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The Importance of Intratidal Salinity Variation**

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1 **Study of lateral flow in a stratified tidal channel-shoal system: the importance of**
2 **intra-tidal salinity variation**

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15 **Key Points:**

- 16 • Intra-tidal salinity variation (ISV) and a strong lateral flow in a channel-shoal system
17 were identified by an integrated tripod system.
- 18 • A 3D hydrodynamic model has been developed to analyze the relationship between ISV
19 and the lateral flow, revealing that lateral flows were primarily driven by the salinity-
20 induced pressure gradient.
- 21 • Lateral flows can be generated or enhanced by human interventions such as groyne fields.
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31 **Abstract**

32 Lateral flow significantly contributes to the near-bottom mass transport of salinity in a channel-
33 shoal system. In this study, an integrated tripod system was deployed in the transition zone of a
34 channel-shoal system of the Changjiang Estuary (CE), China, to observe the near-bottom physics
35 with high temporal/spatial resolution, particularly focusing on the lateral-flow-induced mass
36 transport. These in-situ observations revealed a small-scale salinity fluctuation around low water
37 slack during moderate and spring tidal conditions. A simultaneous strong lateral current was also
38 observed, which was responsible for this small-scale fluctuation. A high-resolution unstructured-
39 grid Finite-Volume Community Ocean Model (FVCOM) has been applied for the CE to better
40 understand the mechanism of this lateral flow and its impact on salinity transport. The model
41 results indicate that a significant southward near-bed shoal-to-channel current is generated by the
42 salinity-driven baroclinic pressure gradient. This lateral current affects the salinity transport
43 pattern and the residual current in the cross-channel direction. Cross-channel residual current
44 shows a two-layer structure in the vertical, especially in the intermediate tide when the lateral
45 flow notably occurred. Both observation and model results indicate that near-bottom residual
46 transport of water moved consistently southward (shoal to channel). Mechanisms for this intra-
47 tidal salinity variation (ISV) and its implications can be extended to other estuaries with similar
48 channel-shoal features.

49

50 **1 Introduction**

51 As transition zones between riverine and marine environments, estuaries experience a
52 wide range of physical, chemical and biological processes, e.g. suspended material transport,
53 chemical reaction of dissolved ions, variation of biological productivity. These processes
54 influence not only ecology but also have socio-economic impacts because many large cities are
55 located in the vicinity of estuaries (Woodroffe et al., 2006).

56 Salinity is a significant driver for sediment trapping, e.g. sediment accumulation at the
57 front of salt wedge where the convergence of fresh and saline water occurs (Postma, 1967). The
58 upstream limit of salt intrusion is an indicator for location of estuary turbidity maximum (Dyer,
59 1986), the dynamics of which in itself strongly influences the navigability. Moreover, salinity
60 also has an effect on ecology, e.g. its impact on temporal distribution of planktons (Dube et al.,
61 2010).

62 The long-term, large-scale salt balance is primarily the result of landward transport by
63 estuarine circulation and tidal dispersion (Bowen, 2003; Fischer, 1976) balancing seaward
64 transport by riverine residual flow (Dronkers, 2018). These salt transport mechanisms depend on
65 the longitudinal salinity gradient and (in combination with hydrodynamic mixing) give rise to
66 vertical salinity gradients. Vertical salinity gradients may cause salinity stratification which
67 controls the intensity of momentum exchanges from the surface to the bottom (Simpson et al.,
68 1990).

69 In addition to these longitudinal and vertical processes, lateral processes in estuaries also

70 have effects on mass transport. Cross-channel (lateral) currents are relatively small (about 10%
71 in magnitude) compared to the along-channel currents (Lerczak & Geyer, 2004). However, their
72 influences on dynamics of estuaries may be significant by generating considerable cross-channel
73 gradients in salt, turbidity and other constituents.

74 Lerczak & Geyer (2004) set up a model for an idealized straight estuary. Their results
75 suggest that lateral circulation is stronger during flood tides than ebb tides. This asymmetry is
76 caused by the interaction between lateral circulation, stratification and differential advection by
77 along-channel tidal currents. The flood-ebb asymmetry of lateral flow patterns may lead to an
78 asymmetric cross-sectional shape of a straight channel. However, in a channel which already is
79 asymmetrically shaped, lateral flow will also be affected by bathymetric features, as well as
80 hydraulic structures. Huijts et al. (2006) studied the mechanisms for lateral sediment entrapment
81 using an idealized model. They examined mechanisms that could lead to the sediment
82 accumulation, including Coriolis forcing and lateral density gradients. They found that because
83 of the difference of along-channel flow velocity from bed to surface, the intensity of the Coriolis
84 deflection varies from bed to surface. Therefore, vertical circulation in the lateral direction can
85 be induced by the Coriolis deflection. Fugate et al. (2007) conducted an observation in upper
86 Chesapeake Bay, USA, focusing on the impact of lateral dynamics on sediment transport. Using
87 a lateral momentum balance they observed that stronger cross-channel circulation by rotational
88 effects (Coriolis and channel curvature) is larger during the ebb. A modelling study by Zhu et al.
89 (2018) in a strongly anthropogenically impacted estuary (the Changjiang Estuary) suggest that
90 lateral circulations are strong near the main navigation channel, and peak close to slack tide
91 conditions. This has great implications for siltation rates in the channel, in which more than 80
92 million m³/year (by 2011) needs to be dredged annually (Wang et al., 2015). This fuels a great
93 interest in the dynamics of lateral flows in general, but in the CE in particular.

94 Cross-sectional flows influence the salinity, leading to an intra-tidal salinity variation
95 (ISV). Many studies have identified the salinity's response to the longitudinal and vertical
96 process (Bowen, 2003; Dronkers, 2018; Dyer, 1986; Simpson et al., 1990). However, few
97 observations directly relate salinity fluctuations to lateral flows. Yet, salinity variations may
98 provide an indicator for lateral flows, which is much easier to measure than transverse flows
99 themselves. The objective of this study is to detect evidence for lateral flow using a combination
100 of flow velocity and salinity observations, and quantitatively analyze and explain their driving
101 mechanisms and effects. This paper is organized as follows. Section 2 introduces the study area,
102 methods for observation and data analysis, and the numerical model. Section 3 presents the main
103 results of observations and simulations, focusing on formation of ISV and the mechanism for the
104 lateral flow. Section 4 gives discussion on momentum balance, stratification and mixing, effects
105 of dikes and groynes and the implication of ISV. Main conclusions are summarized in section 5.
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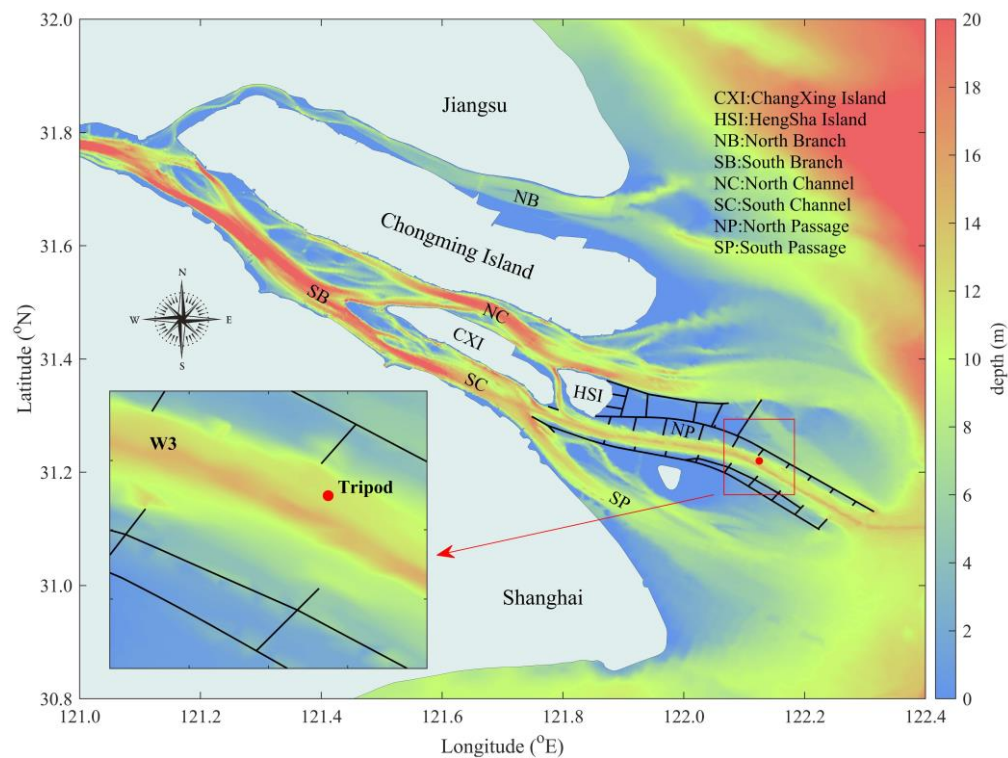
107 2 Study site, observational and numerical method

108 2.1 Study site

109 Field observations were carried out in one of the four outlets of the CE (Fig. 1). The
 110 discharge of the Changjiang River, measured at the Datong Gauging Station, is approximately
 111 40000 m³/s in wet season, about 10000 m³/s in dry season. The mean and maximum tidal ranges
 112 are approximately 2.7 m and 4.6 m, respectively, measured at Zhongjun Gauging Station.

113 The CE is characterized as a channel-shoal system with multiple outlets and shallow
 114 shoals. The South Channel (SC), North Passage (NP) and South Passage (SP) are the major tidal
 115 channels in the turbidity maximum zone of the CE. The NP is deepened and protected, resulting
 116 in the so-called Deepwater Navigation Channel (DNC), aiming at improving the shipping
 117 capacity in and out of the CE. The protection works include two 50 km long dikes parallel to the
 118 flow, to which 19 groynes perpendicular to the flow are attached (Fig. 1).

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122 Figure 1: Bathymetry of the Changjiang Estuary and adjacent regions. The black lines in the river mouth indicate
 123 dikes and groynes around the North Passage. The red dot indicates the measuring site. W3 is the bending corner of
 124 the main channel.

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126 The estuarine turbidity maximum of the CE is located in the North Passage (NP),
 127 resulting in strong sedimentation rates (Ge et al., 2015, 2018; Liu et al., 2011). Many groynes
 128 have been buried since the construction of the DNC. This has led to shallow shoals in the groyne

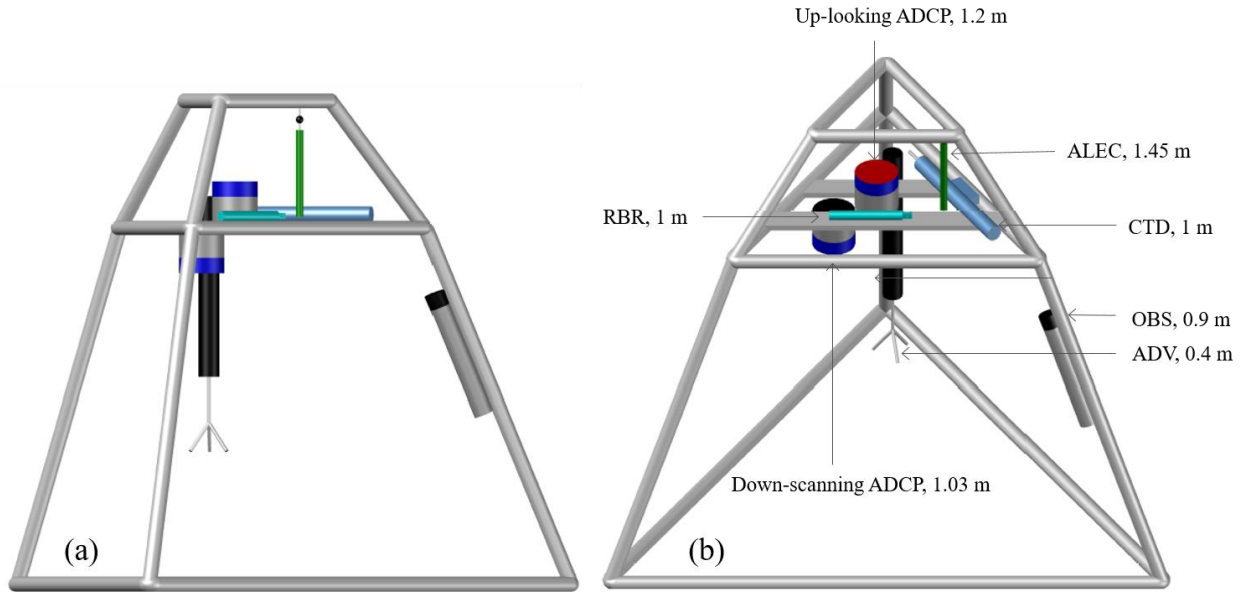
129 region and a deeper main channel in the middle of the NP, forming a typical channel-shoal
130 system. Cross-channel flows have been observed in the NP through observational and numerical
131 studies (Liu et al., 2011; Zhu et al., 2018). The NP is also the main mixing front of freshwater
132 and saline water, resulting in strong horizontal salinity gradient and vertical stratification (Ge et
133 al., 2012, 2018; Wu et al., 2012).

134 2.2 Observation methods

135 The middle section of the NP, located in the center of the turbidity maximum, was
136 selected to conduct our observations (red rectangle in Fig. 1). A tripod system was deployed on
137 the north side of the main channel, an area with pronounced saltwater intrusion and a strongly
138 stratified water column (Ge et al., 2018). The observation site was in the middle of the shallow
139 shoal and the main channel to observe lateral flows between the shoal and the deep channel.

140 The tripod system was designed to measure flow velocity and direction, salinity,
141 suspended sediment concentration and temperature near the bottom. To achieve this goal, the
142 tripod integrated multiple instruments (Fig. 2). An upward-looking 600 kHz RDI Acoustic
143 Doppler Current Profilers (ADCP-up) was mounted 1.2 m above the sea bed (abbreviated as
144 ‘mab’) with a resolution of 0.5 m for each cell. A downward-looking 1200 kHz RDI ADCP
145 (ADCP-down) was placed 1.03 mab to measure velocities from 0.2 to 0.7 mab at high resolution
146 (0.1 m) near the bed. Based on earlier observations (Liu et al., 2011), the ADCP with 1200 kHz
147 sensor frequency works well in the near-bottom area under typical high sediment concentration
148 for this area. A Nortek Acoustic Doppler Vector was mounted at 0.4 mab to measure the near-
149 bed current velocities at a sampling frequency of 16 Hz, which means valid data were collected
150 at the height of about 0.25 mab. The tripod system also included a Point Current Meter (ALEC,
151 JFE ALEC CO., LTD, JAPAN) at 1.45 mab to obtain flow velocities in the blanking range of
152 ADCP-up, a Tide/wave Logger (RBR, RBR Ltd., Canada) at 1.0 mab to record wave conditions
153 and a Conductivity, Temperature, and Pressure Recorder (CTD, Sea-Bird Electronics, Inc., USA)
154 at 1.0 mab to record temperature and salinity continuously. An Optical Backscattering Sensor
155 (OBS, D&A Instruments CO, type: 3A, USA) was also fixed to the side edge of the tripod.
156 Detailed configurations of instruments installed on the tripod such as burst interval, sampling
157 duration and sampling frequency are listed in Table. 1.

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Figure 2: Side-view and top-view of the tripod system.

Table 1: Instruments mounted on the tripod and their sampling configurations.

Instrument deployed	Distance above bed (m)	Sampling interval (min)	Sampling configuration	Survey parameter
ADCP-up	1.2	/	120s	Profile velocity
ADCP-down	1.03 (down)	/	120s	Profile velocity
ADV	0.25	10	16Hz (every first 70s)	Near-bed velocity
RBR	1	10	4Hz	Wave conditions
ALEC	1.45	2	0.2Hz (every first 50s)	Velocity
OBS	0.9	/	100s	Salinity, temperature, turbidity, pressure
CTD	1	/	120s	Salinity, temperature, pressure

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The tripod was deployed on December 6, 2016 (abbreviated as 12/06), in the dry season of the year. The observations lasted for about 12 days, and the tripod was recovered on 12/18, 2016. The 12 days covered a whole period from the neap tide to the spring tide. The neap tide was from 12/07 to 12/09, and the spring tide was on 12/14 - 12/16.

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2.3 Numerical Model

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To resolve the irregular geometries of the channel, shoals and submerged/exposed dikes and groynes, a hydrodynamic numerical model based on the Finite-Volume Community Ocean Model has been applied for the CE (CE-FVCOM). FVCOM is a three-dimension, unstructured-grid coastal ocean model. A triangle mesh is used horizontally, and σ -coordinates in the vertical. A generalized dike-groyne module is implemented in FVCOM to resolve submerged and

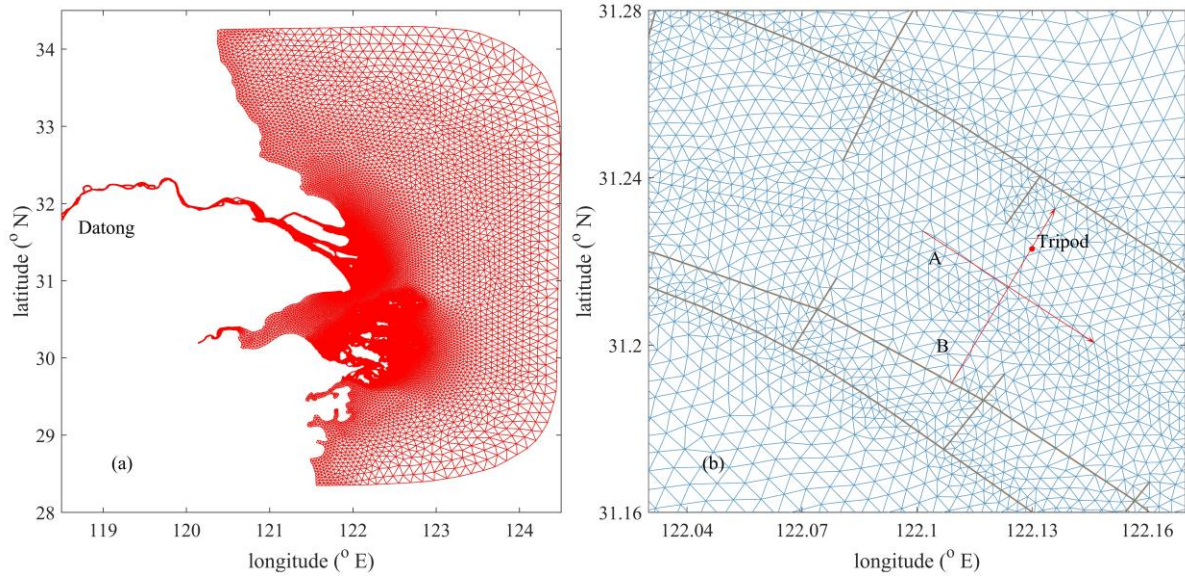
176 exposed man-made coastal and offshore engineering works (Ge et al., 2012). Previous
177 observations showed that the water column in the NP is occasionally strongly salinity stratified,
178 and opposite flows form a two-layer flow structure in the vertical (Ge et al., 2013, 2018). As a
179 mode split model, the adjustment between 3-D internal mode and 2-D external mode could be
180 problematic for capturing this two-layer flow structure (Lai et al., 2010). Therefore, we use a
181 semi-implicit scheme which is capable of well simulating the two-layer-structure flows.

182 The geographically unstructured mesh of FVCOM covers the whole CE as well as the
183 inner shelf of the East China Sea, Hangzhou Bay, and Zhoushan Archipelago (Fig. 3). Different
184 from the original mesh configuration (Ge et al., 2012, 2013), the river boundary in this model is
185 extended to ~ 600 km upstream Datong Gauging Station to better resolve the river-estuary
186 interaction (Fig. 3a). It provides flexible resolution from the open boundary in the inner shelf to
187 channels and shoals at the river mouth. The horizontal resolution of this model is down to ~200
188 m in the channel (Fig. 3b).

189 The model is forced by 8 major astronomical tidal constituents specified at the open
190 boundaries, including four diurnal tides (K_1 , O_1 , P_1 and Q_1), four semi-diurnal tides (M_2 , S_2 , N_2
191 and K_2). The data for the tidal constituent sources are from TPXO 8 (Egbert & Erofeeva, 2002).
192 Daily river discharge of the CJ (data source: www.cjh.com.cn) is considered at the upstream
193 boundary at Datong Gauging Station. The atmospheric forcing is the ERA-Interim data from the
194 European Centre for Medium-Range Weather Forecasts (ECMWF) with 0.125° spatial resolution
195 and 3-hour temporal resolution. This model is discretized into 20 uniform terrain-following
196 sigma layers, which provides sufficient vertical resolution for the bathymetry of 5 - 13 m in this
197 channel-shoal system. The time step is set to 10 seconds with a spin-up time of 15 days. For the
198 turbulence scheme, a Mellor and Yamada level-2.5 turbulent closure scheme with Galperin
199 modification is applied (Chen et al., 2003; Galperin et al., 1988; Mellor & Yamada, 1982). The
200 horizontal and vertical Prandtl number defined as the ratio of turbulent eddy viscosity to the
201 turbulent diffusivity (Chen et al., 2013), are 1.0 and 0.4 respectively.

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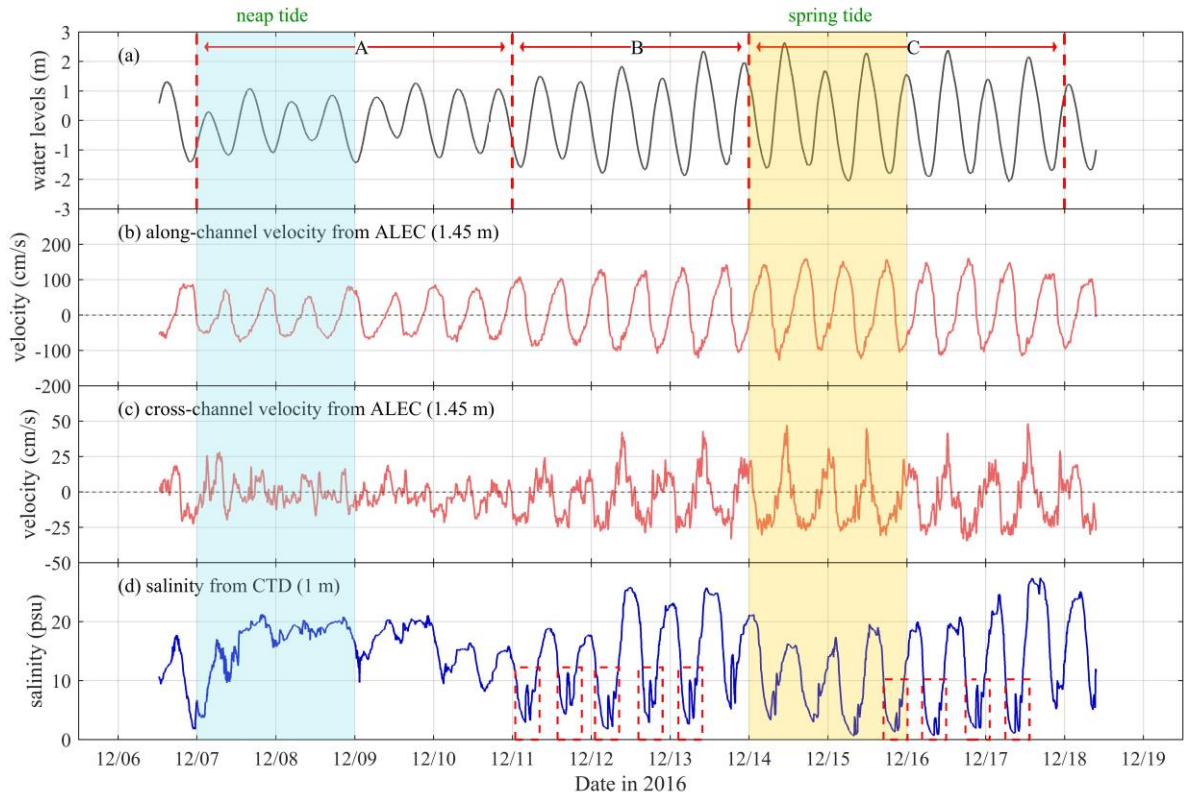
Figure 3: (a) Unstructured mesh for the Changjiang Estuary and adjacent regions. (b) An enlarged view within the NP. Section A, B is selected along- and cross-channel sections for analysis.

209 **3 Results**

210 **3.1 General physics (flow velocity, salinity)**

211 The main data measured by various instruments of the tripod, including water levels,
212 fixed-point velocity and salinity are shown in Fig. 4. The whole observation period can be
213 divided into three phases according to the tidal condition: Phase A from 12/07 to 12/11 (neap
214 tidal conditions), Phase B from 12/11 to 12/14 (intermediate tidal conditions) and Phase C from
215 12/14 to 12/18 (spring tidal conditions).

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219 Figure 4: Time series of (a) water levels, (b) horizontal along-channel and (c) cross-channel velocity from ALEC,
 220 (d) water salinity from CTD by the tripod. An orthogonal coordinate according to channel direction is used for
 221 velocity decomposition. For along-channel flow, the positive value indicates direction of about 120 degrees from the
 222 north, for cross-channel flow, the positive value indicates direction of about 30 degrees from the north.

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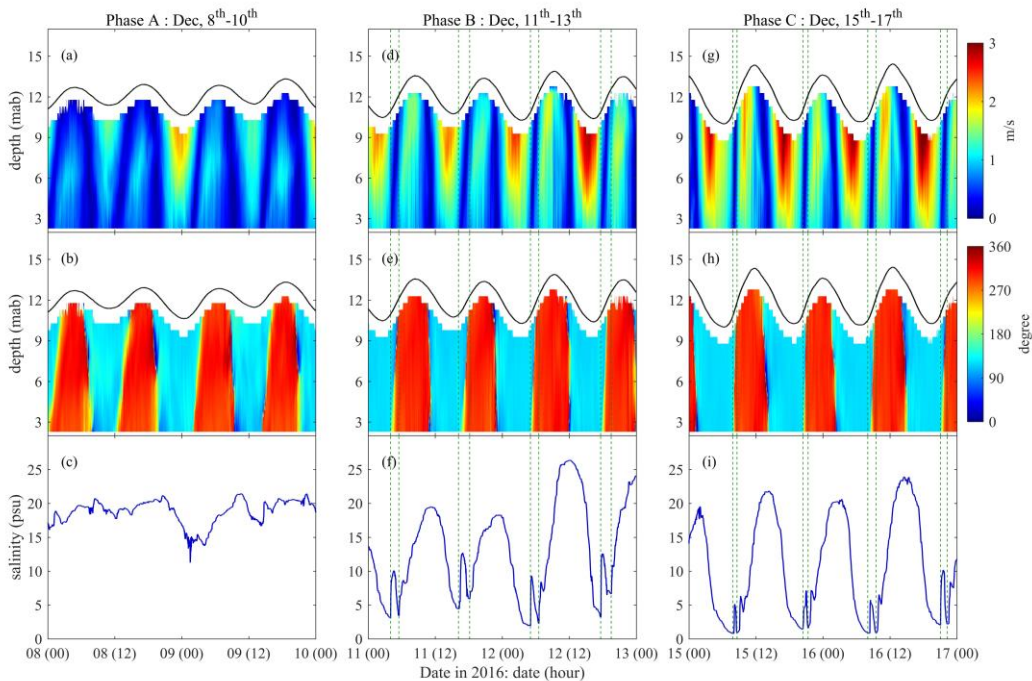
224 The tidal range during neap tides was ~ 2 m, and increased to ~ 4 m during spring tides
 225 (Fig. 4a). The average bottom current velocity at 1.45 mab during neap tides was 44.5 cm/s, and
 226 almost double during spring tides, reaching 86.1 cm/s (Fig. 4b, c). The maximum current velocity
 227 exceeded 150.0 cm/s during ebb tide. The mean flood duration during neap tides was 7.29 h,
 228 whereas the mean ebb duration was 5.25 h.

229 The cross-channel velocity component was irregular during neap tides (Fig. 4c) with
 230 weak-amplitude fluctuation. The maximum cross-channel velocity during neap tides is ~ 25 cm/s
 231 (06:45 LST on 12/07). During intermediate and spring tidal conditions, the cross-channel
 232 velocity had a pronounced flow asymmetry. During intermediate tides, southward cross-channel
 233 flow occurred for 57% of time (with the remaining 43% northward flow), increasing to 68% of
 234 time during spring tides. Also the average strength of southward flow was greater: 15.1 cm/s
 235 during intermediate tides and 17.0 cm/s during spring tides (11.6 cm/s and 15.6 cm/s for
 236 northward flow for the respective tidal conditions). Consequently, there is a pronounced (shoal-
 237 to-channel) cross-channel velocity component, both in magnitude and duration, resulting from
 238 lateral flows.

239 The bottom salinity remained above ~ 20.0 PSU for several days with only small
 240 oscillations during neap tides. This indicates that the area remains in the range of the salt wedge
 241 during neap tides, as observed earlier by Ge et al., (2018). During intermediate and spring tidal
 242 conditions, an oscillating salinity pattern can be observed, with values ranging from nearly 0
 243 PSU around the low water slack to 30 PSU around the high water slack (Fig. 4d). In addition,
 244 from 12/11 to 12/18 (phases B and C), a series of significant oscillations lasting for about 2 hours
 245 with magnitude of about 5 PSU occurred in each trough of salinity curve (Fig. 4d). We refer to
 246 these oscillations as intra-tidal salinity variations (ISV), and will be elaborated on in more detail
 247 in the next section.
 248

249 **3.2 Identification of intra-tidal salinity variation (ISV)**

250 The ISVs were highly consistent with the tidal phase, always occurring during low water
 251 slack (marked by red rectangles in Fig. 4d). During an ISV, the salinity first increases for about
 252 20 minutes, and then drops for about 51 minutes (averaged value of 5 ISVs during intermediate
 253 tides). Although the ISVs were identified during intermediate and spring tidal conditions, the
 254 amplitude was larger and the duration was longer during intermediate tides. The ISVs
 255 disappeared shortly before saline flood currents entered the North Passage close to the
 256 observation area. During this period, the upper water column was dominated by ebb currents
 257 while the bottom area was dominated by flood currents (Fig. 5e, h). This typical two-layer
 258 current structure occurred during low water slack, with weak currents in the whole water column.
 259 Importantly, during a period when ISVs occurred (green dashed lines in Fig. 5f, i), the flow
 260 direction was ~ 210 degrees from the north (Fig. 5e, h).
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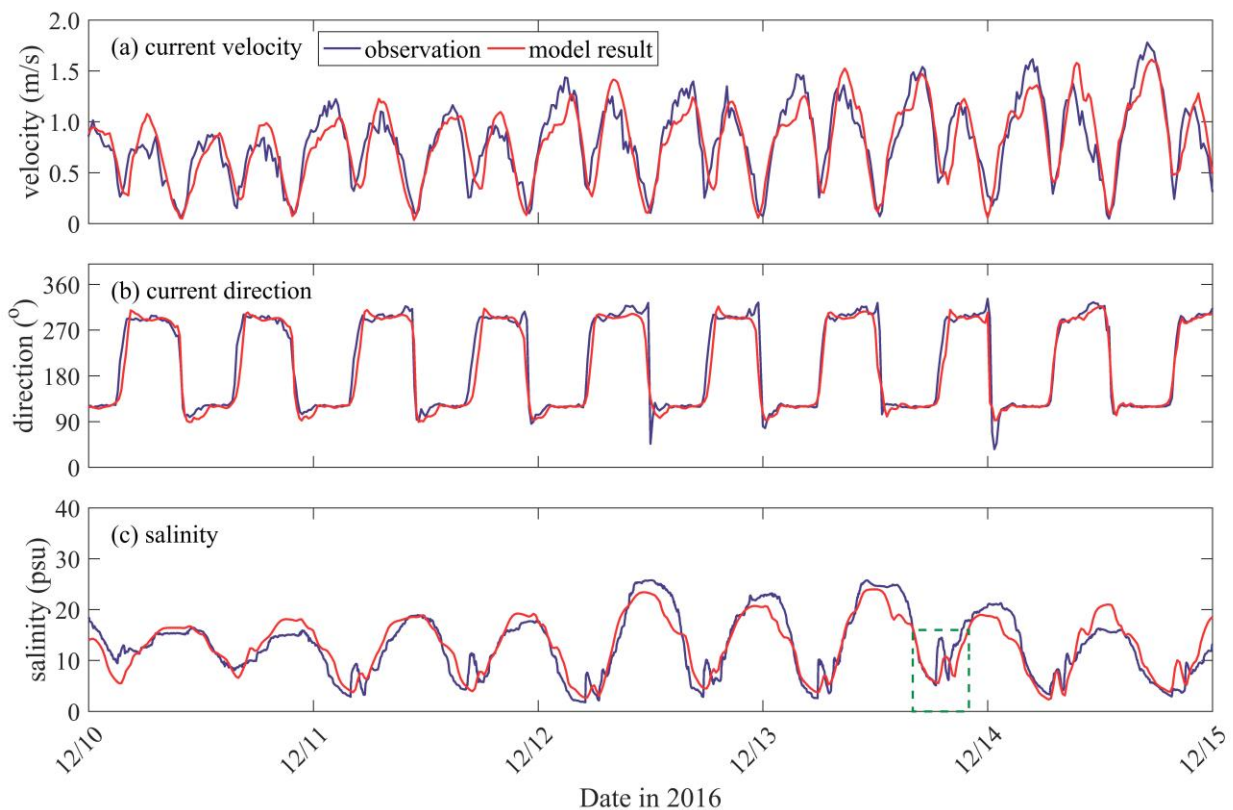
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264 Figure 5: Time series of vertical profiles of flow velocity (row 1), flow direction (row 2), near-bed salinity (row3)
 265 during neap tides (left column), intermediate tides (middle column) and spring tides (right column). The green
 266 dashed lines mark the occurrence of ISVs.

267

268 The observation indicates the formation of ISV is related to lateral flows. However, this
 269 measurement only demonstrates the ISV in a local perspective. It is unable to resolve the
 270 horizontal and vertical propagation of the water and salinity mass under the effect of the lateral
 271 flow. Therefore, the mechanism for the formation of ISV needs to be examined with a numerical
 272 model, allowing a more detailed quantitative analysis of the lateral flows.

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276 Figure 6: Model-data comparison between observed (blue) and simulated (red) results for near-bed tide current
 277 velocity (a), current direction (b) and salinity (c) at the observation site. The green dashed rectangle in (c) shows a
 278 typical intra-tidal salinity variation (ISV), and this ISV will be discussed in details in the following sections.

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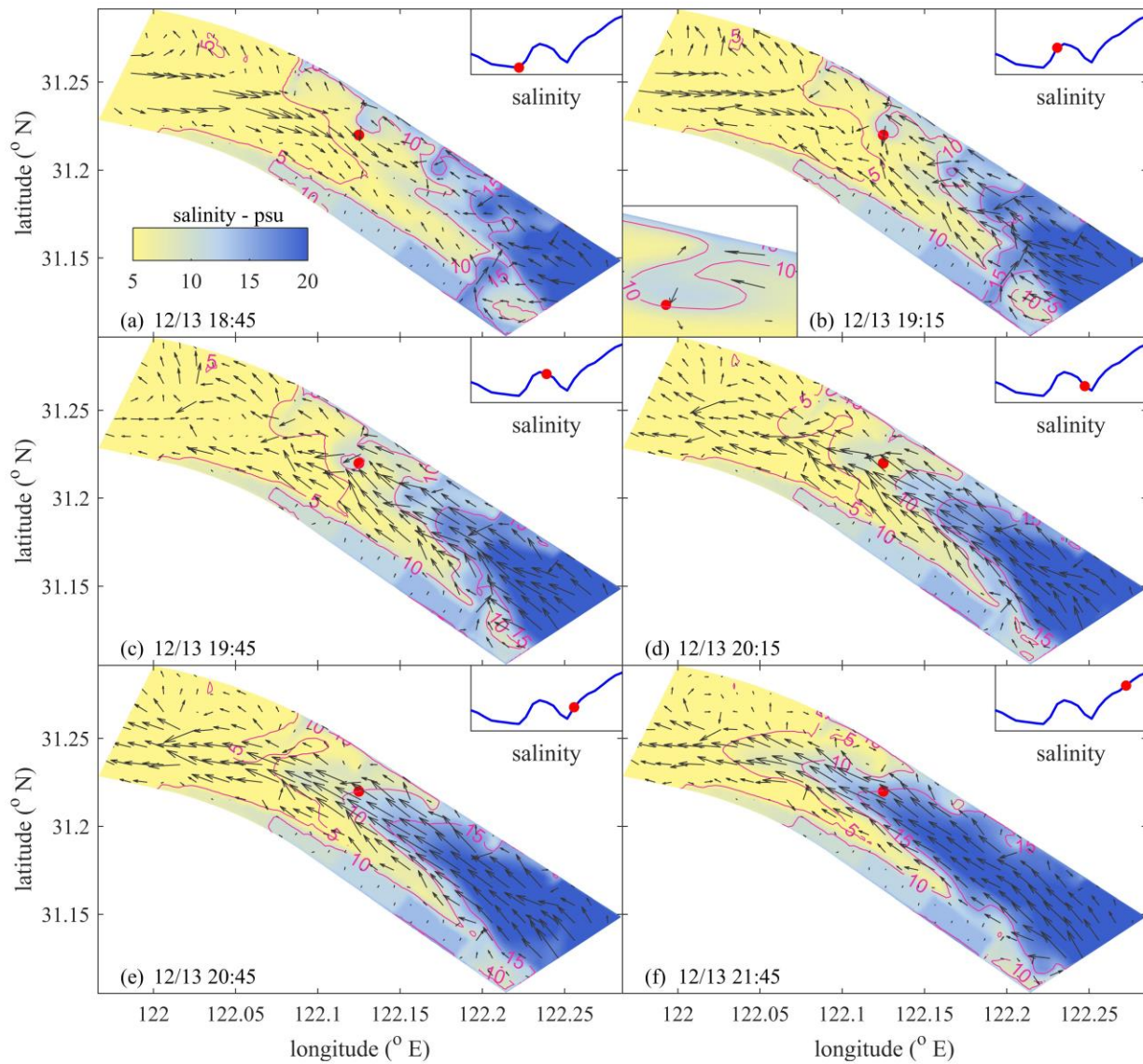
280 The applied model (CE-FVCOM) has been fully validated against hydrodynamics, tide,
 281 salinity in previous studies (Ge et al., 2012, 2013, 2014, 2015; Guo et al., 2018). In this study,
 282 model results are only compared to our tripod data for further validation. A comparison of near-
 283 bed flow velocity, direction and salinity (shown in Fig. 6) shows that the model captures the
 284 magnitude and direction of the upward-looking ADCP (data of the first cell, 2 mab), and the

285 variation of salinity collected by CTD. The overall root mean square errors (RMSEs) for velocity
286 magnitude, velocity direction and salinity are 0.19 m/s, 15.1 degrees and 2.6 PSU, respectively.
287 Even more, the ISVs occurring during intermediate and spring tidal conditions are also well
288 resolved by the model (an example is marked with green dashed box in Fig. 6c). Since the model
289 predicts the occurrence and correct timing of the ISVs, the model can be used to further explore
290 formation mechanisms in more detail.
291

292 3.3 Formation and breakdown of the ISV

293 To understand the formation and breakdown of ISV, a typical ISV during neap-to-spring
294 transition tides is selected (marked by a green dashed rectangle in Fig. 6c) for analysis with the
295 numerical model. At the end of ebb, the salinity in the channel (including the observation site,
296 Fig. 7a) was low while the shoals are more saline. The higher salinity on the shoals generates a
297 near-bed salinity-driven density current peaking at 0.5 m/s (Fig. 8e, f). The transport of more
298 saline water from the shoals towards the channel leads to an increase in near-bed salinity (Fig 7a-
299 c), giving rise to the formation of ISV.

300 The amount of saline water stored over the shoals is limited, however. The high-salinity
301 patch flows 1.5 km into the channel (Fig. 8f, g) after which it dilutes or is advected up-estuary
302 (Fig. 8g-i). During the period of maximum cross-channel flow (Fig. 8e), the near-surface ebb
303 currents are close to 1 m/s (Fig. 9e). Such a large velocity gives rise to mixing, which is
304 illustrated with the local salinity distribution. Near the observation point, the near-surface salinity
305 increases (Fig. 9f-h) when the near-bed salinity decreases (Fig. 8f-h, 9f-h). The salinity rapidly
306 rises after flow reversal when the saline flood propagates into the channel (Fig. 9i, Fig. 7e, f).
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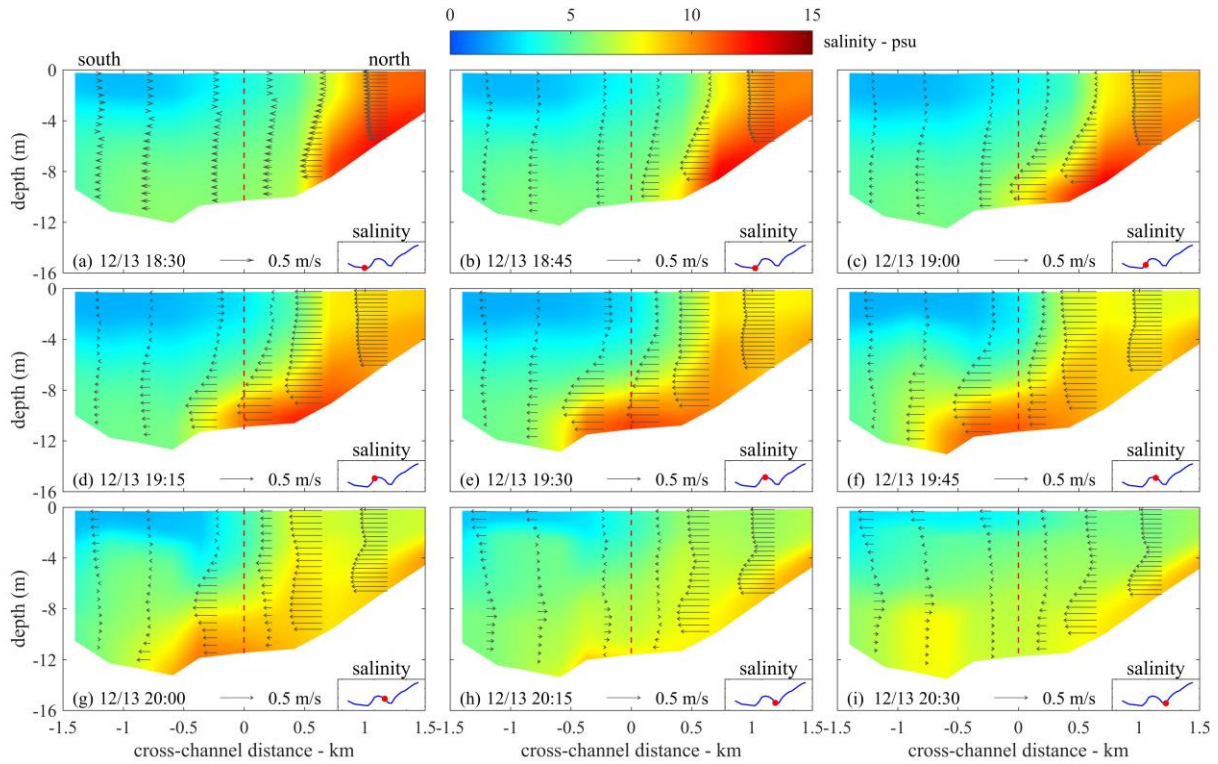


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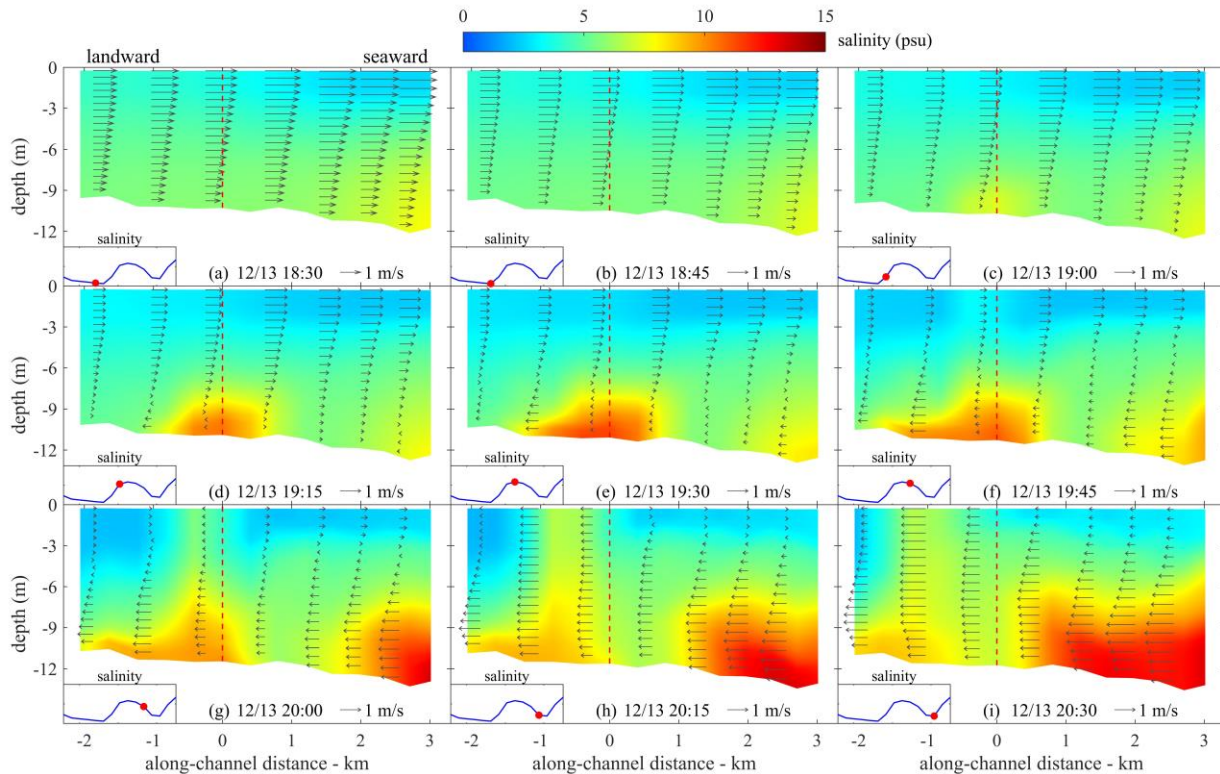
310 Figure 7: Spatial distributions of near-bed salinity and flow velocity at selected times (model result). The red dot
 311 indicates the observation site. The up-right inset shows the process of ISV and the red dot shows salinity of the site
 312 during this typical ISV period.

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316 Figure 8: Vertical distribution of salinity and cross-channel flow velocity (arrows) at selected times (a typical intra-
317 tidal salinity variation period). Red dashed line indicates the projection position of the observation site in a south-
318 north cross-channel section (section B in Fig. 3b). The down-right insets show the process of ISV and the red dot
319 shows salinity of the site during this typical ISV period.
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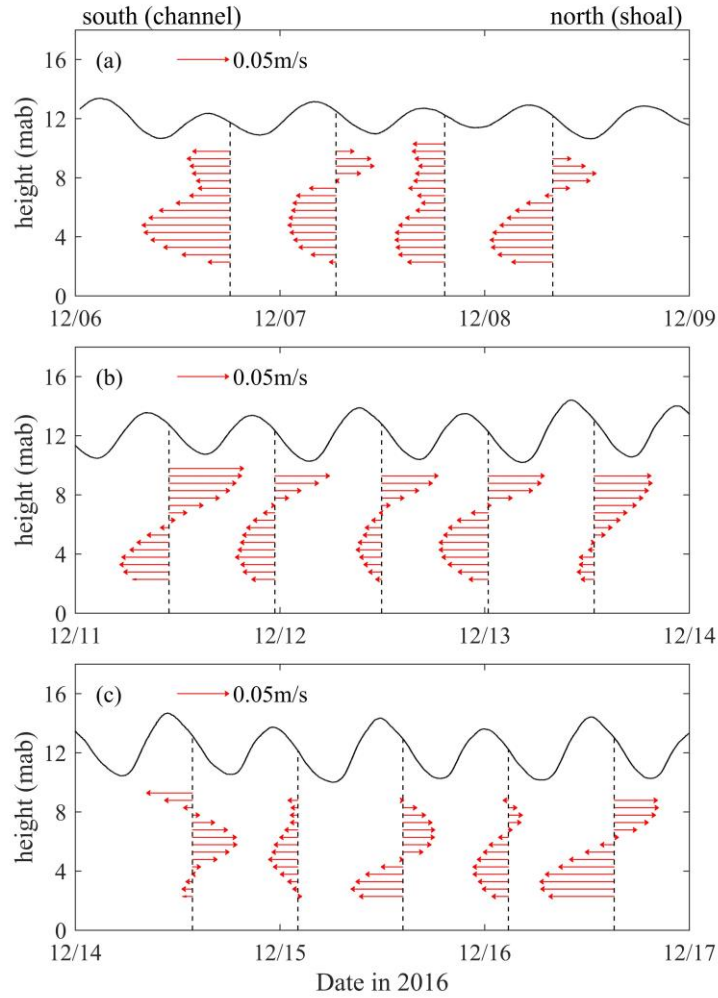


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Figure 9: Vertical distribution of salinity and along-channel flow velocity (arrows) at selected times (a typical intra-tidal salinity variation period). Red dashed line indicates the projection position of the observation site in a west-east along-channel section (section A in Fig. 3b). The down-left inset shows the process of ISV and the red dot shows salinity of the site during this typical ISV period.

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Averaged over the tide, the lateral flows discussed above contribute to a tide-averaged cross-channel residual current (Fig. 10). The variation of this tidally averaged cross-current over the spring-neap tidal cycle can be evaluated with the upward-looking ADCP observations. During the late neap tide and the intermediate tide when pronounced ISV was detected, the cross-channel residual current showed a clear vertical variability (Fig. 10a, b). For the majority of time, the tidally averaged cross-channel component near the bed is directed towards the channel whereas it is directed towards the shoals near the water surface.



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338 Figure 10: Tidally averaged cross-channel residual current calculated from ADCP results in (a) neap tide, (b)
 339 intermediate tide and (c) spring tide at the observation site.

340

341 **4 Discussion**

342 4.1 Mechanism for the lateral flow: momentum balance analysis

343 Interpretation of the data and numerical model results suggests that the density-induced
 344 gradient resulting from the salinity difference between the channel and the shoal is the driving
 345 force for the lateral flow. We will further quantify this hypothesis using a momentum balance
 346 analysis.

347 We define the along-channel direction as the x-axis (the positive value indicates a
 348 direction of 120 degrees from the north) and the cross-channel direction as the y-axis (the
 349 positive value indicates a direction of 30 degrees from the north, away from the main channel),
 350 over which the governing equations for horizontal motions are given as follows:

$$351 \quad \frac{\partial u}{\partial t} + \underbrace{u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} + w \frac{\partial u}{\partial z}}_B - \underbrace{fv}_C = - \underbrace{\frac{1}{\rho} \frac{\partial (P_H + P_a)}{\partial x}}_D + \underbrace{\frac{\partial}{\partial z} \left(K_m \frac{\partial u}{\partial z} \right)}_E + \underbrace{F_x}_F \quad (1)$$

$$352 \quad \frac{\partial v}{\partial t} + \underbrace{u \frac{\partial v}{\partial x} + v \frac{\partial v}{\partial y} + w \frac{\partial v}{\partial z}}_B + \underbrace{fu}_C = - \underbrace{\frac{1}{\rho} \frac{\partial (P_H + P_a)}{\partial y}}_D + \underbrace{\frac{\partial}{\partial z} \left(K_m \frac{\partial v}{\partial z} \right)}_E + \underbrace{F_y}_F \quad (2)$$

353 where x , y and z are the horizontal and vertical axes of the Cartesian coordinate; u , v and w
 354 are the x -, y -, and z -component velocities, respectively; ρ is density; P_H is hydrostatic
 355 pressure; P_a is the air pressure at sea surface; f is the Coriolis parameter and K_m is vertical
 356 eddy viscosity coefficient. Here, F_x and F_y represent the horizontal momentum diffusion terms
 357 in the along- and cross-channel directions, respectively.

358 The air pressure is omitted due to its weak contribution to the local hydrodynamics in the
 359 channel-shoal system. The hydrostatic pressure P_H satisfies:

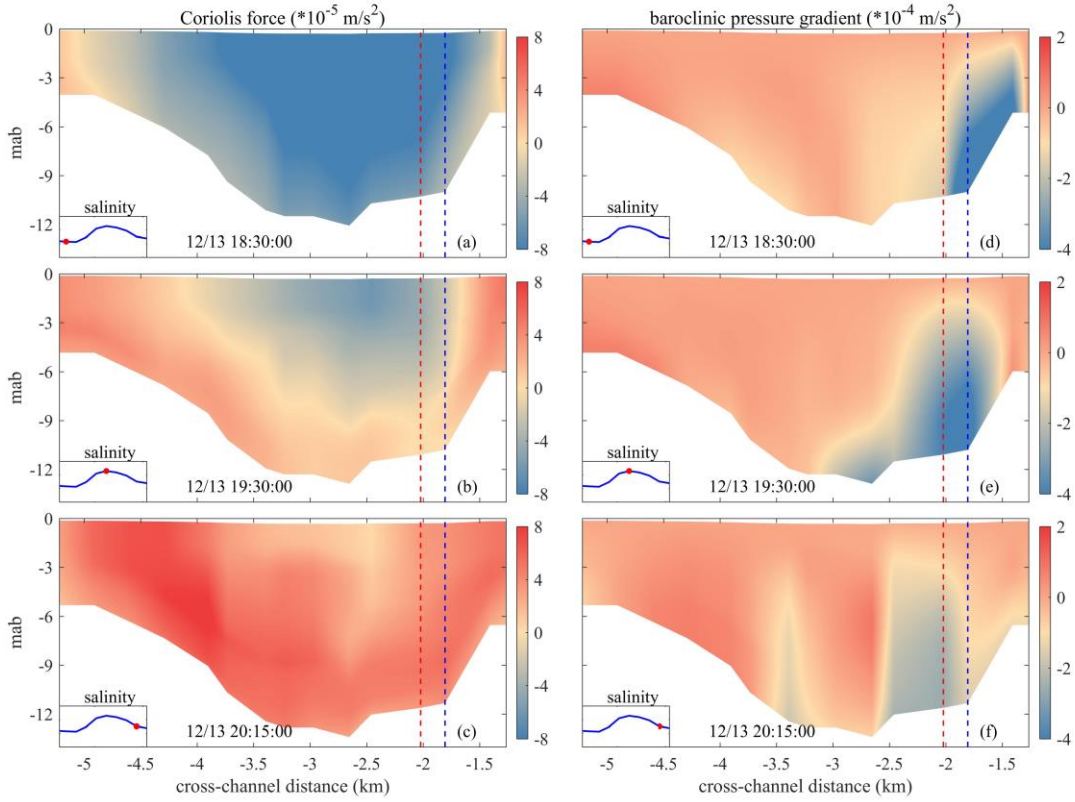
$$360 \quad \frac{\partial P_H}{\partial z} = -\rho g \Rightarrow P_H = \rho_0 g \zeta + g \int_z^0 \rho dz \quad (3)$$

361 where g is the gravitational acceleration.

362 In Eqs. (1) and (2), $A - F$ denote the local acceleration, advection, Coriolis force,
 363 pressure gradient, and vertical and horizontal momentum diffusion terms, respectively. The
 364 pressure gradient force term includes the barotropic (surface elevation) pressure gradient force
 365 and baroclinic (density) pressure gradient force as described in Eq. (3).

366 Figure 11 shows the vertical distribution of momentum terms of the Coriolis force and
 367 baroclinic pressure gradient (BPG) along the selected section B at selected times at before
 368 (18:30), right on the crest of (19:30) and after (20:15) a significant ISV on 12/13. The BPG was
 369 $\sim -3 \times 10^{-4} \text{ m/s}^2$ at 18:30 LST 12/13 in the shoal area (Fig. 11d). The Coriolis force was $\sim -$
 370 $4 \times 10^{-5} \text{ m/s}^2$, one order of magnitude smaller than the BPG since tidal currents are weak during
 371 low water slack (Fig. 11a). The BPG and Coriolis force jointly drove the flow southward (from
 372 the north shoal to the main channel), generating the lateral current. In this stage, the BPG was the
 373 major contributor among all dynamical forces. At 19:30 LST 12/13, the BPG continued to drive
 374 the southward lateral flow, counteracted by the Coriolis force because near bottom this Coriolis
 375 force was in the northward direction (Fig. 11b). Similarly, the Coriolis force was one order of
 376 magnitude smaller than the BPG. Therefore, this hindrance effect could not substantially inhibit
 377 the development of the lateral flow. At the end of the ISV period when flood currents flowed into
 378 the channel, and the Coriolis force was in the northward direction in the whole section (Fig. 11c),
 379 the BPG in the deep channel increased and had a northward component because of the
 380 movement of saline water (Fig. 11f). Later, BPG in the deep channel would continue to increase
 381 due to seawater intrusion.

382



383

384

385 Figure 11: Vertical distributions of Coriolis force (left column) and baroclinic pressure gradient (right column) along
 386 the selected cross-channel section B in the intra-tidal salinity variation (ISV) period. The red dashed line indicates
 387 the location of the observation site, and the blue dashed line shows an adjacent location north to the site. The down-
 388 left inset shows the process of ISV in salinity and the red dot shows salinity of the site during this typical ISV
 389 period.

390

391 The momentum term due to flow curvature is also estimated. Eqs. (1) and (2) are written
 392 in Cartesian coordinate, which do not contain separate momentum term induced by channel
 393 curvature. The momentum equation can be written in curvilinear coordinate as follows (Chant,
 394 2010) :

$$395 \quad \frac{\partial u_n}{\partial t} + u_s \frac{\partial u_n}{\partial s} - \frac{u_s^2}{R} + f u_s + g \frac{\partial \eta}{\partial n} - \frac{\partial \tau}{\partial z} = 0 \quad (4)$$

396 where s , n , z represent main-flow, lateral and vertical directions, u_s , u_n are velocity
 397 components in s , n directions, R is the radius of curvature, f is the Coriolis parameter, η is

398 water level, τ is stress. In Eqs. (4), $\frac{u_s^2}{R}$ is momentum term of channel curvature, which is also

399 named centrifugal acceleration. The longitude and latitude information of the north dike is
 400 transferred into a Cartesian coordinate, using a map projection method. Then the new position of
 401 the north dike in the Cartesian coordinate is fitted with a quadratic curve to calculate R . For
 402 cross-channel section B (Fig. 3b) which passes the observation site, $R \approx 49.7$ km, maximum

403 magnitude of u_s was about 1.9 m/s (Fig. 5b), thus maximum value of $\frac{u_s^2}{R}$ was about 7.3×10^{-5}
 404 m/s^2 . At the beginning of ISV (18:30 LST 12/13), u_s was about 0.2 m/s, and the effect of
 405 channel curvature was several orders of magnitude smaller than that of the BPG. Therefore, the
 406 baroclinic pressure gradient caused by the horizontal salinity gradient was the main driving force
 407 for the formation of the lateral flow.

408 Closer to the channel bend (Fig. 1), $R \approx 28$ km, and here curvature effects could be
 409 larger. During the ebb tide, the maximum value of $\frac{u_s^2}{R}$ is about $1.3 \times 10^{-4} m/s^2$, towards the
 410 outside of the channel. While the Coriolis force is about $-1.1 \times 10^{-4} m/s^2$. These two forces were
 411 close and the joint effect is small. However, during the flood tide, maximum centrifugal
 412 acceleration is about $1.0 \times 10^{-4} m/s^2$ while the Coriolis force is about $1.3 \times 10^{-4} m/s^2$. The
 413 combined effect of channel curvature and Coriolis force (about $2.3 \times 10^{-4} m/s^2$ in total) is
 414 comparable to that of the baroclinic pressure gradient which generates the lateral flow. In the
 415 other word, this combined effect could also lead to an opposite lateral flow during flood tide
 416 period close to the river bend.

417 Because u_s in the channel curvature term is squared, this term always has the same sign
 418 within a whole tidal cycle. It may therefore strengthen or weaken the Coriolis term in different
 419 stages of a tidal cycle, giving rise to tidal asymmetry in lateral flows during flood and ebb tide
 420 (Chant, 2010). In many cases, lateral flow or lateral circulation could be dominated by the
 421 centrifugal acceleration for a period in a tidal cycle (Kim & Voulgaris, 2008; Lacy & Sherwood,
 422 2004; Nidzieko et al., 2009) as the radius of curvature R is much smaller than that in our study.
 423 Although R is relatively large and the curvature induced term appear to be less important than
 424 other terms, variation of R leads to different lateral circulation patterns in different cross-
 425 channel sections of the North Passage. In this study, the maximum joint effect of Coriolis term
 426 and centrifugal term is still smaller than the BPG when ISV was generated, indicating the
 427 significance of the trapped saline patch in the shallow shoal and the intensity of this lateral flow.
 428

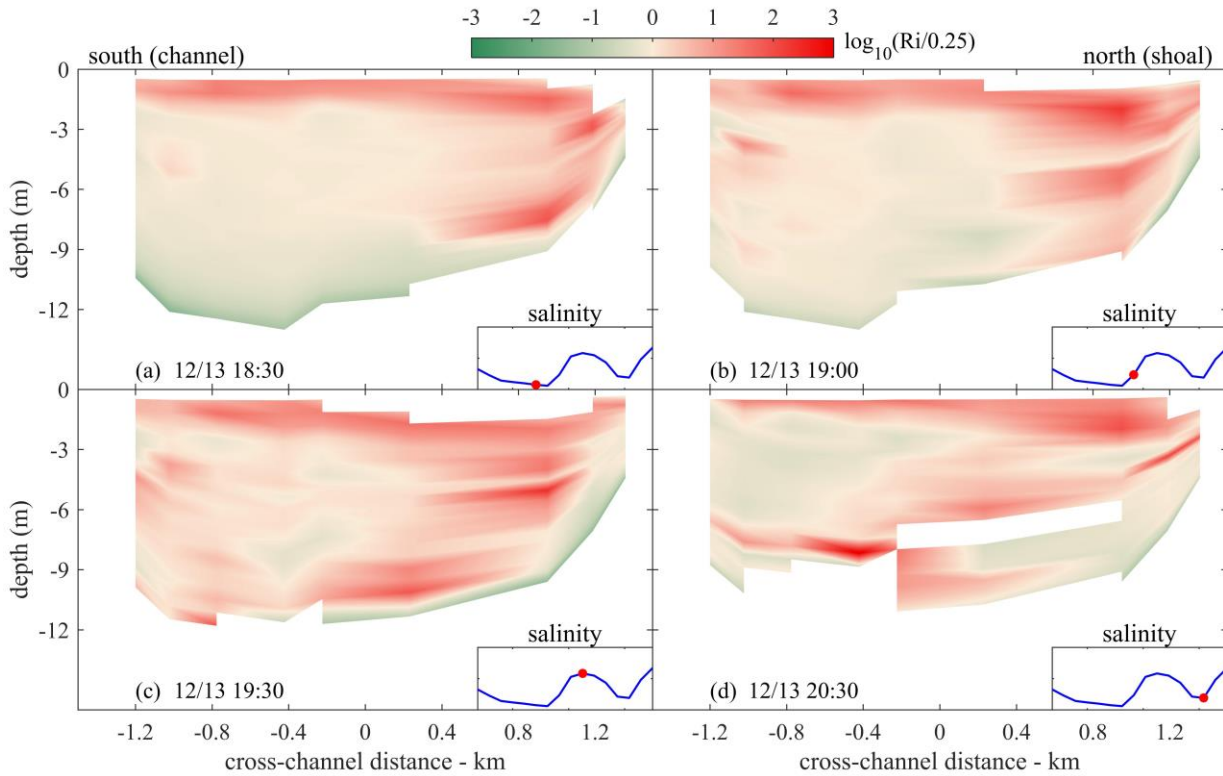
429 4.2 Stratification and mixing

430 Differential advection of the water mass leading to ISV greatly influences stratification
 431 and mixing process in the channel-shoal system. During the ISV period, the cross-channel and
 432 along-channel salinity distribution and velocity showed a pronounced vertical variability (Fig.8,
 433 Fig. 9). To what extent feedback mechanisms exist between salinity-induced stratification and
 434 the flow is further investigated with the gradient Richardson number (R_i):

$$435 R_i = -\frac{g}{\rho_w} \frac{\partial \rho / \partial z}{\left[\left(\frac{\partial u}{\partial z} \right)^2 + \left(\frac{\partial v}{\partial z} \right)^2 \right]} \quad (5)$$

436 where ρ_w is water density, $\partial \rho / \partial z$ is density gradient, u, v are velocity components in x-, y-
 437 direction. The water column stratifies when R_i exceeds 0.25. Therefore $\log_{10}(R_i/0.25)$ is used
 438 as an index for the degree of stratification (Fig. 12, with positive values indicating stratifying
 439 conditions). The density of water did not always increase from the surface to the bottom,

440 resulting in negative values of R_i which give rise to the blank areas in Fig.12 when taken as a
 441 logarithm. As explained earlier, the saline water remaining in the shallow areas flowed
 442 downslope at the end of the ebb due to horizontal density differences, giving rise to the ISV. As a
 443 result of this density current, the near-bed channel gradually became more stratified (Fig. 12b, c)
 444 by differential advection of salinity. During the peak of the ISV (Fig. 12c), the larger part of the
 445 water column was salinity-stratified. However, the layer very close to the bed had a weak
 446 stratification (Fig. 12c). This is because of the large gradient of velocity in the near-bed area due
 447 to the friction (Fig. 8e).
 448



449
 450
 451 Figure 12. Distribution of $\log_{10}(R_i/0.25)$ in cross-channel section B (a) before the ISV, (b) in the increasing stage
 452 of ISV, (c) at the peak of ISV and (d) at the end of the ISV. The zero position in x-axis indicates the position of the
 453 observation site in the section. The down-right inset shows variation of salinity, and the time-stamp salinity at the
 454 site.

455
 456 The gradient Richardson number (R_i) indicates the stratification at a specific height in the
 457 whole vertical column while the bulk Richardson number (R_{ib}) provides information of the
 458 whole water column (bottom to surface). R_{ib} can be calculated as (Hoitink et al., 2011; Lewis,
 459 1997):

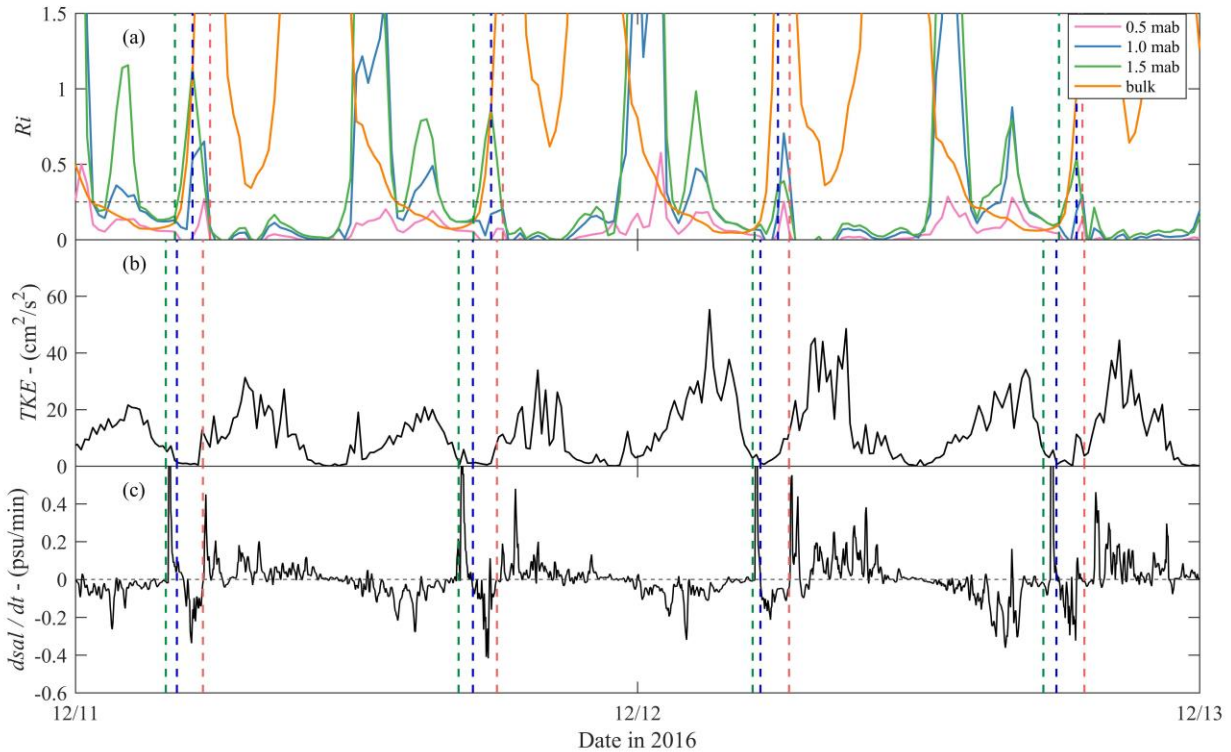
$$460 \quad R_{ib} = \frac{gD\Delta\rho}{\overline{\rho_w} |u_s|^2} \quad (6)$$

461 where D is the water depth, $\Delta\rho$ is the density difference between surface and bottom, $\overline{\rho_w}$ is the

462 averaged density of the water column, $\overline{u_s}$ is the horizontal velocity at the surface. The gradient
 463 Richardson number of different layers and the bulk Richardson number are shown in Fig. 13.
 464 Turbulent kinetic energy (*TKE*) per unit mass is also used for further interpretation of vertical
 465 mixing. *TKE* can be calculated using the high-frequency velocity data collected by ADV as:

$$466 \quad TKE = \frac{1}{2}(\overline{u'^2} + \overline{v'^2} + \overline{w'^2}) \quad (7)$$

467 Green, blue and red dashed lines in Fig. 13 indicate the start, peak (salinity maximum)
 468 and end of the ISV, respectively. Note that Fig. 13a is based on model results while Fig. 13b, 13c
 469 are based on results of the tripod observation. Therefore, a small time lag of ISV process
 470 between these two methods can be found in Fig. 13 (vertical dashed lines are different in three
 471 sub-graphs). Shortly before the start of the ISV (green dashed lines), the near-bottom R_i and R_{ib}
 472 were below 0.25 (Fig. 13a), indicating well-mixed condition. Both the time-variation in near-bed
 473 salinity ($dsal/dt$, Fig. 13c) and vertical salinity variation (Fig. 8a) were small. All information
 474 therefore suggests a steady, well-mixed water column near the end of the ebb. During the first
 475 stage of the ISV (between green and blue dashed lines), the salinity and R_{ib} increased rapidly
 476 (Fig. 13). R_i at 1.5 mab showed a different variation from R_i at 0.5 and 1.0 mab. At 1.5 mab,
 477 there was a pronounced stratification as it was close to the interface between water in the salinity
 478 patch and ambient water. During the decreasing stage of the ISV (between blue and red lines),
 479 the whole water column (as indicated with the bulk Richardson number) remained stably
 480 stratified ($R_{ib} > 0.25$). The near-bed gradient Richardson numbers suggest a non-stratified water
 481 column, but this was because the halocline was located higher up in the water column. The
 482 observation that the water column was stratified during the decline of the ISV suggests that
 483 longitudinal advection (and not mixing) was the main mechanism responsible for the salinity
 484 decrease. Well-mixed conditions ($R_{ib} < 0.25$) re-established halfway the flood, during periods of
 485 high *TKE*.
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Figure 13: Time series of (a) gradient Richardson number at 0.5 mab (pink), 1.0 mab (blue), 1.5 mab (green) and bulk Richardson number (orange), (b) turbulent kinetic energy (TKE) at the observation site, (c) rate of salinity change in time. Green and blue dashed lines indicate the increasing stage in the ISV, blue and red dashed lines indicate the decreasing stage in the ISV.

494

4.3 Effects of dikes and groynes

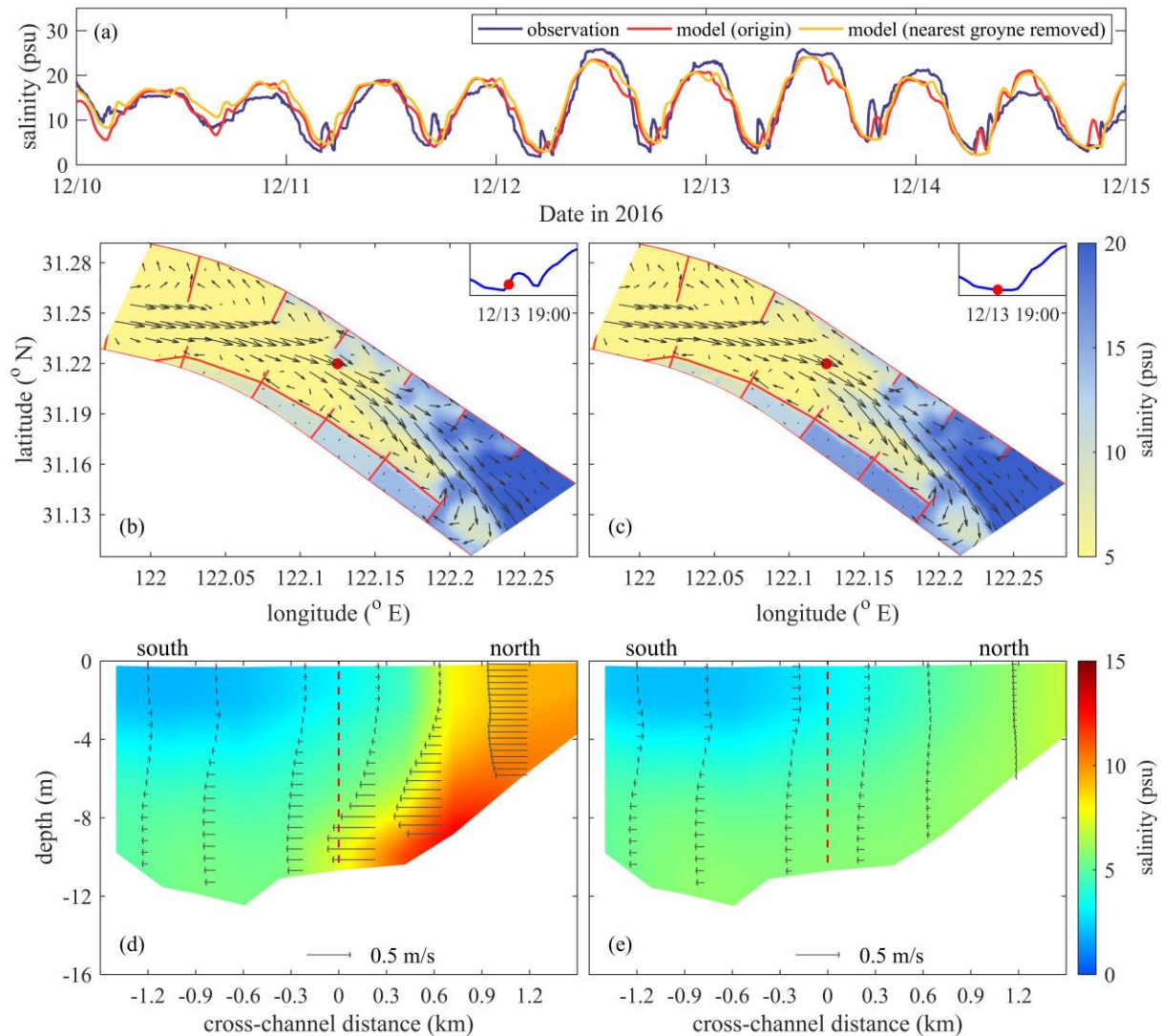
495

The investigated tidal channel was surrounded both towards the south and north with dikes and overflowing groynes (Fig. 1). Although the main parts of the groynes from the tip are submerged by sediment deposition, the remaining parts near dikes are still exposed during low tide, which shapes the groyne-sheltered area as a semi-enclosed region. The groynes create nearly stagnant water masses, and therefore strengthen differential advection of salt water. As a result, these groynes may greatly contribute to the transverse flows (and therefore ISV).

501

The effect of groynes is quantified by rerunning the model with one groyne (nearest to the observation site) removed. As a result, the ISV is no longer predicted at the observation site (Fig. 14a). The bottom salinity distribution clearly shows that no high-salinity water is trapped in the north shoal without the groyne (Fig. 14c). This is also clear in vertical salinity distribution in the cross-channel section and cross-channel velocity is much smaller than the original case (Fig. 14e). For downstream areas where groynes are still present, salinity is still influenced by the groynes (Fig. 14c).

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511 Figure 14: Model-data comparison with groyne-removed case added (a), comparison between original case (b, d)

512 and groyne-removed case (c, e) at the same time in bottom salinity distribution (b, c) and vertical salinity

513 distribution (d, e).

514

515 The numerical model results suggest that the groyne fields north of our observation site
 516 contribute much stronger to the generation of ISV's than their southern counterparts. Due to the
 517 retention effect all along the northern groynes lateral flows developed, and the salinity varied
 518 over the tidal cycle (Fig. 7). This was not true, however, for the south groynes. The southern
 519 groynes were connected by an additional along-estuary dike (Fig. 14b) which limited exchange
 520 flows between the groyne fields and the estuary. As a result, their salinity remained fairly
 521 constant in time (Fig. 7) and cross-channel flows did not develop from the south (Fig. 8).

522 The impact of groynes on lateral flows reported here has, to our knowledge, not been
 523 published elsewhere in the scientific literature, but constitutes an important contribution. Many

524 researchers have reported groynes' impacts on flow hydrodynamics (Brevis et al., 2014; McCoy
525 et al., 2008; Sukhodolov, 2014) and mass exchange (McCoy et al., 2007; Uijttewaai et al., 2001;
526 Weitbrecht et al., 2008). These studies do not include salinity-induced density effect which is
527 closely related to groynes. Groynes are used throughout the world to channelize flow, preventing
528 siltation rates in fairways maintained to provide access to ships. Lateral flows may significantly
529 contribute to near-bed sediment dynamics and siltation rates, and therefore resulting in the exact
530 opposite effect they were originally designed for. As a result, comprehensive understanding of
531 the impact of such engineering works are required for sustainable development of human-
532 impacted estuaries.
533

534 4.4 Implications of ISV in other studies

535 Earlier studies in the North Passage of the Changjiang Estuary also identified the
536 existence of ISV, but was not paid much attention to. Song et al. (2013) conducted a quadrupod
537 observation at the slightly south of the deep channel. They observed a similar intra-tidal salinity
538 variation during the low water slack (in their Fig. 2a). They used velocity skew and flux
539 skewness to examine the lateral sediment transport, identified a net sediment transport from the
540 south groyne to the deep channel with an opposite lateral transport from the north to the south
541 side of the channel in near-bottom area (below 1 mab). This opposite lateral flow and the ISV
542 was likely generated with the same mechanisms discussed in our study. We believe that the ISV
543 in their study was also generated by saline water trapped in the north groyne-sheltered area as
544 this saline water mass could move across the deep channel to where Song et al. (2013) conducted
545 the observation. Actually, the ISV in their study should be a further dispersion of ISV generated
546 from the north shoal and that is why the magnitude of ISV at their observation site was much
547 weaker. The lateral flow discussed in our study should be taken into account when considering
548 sediment transport in the North Passage, as its movement affects the whole deep channel in the
549 cross-channel direction. This will improve the knowledge of severe siltation issues in the
550 channel.

551 ISV was also detected in other estuaries. Ralston et al. (2012) deployed fixed instruments
552 frame in the channel and on the shoal of Hudson River estuary, which also revealed clear
553 evidence for ISV (in their Fig. 3a). The bottom salinity measured in the channel also varied
554 within the tidal cycle during low water slack, and was also the result of lateral flows. They
555 concluded their lateral flow was primarily related to topography features in a channel-shoal
556 system. We believe that more estuaries and channels all over the world are characterized by
557 lateral flows resulting in ISV's albeit that the responsible mechanism (in our case the baroclinic
558 pressure gradient, in Ralston's case the topography) may differ.

559 The ISV directly gives a strong indication for lateral flows in a channel-shoal system.
560 ISV can be a simple indicator of lateral flow and lateral processes which should be considered
561 when investigating the hydrodynamics and sediment dynamics of estuaries with extensive shoals.
562 Actually, the ISV is just one typical feature under the modulation of lateral flow. Besides,
563 particular small-scale fluctuation in other variables (for example temperature) also occurs

564 concurrently with ISV. Generally speaking, these small-scale fluctuations in multiple variables
565 all indicate the importance of the lateral flow caused by many different reasons in a tidal
566 channel-shoal system.

567

568

569 **5 Conclusions**

570 In this study, a tripod system integrated with multiple instruments was deployed in the
571 North Passage of the Changjiang Estuary to measure lateral flow and its impacted salinity
572 transport. Observation results revealed periodic formation of intra-tidal salinity variation during
573 low water slack. The tripod observations indicated the intra-tidal salinity variation (ISV) was
574 generated by a near-bed lateral flow from the shallow shoal to the deep channel. A high-
575 resolution unstructured-grid model for the Changjiang Estuary has been applied to simulate this
576 lateral flow and the salinity transport. The numerical simulation showed that the high-salinity
577 water trapped over the shoal was transported to the deep channel during ebb tide. The
578 momentum balance analysis based on model results identified the density-induced baroclinic
579 pressure gradient was the dominant physical mechanism for the generation of this lateral flow.
580 Cross-channel residual current had a consistent near-bed shoal-to-channel component, which was
581 closely connected with the shoal-to-channel lateral flow. Salinity transport also showed a
582 southward net transport pattern.

583 This mechanism produced intra-tidal peaks in salinity and influenced pattern of
584 stratification and mixing in the cross-channel section. Strong stratification occurred with and
585 enhanced by the lateral flow, which can potentially affect sediment behavior (for example
586 hindered settling) and modulate the pattern of sediment transport.

587 The ISV can be a signal of active lateral process in a tidal estuary, especially in a human-
588 impacted channel-shoal system. Our simulations show that dikes and groynes greatly influence
589 the flow field in the channel and the retention effect of groynes predominantly contributes to this
590 lateral flow. This conclusion reminds us of more caution when designing such engineering
591 structures. The findings reported here provide a key element for future work on sedimentation
592 issues in general, but in the North Passage in particular, since our work suggests that groynes
593 generate a residual current directed from the shoals to the channel.

594 In particular, the role of salinity-induced currents on sediment dynamics as well as
595 sediment-induced density effect on lateral flows needs to be further investigated. Since the
596 results reported in this study were observed and simulated under dry-season freshwater discharge
597 from the upstream Changjiang, caution should be taken when applying these results to wet
598 season when freshwater discharge can be three to four times larger. Additional observations and
599 model simulations are needed in that case.

600

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