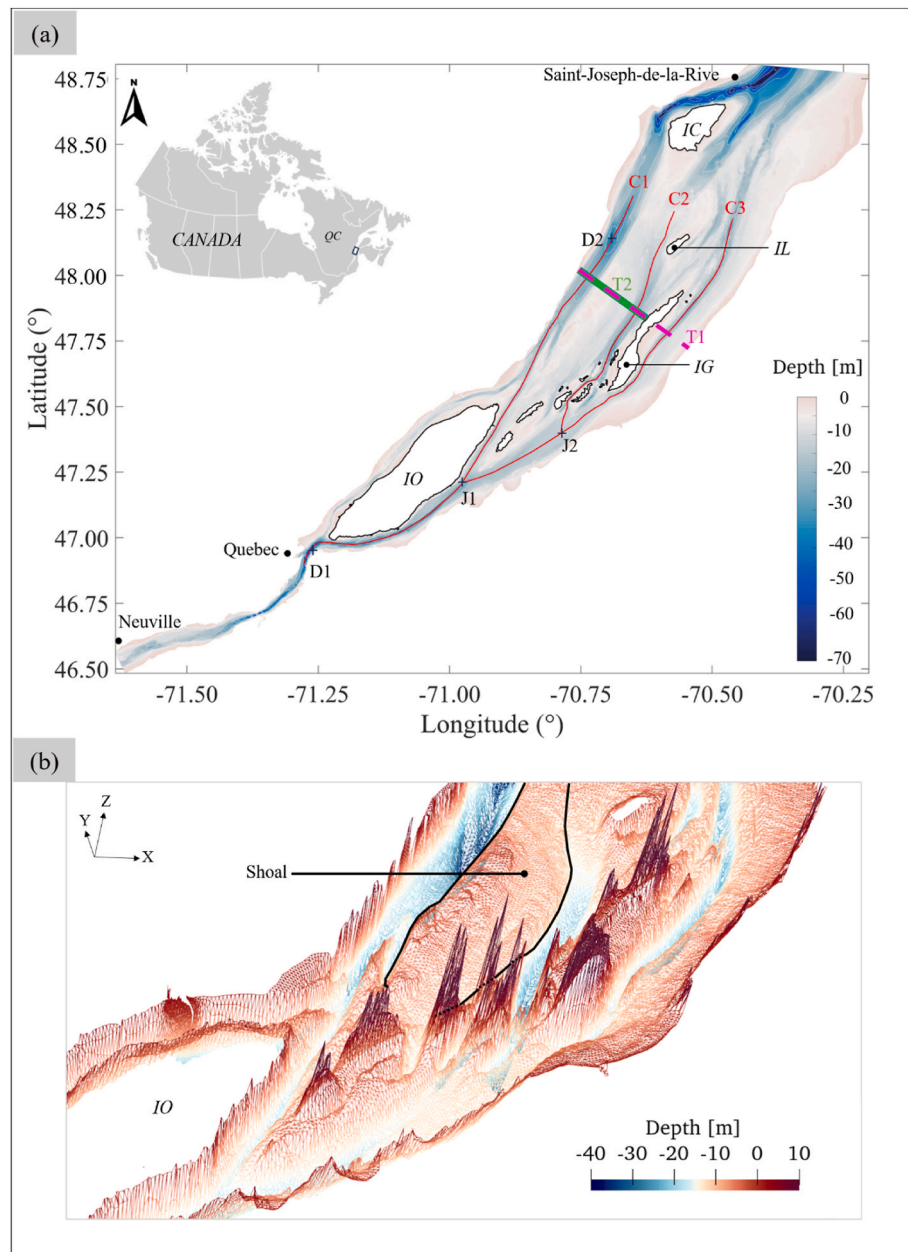


Fig. 1. (a) Bathymetry of the fluvio-intermediate St. Lawrence estuary. (b) 3D geometry with a zoom on the shoal separating C1 and C2. C1, C2, C3: Navigation, intermediate and south channels. IO: Ile d'Orléans. IG: Ile-aux-Grues. IC: Ile-aux-Coudres, IL: Ile-aux-Loups-Marins. D1, D2: local depressions. J1, J2: C1-C3 and C2-C3 junctions. T1: C1-C2-C3 transect (magenta). T2: C1-shoal-C2 transect (green). (For interpretation of the references to colour in this figure legend, the reader is referred to the Web version of this article.)

resolution varies from 25 to 250 m horizontally, with eight sigma layers used to discretize the vertical structure of the water column. A 5 s time step was used to ensure numerical stability and accurate tidal dynamics. Tidal forcing was applied at upstream and downstream nodes using DFO-observed water levels (DFO Fisheries and Oceans Canada, 2009). Simulations covered the spring-tide period 20–26 June 2009. During the simulated period, observed river flows at Québec City ranged between 11,070 and 11,500 m³ s⁻¹.

The Manning friction coefficients were assigned based on the pre-calibrated values reported by Matte et al. (2017b), ranging from 0.018 to 0.0315. These values were initially estimated from bed sediment characteristics and vegetation cover following the approach of Morin et al. (2000), then interpolated onto the 2D finite-element mesh. A slight adjustment of the interpolated values was subsequently performed during calibration against hydrodynamic observations, including water

levels and flow discharge measurements from Matte et al. (2017a).

To isolate barotropic hydrodynamic processes, density-driven currents associated with salinity, temperature, and suspended sediment were neglected (preliminary results for salinity indicate that their contribution remains secondary under the investigated conditions). Tributary inflows were also excluded, as the largest tributary contribution (the Saint-Charles River) is more than an order of magnitude smaller than the St. Lawrence River discharge, even under low-flow conditions. Wind forcing was not considered, as the study focuses on intra-tidal hydrodynamic variability dominated by tidal forcing.

Calibration and validation were conducted using water-level records from two tidal gauges, pressure sensor measurements at three locations, and velocity data obtained from six full three-dimensional transects collected with an Acoustic Doppler Current Profiler (ADCP). During spring tide, water-level errors at the tested stations range between 10

and 20%. For ADCP velocities, errors generally fall between 14 and 23%, except in the upstream part of the northern channel of Île d'Orléans, where they reach 40%. In fact, around LW+45 min, an early flow reversal in the southern branch generates a discharge peak in the northern one, producing the transient jet responsible for the larger discrepancies. Although these discrepancies peak during transition periods and on a localized transect, the ADCP transects do not provide a sufficiently resolved estimate of the timing error associated with velocity reversals. For this reason, the inter-layer lags discussed in this paper should be interpreted as indicative. Further details on the calibration procedure and the observational datasets can be found in Le Mouel et al. (2025); Matte et al. (2014a), respectively.

2.3. Analysis criteria

Reference water levels (HW: high water, LW: low water) are computed from D1 (Fig. 1a). The longitudinal velocity U_n is positive seaward (ebb) and negative landward (flood) and is reference for calculating current inversion fronts. The transverse velocity U_t is positive in the C1→C2 direction, and the vertical velocity W is positive upward.

The dominance between U_n and U_t is assessed using the lateral-to-normal velocity ratio $R = U_t / \sqrt{U_t^2 + U_n^2}$. Values close to zero indicate dominance of U_n , whereas $|R| > 0.7$ indicates transverse dominance.

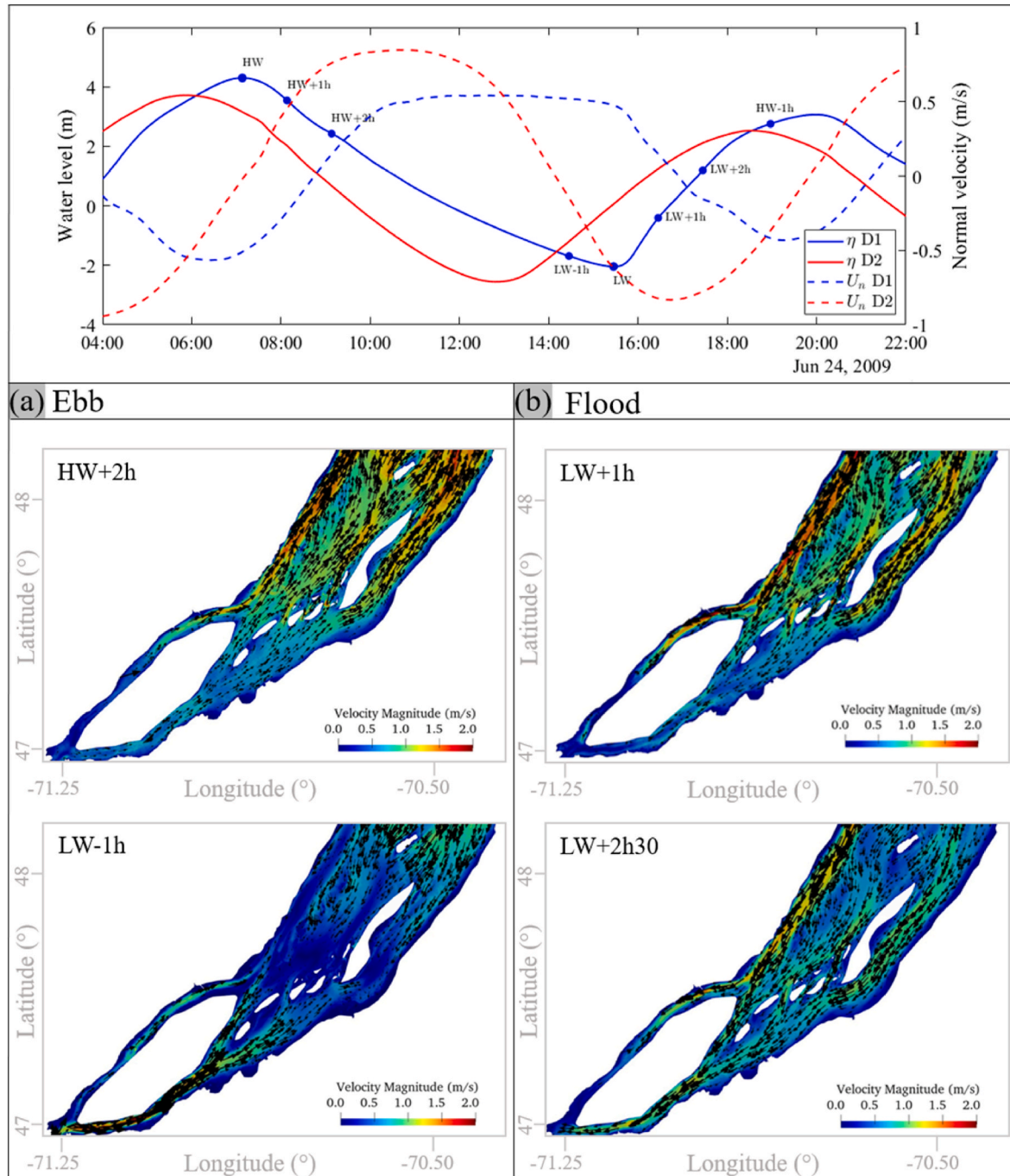


Fig. 2. 2D current propagation during (a) ebb and (b) flood. D1 and D2 denote upstream and downstream reference points. The time reference is defined on the water-level plot η (solid lines) and on the velocity plot U_n (dashed lines) for D1 (blue) and D2 (red). (For interpretation of the references to colour in this figure legend, the reader is referred to the Web version of this article.)

Lateral connectivity between channels is quantified using the lateral advection of longitudinal momentum $Ut^*\partial U_n/\partial l$, where l is the unit-normal tangent direction of the transect. It highlights the dynamic coupling between cross-channel exchange and along-channel circulation. $\partial U_n/\partial l$ positive values denote increasing velocity in the C1→C2 direction. Weak gradients indicate a homogeneous system, while strong gradients reflect pronounced inter-channel velocity differences at dynamic interfaces or shear zones. This gradient also indicates lateral exchange potential: weak gradients imply limited forcing, while strong gradients favor inter-channel mixing.

The tangential convective inertial term ($Ut^*\partial U_t/\partial l$) describes the internal evolution of the lateral cell itself and complements the analysis of inter-channel connectivity. It reveals evolution cycles of compression (positive values), dilatation (negative values), and oscillatory adjustments within the transverse flow.

Vertical analyses are conducted on the sigma layers L2, L4, and L7, which respectively represent bottom, mid-depth, and surface layer dynamics.

3. Results

3.1. Large-scale hydrodynamic organization

To characterize the large-scale tidal circulation, depth-averaged velocity fields were analyzed to identify dominant flow pathways and large-scale asymmetries before examining the three-dimensional dynamics.

3.1.1. Spatial partitioning of the tidal circulation

The system exhibits a recurrent three-phase hydrodynamic cycle: (i) a dynamically established ebb or flood regime with coherent velocity structures (Fig. 2); (ii) a transition phase marked by flow reorganization and localized recirculations; and (iii) a recirculation-dominated phase (Figs. 3 and 4) during which vortical structures intensify and decay before the next tidal state.

During fully developed conditions, a hierarchy emerges between channels (Fig. 3). Peak flood instantaneous tidal discharge reaches $\approx 95,000 \text{ m}^3 \text{ s}^{-1}$ in C1, compared with $60,000 \text{ m}^3 \text{ s}^{-1}$ in C2 and $55,000 \text{ m}^3 \text{ s}^{-1}$ in C3. During ebb, C1 is still dominant ($\approx 90,000 \text{ m}^3 \text{ s}^{-1}$), exceeding C2 ($\approx 55,000 \text{ m}^3 \text{ s}^{-1}$) and C3 ($\approx 50,000 \text{ m}^3 \text{ s}^{-1}$). Lateral connection is strongest between C2 and C1, with C2→C1 transfer during developed ebb (Fig. 3a) and C1→C2 flow during early flood (Fig. 3b).

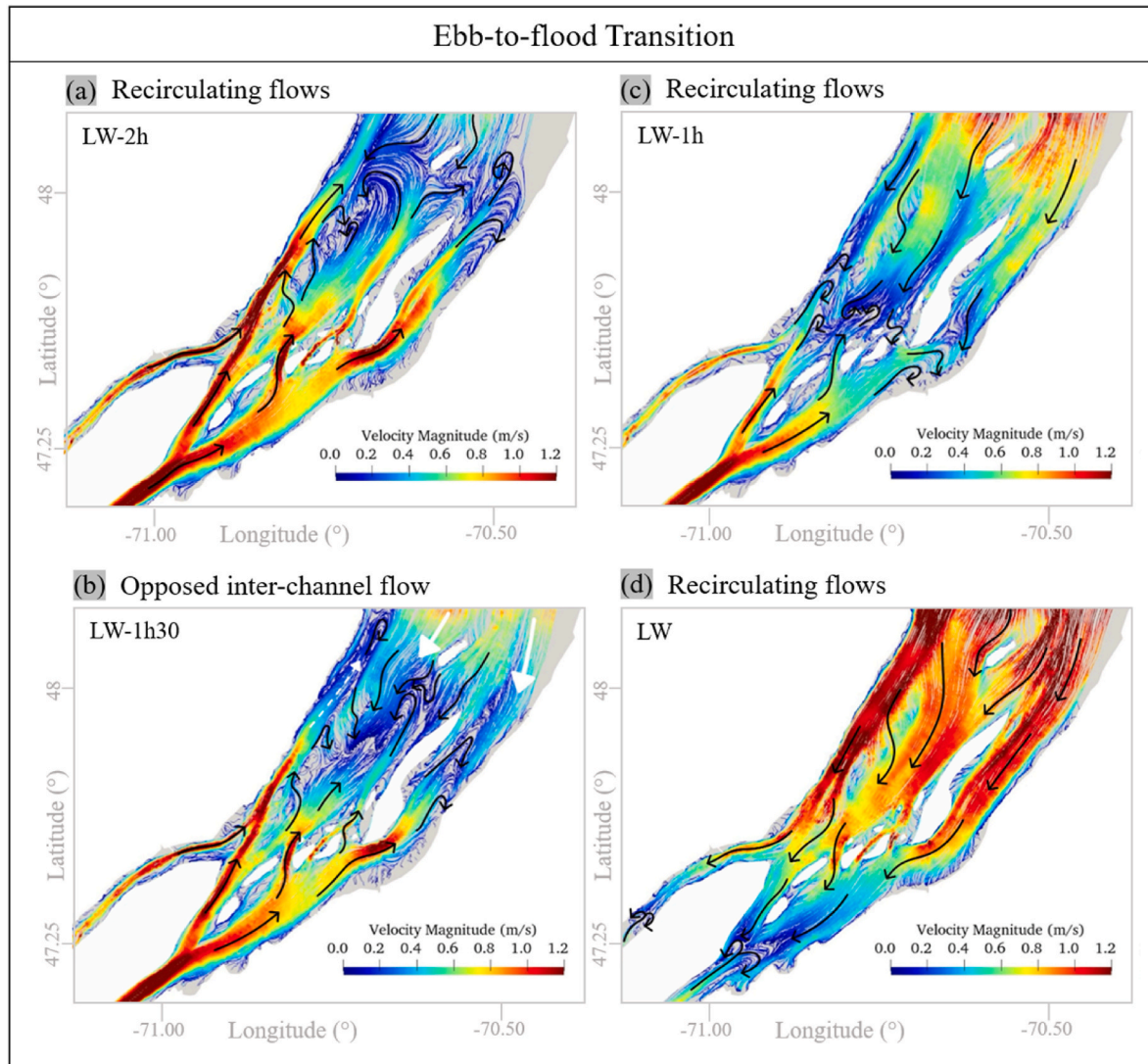


Fig. 3. 2D currents and streamlines during flood-to-ebb, illustrating (a-c-d) recirculation (b) inter-channel flow opposition. Solid white arrows indicate main flow, dashed arrows the opposite direction.

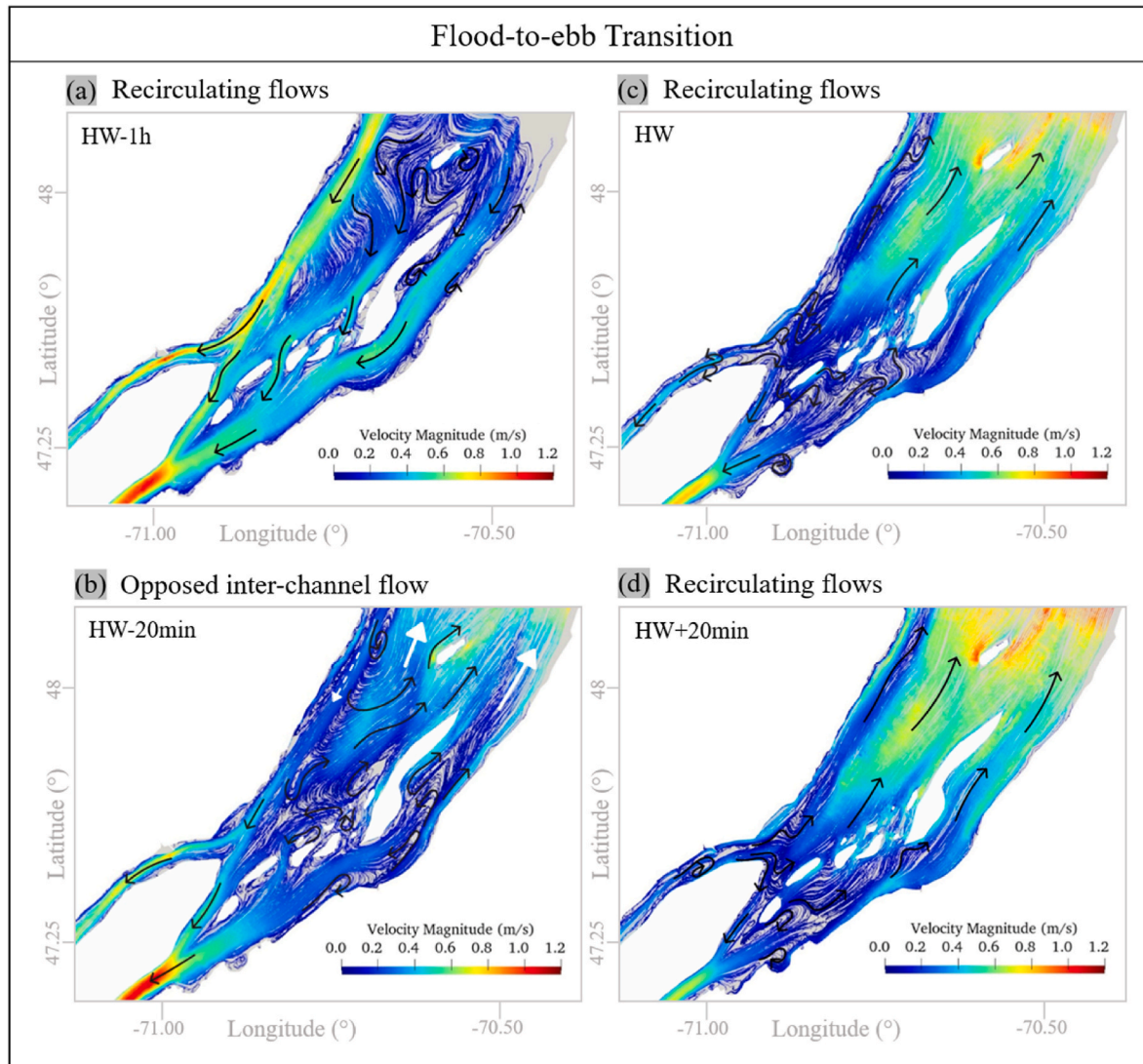


Fig. 4. 2D currents and streamlines during ebb-to-flood, illustrating (a-c-d) recirculation (b) inter-channel flow opposition recirculation. Solid white arrows indicate main flow, dashed arrows the opposite direction.

This aspect is further examined in Sections 3.2 and 3.3. C3 remains more confined due to the IG archipelago.

3.1.2. Transitional flow structures and recirculation zones

On the whole domain, in 2D, reversal remains asynchronous across channels. C2 and C3 reverse earlier than C1, with sequences depending on tidal phase. During ebb-to-flood (Fig. 4), reversal initiates in C3, followed by C2 (≈ 10 min) and C1 (≈ 30 min). During flood-to-ebb (Fig. 5), C2 reverses first, followed by C3 (≈ 10 min) and C1 (≈ 20 min). Temporary flow opposition systematically occurs between channels: C1 may still flow in the previous tidal direction while C2 and C3 have already reversed (Figs. 3b–4b).

During ebb-to-flood, the three channels reverse nearly simultaneously, producing no persistent dominance pattern. Local dominance shifts as the reversal front propagates upstream: C1 initially remains dominant due to its strong ebb remanence, then C2 becomes dominant after its earlier reversal, before C1 regains dominance near LW. These alternations reflect the collapse of ebb momentum and weak transverse pressure gradients at end ebb. However, flood-to-ebb exhibits a more structured and asymmetric dominance sequence. C1 maintains a strong remanent flood flow until HW-1 h, indicating hydraulic dominance. Around HW-20 min, return currents develop from C1 toward C2

downstream, marking a shift toward C2 dominance. This dominance persists around HW, until the ebb reversal propagates upstream and re-establishes large-scale ebb regime.

During these transitions, return currents and recirculation structures develop. Shear zones form between the northern bank and C1, and between C1 and C2, while interface vortices emerge along strong shoal-channel bathymetric gradients (Figs. 3b–4b). These patterns are consistent with interface vortices induced by difference of velocities and steep bathymetric gradient described by Shiono and Knight (1991). As the transition progresses (Fig. 4b), recirculation cells develop at J1 and J2 (near HW and LW+1h), reflecting the competition between remnant and newly established flows.

3.1.3. Propagation of tidal reversal across channels

Tidal reversal fronts ($Un \approx 0 \text{ m s}^{-1}$) exhibit strong spatial heterogeneity driven by channel geometry, vertical shear, and inter-channel connectivity. Propagation is systematically faster during ebb-to-flood (mean velocities of 23.6, 43.4 and 46.5 km h^{-1} across C1–C2–C3) than during flood-to-ebb (16.7, 27.9 and 29.7 km h^{-1}), confirming a strong asymmetry in large-scale adjustment. Asynchronous reversal across channels indicates partial hydraulic decoupling, with flood-to-ebb propagation remaining slowest in C1 and most rapid in C3, while

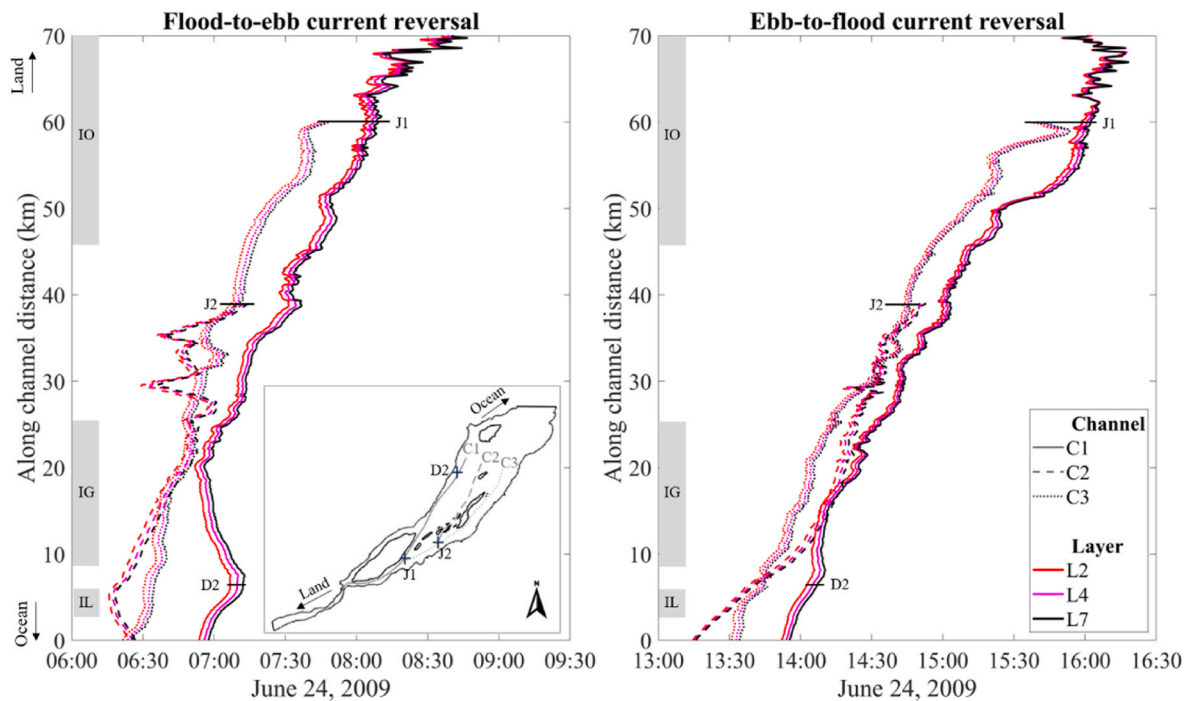


Fig. 5. Propagation of the current reversal front along C1, C2, and C3 thalwegs ($Un \approx 0 \text{ m s}^{-1}$).

ebb-to-flood propagation accelerates markedly in all channels, particularly in C2 and C3 (Table 1). Propagation reversals also exhibit a weak but systematic vertical structure (Table 1). Ebb-to-flood reversals propagate more rapidly than flood-to-ebb reversals, particularly in the deeper channels. During flood-to-ebb, the lower layer (L2) generally propagates faster than the upper layer (L7), with the strongest vertical differences occurring in C1, indicating a delayed response of the deeper flow. During ebb-to-flood, reversal propagation accelerates substantially, and the vertical structure changes: L7 becomes slightly faster than L2 in C1, whereas inter-layer propagation remains nearly synchronized in C2 and C3 (Table 1). On hydraulic control points, vertical phase shifts between layers are small but persistent ($\approx 1\text{--}3 \text{ min}$). At D2, L2 leads L4 by 2–3 min and L7 by 2–3 min depending on the transition. At J1, L2 leads L4 by $\sim 1 \text{ min}$ and L7 by 2–3 min. Alternating zones of hydraulic coupling and decoupling are observed along the domain (Fig. 5). Synchronized fronts indicate lateral coupling, whereas divergence reflects partial hydraulic independence. J1 and J2

Table 1
Mean front propagation velocities and phase lags for the flood-to-ebb and ebb-to-flood transitions, computed for each channel (C1–C3) and layer (L2, L4, L7). Front-propagation velocities are calculated considering $Un = 0 \text{ m s}^{-1}$. Phase lags are expressed relative to layer L7. Positive values indicate an earlier arrival of the inversion front compared to L7.

Channel	Layer	Front-propagation velocities flood-to-ebb (km.h ⁻¹)	Front-propagation velocities ebb-to-flood (km.h ⁻¹)	Phase lag flood-to-ebb	Phase lag ebb-to-flood
C1	L2	18.1	17.3	4min37s	5min56s
	L4	17.3	26.8	2min32s	2min13s
	L7	14.6	26.7	00min00s	00min00s
C2	L2	28.0	43.8	2min26s	3min13s
	L4	27.9	43.3	1min19s	2min56s
	L7	27.8	43.0	00min00s	00min00s
C3	L2	29.7	46.7	4min37s	4min16s
	L4	29.7	46.5	2min33s	2min24s
	L7	29.7	46.3	00min00s	00min00s

act as control points where the front accelerates or decelerates because of lateral momentum redistribution. D2 is a dissipative zone, particularly slowing propagation in C1 during flood-to-ebb. Localized front retreats occur in all channels, reflecting transient oscillations around zero velocity, localized turbulence, and hydrodynamic adjustments linked to channel interactions. These features are particularly pronounced in the central sector between IO and IG with strong irregularity (25–38 km downstream J2), where C2 and C3 alternate between synchronized and desynchronized propagation, while C3 generally maintains a more linear trajectory. Smaller retreats also appear downstream D2 in C1 (8–18 km ebb-to-flood; 8–21 km flood-to-ebb), preceding the high variability observed in C2 (28–40 km), highlighting inter-channel interactions and delayed responses (see Section 3.3). Additional deviations occur in C3 upstream J1 and J2 (32–35 km and 56–59 km), particularly during flood-to-ebb, with a similar feature in C1 (53–56 km). During ebb-to-flood, the weaker retreat (59–60 km) reflects a local imbalance in C3–C2 reversal dynamics (Fig. 5). The influence of IO is significant beyond $\sim 56 \text{ km}$, where C1 and C3 converge into the south branch (downstream J1). Upstream ($>70 \text{ km}$), increased variability in C1 reflects the influence of the tidal junction and associated bathymetric gradients near D1 (Fig. 5).

3.2. 3D hydrodynamic connectivity

Three-dimensional velocities are analyzed to explain inter-channel connectivity during the tidal cycle, combining high-resolution spatio-temporal analysis. 3.2.1. Vertical structure of inversion fronts Phase lags between channels are examined through the vertical structure of inversion fronts along C1, C2, and C3 (Fig. 6). Bathymetry exerts a first-order control on the longitudinal and vertical flow structure. Along C1, D2 (0–25 km) enhances acceleration and vertical shear, especially around LW and LW+1h. The intermediate reach (30–55 km), shallower (–20 to –15 m), is vertically homogeneous at HW but develops strong shear and a confined low-velocity core at LW+1h. Upstream (55–80 km), highly variable bathymetry (–15 to –55 m) across the IO southern branch and D1 produces the most energetic conditions,

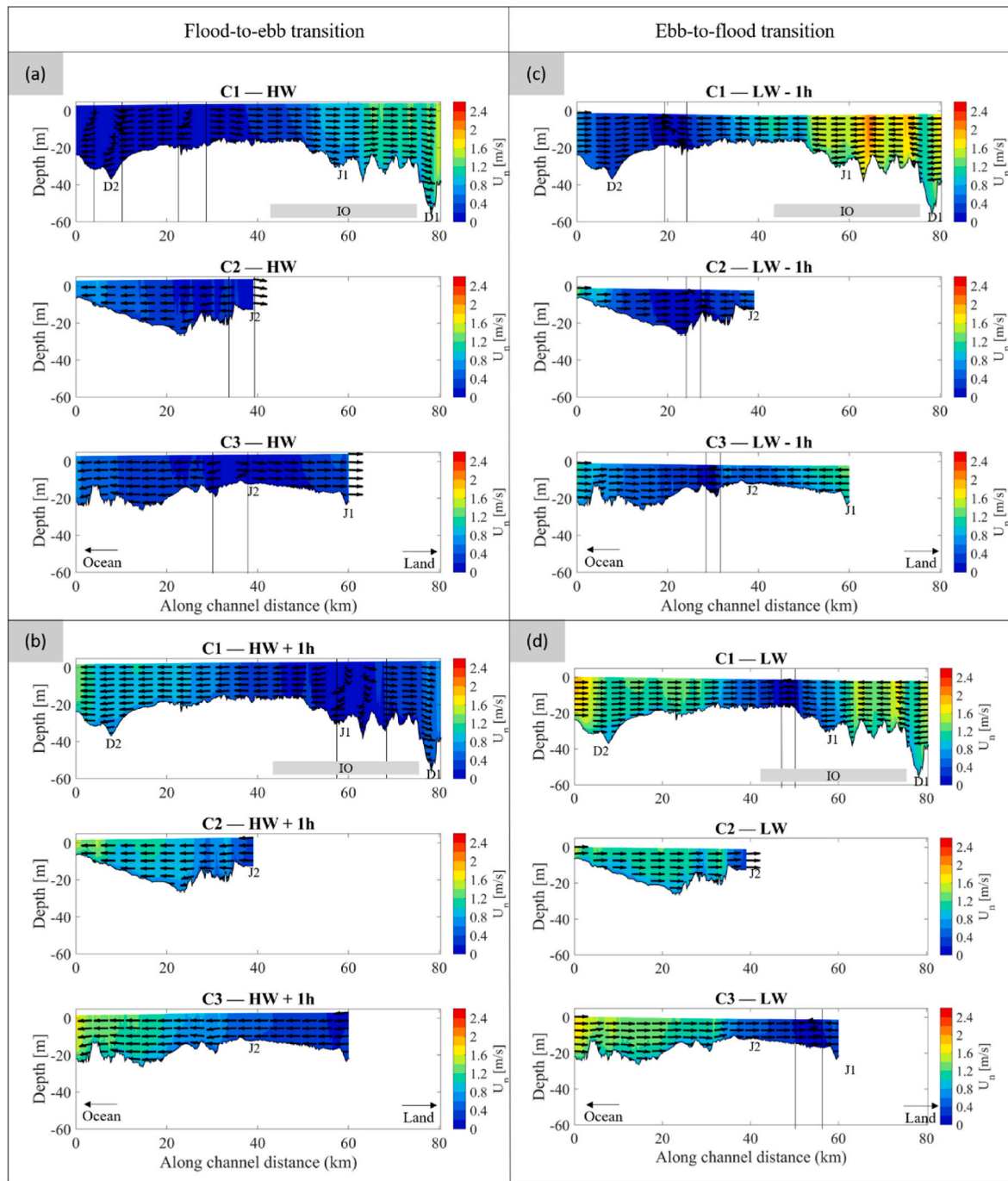


Fig. 6. Longitudinal vertical velocity sections in C1, C2 and C3 during (a)-(b) flood-to-ebb and (c)-(d) ebb-to-flood. Arrows indicate along-channel velocity direction.

with velocities reaching 2.2 m s^{-1} during fully developed ebb. Vertical velocities remain significant upstream ($>65 \text{ km}$) and near flow reversals, consistent with intensified three-dimensional adjustment.

Tidal reversal fronts exhibit a strongly non-uniform and phase-lagged propagation across the three-channel system. Rather than a continuous progression, reversals occur at discrete locations and reorganize progressively in space and time, with C2 and C3 systematically preceding C1. During ebb-to-flood (LW–1h to LW+1h), multiple coexisting fronts indicate a stepwise propagation of the inversion, associated with return currents and recirculations (Figs. 4 and 5). C2 reverses first ($\approx 20\text{--}40 \text{ km}$), followed by C3 near J2 and C1 near J1 ($40\text{--}70 \text{ km}$), producing a spatially distributed transition with temporal offsets ranging from $\sim 10 \text{ min}$ to $\sim 1 \text{ h}$. However, flood-to-ebb (HW to HW+1h)

exhibits a more continuous reversal, with reduced inter-channel phase differences and fewer convergence zones. At HW–1h, upstream C1 remains in flood ($\sim 1.6 \text{ m s}^{-1}$) while downstream areas are already weakening, and the inversion front propagates upstream from $\sim 30 \text{ km}$ to $\sim 65 \text{ km}$ by HW+1h.

J1 and J2 act as cross-channel synchronization points. During ebb-to-flood, initially fragmented fronts become locally aligned at these junctions, forming transient convergence zones that partially synchronize the inversion. During flood-to-ebb, fronts are more continuous and diffuse, with reduced spatial heterogeneity across channels.

3.2.2. Lateral circulation pattern

This section investigates inter-channel connectivity shaped by shoals

and island-induced channelization. Transect T1 (20 km) crosses the three main channels separated by a shoal and IG, defining five units: C1 (0–3 km), the shoal (4–10 km), C2 (10–12 km), IG (12–15 km), and C3 (15–20 km) (Fig. 8).

Lateral instantaneous flow across the shoal reaches $\approx 8900 \text{ m}^3 \text{ s}^{-1}$ during flood ($27,400 \text{ m}^3 \text{ s}^{-1}$ longitudinal) and $\approx 6400 \text{ m}^3 \text{ s}^{-1}$ during ebb ($32,100 \text{ m}^3 \text{ s}^{-1}$ longitudinal), indicating strong and persistent inter-channel coupling. During transitions, lateral flow remains significant,

with $\approx 6300 \text{ m}^3 \text{ s}^{-1}$ from flood-to-ebb ($4300 \text{ m}^3 \text{ s}^{-1}$ longitudinal) and $\approx 1200 \text{ m}^3 \text{ s}^{-1}$ from ebb-to-flood ($13,400 \text{ m}^3 \text{ s}^{-1}$ longitudinal). Lateral fluxes are generally stronger near the surface.

U_n exhibits strong spatial structuring across the system. Peak flood and ebb velocities reach $1.2\text{--}1.6 \text{ m s}^{-1}$ in C1 and $1.0\text{--}1.2 \text{ m s}^{-1}$ in C2 and C3, while remaining much lower over the shoal due to reduced depth. Around LW–1h and HW–1h (Fig. 7a–c), U_n decreases to near-zero across most of the section, creating near-slack conditions that enhance

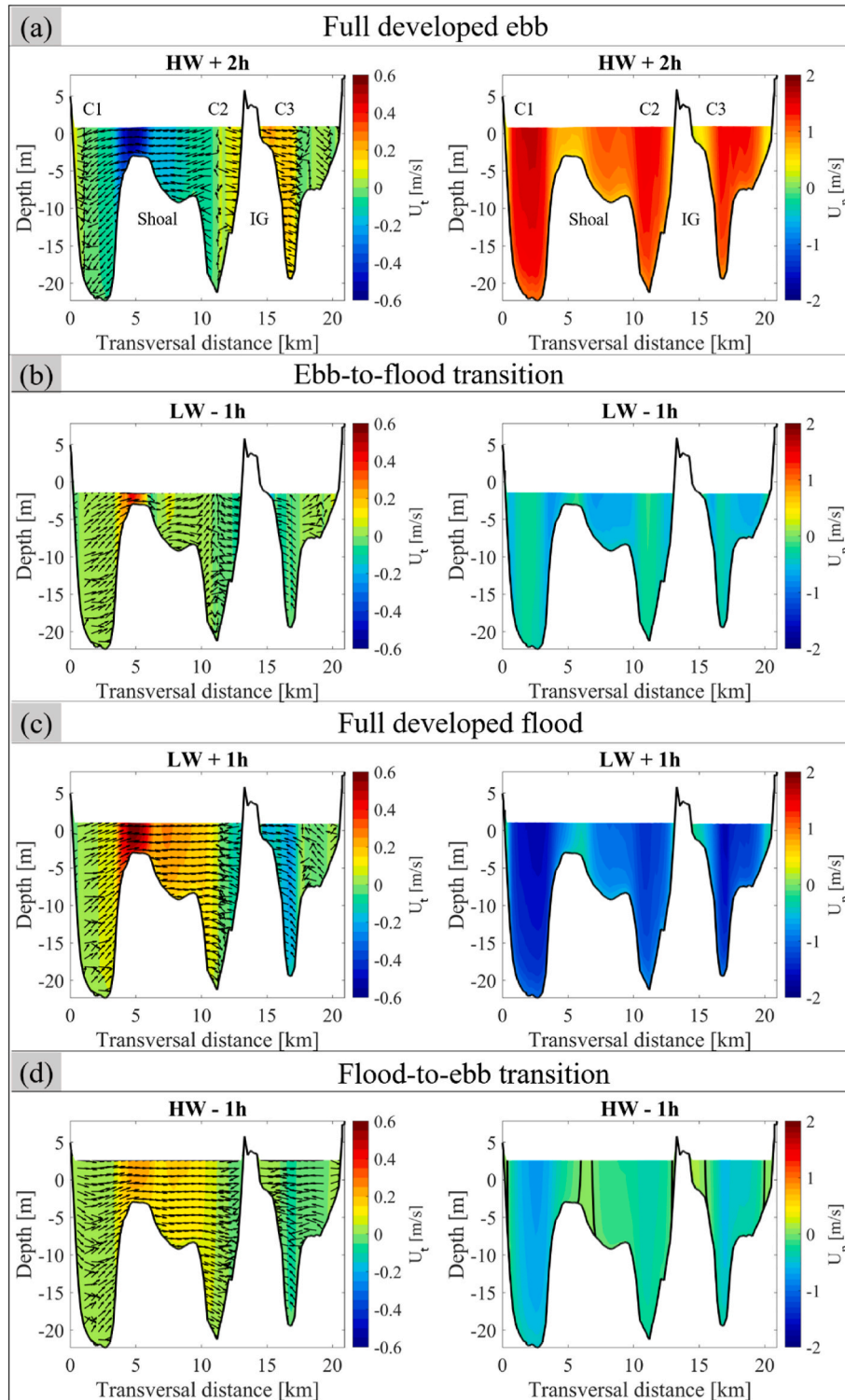


Fig. 7. Cross-sectional velocity fields of T1 at (a) developed ebb (b) ebb-to-flood, (c) developed flood and (d) flood-to-ebb. U_t denotes the lateral velocity component along the transect; black arrows indicate U_t direction. U_n denotes the velocity component normal to the transect, and the thick black line indicates $U_n = 0 \text{ m s}^{-1}$.

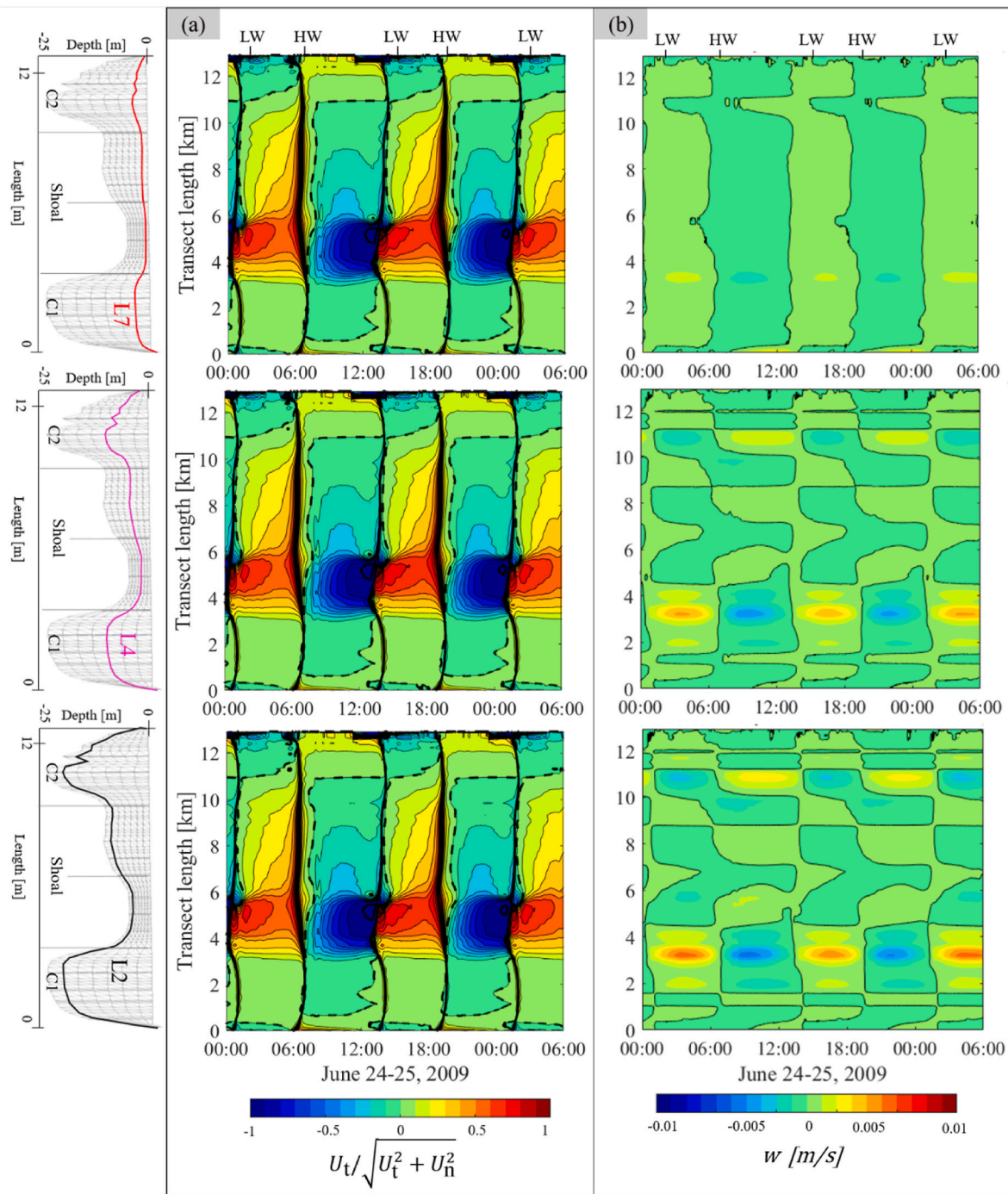


Fig. 8. Temporal evolution of (a) lateral-normal velocity ratio and (b) vertical velocity at T2 (L2, L4 and L7). In (a), dashed line: $U_t = 0 \text{ m s}^{-1}$; thick black line: Un reversal. In (b), black line: $W = 0 \text{ m s}^{-1}$.

transverse circulation. At LW–1h (Fig. 7c), the shoal is already in flood while the C1 and C3 thalwegs still sustain ebb velocities $>0.5 \text{ m s}^{-1}$. This non-synchronous reversal induces transient longitudinal gradients and enhances cross-channel exchanges during tidal transitions.

U_t reveals structured, tide-dependent exchanges between C1 and C2 across the shoal. During ebb development (HW+1h to HW+2h), U_t intensifies along the shoal crest, reaching 0.5 m s^{-1} toward C1, confirming the shoal as a preferential pathway for lateral exchanges (Fig. 7a). During flood, a stronger reversed pattern emerges: U_t reaches 0.6 m s^{-1} at LW+1h, with C1→shoal→C2 flow (Fig. 7c), indicating a flood-dominant asymmetry ($\sim 20\%$) and more efficient lateral exchanges than during ebb.

Lateral circulation exhibits channel-dependent complexity due to interactions between longitudinal flows and local bathymetry. In C3, circulation is rotational coherent, with counter-clockwise flow during flood-to-ebb and clockwise flow during ebb-to-flood, reaching $U_t \sim 0.3\text{--}0.4 \text{ m s}^{-1}$ (Fig. 7), reflecting the hydraulic isolation the channel. C1 and C2 display more complex patterns: in C2, lateral inflow from C1 ($U_t \sim 0.5\text{--}0.6 \text{ m s}^{-1}$ at LW+1h) interacts with bathymetry-induced shear near IG, producing competing effects between externally forced and locally induced circulation.

These patterns reveal a strongly time-dependent connectivity, motivating a refined temporal analysis of longitudinal-lateral interactions over the tidal cycle.

3.2.3. Temporal evolution of lateral inter-channel connectivity

Figs. 8 and 9 show the temporal redistribution of lateral momentum along T2, spanning C1 (0–3 km), the shoal (4–10 km), and C2 (10–12 km). The thalwegs of C1 and C2 lie at 2.3 km and 10.8 km, and the shoal slope between 5.6 and 8 km reaches 0.25% (Fig. 8).

The lateral-to-normal velocity ratio reveals strong spatial differences between the shoal and the two channels (Fig. 8a). Connectivity is confined to the shoal, while both channels remain dominated by U_n . Over the shoal, the ratio often exceeds 0.6 and approaches 0.8–0.9 during tidal transitions, with durations of ~35 min (end flood) and 3h20 (ebb-to-flood). This dominance persists for 70% of the tidal cycle, showing that inter-channel connectivity is primarily controlled by transverse processes.

Flood transitions propagate spatially along the shoal slope, whereas ebb transitions occur more synchronously across T2. During flood, oblique ratio contours (6–10 km) indicate a progressive, space-dependent shift between U_t - and U_n -dominated regimes, extending from the shoal into C2 (Fig. 8a). During ebb, contours become quasi-perpendicular to T2, reflecting a near-simultaneous transition and a narrower adjustment window.

The strongest cross-channel exchanges occur during tidal transitions (Figs. 7 and 8). At LW–1h, U_t reaches 0.4 m s^{-1} over the shoal, although opposing longitudinal flows may limit connectivity (Fig. 7b). At HW–1h, as weakens U_n across the section, allows effective lateral circulation, with U_t of $0.2\text{--}0.3 \text{ m s}^{-1}$ from C2 to C1 (Fig. 7d). During fully developed phases, U_t remains weak and the lateral-to-normal velocity

ratio decreases, making inter-channel exchanges secondary to longitudinal flow.

Tidal reversals exhibit a clear channel-dependent asymmetry. On the T2 transect, C2 leads C1 by 20–25 min during flood-to-ebb, while C1 leads by ~5 min during ebb-to-flood. This desynchronization generates transient shear zones that enhance lateral exchanges, producing a smoother flood-to-ebb transition and a sharper ebb-to-flood reversal. During ebb-to-flood, strong lateral shear develops between already-reversed flow in C2 and still-ebbing flow in C1, especially over the shoal crest (Fig. 9b). The intensity and timing of lateral exchanges therefore differ between flood and ebb. During fully developed flood conditions, increasing longitudinal velocities progressively constrain transverse circulation, whereas during tidal reversals, the weakening of U_n allows lateral jets to expand across the shoal. The delayed adjustment between C1 and C2 enhances cross-channel exchange before the flow becomes longitudinally dominated again.

Vertical exchange is strongest near the bottom and driven by interactions between lateral flow and the steep C1–shoal slope (~1%), with vertical velocities reaching 0.007 m s^{-1} in L2. During flood, C1→C2 flow generates upward motion and compensating downward velocities near 10–11 km. During ebb, reversed C2→C1 flow produces downward motion that reinforces lateral velocities over the shoal crest (Fig. 8b).

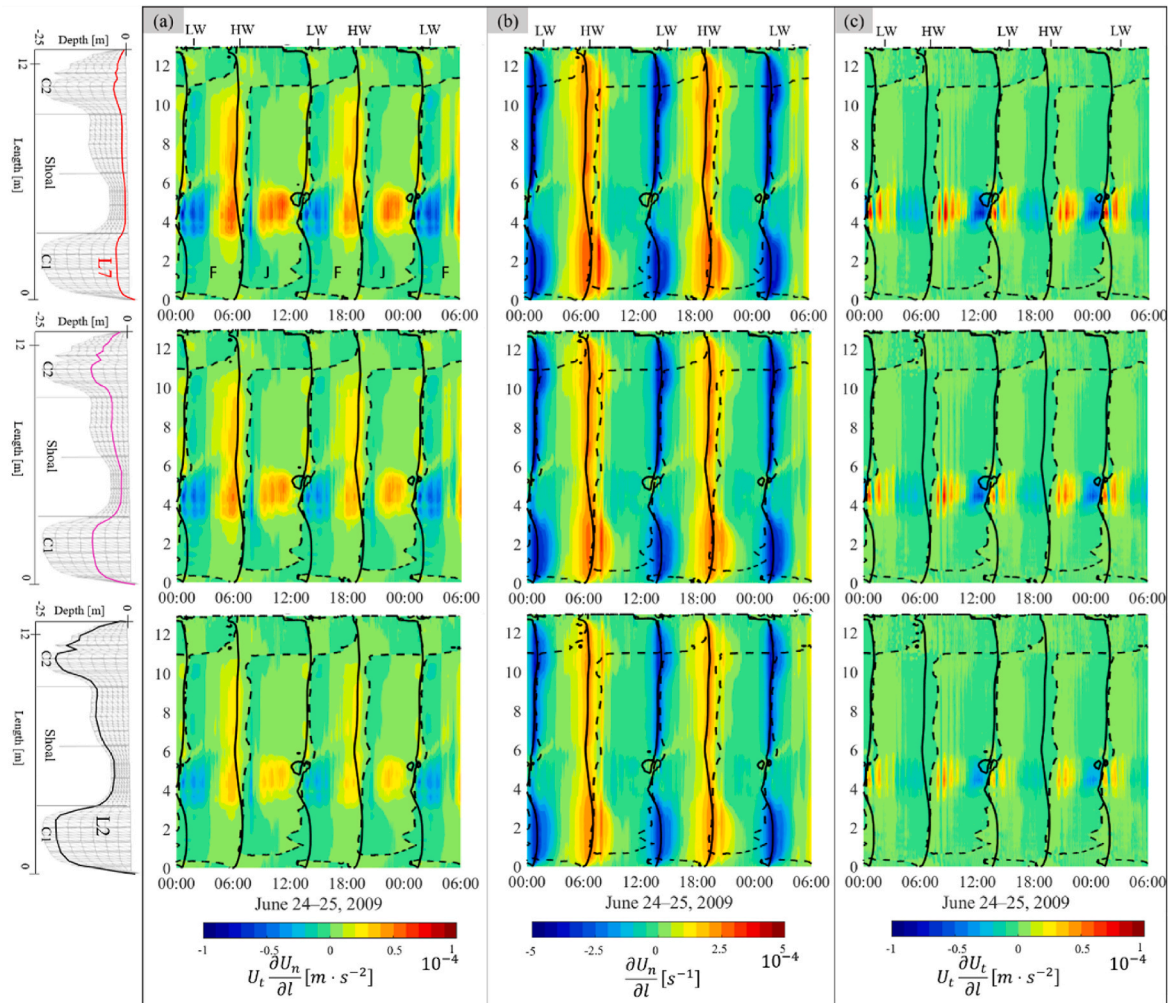


Fig. 9. Temporal evolution of (a) lateral advection of longitudinal velocity momentum (b) horizontal shear of the normal velocity component, and (c) tangential convective inertial term at T2 (L2, L4 and L7). Dashed line: U_t reversal; thick black line: U_n reversal.

3.3. Mechanisms of lateral momentum exchange

3.3.1. Lateral advection of transverse-channel velocity momentum

Alternating C1→C2 (ebb) and C2→C1 (flood) dominance directly results from asymmetric channel reversibility. During ebb and tidal transitions, the flow is in the C1→C2 direction and reinforces the main-channel circulation. Whereas a C2→C1 flow is primarily observed during flood.

The system is controlled by two competing lateral processes: dominant channel reinforcement versus slower channel compensation. When U_t and $\partial U_n/\partial l$ share the same sign, lateral advection strengthens the dominant branch (reinforcement regime). When they have opposite signs, momentum is redistributed toward the slower branch (compensation regime).

During the first half of the flood (00h20–03h10), lateral advection ($U_t \partial U_n/\partial l$) is $-0.5 \cdot 10^{-4} \text{ m s}^{-2}$ between 4 and 6 km, with $U_n < 0 \text{ m s}^{-1}$, $U_t > 0 \text{ m s}^{-1}$, and $\partial U_n/\partial l < 0 \text{ s}^{-1}$, producing a flux from C1 toward the slowest channel C2, opposed to the lateral gradient of longitudinal velocity (Fig. 9a–b). In the reinforcement regime, just before flood-to-ebb (04h30–06h30), lateral advection becomes positive ($0.3 \cdot 10^{-4}$ to $0.5 \cdot 10^{-4} \text{ m s}^{-2}$) over 4–6 km and extends downslope to 10 km to C2. During this period, $U_n < 0 \text{ m s}^{-1}$, $U_t > 0 \text{ m s}^{-1}$, and $\partial U_n/\partial l > 0 \text{ s}^{-1}$, indicating a flux from C1 toward the fastest channel C2 along the longitudinal gradient of normal velocity (Fig. 9a–b). Immediately after flood-to-ebb (06h20–07h40), U_n has reversed ($U_n > 0 \text{ m s}^{-1}$), $U_t > 0 \text{ m s}^{-1}$, and $\partial U_n/\partial l > 0 \text{ s}^{-1}$, resulting in positive lateral advection on the shoal tip and continued reinforcement of the flow toward C2 (Fig. 9a–b).

In the compensation regime, during the short period following inversion of U_t at the shoal crest (07h43–08h00), lateral advection becomes negative as $U_n > 0 \text{ m s}^{-1}$, $U_t < 0 \text{ m s}^{-1}$, and $\partial U_n/\partial l > 0 \text{ s}^{-1}$, resulting in a flux from C2 to C1 and reinforcing the flow in C1 (Fig. 9a–b). During ebb (08h45–12h15), lateral advection remains positive, with $U_n > 0 \text{ m s}^{-1}$, $U_t < 0 \text{ m s}^{-1}$, and $\partial U_n/\partial l < 0 \text{ s}^{-1}$, corresponding to a C2→C1 lateral redistribution along the longitudinal gradient of normal velocity. Finally, during ebb-to-flood, the lateral flux is non-uniform along the transect. Between 4 and 10 km, U_t is near zero and lateral advection is minimal, except at 6 km, where $U_n > 0 \text{ m s}^{-1}$, $U_t < 0 \text{ m s}^{-1}$, and $\partial U_n/\partial l < 0 \text{ s}^{-1}$, indicating C2→C1 flow along the gradient, reinforcing the slowest channel, C1 (Fig. 9a–b).

Vertical structure systematically enhances exchanges, regardless of tidal phase. The intensity of $U_t \partial U_n/\partial l$ exhibits a clear vertical organization, with L7 reaching $0.6 \cdot 10^{-4} \text{ m s}^{-2}$, L4 around $0.4 \cdot 10^{-4} \text{ m s}^{-2}$, and L2 approximately $0.2 \cdot 10^{-4} \text{ m s}^{-2}$, highlighting the enhanced lateral connectivity near the surface (Fig. 9a).

3.3.2. Lateral gradient of longitudinal velocity

Maximum lateral shear occurs during tidal transitions, highlighting their importance for lateral reorganization. During ebb-to-flood, $\partial U_n/\partial l$ is predominantly negative, with velocities decreasing in C1→C2 direction. This regime persists from 13h10 to 15h10 in C1, 13h30–14h55 in C2, and 14h00–14h10 over the shoal (L7 reference) (Fig. 9b).

Conversely, during flood-to-ebb, the gradient becomes positive, indicating that velocities increase from C1 to C2 ($U_n, C1 < U_n, C2$), with this phase extending from 05h20 to 08h50 in C1, 05h25 to 08h10 in C2, and 05h55 to 07h02 over the shoal. Depth effects shorten these phases (L4 and L2), showing that lateral shear is more persistent near the surface and more transient at depth due to frictional damping and the delayed response of the lower water column (Fig. 9b).

The sign of the lateral velocity gradient provides a direct indication of transient channel dominance, not only in terms of velocity magnitude but also in terms of dynamical control over the cross-channel structure. When $\partial U_n/\partial l < 0 \text{ s}^{-1}$, higher velocities in C1 ($U_n, C1 > U_n, C2$) indicate that C1 carries greater longitudinal momentum and imposes the shear structure, thereby shaping the lateral organization of the flow. Conversely, when $\partial U_n/\partial l > 0 \text{ s}^{-1}$, dominance shifts to C2, which then

controls the velocity gradient and the associated shear (Fig. 9b).

3.3.3. Tangential convective inertial term

In general, the tangential convective inertial term is largest at L7 ($1.10 \cdot 10^{-4} \text{ m s}^{-2}$) and damped at L2 ($0.4 \cdot 10^{-4} \text{ m s}^{-2}$), and it is also more temporally extended in the upper layers, highlighting the importance of depth-resolved measurements for capturing three-dimensional connectivity (Fig. 9c).

During the first phase of flood (07h30–10h50), positive tangential convective inertial term corresponds to self-reinforcement of the C1→C2 flow, with transverse gradients concentrating the flow toward C2. During the second phase (10h50–13h20), the term becomes negative, indicating lateral redistribution toward C1, weakening the lateral flow (Fig. 9c).

During ebb, the first phase (13h20–16h10) shows positive tangential convective inertial term as upslope motion compresses streamlines, strengthening C2→C1, while the second phase (16h10–19h55) becomes negative, reflecting redistribution toward C2. Weakly structuring phases occur when slopes are small, flows reverse, or compression and dilatation effects compensate. The shoal crest emerges as a lateral adjustment zone, where C1↔C2 cells alternately strengthen, weaken, and reorganize under tidal forcing and shoal geometry (Fig. 9c).

Even within a given regime (i.e., same sign configuration of U_t and $U_t \partial U_t/\partial l$), the intensity of the tangential convective inertial term oscillates. The pulsations are stronger, especially during the early stages of flood and ebb when the lateral cell is forming. During these phases, rapid increases in U_n , continuous lateral adjustment of streamlines, and oscillations of the transverse gradient produce successive peaks of positive tangential convective inertial term. During mid-tide conditions, the amplitude of these oscillations decreases.

The efficiency of lateral inter-channel exchange depends on tidal transitions shoal-induced lateral inertia. During flood-to-ebb, C1 lateral currents reverse first, dissipating the C1→C2 cell. However, during ebb-to-flood, U_t and U_n in C2 reverse almost simultaneously, sometimes with a slight lead of U_t , so the C2→C1 cell decays more gradually. Over the shoal (~6 km), U_t consistently lags behind U_n in both transitions due to the slope-induced lateral inertia. While longitudinal flow reverses quickly, the lateral cell requires extra time to traverse the shoal slope and reestablish a coherent transverse gradient, propagating adjustments along the shoal.

4. Discussion

Inter-channel connectivity emerges from a sequence of coupled hydrodynamic processes involving longitudinal forcing, lateral exchanges, and three-dimensional flow reorganization. Bathymetric contrast between deep channels and shallow shoals first controls the propagation of tidal reversals, generating phase lags and transient synchronization/desynchronization between channels. These phase lags enhance lateral shear and promote cross-channel momentum exchanges, which become the dominant connectivity mechanism during tidal transitions. The resulting lateral circulation is not a passive response but a dynamically reorganized three-dimensional structure, where the tangential convective inertial term governs the alternation between reinforcement and redistribution of transverse flow. This sequence highlights that hydrodynamic coupling depends not only on instantaneous velocity magnitude, but also on spatial gradients, channel connectivity, and morphologically constrained pathways.

4.1. Bathymetric control and longitudinal forcing

Bathymetric contrast between deep channels and shallow shoals first controls the propagation of tidal reversals by modifying inertia, frictional dissipation, and momentum redistribution. These effects generate phase lags between channels, which subsequently regulate the timing and intensity of lateral exchanges.

The results presented in Section 3.2.1 allow us to interpret the tidal transition asymmetry along C1 as resulting from the combined effects of hysteresis, bathymetry, and bottom friction (Section 3.2.1). The ebb-to-flood transition is slower and more heterogeneous due to three mechanisms: (i) the shallow portion of the talweg at low water increases bottom friction, reducing inertia and delaying flood propagation; (ii) this low-velocity buffer dissipates flood energy, delaying reversal front migration; (iii) upstream ebb flow (at IO and upstream) maintains residual momentum, with velocities remaining high until LW-1h, providing additional resistance to the incoming flood and enhancing phase lag (Fig. 6). By contrast, flood-to-ebb is faster and dominated by the pressure gradient associated with seaward estuarine drainage. Higher water depths reduce frictional effects, and the absence of a low-velocity structure allows the ebb reversal signal to propagate upstream across the shallow area of C1. These mechanisms generate stronger phase lags and enhance lateral momentum transfer.

Beyond controlling tidal propagation, bathymetry also structures the vertical organization of the flow. High-velocity cores and sharp vertical shear develop within bathymetric depressions (D1 and D2). As the current enters these deep regions, the combined effects of slope, depth variations, and longitudinal pressure gradients intensify velocities and enhance vertical shear, reinforcing the three-dimensional flow structure. When the current exits toward shallower areas, increased friction promotes rapid momentum dissipation, reducing velocity gradients and weakening the three-dimensional flow organization. In C1 the shallow reach amplifies this effect by reducing inertia and forming low-velocity zones that act as hydraulic dampers and redistribute the flow (Fig. 6).

Phase lags between C1 and C2 arise from the asymmetric propagation of inter-channel reversal fronts, where one channel reverses while the other still conveys ebb flow. This generates strong lateral shear and drives momentum transfer across the shoal. Under these conditions, the shallow inter-channel zone acts as an active mixing interface, where friction and turbulence dissipate momentum and promote exchange. As described in Section 3.2.2, during flood-to-ebb, the double front of propagation induced by recirculation in C1 reflects these phase-lagged inversions, indicating that shoal-driven lateral circulation shapes transient velocity structures. These interactions generate return flows and recirculation patterns controlled by inter-channel dominance, bathymetry, and lateral connectivity (see Sections 3.1.2 and 3.2).

Once generated, these phase lags evolve according to the efficiency of lateral connectivity between channels. At the C1-shoal-C2 connection, just upstream D2 (Figs. 5 and 6), reversal fronts progressively resynchronize during flood-to-ebb as momentum is transferred from C1 toward C2. During ebb-to-flood, lateral connectivity adjusts more slowly, as phase lags persist due to the inability of lateral fluxes to compensate for longitudinal asymmetry and channel inertia. The inversion initiates over the shallow shoal and propagates toward the deeper channel (Figs. 7 and 8). Building on Le Mouel et al. (2025), which highlighted the role of phase lags and asynchronous reversals in inter-channel dynamics, these results suggest a more general mechanism in which tidal inversions propagate from shallow to deeper regions, controlled by bathymetry and inertia gradients, providing a unified framework for inter-channel connectivity.

In Section 3.1.3 and 3.2.1, we show junctions behave as hydraulic nodes where tidal forcing, channel geometry, and lateral asymmetry jointly determine whether flows synchronize or desynchronize. They partition flow and redistribute momentum between branches. In diffuence configurations, flow partitioning leads to differential propagation velocities and promoting progressive phase divergence between channels. In contrast, confluence regions tend to favor flow resynchronization through mass and momentum continuity. However, this resynchronization depends on the phase lags accumulated upstream. These nodal regions therefore represent zones of intense three-dimensional momentum redistribution, with large velocity fluctuations during tidal reversals ($>1 \text{ m s}^{-1}$ in the present case). Shallow inter-channel shoals further act as buffering zones, where reduced inertia and

enhanced friction progressively damp phase differences before junction convergence.

Inter-channel dynamics are controlled by geometric, forced, and dynamic asymmetries, particularly during transitional states. Geometric asymmetry, driven by differences in depth and width between C1, C2, and C3, structures lateral shear and recirculation zones (Yuan et al., 2025). Forced asymmetry, linked to inter-channel differences in tidal propagation, contributes to temporal lags in flow response (Buschman et al., 2013; Iwamoto et al., 2020; Ragno et al., 2020). Dynamic asymmetry emerges internally from nonlinear hydrodynamic interactions (Dai et al., 2020; Le Mouel et al., 2025; Xiao et al., 2021). The relatively continuous reversal occurring during HW-HW+1h indicates limited control of junctions on lateral connectivity under smooth propagation conditions. However, dynamic asymmetry dominates during flood-to-ebb (LW-LW+2h), when phase lags develop between channels, particularly at the IO junction and within inter-channel recirculation zones (Fig. 4).

4.2. Lateral momentum exchanges and tidal asymmetry

Asynchronous current reversals between channels enhance lateral shear and promote cross-channel momentum exchanges, transforming longitudinal asymmetry into transient lateral connectivity. Lateral circulation intensifies during flow-reversal and plays a central role in shoal connectivity, with lateral advection occasionally surpassing longitudinal currents. This dynamic depends on three main physical mechanisms: differential advection, rotation, and diffuse bottom boundary layers. This behavior is consistent with the mechanisms identified by the study of Lerczak & Geyer (2004), which demonstrated that lateral circulation can dominate the momentum budget in quasi-homogeneous estuaries. Lateral circulation becomes a key driver of inter-channel connectivity during slack conditions, in agreement with Zhou et al. (2025); Zhou et al. (2019); Zhu et al. (2018).

As depicted in Section 3.1.2, the opposing C1 and C2 flows occurring during short periods at the onset of tidal transitions generate convergence zones, stagnation regions and interface vortices over the shoal. This transient phenomenon differs from systems characterized by persistent ebb- or flood-dominated channels (Van Veen et al., 2005). Their results also support our interpretation that bathymetric transitions promote shoal connectivity through convergence fronts and horizontal shear (Collignon and Stacey, 2012), consistent with the stagnation zones and interface vortices induced by brief opposing-flow phases. These transient structures also illustrate that inversion fronts are difficult to characterize, as local reversal patterns may be different from the system-scale behavior and produce distinct local and global signatures (Section 3.1 versus Section 3.2).

The asymmetry of tidal transitions reflects the shoal's dynamical control. During flood-to-ebb, C2 reverses earlier and the inversion front propagates progressively along the shoal slope. Un reverses first, followed by the sign change of $Ut^*\partial Un/\partial l$, while Ut adjusts later, producing a multi-stage reorganization. The ebb-to-flood transition is more abrupt: lower water levels amplify shoal control, Un and Ut reverse almost simultaneously, and $Ut^*\partial Un/\partial l$ adjusts immediately with no spatial propagation. These patterns structure the alternation between reinforcement and redistribution. Flood-to-ebb begin with C1→C2 reinforcement, followed by C2→C1 redistribution and downslope propagation before stabilizing into an ebb-dominated regime. Ebb-to-flood are shorter and occur in a single stage: weak exchange, a brief redistribution event, and immediate C1→C2 reinforcement.

The shoal therefore acts as a nonlinear filter within the C1-shoal-C2 system. It amplifies or attenuates the momentum transfer depending on tidal phase and the evolving three-dimensional flow structure. Its response depends on the magnitude and direction of the longitudinal gradient, the previous half-cycle tidal state (hysteresis effect), and the relative timing of Un and Ut reversals. When the lateral flux aligns with $\partial Un/\partial l$, the shoal amplifies cross-channel exchanges and reinforces the

dominance of the faster channel. When flux and gradient oppose each other, momentum transfer is redistributed toward the weaker channel (see Section 3.2.2, 3.2.3 and 3.3.3).

Inter-channel hierarchy and dominance remain difficult to characterize because they exhibit hysteresis signatures controlled by bathymetry, phase lags, and nonlinear momentum exchange. Dominance alternates between C1 and C2: C1 prevails during fully developed ebb and flood conditions, whereas C2 temporarily dominates during the flood-to-ebb transition. This apparent inversion results from differential channel inertia: during flood-to-ebb, C1 maintains momentum longer, while C2 reverses earlier and establishes the initial lateral gradients. By contrast, during ebb-to-flood, near-simultaneous reversals generate weaker and shorter-lived gradients, limiting the emergence of a clear dominance.

During tidal transitions, inter-channel flux temporarily becomes the dominant mechanism of connectivity, revealing a three-dimensional reorganization of the flow that cannot be inferred from depth-averaged velocities alone.

4.3. Three-dimensional reorganization through reinforcement-redistribution cycles

Within the asymmetric C1-shoal-C2 connectivity configuration, the flow undergoes a recurrent four-stage evolution during each tidal cycle: fully developed states, recirculation phases, opposing-flow transitions, and progressive reorganization toward a new fully developed state as described in Section 3.1.2. Throughout this sequence, each stage is characterized by internal micro-adjustments of the lateral circulation, involving shifts in flow structure, intensity, and momentum redistribution. This evolution, addressed in Section 3.3.3, highlights that inter-channel connectivity is dynamic and results from continuous three-dimensional reorganization of the flow.

On the shoal crest, the tangential convective inertial term reaches its maximum and alternates between reinforcement and redistribution, showing that transverse circulation continuously adapts to tidal forcing. This behavior results from the interaction between lateral flow direction and shoal slope, which periodically enhances or attenuates transverse momentum transfer. Even when the direction of U_t remains unchanged, the lateral cell oscillates around a quasi-equilibrium configuration (Fig. 9).

In asymmetric channel geometry, these micro-adjustments generate oscillations in the tangential convective inertial term, which reflects the transient reorganizations associated with the establishment and stabilization of the lateral structure (see Section 3.3.3). During early flood and early ebb (first-half period), the lateral cell is still forming and undergoes rapid adjustments driven by transverse shear and evolving streamline curvature. This constructive phase produces pronounced pulsations of the inertial term. As the tide progresses, the circulation becomes more organized, transverse gradients weaken, and the system approaches a quasi-steady configuration near mid-tide (Fig. 9).

Three mechanisms shape these reinforcement-redistribution cycles: (i) the transverse structure is intrinsically unsteady, with the lateral cell contracting, relaxing, and shifting as the shoal slope induces alternating compression and dilatation; (ii) variations in, driven by tidal acceleration and deceleration, reorganize the three-dimensional flow and transiently modify the strength and position of the lateral cell; and (iii) with upslope flows (C2→C1) enhancing transverse gradients and downslope flows (C1→C2) weakening them. Because the slope contains small-scale irregularities, this amplification also oscillates (Fig. 9).

The tangential convective inertial term confirms that reinforcement-redistribution cycles are structured and constitute key mechanisms for interpreting inter-channel connectivity. Strong shear zones act as drivers of inter-channel connectivity, while internal acceleration-deceleration reorganizes the lateral cell. Finally, lateral shear and tangential convective inertial term reveal that connectivity emerges from a interactions between gradient-driven exchange, internal flow

reorganization, and three-dimensional bathymetric constraints.

4.4. Limitations

This study focuses exclusively on barotropic hydrodynamic processes, which represent the dominant first-order controls of the circulation patterns analyzed at the locations of interest. Although baroclinic effects associated with salinity gradients are not explicitly simulated, previous results indicate that their contribution remains secondary the June 2009 simulation period, with mean errors over the study area, including and excluding salinity, remaining below 3% for each component of velocity. Therefore, the main mechanisms identified here, including phase lags, lateral exchanges, and three-dimensional flow reorganization, primarily reflect the response of the system to tidal forcing and bathymetric constraints. A complete assessment of salinity intrusion in and density-driven circulation extreme conditions (HW and low discharge) would nevertheless require variable-density modelling and the explicit representation of baroclinic processes, which constitute a natural extension for future work, particularly under conditions where density gradients may significantly influence circulation patterns.

The interpretation of the three-dimensional flow structure is inherently constrained by the vertical resolution of the sigma-coordinate system. Although the comparison between L2, L4, and L7 reveals a clear vertical organization of lateral dominance, these results should be interpreted as layer-integrated behaviors rather than a fully resolved vertical shear profile. As a result, the magnitude and sharpness of vertical shear may be smoothed by the water column discretization.

The bathymetry of the system is highly heterogeneous, and some shoal areas become partially emergent during parts of the tidal cycle. This intermittently alters the effective cross-sectional area and restricts lateral pathways, meaning, some of the mechanisms identified, particularly those involving nonlinear filtering by the shoal, may vary with the degree of submergence. These constraints should be considered when generalizing the results to other morphologies or tidal conditions.

In addition, the archipelagos are represented in the model, and there are effects associated with intertidal shoals (wet and dry). We assume these effects are limited and do not significantly alter the main flow. However, for the circulation in the immediate vicinity of the islands, these processes may be more important and would need further examination.

The simulated flow dynamics may be sensitive to the turbulence closure scheme used to represent vertical mixing. While the selected parameterization provides stable and physically consistent results, turbulence models could modify vertical shear and inter-layer exchanges, potentially influencing the distribution and persistence of lateral circulation.

Finally, the present analysis focuses on a limited range of river discharge conditions. Variations in freshwater input, particularly during high-flow events, may significantly change the structure of lateral circulation and inter-channel connectivity. Previous studies suggest that increased discharge can weaken lateral exchanges and induce longer recovery times compared with longitudinal flow adjustments (Yan et al., 2020). Extending the analysis to a broader range of hydraulic conditions would therefore provide further insight into the robustness and transferability of the mechanisms identified.

5. Conclusion

This study improves our understanding of inter-channel estuarine connectivity in the macrotidal St. Lawrence Estuary, a system characterized by strong three-dimensional asymmetry and pronounced spatial heterogeneity. This study demonstrates that connectivity is not solely controlled by fully developed flood or ebb conditions, but emerges most strongly during transient flow-reversal periods. During these short-lived transitions, longitudinal velocities weaken, transverse gradients intensify, and lateral exchanges across the shoal become the dominant

mechanism driving momentum redistribution between channels.

The three channels respond to tidal forcing with distinct phase lags, generating longitudinal asymmetries that interact with the shoal separating the navigation and intermediate channels. The south channel is more hydraulically isolated due to the Île-aux-Grues archipelago and the limited width of the junctions, resulting in reduced contributions to cross-channel exchanges that are mainly concentrated near the junction areas.

By resolving the propagation of tidal reversals, the development of lateral shear, and the three-dimensional structure of cross-channel exchanges, this study identifies mechanisms that have not previously been documented in deep, tidally dominated multichannel systems. In this framework, inter-channel connectivity emerges as a state-dependent process governed by the interplay between tidal phase lags, bathymetric confinement, and three-dimensional shear. Rather than reflecting a fixed morphological configuration, connectivity develops through a recurrent sequence of spatially phased tidal reversals during which channels alternately act as sources and sinks of momentum.

A key result is the dominant role of transitional phases, when longitudinal velocities weaken and transverse gradients intensify. These short-lived intervals activate lateral pathways, generate transient shear zones, and reorganize momentum distribution across the estuary. During these periods, lateral velocities can exceed longitudinal components, indicating that inter-channel exchanges are maximized under near-slack, non-equilibrium conditions rather than during fully developed flows.

Within this framework, the central shoal behaves as a nonlinear hydrodynamic filter. Depending on the alignment between longitudinal velocity gradients, transverse fluxes, and channel-specific phase lags, it selectively amplifies, redistributes, or suppresses momentum transfer. This filtering generates intermittent and directional connectivity, leading to alternating regimes of hydraulic coupling and partial isolation between deep channels over the tidal cycle.

Channel junctions act as dynamical nodes that regulate the propagation and synchronization of tidal reversals. By concentrating shear, flow opposition, and recirculation, they control inversion timing and determine whether the system behaves as a coherent hydraulic unit or as partially decoupled subsystems during transitions. In addition, shallow inter-channel zones play a key role in phase adjustment, facilitating the resynchronization of flows between channels.

These processes reveal a form of lateral self-organization in which channels do not respond independently to tidal forcing but evolve through alternating dominance regimes shaped by bathymetric constraints and the balance between longitudinal forcing and transverse exchanges.

Beyond site-specific insights, this work provides a process-based

framework for interpreting dynamic connectivity in multichannel estuaries. Because connectivity emerges from nonlinear, state-dependent interactions, morphological changes such as dredging, channel deepening, or sea-level rise can shift the system between reinforcement and redistribution regimes. Such sensitivity is consistent with observations in systems such as the Weser estuary, where channel deepening combined with sea-level rise has enhanced tidal propagation and increased salt intrusion (Kolb et al., 2022). Capturing these transitions requires fully three-dimensional approaches capable of resolving transient, hysteretic, and spatially confined exchanges that structure estuarine flow.

CRediT authorship contribution statement

M. Le Mouel: Writing – review & editing, Writing – original draft, Visualization, Validation, Software, Methodology, Investigation, Data curation. **O. Lelandais:** Writing – review & editing, Visualization, Investigation. **A. Hammouti:** Writing – review & editing, Supervision, Software, Resources, Methodology. **D. Pham Van Bang:** Writing – review & editing, Supervision, Resources, Methodology, Funding acquisition.

Declaration of competing interest

The authors declare the following financial interests/personal relationships which may be considered as potential competing interests: Damien Pham Van Bang reports a relationship with Natural Sciences and Engineering Research Council of Canada that includes: funding grants. If there are other authors, they declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

Acknowledgements

The authors wish to thank Pascal Matte for providing the model validation ADCP dataset. The authors gratefully acknowledge the support of the Natural Sciences and Engineering Research Council of Canada (RGPIN-2024-03882), the Fonds de Recherche du Québec, and the Agence Nationale de la Recherche (FRQ-ANR, project EMPHASE, No. 280266), whose funding made this work possible. We also extend our sincere thanks to Compute Canada (ESPRILE No. 4516 and GAIA No. 5625) for providing access to high-performance computing resources. The first author further acknowledges the support of the FRQ and the CCCC through the Climate Action Scholarship (No. 364826). The authors also acknowledge the anonymous reviewers for their valuable comments and suggestions, which helped improve the manuscript.

Appendix: 3D. Hydrodynamic Equations and Turbulence Closure Models

Navier-Stokes Equations

$$\frac{\partial U}{\partial x} + \frac{\partial V}{\partial y} + \frac{\partial W}{\partial z} = 0 \quad (1)$$

$$\frac{\partial U}{\partial t} + U \frac{\partial U}{\partial x} + V \frac{\partial U}{\partial y} + W \frac{\partial U}{\partial z} = -\frac{1}{\rho} \frac{\partial p}{\partial x} + \nu \Delta U + F_x \quad (2)$$

$$\frac{\partial V}{\partial t} + U \frac{\partial V}{\partial x} + V \frac{\partial V}{\partial y} + W \frac{\partial V}{\partial z} = -\frac{1}{\rho} \frac{\partial p}{\partial y} + \nu \Delta V + F_y \quad (3)$$

$$\frac{\partial W}{\partial t} + U \frac{\partial W}{\partial x} + V \frac{\partial W}{\partial y} + W \frac{\partial W}{\partial z} = -\frac{1}{\rho} \frac{\partial p}{\partial z} + \nu \Delta W + F_z \quad (4)$$

where (U, V, W) are the Reynolds averaged component of velocity (m s^{-1}), ρ (kg m^{-3}) the density of water, p (Pa) the pressure, g ($=9.81 \text{ m s}^{-2}$) the

gravitational acceleration, ν ($\text{m}^2 \text{s}^{-1}$) the effective viscosity, (F_x, F_y, F_z) the external Force component (m s^{-2}).

Pressure and density formulation

$$p = p_{atm} + \rho_0 g (Z_s - z) + \rho_0 g \int_z^{Z_s} \frac{\Delta \rho}{\rho_0} d\zeta \quad (5)$$

$$\rho = \rho_0 [1 - (7(T - T_{ref})^2 - 750S)10^{-6}] \quad (6)$$

where p_{atm} (Pa) is the atmospheric pressure, Z_s the free-surface elevation (m), ρ_0 the density of pure water (kg m^{-3}), $\Delta \rho$ the density difference between the fluid and pure water, T ($^{\circ}\text{C}$) the temperature, S (g L^{-1}) the salinity and $T_{ref} = 4$ $^{\circ}\text{C}$ the reference temperature. The water temperature is fixed at $T = 18$ $^{\circ}\text{C}$ and the salinity at $S = 0$ g L^{-1} throughout the entire domain and for the whole simulation period.

Reynolds decomposition

The Navier-Stokes equations describe the instantaneous evolution of the velocity field $u(x,t)$ and the pressure $p(x,t)$ in an incompressible fluid. They can be written as follows:

$$\nabla \cdot u = 0 \quad (7)$$

$$\rho \left(\frac{\partial u_i}{\partial t} + u_j \frac{\partial u_i}{\partial x_j} \right) = - \frac{\partial p}{\partial x_i} + \mu \nabla^2 u_i + \rho g_i \quad (8)$$

where μ is the dynamic viscosity ($\text{kg m}^{-1} \text{s}^{-1}$).

Instantaneous quantities are decomposed into mean and fluctuating components:

$$u_i = \bar{u}_i + u'_i \quad (9)$$

$$p = \bar{p} + p' \quad (10)$$

Where $\bar{u}_i$ (m s^{-1}) is the mean velocity and, u'_i (m s^{-1}) is the turbulent velocity fluctuation.

Applying this decomposition to the instantaneous Navier-Stokes equations yields the RANS system:

$$\rho \left(\frac{\partial \bar{u}_i}{\partial t} + \bar{u}_j \frac{\partial \bar{u}_i}{\partial x_j} \right) = - \frac{\partial \bar{p}}{\partial x_i} + \mu \nabla^2 \bar{u}_i - \frac{\partial}{\partial x_j} (\rho \bar{u}'_i u'_j) + \rho g_i \quad (11)$$

The Boussinesq hypothesis is introduced, linking the turbulent stresses to an effective viscous tensor:

$$\bar{u}'_i u'_j = \nu_t \left(\frac{\partial \bar{u}_i}{\partial x_j} + \frac{\partial \bar{u}_j}{\partial x_i} \right) \quad (12)$$

where ν_t ($\text{m}^2 \text{s}^{-1}$) is the turbulent viscosity.

k-ε Turbulence Model

The horizontal turbulent viscosity ν_t ($\text{m}^2 \text{s}^{-1}$) is expressed as:

$$\nu_t = C_\mu \frac{k^2}{\varepsilon} \quad (11)$$

With:

$$k = \frac{1}{2} \bar{u}'_i u'_i \quad (12)$$

$$\varepsilon = \nu \frac{\partial \bar{u}'_i}{\partial x_j} \frac{\partial \bar{u}'_i}{\partial x_j} \quad (13)$$

The first characteristic scale, k , corresponds to the turbulent kinetic energy ($\text{m}^2 \text{s}^{-2}$). The second, ε , represents the turbulent dissipation rate ($\text{m}^2 \text{s}^{-3}$). The transport equations for k and ε are written as:

$$\frac{\partial k}{\partial t} + U \frac{\partial k}{\partial x} + V \frac{\partial k}{\partial y} + W \frac{\partial k}{\partial z} = \frac{\partial}{\partial x} \left(\frac{\nu_t}{\sigma_k} \frac{\partial k}{\partial x} \right) + \frac{\partial}{\partial y} \left(\frac{\nu_t}{\sigma_k} \frac{\partial k}{\partial y} \right) + \frac{\partial}{\partial z} \left(\frac{\nu_t}{\sigma_k} \frac{\partial k}{\partial z} \right) + P - G - \varepsilon \quad (14)$$

$$\frac{\partial \varepsilon}{\partial t} + U \frac{\partial \varepsilon}{\partial x} + V \frac{\partial \varepsilon}{\partial y} + W \frac{\partial \varepsilon}{\partial z} = \frac{\partial}{\partial x} \left(\frac{\nu_t}{\sigma_\varepsilon} \frac{\partial \varepsilon}{\partial x} \right) + \frac{\partial}{\partial y} \left(\frac{\nu_t}{\sigma_\varepsilon} \frac{\partial \varepsilon}{\partial y} \right) + \frac{\partial}{\partial z} \left(\frac{\nu_t}{\sigma_\varepsilon} \frac{\partial \varepsilon}{\partial z} \right) + C_{1\varepsilon} \frac{\varepsilon}{k} [P + (1 - C_{3\varepsilon})G] - C_{2\varepsilon} \frac{\varepsilon^2}{k} \quad (15)$$

where P ($\text{m}^2 \text{s}^{-3}$) is the turbulent kinetic energy production term and G ($\text{m}^2 \text{s}^{-3}$) is the buoyancy-related source term of turbulent kinetic energy.

$$P = \nu_t \left(\frac{\partial \bar{u}_i}{\partial x_j} + \frac{\partial \bar{u}_j}{\partial x_i} \right) \frac{\partial \bar{u}_i}{x_j} \quad (16)$$

$$G = - \frac{\nu_t}{Pr_t} \frac{g}{\rho} \frac{\partial \rho}{\partial z} \quad (18)$$

In TELEMAC-3D, the parameters C_μ , $C_{1\epsilon}$, $C_{2\epsilon}$, $C_{3\epsilon}$, σ_k and σ_ϵ are experimentally determined constants, with respective values of 0.09, 1.44, 1.92, 1, 1.0 et 1.3 (Hervouet, 2007).

Mixing-Length Turbulence Model

The vertical turbulent viscosity ν_t ($\text{m}^2 \text{s}^{-1}$) is expressed as (Hervouet, 2007; Prandtl, 1925):

$$\nu_t = L_m^2 \sqrt{2 D_{ij} \bar{D}_{ij}} \quad (19)$$

Where L_m is the mixing length (m) and $\bar{D}$ is the strain-rate tensor of the mean flow (Hervouet, 2007; Prandtl, 1925):

$$D_{ij} = \frac{1}{2} \left(\frac{\partial \bar{u}_i}{\partial x_j} + \frac{\partial \bar{u}_j}{\partial x_i} \right) \quad (20)$$

The mixing length L_m follows the classical Prandtl formulation (Prandtl, 1925):

$$L_m = \kappa z \quad \text{si } z < 0, 2h \quad L_m = 0, 2\kappa z \quad \text{si } z \geq 0, 2h \quad (21)$$

with $\kappa = 0.41$ the von Karman constant, z the distance from the bed (m), and h the water depth (m).

σ -Coordinate System

The vertical discretization uses a terrain-following σ -coordinate system (Hervouet, 2007)

$$x^* = x \quad (22)$$

$$y^* = y \quad (23)$$

$$\sigma = \frac{z - Zb}{Zs - Zb} \quad (24)$$

$$Zs = Zb + h + \eta \quad (25)$$

where Zb is the bed elevation (m), Zs the free-surface elevation (m), h the water depth (m), η the free-surface variation (m), z the vertical coordinate, (x, y, z) the Cartesian coordinate system, and (x^*, y^*, σ) the σ -transformed coordinate system, with σ ranging from 0 to 1.

Data availability

Data will be made available on request.

References

- Abair, S., Hammouti, A., Noman, J., Daniel, S., Pham Van Bang, D., 2025. Strengthening coastal flood forecasting through event-based data capitalization : a case study from the December 2022 storm surge in Québec City. *Nat. Hazards* 122 (1), 30. <https://doi.org/10.1007/s11069-025-07780-5>.
- Bilal, A., Xie, Q., Zhai, Y., 2020. Flow, sediment, and morpho-dynamics of river confluence in tidal and non-tidal environments. *J. Mar. Sci. Eng.* 8 (8), 8. <https://doi.org/10.3390/jmse8080591>.
- Buschman, F.A., Van Der Vegt, M., Hoitink, A.J.F., Hoekstra, P., 2013. Water and suspended sediment division at a stratified tidal junction. *J. Geophys. Res. Oceans* 118 (3), 1459–1472. <https://doi.org/10.1002/jgrc.20124>.
- Chu, N., Liu, G., Xu, J., Yao, P., Du, Y., Liu, Z., Cai, Z., 2022. Hydrodynamical transport structure and lagrangian connectivity of circulations in the pearl river Estuary. *Front. Mar. Sci.* 9. <https://doi.org/10.3389/fmars.2022.996551>.
- Collignon, A.G., Stacey, M.T., 2012. Intratidal dynamics of fronts and lateral circulation at the shoal-channel interface in a partially stratified Estuary. *J. Phys. Oceanogr.* 42 (5). <https://doi.org/10.1175/JPO-D-11-065.1>, 869–883.
- Dai, W., Bilal, A., Xie, Q., Ahmad, I., Joshi, I., 2020. Numerical modeling for hydrodynamics and near-surface flow patterns of a tidal confluence. *J. Coast Res.* 36 (2), 295–312. <https://doi.org/10.2112/JCOASTRES-D-19-00058.1>.
- DFO Fisheries and Oceans Canada, 2009. Tides, currents, and water levels [Jeu de données]. <https://www.tides.gc.ca/en>.
- Dionne, J.-C., 2002. Érosion récente du schorre supérieur à Sainte-Anne-de-Beaupré, Québec. *Géogr. Phys. Quaternaire* 54 (1), 69–89. <https://doi.org/10.7202/004792ar>.
- Geyer, W.R., Ralston, D.K., Chen, J.-L., 2020. Mechanisms of exchange flow in an Estuary with a narrow, deep channel and wide, shallow shoals. *J. Geophys. Res. Oceans* 125 (12), e2020JC016092. <https://doi.org/10.1029/2020JC016092>.
- Gundlach, J., Zorndt, A., Prooijen, B. C. van, Wang, Z.B., 2021. Two-channel system dynamics of the outer weser estuary—A modeling study. *J. Mar. Sci. Eng.* 9 (4). <https://doi.org/10.3390/jmse9040448>.
- He, C., Yin, Z.-Y., Stocchino, A., Wai, O.W.H., 2023. Generation of macro-vortices in estuarine compound channels. *Front. Mar. Sci.* 10. <https://doi.org/10.3389/fmars.2023.1082506>.
- Hervouet, J.-M., 2007. *Hydrodynamics of Free Surface Flows : Modelling with the Finite Element Method*. Wiley.
- Hiatt, M., Addink, E.A., Kleinhans, M.G., 2022. Connectivity and directionality in estuarine channel networks. *Earth Surf. Process. Landf.* 47 (3), 807–824. <https://doi.org/10.1002/esp.5286>.
- Hiatt, M., Sonke, W., Addink, E.A., van Dijk, W.M., van Krevel, M., Ophelders, T., Verbeek, K., Vlamming, J., Speckmann, B., Kleinhans, M.G., 2020. Geometry and topology of Estuary and braided river channel networks automatically extracted from topographic data. *J. Geophys. Res. Earth Surf.* 125 (1), e2019JF005206. <https://doi.org/10.1029/2019JF005206>.
- Hlaing, N.O., Azhikodan, G., Yokoyama, K., 2024. Effect of topography and tidal variations on the flow dynamics in multi-channel tidal river estuaries. *Estuar. Coast Shelf Sci.* 308, 108923. <https://doi.org/10.1016/j.ecss.2024.108923>.
- Iwantoro, A.P., van der Vegt, M., Kleinhans, M.G., 2020. Morphological evolution of bifurcations in tide-influenced deltas. *Earth Surf. Dyn.* 8 (2), 413–429. <https://doi.org/10.5194/esurf-8-413-2020>.
- Kolb, P., Zorndt, A., Burchard, H., Gräwe, U., Kösters, F., 2022. Modelling the impact of anthropogenic measures on saltwater intrusion in the Weser estuary. *Ocean Sci.* 18 (6), 1725–1739. <https://doi.org/10.5194/os-18-1725-2022>.

- Le Mouel, M., Matte, P., Hammouti, A., Pham-Van-Bang, D., 2025. Investigation of 3D circulation and secondary flows in the st. Lawrence fluvial Estuary at a tidal junction. *Estuar. Coast Shelf Sci.* 313, 109058. <https://doi.org/10.1016/j.ecss.2024.109058>.
- Lerczak, J.A., Geyer, W.R., 2004. Modeling the lateral circulation in straight, stratified estuaries. *J. Phys. Oceanogr.* 34 (6), 1410–1428. [https://doi.org/10.1175/1520-0485\(2004\)034%3C1410:MTLCIS%3E2.0.CO;2](https://doi.org/10.1175/1520-0485(2004)034%3C1410:MTLCIS%3E2.0.CO;2).
- Matte, P., Secretan, Y., Morin, J., 2014. Temporal and spatial variability of tidal-fluvial dynamics in the st. Lawrence fluvial estuary : an application of nonstationary tidal harmonic analysis. *J. Geophys. Res. Oceans* 119 (9), 5724–5744. <https://doi.org/10.1002/2014JC009791>.
- Matte, P., Secretan, Y., Morin, J., 2014a. Quantifying lateral and intratidal variability in water level and velocity in a tide-dominated river using combined RTK GPS and ADCP measurements. *Limnol Oceanogr. Methods* 12 (5), 281–302. <https://doi.org/10.4319/lom.2014.12.281>.
- Matte, P., Secretan, Y., Morin, J., 2017a. Hydrodynamic modeling of the St. Lawrence fluvial Estuary. I : model setup, calibration, and validation. *J. Waterw. Port. Coast. Ocean Eng.* 143 (5), 04017010. [https://doi.org/10.1061/\(ASCE\)WW.1943-5460.0000397](https://doi.org/10.1061/(ASCE)WW.1943-5460.0000397).
- Matte, P., Secretan, Y., Morin, J., 2017b. Hydrodynamic modeling of the St. Lawrence fluvial Estuary. II : reproduction of spatial and temporal patterns. *J. Waterw. Port. Coast. Ocean Eng.* 143 (5), 04017011. [https://doi.org/10.1061/\(ASCE\)WW.1943-5460.0000394](https://doi.org/10.1061/(ASCE)WW.1943-5460.0000394).
- Matte, P., Secretan, Y., Morin, J., 2019. Drivers of residual and tidal flow variability in the st. Lawrence fluvial estuary : influence on tidal wave propagation. *Cont. Shelf Res.* 174, 158–173. <https://doi.org/10.1016/j.csr.2018.12.008>.
- Morin, J., Boudreau, P., Secretan, Y., Leclerc, M., 2000. Pristine Lake saint-François, St. Lawrence River : hydrodynamic simulation and cumulative impact. *J. Great Lake Res.* 26 (4), 384–401. [https://doi.org/10.1016/S0380-1330\(00\)70702-7](https://doi.org/10.1016/S0380-1330(00)70702-7).
- Prandtl, L., 1925. Über die ausgebildete Turbulenz. *Z. Angew. Math. Mech.* 5, 136.
- Ragno, N., Tambroni, N., Bolla Pittaluga, M., 2020. Effect of small tidal fluctuations on the stability and equilibrium configurations of bifurcations. *J. Geophys. Res. Earth Surf.* 125 (8), e2020JF005584. <https://doi.org/10.1029/2020JF005584>.
- Ralston, D.K., Geyer, W.R., Warner, J.C., 2012. Bathymetric controls on sediment transport in the Hudson River estuary : lateral asymmetry and frontal trapping. *J. Geophys. Res. Oceans* 117 (C10). <https://doi.org/10.1029/2012JC008124>.
- Ross, L., Huguenard, K., Sottolichio, A., 2019. Intratidal and fortnightly variability of vertical mixing in a macrotidal Estuary : the Gironde. *J. Geophys. Res. Oceans* 124 (4), 2641–2659. <https://doi.org/10.1029/2018JC014456>.
- Sandbach, S.D., Nicholas, A.P., Ashworth, P.J., Best, J.L., Keevil, C.E., Parsons, D.R., Prokocki, E.W., Simpson, C.J., 2018. Hydrodynamic modelling of tidal-fluvial flows in a large river estuary. *Estuar. Coast Shelf Sci.* 212, 176–188. <https://doi.org/10.1016/j.ecss.2018.06.023>.
- Shiono, K., Knight, D.W., 1991. Turbulent open-channel flows with variable depth across the channel. *J. Fluid Mech.* 222 (1), 617. <https://doi.org/10.1017/S0022112091001246>.
- Simons, R.D., Monismith, S.G., Saucier, F.J., Johnson, L.E., Winkler, G., 2010. Modelling stratification and baroclinic flow in the estuarine transition zone of the st. Lawrence estuary. *Atmos.-Ocean* 48 (2), 132–146. <https://doi.org/10.3137/OC316.2010>.
- Van Veen, J., Van Der Spek, A.J.F., Stive, M.J.F., Zitman, T., 2005. Ebb and flood channel systems in the Netherlands tidal Waters¹. *J. Coast Res.* 216, 1107–1120. <https://doi.org/10.2112/04-0394.1>.
- Wang, J., de Swart, H.E., Dijkstra, Y.M., 2021. Dependence of tides and river water transport in an estuarine network on river discharge, tidal forcing, geometry and sea level rise. *Cont. Shelf Res.* 225, 104476. <https://doi.org/10.1016/j.csr.2021.104476>.
- Xiao, C., Kawanisi, K., Torigoe, R., Al Sawaf, M.B., 2021. Mapping tidal current and salinity at a shallow tidal channel junction using the fluvial acoustic tomography system. *Estuar. Coast Shelf Sci.* 258, 107440. <https://doi.org/10.1016/j.ecss.2021.107440>.
- Yan, Y., Song, D., Bao, X., Ding, Y., 2020. The response of lateral flow to peak river discharge in a macrotidal Estuary. *Water* 12 (12). <https://doi.org/10.3390/w12123571>.
- Yuan, H., Tian, H., Huang, W., Zhang, Z., Hu, R., 2025. Numerical investigation on the hydraulic characteristics of curved open channel confluence with different main channel widths. *Phys. Fluids* 37 (10), 105106. <https://doi.org/10.1063/5.0281510>.
- Zhou, J., Stacey, M.T., Holleman, R.C., Nuss, E., Senn, D.B., 2020. Numerical investigation of baroclinic channel-shoal interaction in partially stratified estuaries. *J. Geophys. Res. Oceans* 125 (4), e2020JC016135. <https://doi.org/10.1029/2020JC016135>.
- Zhou, S., Zhu, C., Lin, J., Xie, W., Zhang, N., Guo, L., He, Q., 2025. Multi-asymmetry on residual sediment transport in the branching channels of the Yangtze Estuary. *J. Hydrol.* 655, 132947. <https://doi.org/10.1016/j.jhydrol.2025.132947>.
- Zhou, Z., Ge, J., Wang, Z.B., van Maren, D.S., Ma, J., Ding, P., 2019. Study of lateral flow in a stratified tidal channel-shoal system : the importance of intratidal salinity variation. *J. Geophys. Res. Oceans* 124 (9), 6702–6719. <https://doi.org/10.1029/2019JC015307>.
- Zhu, L., He, Q., Shen, J., 2018. Modeling lateral circulation and its influence on the along-channel flow in a branched estuary. *Ocean Dyn.* 68 (2), 177–191. <https://doi.org/10.1007/s10236-017-1114-8>.