



Fluid mud impacts on estuarine sediment import

Shiyu Bao¹, Henk Jongbloed¹, Reinier Schrijvershof^{1,2}, Jinhai Zheng³, and Antonius J.F. Hoitink¹

¹Hydrology and Environmental Hydraulics Group, Wageningen University and Research, Wageningen, the Netherlands

²Deltares, Delft, The Netherlands

³Hohai University, Nanjing, China

Correspondence: Shiyu Bao (shiyu.bao@wur.nl)

Abstract. In hyperturbid estuaries, fine sediment accumulation frequently generates fluid mud layers that pose major challenges to ecology and navigation. Few studies have addressed the relation between fluid mud and residual sediment transport mechanisms to date. Here, we present Acoustic Doppler Current Profiler (ADCP) and water-sampling observations from the hyperturbid Emden Fairway (Ems Estuary), collected under three river discharge conditions ($45\text{--}200\text{ m}^3\text{ s}^{-1}$), including two
5 13 h tidal-cycle surveys with full-depth suspended sediment concentration (SSC) profiles. The thickness of the fluid mud layers differed among surveys with contrasting discharge in the research area, exceeding 2 m at high discharge ($200\text{ m}^3\text{ s}^{-1}$), whereas no fluid mud was present at low discharge. When a thin fluid mud layer was present ($\sim 1\text{ m}$), the net landward sediment flux per unit width over around 12 h observation windows reached 200 t m^{-1} , which is one order of magnitude larger than that measured when the fluid mud was absent. The presence of a fluid mud layer coincided with a two-layer estuarine circulation, and
10 enhanced tidal pumping through elevated near-bed SSC, which both promote landward sediment transport. In contrast, when the fluid mud layer has retreated, the residual flow is seaward over the full water depth, likely due to a sediment-induced density gradient opposing that of salinity, weakening estuarine circulation. The net landward transport relies mainly on tidal pumping. During the highest-discharge survey, near-bed sediment-induced density gradients were larger than the salinity gradient, indicating that fluid mud is not merely passively advected but may exert a control on the near-bed residual flow. Our results show
15 that resolving the full near-bed layer is essential for quantifying sediment transport and understanding the discharge-driven redistribution of sediment in hyperturbid estuaries.

1 Introduction

Tidal rivers and estuaries increasingly experience high levels of suspended sediment (Grasso et al., 2025; Zhu et al., 2022), which severely impact water quality and ecosystem functioning (Huang et al., 2022). Under human-induced changes, such
20 as channel deepening and narrowing, many estuaries have shifted from low to high turbidity states. In these systems, surface suspended sediment concentrations (SSC) can increase from roughly 0.1 g L^{-1} to over 10 g L^{-1} (Winterwerp et al., 2013; Van Maren et al., 2015b; Dijkstra and de Goede, 2024). Elevated SSC alters bed dynamics, threatening the navigability of shipping channels, and reduces water clarity, which subsequently degrades primary production and habitat suitability (Wood and Armitage, 1997). A mechanistic understanding of sediment transport under hyperturbid conditions is therefore essential
25 for effective navigation management and the preservation of estuarine ecosystem services.



At the estuary scale, suspended sediment transport is commonly interpreted as the result of competing riverine and marine transport processes. Residual sediment fluxes are generally controlled by the balance between river-driven export and landward transport induced by tidal asymmetry, estuarine circulation, and baroclinic forcing (Burchard et al., 2018). The relative importance of these processes depends on external forcing conditions, including river discharge, tides, waves, and wind, as well as on system morphology, sediment availability, and human interventions such as dredging and freshwater regulation. Over recent decades, substantial progress has been made in understanding the mechanisms responsible for estuarine sediment trapping and transport (Scully and Friedrichs, 2007; Allen et al., 1980; Dronkers, 1986; Jay and Musiak, 1994). However, in hyperturbid estuaries, residual sediment transport may deviate from these classical mechanisms, because the hydrodynamics and sediment dynamics may become coupled. The formation of near-bed high-concentration layers, commonly referred to as fluid mud, can fundamentally modify the estuarine turbulence structure, vertical mixing, and residual circulation.

The Ems Estuary represents one of the most hyperturbid estuarine systems worldwide and is characterized by recurring, large-scale, discharge-dependent occurrences of fluid mud (Winterwerp and Wang, 2013; Papenmeier et al., 2013; Talke et al., 2009). The Emden Fairway and lower Ems River experience persistently high SSC and require intensive maintenance dredging to preserve navigability (Van Maren et al., 2016). However, the persistence of such hyperturbid conditions remains difficult to reconcile with the large-scale sediment transport dynamics inferred from previous studies. The upstream Ems River supplies relatively little sediment, insufficient to sustain the observed turbidity levels (Van Maren et al., 2015b). Furthermore, the mechanism of tidal velocity asymmetry within the Emden Fairway alone does not provide a sufficient explanation for the observed dynamics (Pein et al., 2014; Van Maren et al., 2015b). Compounding this paradox, synoptic observations from the Ems-Dollard Measurements (EDoM) campaign also revealed substantial vertical variability in sediment transport and suggested net seaward sediment transport over much of the water column (Van Maren et al., 2023). Together, these hydrodynamic indicators and water-column observations raise the unresolved question of how sediment is maintained and transported upstream within the Emden Fairway despite apparently export-favorable conditions.

One possible explanation is that existing observations insufficiently resolve transport processes within the near-bed layer, where a missing landward flux may be present. In hyperturbid estuaries, near-bed SSC is strongly influenced by concentration-dependent settling and resuspension processes, which can support a very high suspended sediment concentration layer (Dijkstra et al., 2018; Defontaine et al., 2024). At the same time, fluid mud can strongly damp near-bed turbulence and enhance vertical stratification (Winterwerp, 2001; Becker et al., 2018; Woods, 1969; Winterwerp, 2006). These processes may alter tidal deformation, effective hydraulic drag, and estuarine circulation, thereby enhancing net landward sediment transport (Zhu et al., 2021; Moodie et al., 2022; Van Maren et al., 2015b). Numerical studies have demonstrated the potential importance of these feedbacks (Dijkstra et al., 2018; Winterwerp, 2011), yet field observations capable of resolving both velocity and SSC within near-bed high-concentration layers remain scarce because of the technical challenges associated with measurements in fluid mud (Sottolichio et al., 2011). As a result, the contribution of the near-bed layer to the depth-integrated residual sediment flux remains poorly constrained, leaving the apparent sediment budget unbalanced.

To advance our understanding of sediment transport mechanisms under hyperturbid conditions, we investigate the seasonal variability of fluid mud thickness, flow structure and sediment transport within the landward section of the hyperturbid Emden



Fairway in the Ems Estuary, with particular focus on the near-bed layer. Using vessel-based downward-looking Acoustic Doppler Current Profiler (ADCP) measurements combined with direct water sampling, we reconstruct high-resolution vertical profiles of velocity and SSC over full tidal cycles. The resulting observations are used to quantify depth-resolved residual sediment fluxes and to assess how near-bed stratification and fluid mud modify flow structure and net sediment transport. By comparing periods with and without persistent near-bed high-concentration layers, we further evaluate the role of fluid mud in controlling landward sediment transport within the Emden Fairway.

2 Material and Methods

2.1 Study site and data collection

The study area is located in the Emden Fairway, a navigation channel in the German part of the Ems Estuary (Fig. 1). The channel depth at the observed section varies between 10 and 20 m, and is laterally bounded by a large tidal flat area called the Dollard Bay along its southern slope. The channel is characterized by a strong mean horizontal SSC gradient, ranging from $\sim 0.1 \text{ g L}^{-1}$ on the western (seaward) side to several grams per liter near the eastern (landward) side within several kilometers (Van Maren et al., 2023). In the lower Ems River, a highly concentrated benthic suspension (fluid mud) frequently develops close to the bed (Talke et al., 2009). Approximately 1.6 million tons of fine-grained sediments are annually dredged from the port of Emden and 3.6 million tons from the fairway (Van Maren et al., 2015a, 2023). The mud content of these deposits is close to 100% near Emden, with an organic fraction reaching 5% (Van Maren et al., 2016). River discharge from upstream varies seasonally from several tens to hundreds of $\text{m}^3 \text{ s}^{-1}$, with high discharge typically occurring in winter (driven by precipitation in the catchment) and low discharge prevailing in summer; the three surveys were timed to capture contrasting phases of this cycle ($200 \text{ m}^3 \text{ s}^{-1}$ in February 2022, $125 \text{ m}^3 \text{ s}^{-1}$ in January 2025, and $45 \text{ m}^3 \text{ s}^{-1}$ in September 2024, Fig. 1c, data from BSH).

The boat surveys were executed under neap-tide conditions on 10 February 2022, 29 September 2024, and 24 January 2025 using a vessel-mounted ADCP and a rigid, custom-made mobile sampling frame (Fig. 1e). The ADCP is a downward-looking Nortek Signature 500 kHz ADCP, featuring four slanted beams (25° from the vertical) for velocity measurements and one dedicated vertical beam for echo sounding. All beams operate at an acoustic frequency of 500 kHz. The vertical beam is designed for high-resolution (0.012 m) echo sounder measurements to capture small-scale vertical variations in SSC, utilizing a smaller bin size compared to the four slanted beams (0.5 m) used for velocity measurements. The instrument blanking range is 0.5 m. Local bathymetry (Fig. 1d) was obtained from the Wasserstraßen- und Schifffahrtsamt (WSA) Ems-Nordsee. The frame housed four sampling nozzles connected to boat-mounted Verderflex isokinetic pumps and a conductivity–temperature–depth (CTD) sensor. The nozzles (4.5 mm diameter) were positioned at varying vertical distances below the cross-beam: 1.9, 1.8, 1.6, and 1.2 m in 2022, and 1.8, 1.6, 1.2, and 0.5 m in the 2024 and 2025 surveys. A pressure sensor was mounted on the cross-beam to determine the location of the water samples. The nozzle inlets were laterally offset from the ADCP measurement location by up to approximately 2 m, as the sampling frame was deployed from the opposite side of the vessel to minimize the influence of frame-induced disturbance on the ADCP-derived velocity profiles. This lateral offset is assumed to have a negligible effect on

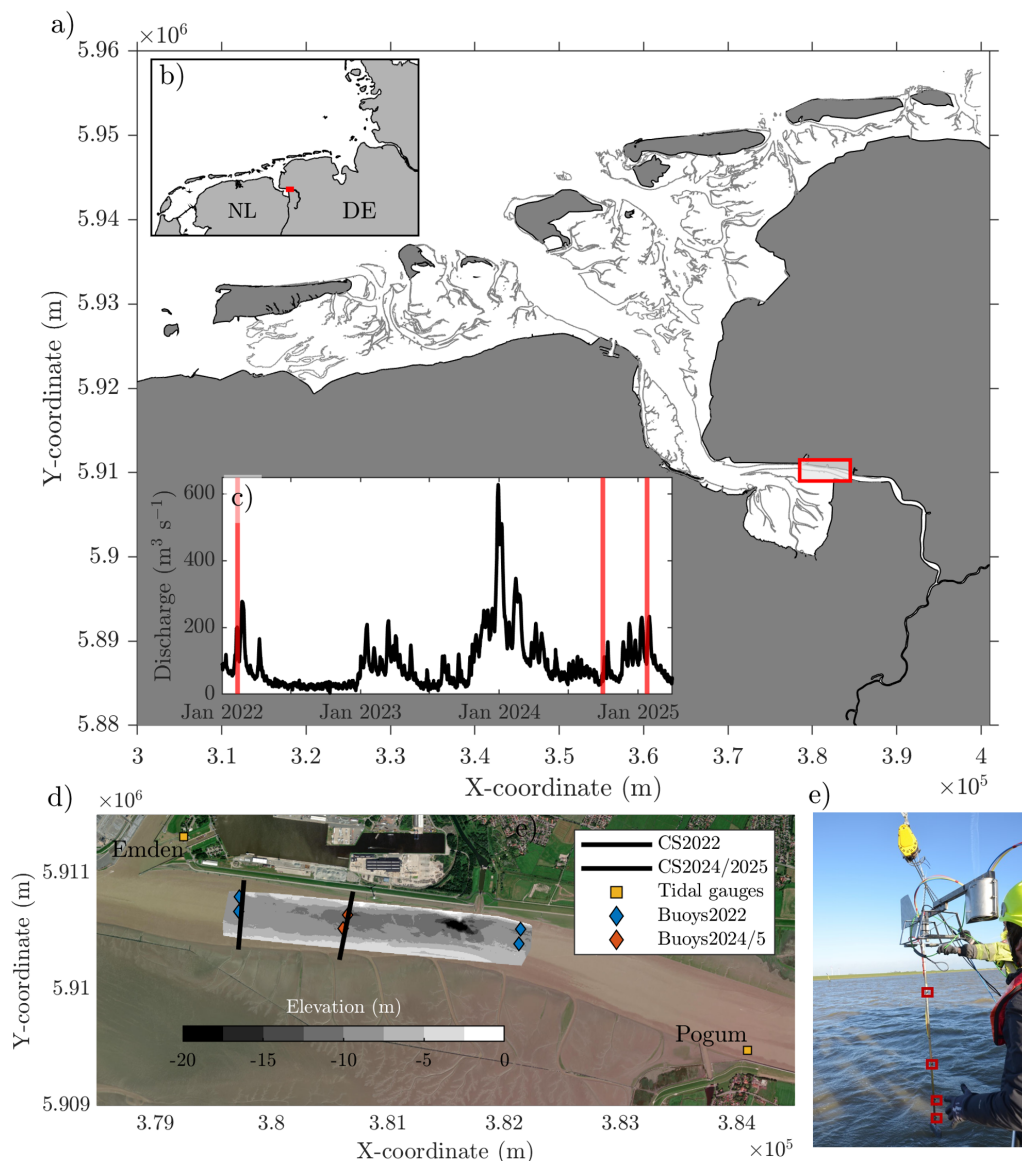


Figure 1. Study area and a photo of the mobile frame on the survey boat. (a) Study area in the Ems Estuary. (b) Map of the estuary location at the Dutch-German border. (c) River discharge at Versen station. Red lines indicate the timing of the three surveys. (d) Detail of the survey region in the Emden Fairway, displaying the local bathymetry and measurement locations. Blue dots represent sampling locations in the 2022 survey and red dots are sampling locations in the 2024/2025 survey. (e) Nozzles of tubes for sampling on the mobile frame. Map source of (a) and (b): Rijkswaterstaat servicedesk (<https://www.rijkswaterstaat.nl/formulieren/contactformulier-servicedesk-data>). In (d), the background satellite imagery is from Google Maps (Imagery © 2025 Google), and the bathymetry was provided by WSA Ems-Nordsee.



Table 1. Summary of measurements during the three campaigns

Campaign	Fixed positions	Fixed-position profiles	Water samples	Moving-boat transects
2022	NW, NE, SW, SE	4×4	64	8
2024	NB, SB	14×2	168	28
2025	NB, SB	13×2	153	26

the calibration, as lateral differences in sediment concentration are negligible at this scale, with the dominant gradient occurring in the vertical direction. The pumped water samples were collected in 1-liter bottles, each taking about 2 min to fill completely after a 3-min flushing of the tubes. This yields a 2-min averaged SSC value.

The field surveys combined two complementary measurement strategies: moving-boat transects and fixed-position measurements. During the moving-boat surveys, the vessel repeatedly navigated across the channel (indicated in Fig. 1d) to obtain cross-sectional profiles of velocity and acoustic backscatter intensity throughout the tidal cycle. These transects provided spatial information on the flow structure and fluid mud distribution across the Emden Fairway. In addition, fixed-position measurements were conducted by alternately anchoring the vessel to designated buoys, allowing detailed observations of the temporal evolution of velocity, SSC and salinity profiles at specific locations.

During the fixed-position measurements, a standardized sampling procedure was followed for each vertical profile at the fixed buoys (in 2022: northwest (NW), northeast (NE), southwest (SW) and southeast (SE); in 2024 and 2025: the North Buoy (NB) and South Buoy (SB); Fig. 1d). First, the mobile sampling frame was lowered into the water column until it hovered approximately 2 m above the bed. After deployment, the frame remained at this elevation for several minutes to allow local flow velocities and suspended sediment distributions to recover from deployment-induced disturbances. During this stabilization period, typically lasting around 5 min, the pumps were simultaneously activated to flush residual water from previous samplings. Subsequently, continuous pumping was performed for approximately 2 min to fill the sample bottles, yielding four simultaneous water samples within the near-bed layer. The enhanced sampling resolution near the bed aims to resolve the strong vertical concentration gradients expected under hyperturbid conditions. Following the near-bed sampling, the frame was raised to mid-depth, where an additional water sample was collected to characterize suspended sediment concentrations in the central water column. At the same time, a sixth sample was manually collected near the free surface at approximately 0.1 m water depth. This six-sample protocol was followed during the 2024 and 2025 surveys. During the 2022 survey, only the four near-bed nozzle samples were collected at each profile, without raising the frame to mid-depth or sampling near the free surface.

Between successive fixed-position measurements, the sampling frame was lifted entirely out of the water and the vessel resumed moving-boat operations to perform cross-sectional transects of velocity and acoustic backscatter intensity. Measurement efforts for the three field campaigns are summarized in Table 1.



120 Mass concentrations were determined through laboratory analyses of the water samples collected during the sampling campaigns in 2022 (16 profiles; 64 samples), 2024 (28 profiles; 168 samples), and 2025 (26 profiles; 153 samples) (Table 1). Filtering failed for three samples collected in the 2025 campaign because of extremely high concentrations.

The SSC of all water samples was determined using a two-stage filtration procedure to prevent filter clogging, and a vacuum pump was employed to accelerate processing. The water samples were sequentially passed through a coarse pre-filter (pore size $\sim 2.5 \mu\text{m}$) and subsequently through a $0.45 \mu\text{m}$ filter. Both filters were then oven-dried and weighed to calculate the total cumulative sediment mass concentration. These in situ SSC measurements were subsequently used to calibrate and invert the ADCP-derived acoustic backscatter data, enabling reconstruction of continuous SSC profiles throughout the water column and calculation of depth-integrated sediment fluxes. The ADCP backscatter conversion exploits a recent processing approach described in Bao et al. (2026) and Sassi et al. (2012), which is briefly summarized in Appendices A and B.

130 2.2 Data processing and sediment flux calculation

High SSC in hyperturbid environments introduces strong attenuation and non-linear scattering, causing structural slope breaks in the measured acoustic backscatter profiles. These discontinuities invalidate standard single-relation acoustic inversion across the full water column (Sassi et al., 2012). To overcome this limitation, we developed a hybrid processing framework that combines automated changepoint detection, segmented acoustic inversion, and direct water sample measurements to reconstruct continuous SSC profiles and compute sediment fluxes.

2.2.1 Data pre-processing and changepoint detection

Prior to analysis, acoustic backscatter signals below the detected bathymetric boundary were removed to exclude bed reflections. Vertical profiles of backscatter (specifically, $S'_{v,\alpha_s=0}$, as defined in Appendix B) from moving-boat transects were averaged over 5 s intervals to suppress short-timescale acoustic turbulence while preserving structural interfaces. The backscatter profiles from the fixed-location measurements were averaged over 2-min windows to align with the 2-min averaged physical water samples, and minimize non-sediment-related noise.

To identify structural breaks induced by high near-bed SSC gradients or environmental noise (e.g., near-surface disturbance from boat traffic or wind), we applied an automated piecewise linear segmentation approach (Appendix A) to each vertical backscatter profile, following the optimal partitioning framework of Killick et al. (2012) with a "MinThreshold" penalty of 400. The detected changepoints were used to identify the depth interval of each profile suitable for conventional acoustic inversion, as described in Sect. 2.2.2. Using this method, we further define the changepoint corresponding to the maximum backscatter value in the moving-boat transects as the candidate fluid mud signature around high water slack (HWS). Such patterns are typically associated with the enhanced acoustic sensitivity to denser sediment aggregates at the interface between dilute fluid and fluid mud, corresponding to the lutocline (Ha et al., 2011).

150 For each survey, eight representative transects were selected at nominal phases relative to high water level (HW) and low water level (LW), identified independently from the corresponding water level. The water levels at the observation transects were interpolated from observations at the nearby Emden and Pogum tidal stations to account for spatial variations associated



with tidal propagation. Because the survey duration (about 13 h) exceeded one full semidiurnal tidal cycle but began at a different absolute tidal phase in each of the three surveys, the chronological order in which these eight transects were measured does not necessarily follow their nominal phase order.

2.2.2 SSC profiles

Continuous SSC profiles were reconstructed using a three-layer hybrid approach. The three layers are separated by the first and second changepoints in the backscatter profile below the surface, which mark where the backscatter–SSC relationship may change. The first changepoint typically occurs about 2 m below the surface, reflecting near-surface disturbances from vessel motion and wind. Between the two changepoints, SSC was derived from 2-min averaged vertical-beam backscatter of the ADCP, using an explicit inversion that corrects for sediment-induced attenuation (Thorne and Hanes, 2002; Sassi et al., 2012). Within this segment, the mid-depth water sample was used to determine one of the two calibration coefficients ($K(R_{ref})$, see Appendix B) and the lowest sample above the second changepoint to determine the other (ξ_{total} , Appendix B); any samples in between were used for independent validation. Further details of the inversion and calibration are given in Appendix B. The surface and near-bed SSCs outside the inversion range are estimated as follows. Above the first changepoint, a linear line was fitted to the surface water sample and the uppermost inverted value, and extrapolated to the surface. Below the second changepoint, SSC was linearly fitted to the lowest inverted value and the near-bed water samples, the lowest of which was collected about 0.2 m above the consolidated bed (in the 2024/2025 surveys). The fitted line was then extrapolated to the bed. For profiles without a second changepoint, the inversion was applied down to the lowest sampling level.

2.2.3 Velocity profile and sediment flux

Velocity data from both the moving-boat transects and the fixed-position measurements were processed with the `adcptools` toolbox (<https://github.com/bartverm/adcptools>), which implements the method of Vermeulen et al. (2014). The moving-boat transect data were processed following Jongbloed et al. (2025), who developed a physics-informed spatial interpolation approach, based on Vermeulen et al. (2014), to reconstruct turbulence-averaged three-dimensional residual flow velocities across the channel. In both data sets, measurements affected by side-lobe interference were removed per beam. Within the affected near-bed layer, velocity was linearly interpolated from the lowest reliable bin to zero at the consolidated bed. The fixed-position velocity profiles were then averaged over the same 2-min intervals as the backscatter data and used for the sediment flux calculations. Water levels at the buoy locations were interpolated from observations at the nearby Emden and Pogum tidal stations.

Depth-resolved sediment fluxes were calculated from the product of SSC and velocity based on the 2024 and 2025 surveys. Because velocity measurements were unavailable within the lowest acoustic bin (~ 0.5 m) immediately above the consolidated bed, sediment flux in this unresolved layer could not directly be quantified. We therefore prescribed a zero sediment flux at the consolidated bed. This assumption is supported by observations of fluid mud dynamics in the Ems River, which indicate negligible transport close to the consolidated bed where velocities within the fluid mud layer are strongly reduced and the layer remains roughly stationary (Becker et al., 2018).



To standardize the vertical grid for profile analysis and flux decomposition, the high-frequency velocity and sediment data were analyzed in a vertical sigma coordinate system ($\sigma = z/H$, where H is the instantaneous water depth). Specifically, velocity profiles were vertically interpolated onto the exact grid levels where valid SSC observations were available, ensuring a fully synchronized and data-driven grid within each σ -layer (Klingbeil et al., 2019). To separate the contributions of residual
190 circulation and tidal pumping to the net sediment transport, the tidally averaged sediment flux was decomposed according to:

$$\langle uC \rangle = u_0 C_0 + \langle u' C' \rangle \quad (1)$$

where u_0 and C_0 denote the tidal means computed over an M_2 tidal cycle, and the primes indicate tidal fluctuations relative to the mean. The first term ($u_0 C_0$) represents transport associated with residual circulation, whereas the second term ($\langle u' C' \rangle$) represents tidal pumping driven by covariance between velocity and concentration fluctuations.

195 To examine the vertical structure of residual sediment transport, the water column was partitioned into an upper layer and a near-bed layer separated at 30% of the total water depth ($z/H = 0.3$). This threshold was selected based on preliminary inspection of tidally averaged sediment flux profiles, which consistently showed enhanced transport within the lower 20–40% of the water column. Because only near-bed water samples were collected during the 2022 survey, full-depth SSC reconstruction and depth-integrated sediment flux calculations were performed exclusively for the 2024 and 2025 campaigns.

200 2.2.4 Richardson number

To assess the impact of stratification on flow structure, we calculated the gradient Richardson number (Richardson, 1920):

$$Ri_g = -\frac{g}{\rho} \frac{\partial \rho / \partial z}{(\partial u / \partial z)^2 + (\partial v / \partial z)^2} \quad (2)$$

where z is positive upward, u and v are the two horizontal velocity components measured by the ADCP. The density ρ is approximated as a linear function of suspended sediment mass concentration M_s and salinity S , neglecting temperature effects:

205

$$\rho = \rho_w + nM_s + mS \quad (3)$$

Herein, $n = 1 - \rho_w / \rho_s = 0.62$ accounts for the volume of water displaced by suspended sediment, with $\rho_w = 1000 \text{ kg m}^{-3}$ and $\rho_s = 2650 \text{ kg m}^{-3}$. m is a haline contraction coefficient of $m \approx 0.82 \text{ kg m}^{-3} \text{ psu}^{-1}$ (Talke et al., 2009). M_s is obtained from the reconstructed SSC profile and S from CTD casts (Appendix C). The Richardson number accounting only for sediment-
210 induced density variations, with $m = 0$ in Eq. (3), is denoted as Ri_{SSC} . $Ri < 0.25$ indicates that the flow is susceptible to shear instability and turbulent mixing (Miles, 1961).

2.2.5 Horizontal density gradients

To quantify the contribution of horizontal density gradients to the residual circulation, near-bed water samples collected during the 2022 survey at the four fixed buoys (Fig. 1d) were grouped into a western pair and an eastern pair, separated by approxi-
215 mately 2 km along the fairway. For each pair, the near-bed mass concentration M_s was averaged over all available samples (32



samples per pair, pooled across tidal phases to represent a tidally averaged state) and converted to density using Eq. (3). The near-bed horizontal density gradient induced by sediment ($\partial\rho_{ssc}/\partial x$) was then estimated as the difference between the tidally averaged near-bed densities at the eastern and western buoy pairs, divided by their along-channel separation.

Because only near-bed water samples were collected during the 2022 survey, the near-surface density gradient could not be estimated from the buoy data. Instead, historical surface SSC records at the Emden and Pogum tidal stations (around 5 km apart) were obtained from the Lower Saxony Water Management, Coastal Defence and Nature Conservation Agency (NLWKN) and used to estimate the near-surface sediment concentration gradient. The background horizontal salinity gradient was estimated analogously from the salinity records at the same two stations.

3 Results

3.1 Seasonal and intra-tidal variability

Figure 2 presents the acoustic backscatter transects obtained during the three field campaigns at eight nominal tidal phases. Several transects exhibit a strong lutocline in the backscatter intensity, corresponding to the upper boundary of the fluid mud layer.

Pronounced seasonal differences in fluid mud occurrence were observed among the three campaigns. During the 2022 survey, which coincided with the highest river discharge ($200 \text{ m}^3 \text{ s}^{-1}$, compared to $45 \text{ m}^3 \text{ s}^{-1}$ in 2024 and $125 \text{ m}^3 \text{ s}^{-1}$ in 2025), an extensive fluid mud layer persisted throughout much of the tidal cycle. The thickness around HWS exceeded 2 m in the 2022 survey, while a thinner fluid mud layer existed, confined to the deep channel in the 2025 survey (about 1 m). The bottom signal was strongly attenuated, and smooth near-bed deposits covered large portions of the channel, particularly around HW. In contrast, no persistent fluid mud layer was detected during the low-discharge 2024 survey. During the 2025 survey, a thinner and more intermittent fluid mud layer was observed, with changepoints alternating between elevated positions in the water column and locations close to the bed. This pattern is consistent with a discharge-driven seaward intrusion of the fluid mud pool.

In addition to the seasonal variability, the transects reveal systematic intra-tidal changes in the vertical sediment structure. Around HW, the upper water column remained relatively clear while a compact fluid mud layer settled near the bed, forming a relatively flat and well-defined lutocline. During periods of increasing tidal currents, the fluid mud layer expanded vertically upward into the water column and the top of the fluid mud layer became progressively less distinct. Around LW, suspended sediments settled again toward the bed, although the interface between the upper water column and the near-bed layer remained more diffuse than during HW. By the end of the tidal cycle in the 2022 survey, the thickness of the fluid mud layer was largely restored.

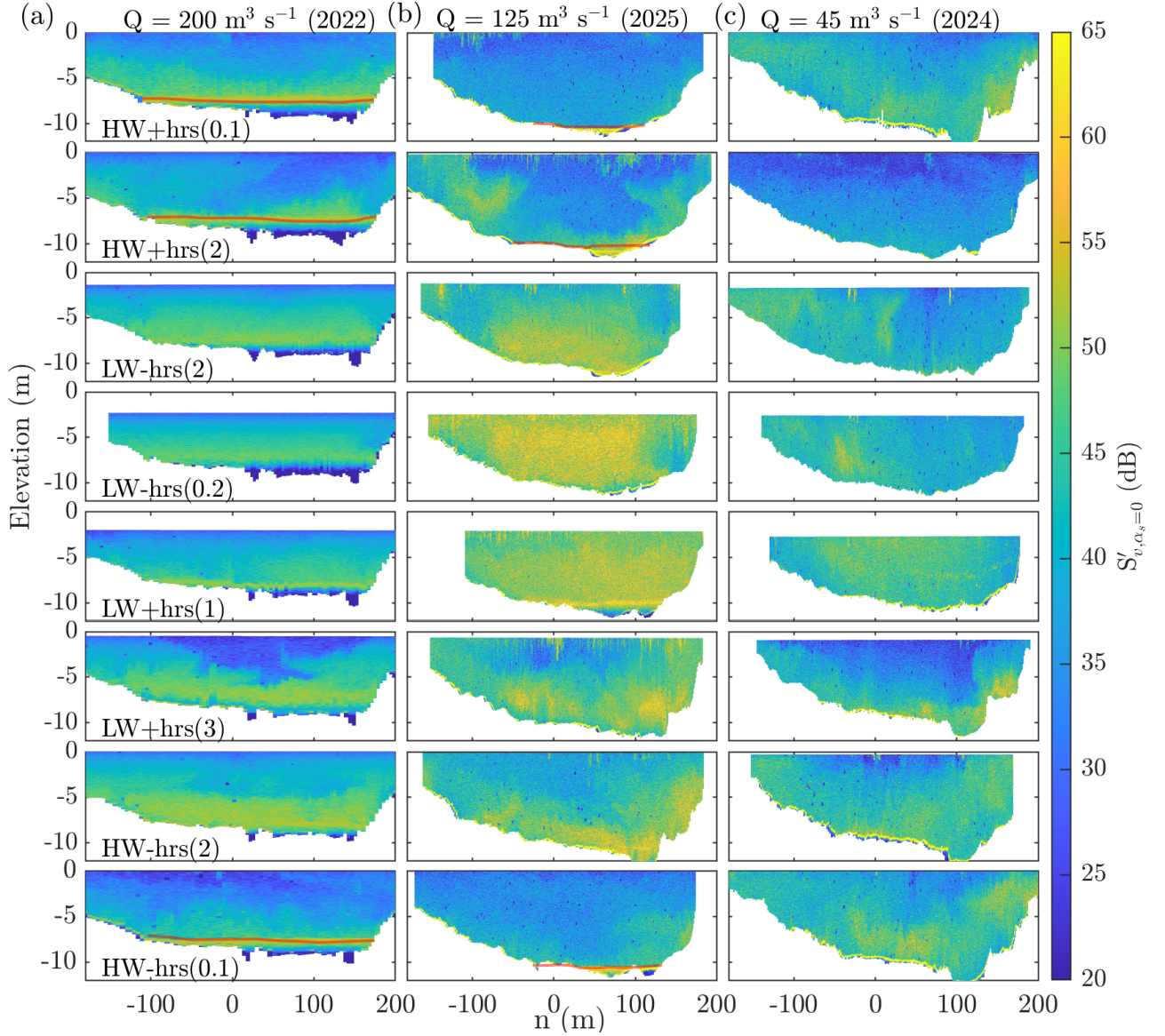


Figure 2. Acoustic backscatter transects from the 2022 (a), 2025 (b) and 2024 (c) field survey campaigns when discharge was 200, 125 and $45 \text{ m}^3 \text{ s}^{-1}$. n is distance from center of the transects (m). Y-axis is the elevation referring to NAP, which is the Dutch ordnance level, close to mean sea level. The location of the monitored transects is indicated in Fig. 1d. Data shown in the transects are bounded by the measured bathymetry. The red line indicates the top of the fluid mud layer, identified through the changepoint detection. HW: high water; LW: low water. Rows for the 2025 and 2024 surveys are ordered by nominal tidal phase relative to HW and LW for each survey independently, rather than by chronological measurement sequence.

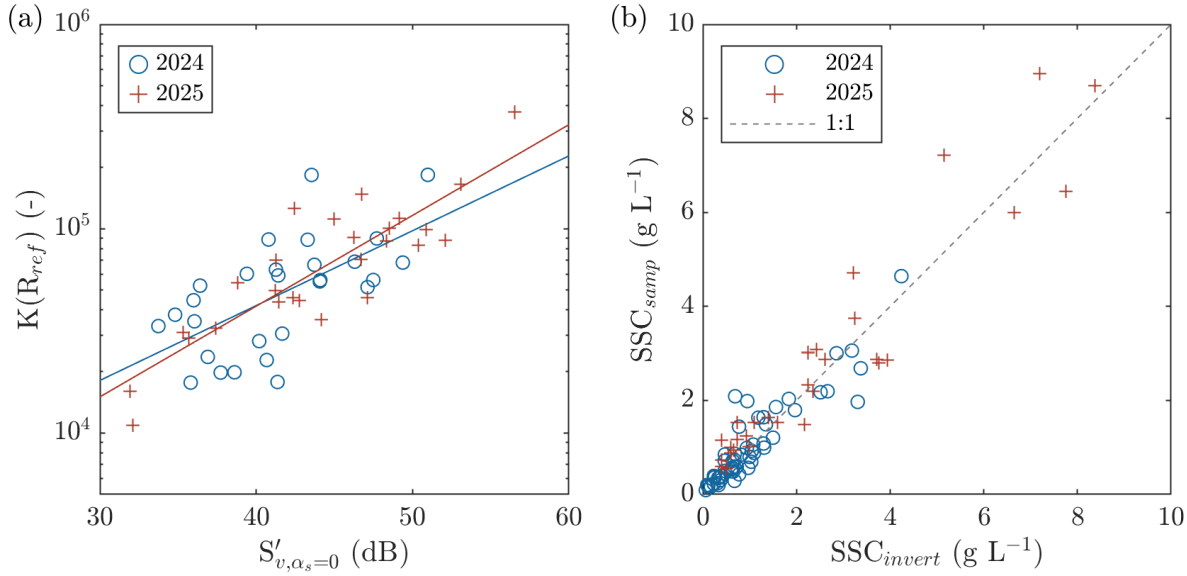


Figure 3. Calibration and validation. (a) Linear regression between $K(R_{ref})$ and $S'_{v, \alpha_s=0}$ using Eq. (B5). (b) Validation results by comparison of the acoustically derived concentration (SSC_{invert}) and measured concentration (SSC_{samp}) using the water samples within the inversion range detected by changepoints.

245 3.2 Acoustic inversion and validation

In Fig. 3a, we apply the calibration based on Eq. (B5), including data from two sites of each survey separately. The regression lines in Fig. 3a determine the calibration parameter ($K(R_{ref})$) based on backscatter strength ($S'_{\alpha_s=0}$) at the reference range for each profile. The calibration suggests the suspensions at the mid-depth water samples generally fit the homogeneity requirement, as expected. Values of the regression slope between $K(R_{ref})$ and $S'_{\alpha_s=0}$ amount to 0.38 ± 0.09 and 0.42 ± 0.05 for
 250 2024 and 2025 surveys, respectively. The estimates of the linear relationship show limited variation between seasons.

Figure 3b shows the validation of the ADCP-derived estimates of SSC using Eq. (B3). The ADCP-derived concentrations are compared with those from water samples located within the inversion range with a coefficient of determination $r^2 = 0.94$, which is satisfactory for the field setting and for the following hybrid reconstruction of SSC profiles.

3.3 SSC and sediment fluxes

255 Figure 4 summarizes the water level, velocity structure, reconstructed SSC profiles, and stratification during the 13-h observation periods in 2024 and 2025. Water levels were interpolated from the Emden and Pogum tidal stations, while velocity and SSC profiles correspond to measurements collected at the North Buoy (Fig. 1d). Slack water is defined as $|u| < 0.1 \text{ m s}^{-1}$ in this study. Figure 4a and b show that slack water lagged high and low water in both surveys, with a considerably longer lag in 2024 (1–2 h) than in 2025 (shorter than the $\sim 1 \text{ h}$ sampling interval). Although peak flood and ebb velocities were comparable

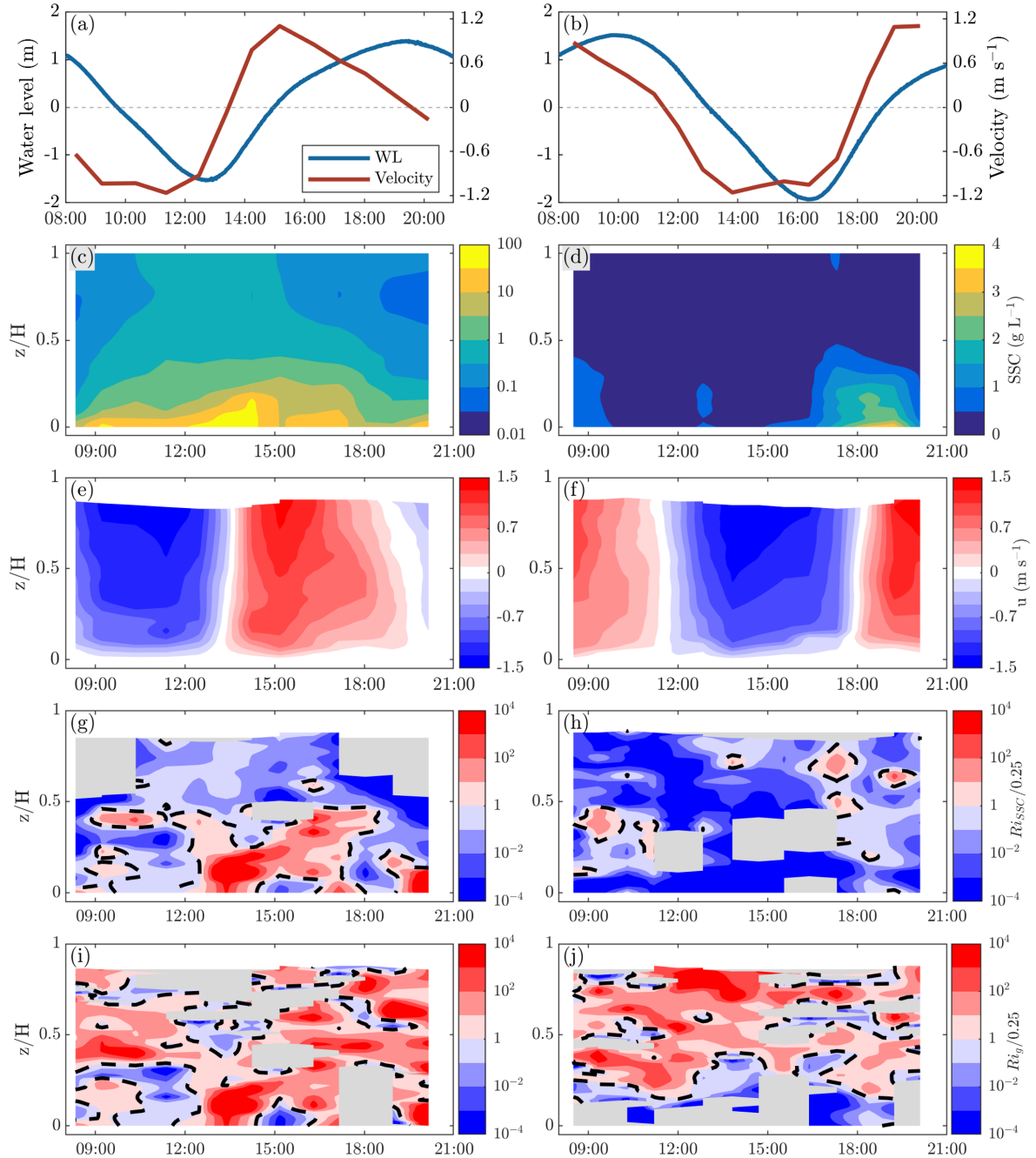


Figure 4. Water level, velocity, SSC, and Richardson number during the ~ 13 h surveys in January 2025 (a,c,e,g,i) and September 2024 (b,d,f,h,j). (a) and (b) Water level and depth-averaged along-channel velocity, positive value means landward flow. (c) and (d) Suspended sediment concentration (SSC). (e) and (f) Tidal velocity as a function of normalized height z/H . Red (positive) indicates landward direction. (g) and (h) Richardson number induced by SSC. (i) and (j) Gradient Richardson number, including both salinity and SSC. In (g)–(j), colours show $Ri/0.25$ on a logarithmic scale, and the dashed black contours mark $Ri = 0.25$; above this value, turbulent mixing is suppressed. Grey areas indicate negative Richardson numbers. Salinity is shown in Fig. C1.



260 in magnitude, the transitions between ebb and flood flow differed substantially in shape, particularly around slack water. In both surveys, the duration of slack water was longer at HWS than at low water slack (LWS).

During the 2025 survey, a fluid mud layer with near-bed SSC exceeding 70 g L^{-1} was present. The corresponding SSC, velocity, Ri_{SSC} and Ri_g are shown in Fig. 4c, e, g and i. During ebb acceleration, near-bed SSC briefly peaked ($>32 \text{ g L}^{-1}$) at around 09:00, roughly 2 h before the near-bed peak ebb velocity ($\sim 11:00$), and had already declined by then, so that peak ebb
 265 velocity coincided with reduced SSC. Note that the early part of this ebb was not captured by the survey; early-ebb conditions are instead inferred from the following ebb at the end of the record (after $\sim 19:30$). At the onset of the flood tide, the flow reversal appeared first near the bed and propagated upward through the water column, reaching the near-surface layer about 30 min later. This bed-to-surface progression coincided with an abrupt increase in near-bed SSC, which, as during ebb, reached its maximum during the flood acceleration phase and before the near-bed peak flood velocity (around 15:00). At this time, the
 270 regions where both near-bed Ri_{SSC} and Ri_g exceeded 0.25 expanded synchronously near the bed (Fig. 4g,i), indicating that the near-bed stratification, and hence the suppression of turbulent mixing, was dominated by suspended sediment rather than salinity.

Unlike the ebb pattern, near-bed SSC remained elevated through the peak flood velocity, declining only afterward. Around peak flood, near-bed Ri_{SSC} and Ri_g briefly dropped below 0.25 where velocity decreased strongly in the vertical, favoring
 275 turbulence production and allowing entrainment of the fluid mud into the water column above. This is consistent with the flood-phase breakdown of mud-induced stratification reported for the Ems River by Becker et al. (2018). Following peak flood in 2025, a velocity maximum developed at mid-depth ($z/H \approx 0.5$), where the velocity exceeded that at the surface and reached up to 0.9 m s^{-1} (Fig. 4e). As the flood decelerated, the maximum descended and weakened, approximately following the 1 g L^{-1} SSC isoline, and reached about 0.3 m s^{-1} at $z/H \approx 0.25$ before the near-bed flow reversed. The cause of such mid-depth
 280 velocity maxima is not fully understood (Tu et al., 2019; Chant et al., 2007).

At around HWS, the near-surface layer had already begun ebbing while the near-bed layer remained close to slack, not yet reversing to seaward flow. In contrast to early flood, when Ri_g closely followed Ri_{SSC} near the bed, SSC-induced stratification was now confined close to the bed ($z/H < 0.2$), consistent with the low sediment concentration in the rest of the water column. Ri_g , however, remained elevated throughout the water column, indicating that salinity stratification extended the suppression
 285 of turbulent mixing well beyond the SSC-dominated near-bed layer. At the end of the record, during early ebb, the flow accelerated more gradually than during flood, initially only in the upper water column. Near-bed SSC responded only weakly, likely because the near-bed velocity remained low during the delayed transition to seaward flow. Since the near-bed flow turned landward at $\sim 13:00$ and had not yet reversed to seaward flow by the end of the record ($\sim 20:30$), near-bed seaward flow could have persisted for at most $\sim 5 \text{ h}$ of the $\sim 12.4 \text{ h}$ tidal cycle.

290 In comparison, the 2024 survey showed low SSC and weak stratification, with the highest SSC values reaching only 4 g L^{-1} (Fig. 4d,f,h,j). Values of Ri_{SSC} and Ri_g remained mostly below 0.25 (or negative) near the bed during both flood and ebb. Near-bed SSC showed a brief, localized elevation ($0.5\text{--}1.0 \text{ g L}^{-1}$) during early ebb, followed by a stronger and more sustained rise that began before flood onset and continued throughout the flood period. This mirrors the flood-dominant entrainment asymmetry observed in the 2025 survey, but at a much weaker overall magnitude. The weaker entrainment is consistent with



295 the absence of a substantial near-bed fluid mud pool available for resuspension. During ebb onset in 2024, flow in the near-bed layer lagged behind flow in the upper water column. Velocity near bed subsequently grew only modestly over the following 2 h while the upper column rapidly approached its ebb peak. Ri_g remained below 0.25 (or negative) near the bed throughout this period.

Across the two surveys, several patches of negative Ri_{SSC} and Ri_g were observed, but at distinct locations. The negative Ri_{SSC} may be caused by sediment clouds, such as those visible in the 2022 backscatter profiles (Fig. S1 in the Supplement), which can introduce locally reversed SSC gradients. There are time windows of negative Ri_g values occurring at the mid-depth peak velocity, which may relate to the locally reversed salinity gradient (Fig. C1). A similar pattern was observed in the lower Ems River (Becker et al., 2018). Generally, $Ri_{SSC} > 0.25$ occurs mainly near the bed, reflecting the concentration of SSC-driven stratification in the near-bed layer, whereas $Ri_g > 0.25$ extends further into the upper water column once the salinity contribution enhances the density gradient.

Overall, comparison of the 2024 and 2025 surveys shows a consistent tidal asymmetry favoring landward sediment transport: near-bed SSC was higher during flood than during ebb, and HWS lasted longer than LWS, allowing suspended sediment more time to settle before the following ebb. The fluid mud layer provided an abundant near-bed sediment source, consistent with the much stronger near-bed SSC asymmetry in 2025, whereas the flood dominance of this asymmetry was present in both surveys, albeit much weaker in 2024.

The depth-integrated sediment flux exhibited clear semi-diurnal variability (Fig. 5). Sediment transport was predominantly landward during flood and seaward during ebb, with maximum fluxes occurring close to peak tidal currents. Around HWS, the transition from landward to seaward transport occurred gradually, whereas the reversal around LWS occurred over a substantially shorter interval.

315 Strong seasonal differences were observed in the vertical distribution of sediment transport. During the January 2025 survey, when fluid mud was present and SSC reached several tens of grams per liter, sediment transport was strongly concentrated within the near-bed layer, particularly during flood when stratification was strong. In contrast, during the September 2024 survey, flux magnitudes were substantially smaller and distributed more evenly throughout the water column. The near-bed contribution to the total sediment flux was therefore substantially larger during the 2025 campaign than during 2024.

320 The tidal variations of flux are consistent with the velocity and SSC patterns in Fig. 4. Sediment flux buildup during the onset of flood was much larger than during the onset of ebb in both surveys. The transition at the North Buoy around LWS, characterized by a strong landward near-bed flux and a weak surface flux, is consistent with the flow reversal appearing first near the bed (Fig. 4e). The near-bed flux contributes more during flood than during ebb, as fluid mud supplies abundant sediment for resuspension during flood.

325 Table 2 shows the sediment flux of the fixed-position measurements. The near-bed layer ($z/H < 0.3$) accounted for all of the net landward transport during both surveys. The net flux during the 2025 survey, when fluid mud was present, was about one order of magnitude larger than that during the 2024 survey. The near-bed layer also carried a larger share of the gross (flood plus ebb) transport in 2025 (66–68 %) than in 2024 (44 %). These results indicate that the near-bed layer, despite occupying

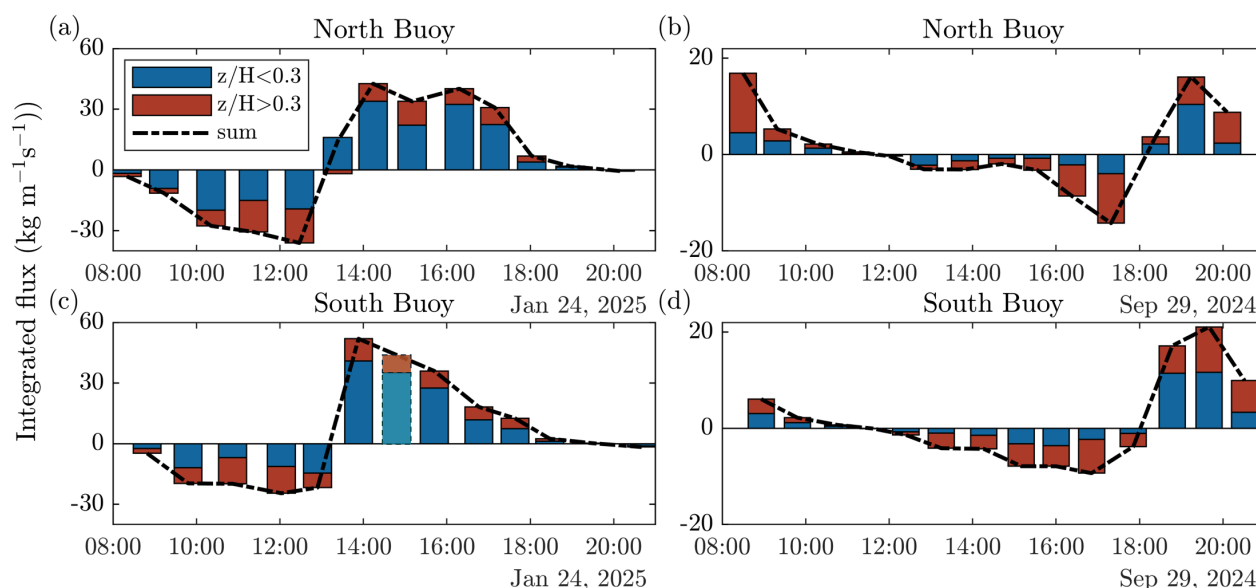


Figure 5. Depth-integrated sediment flux over the tidal cycle at the North Buoy (a, b) and South Buoy (c, d) during the January 2025 (a, c) and September 2024 (b, d) surveys. The flux is separated into the near-bed layer ($z/H < 0.3$; blue bars) and the upper layer ($z/H > 0.3$; red bars), and the dash-dotted black line denotes their sum. Positive values indicate landward transport. The lighter bar in (c) was linearly interpolated from the two adjacent measurements because near-bed data were unavailable at that time.

Table 2. Time-integrated sediment flux per unit width in the near-bed layer ($z/H < 0.3$) and the upper layer ($z/H > 0.3$) during the 2024 and 2025 surveys. The flux is integrated over the measurement period. Positive values indicate landward transport. The near-bed share is the fraction of the gross (flood plus ebb) transport carried by the near-bed layer.

Campaign	Position	Net flux (t m^{-1})			Near-bed share (%)	Duration (h)
		$z/H < 0.3$	$z/H > 0.3$	Total		
2025	NB	+207	−31	+176	68	11.9
	SB	+241	−38	+204	66	12.4
2024	NB	+30	−12	+18	44	11.7
	SB	+41	−25	+17	44	11.5

only the lowest 30 % of the total water depth, forms the primary pathway for residual landward sediment transport in the Emden Fairway, particularly when fluid mud is present.

3.4 Residual flow and entrainment

To separate the mechanisms driving net sediment transport, we first examine the residual flow structure using the ADCP moving-boat data from the 2022, 2024, and 2025 surveys. Clear structural contrasts are observed across the different sea-



sons (high river discharge vs. low discharge), which coincide with major shifts in the net sediment transport. During the high-discharge surveys (February 2022 and January 2025), a pronounced two-layer estuarine circulation developed across the transect, characterized by a landward residual flow near the bed and a seaward flow near the surface (Fig. 6a). This landward residual flow coincides with elevated near-bed SSC, driving net landward sediment transport. In contrast, during the low-discharge survey (September 2024), the residual flow was predominantly seaward throughout the entire water column.

As described previously, the asymmetry in flow acceleration is associated with greater sediment entrainment during flood than during ebb. During the 2024 survey, the vertical structure and magnitude of the sediment flux were broadly consistent between the North Buoy (NB) and the South Buoy (SB), indicating weak cross-sectional variability. For both verticals, the residual advection term ($u_0 C_0$) was seaward and remained small over most of the water column. In contrast, the tidal pumping ($\langle u' C' \rangle$) dominated and was bottom-intensified, aligning with the phase-dependent resuspension documented in Fig. 4d and f. This suggests that net landward transport during the low-discharge season is primarily controlled by tidal pumping rather than by a spatially structured residual circulation.

During the January 2025 survey, the flux distribution became markedly non-uniform between NB and SB. The near-bed transport contribution increased dramatically, concentrating the flux maximum within the near-bed layer, while upper-layer contributions remained comparatively weak. This shift is accompanied by a more pronounced residual flow structure and an enhanced near-bed covariance ($\langle u' C' \rangle$), implying that both residual advection and tidal pumping operate synergistically to drive the landward transport regime in winter. This extreme near-bed dominance is consistent with the rapid flood-onset entrainment of fluid mud, which elevates near-bed SSC precisely when near-bed landward velocities are highest, thereby amplifying $\langle u' C' \rangle$ and the resulting near-bed flux.

3.5 Near-bed baroclinic forcing

To evaluate the relative contribution of sediment- and salinity-induced density gradients to the baroclinic forcing driving residual circulation, near-bed water samples collected at the four fixed buoys during the 2022 survey were examined (Fig. 7a). The tidally averaged near-bed SSC differed across the four buoys, ranging from 37.3 g L⁻¹ (NW) to 26.9 g L⁻¹ (NE), with the two southern buoys (SW: 28.5 g L⁻¹; SE: 30.1 g L⁻¹) falling in between.

Grouping the buoys into a western pair (NW, SW) and an eastern pair (NE, SE) yields a cross-sectional average near-bed concentration difference of approximately 4.4 g L⁻¹ over the 2 km buoy separation, equivalent to $\partial \rho_{ssc} / \partial x \approx 1.4 \times 10^{-3}$ kg m⁻⁴ ($n = 0.62$), with the seaward (western) side exhibiting higher near-bed concentrations. Considering the two north buoys (NW, NE) alone, this difference increases to approximately 10.4 g L⁻¹ ($\sim 3.2 \times 10^{-3}$ kg m⁻⁴), whereas the south buoys (SW, SE) instead show a smaller, oppositely directed difference, indicating pronounced lateral asymmetry in the along-channel SSC gradient during this survey.

This high seaward sediment gradient near the bed is opposite to the typical up-estuary increase described in Sect. 2.1 (Van Maren et al., 2023), and was pronounced at the north buoy pair but not at the south pair. We attribute this anomaly to the seaward migration of fluid mud, which either shifts the fluid mud center of mass into this reach directly, or enhances local sediment deposition at the nearby Port of Emden. The latter pathway is consistent with Van Maren et al. (2015a), whose model

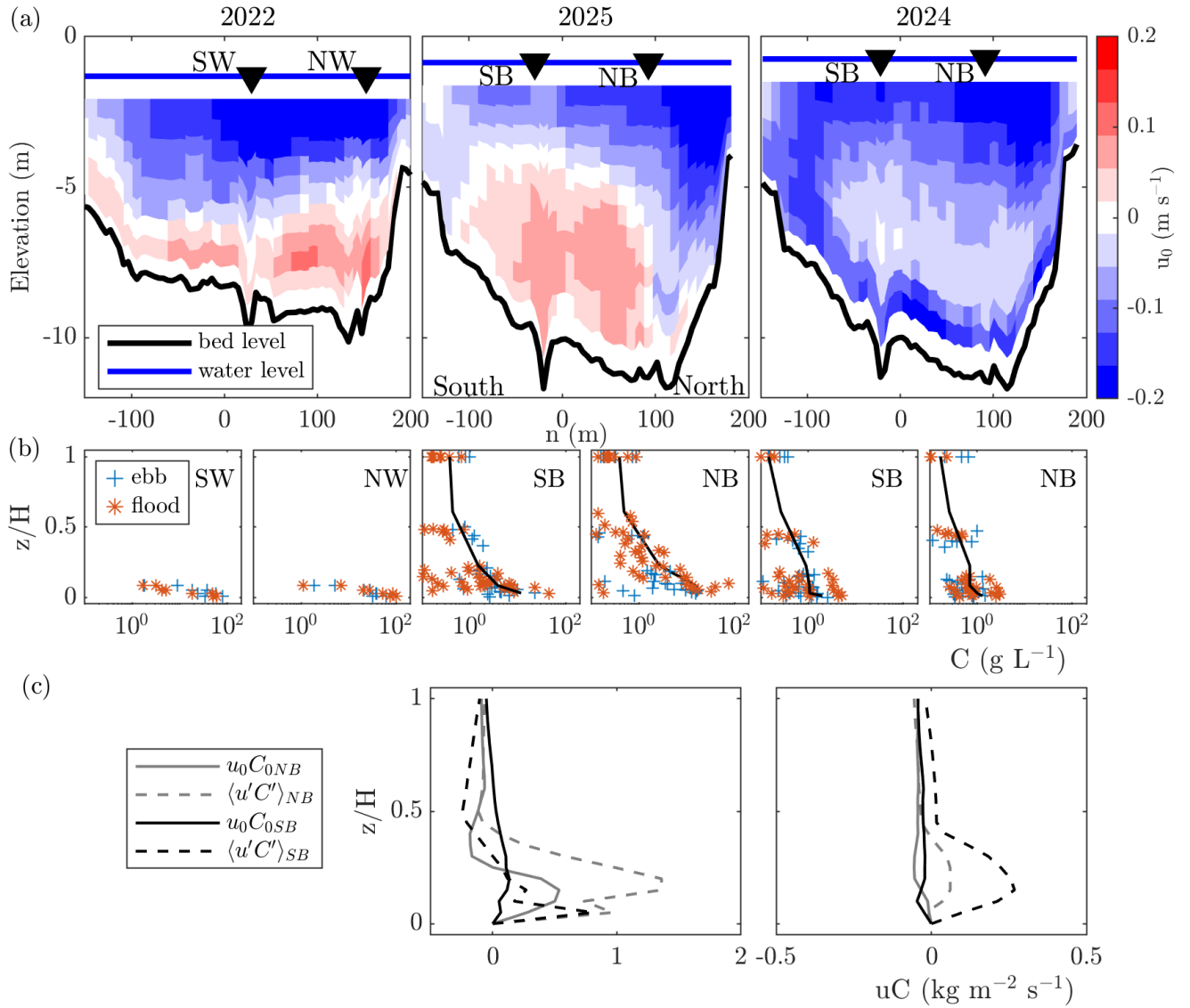


Figure 6. Residual flow structure and the relative contributions of residual advection and tidal pumping. Columns correspond to the 2022, 2025 and 2024 surveys. (a) Cross-sectional transects of residual velocity u_0 (positive landward). The thick blue line denotes the water level and the black line indicates the bathymetry; triangles mark the sampling verticals (SW and NW in 2022; SB and NB in 2024 and 2025). Y-axis is the elevation referring to NAP, which is the Dutch ordnance level, close to mean sea level. (b) Vertical distributions of bottle-sampled SSC at buoys during ebb (blue) and flood (orange). The dark lines represent tidally averaged concentration profiles. (c) Vertical profiles of flux contributions from residual advection ($u_0 C_0$) and tidal pumping ($\langle u' C' \rangle$) at SB and NB.

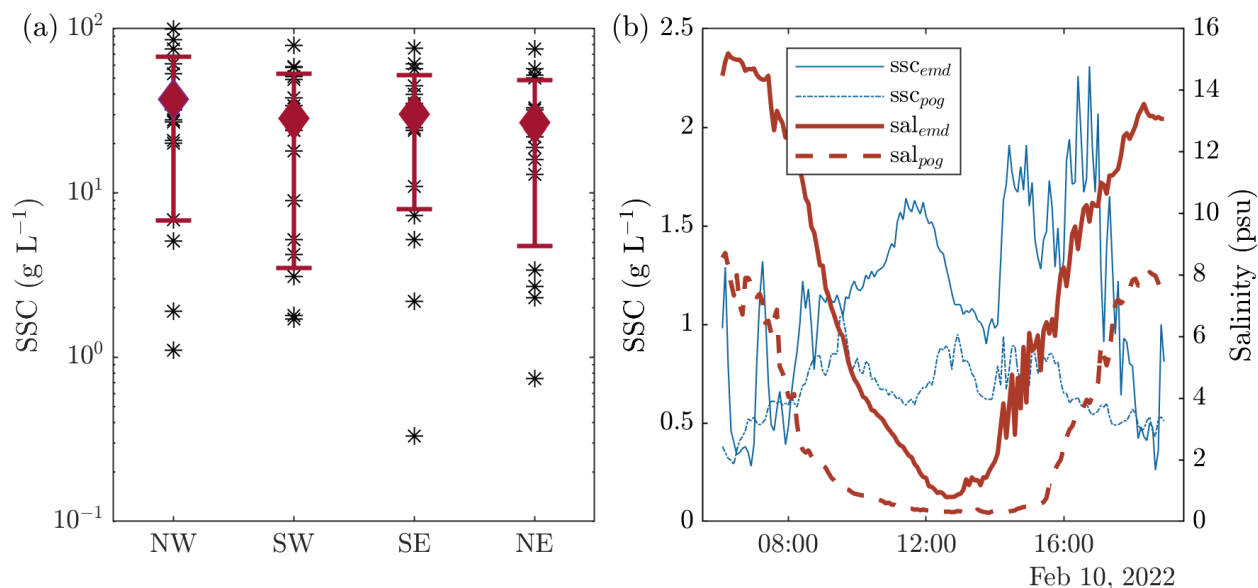


Figure 7. (a) Near-bed SSCs from 2022 survey. Black dots indicate concentrations from water samples. Red diamonds represent the averaged SSC for each Buoy. NW, SW, SE and NE represent the buoys at the northwestern, southwestern, southeastern and northeastern corners of the study area (Fig. 1d). (b) Historical records of SSC and salinity from Emden and Pogum stations during the 2022 survey.

reproduced long-term deposition rates well at the more seaward ports of Eemshaven and Delfzijl, but substantially underestimated deposition at the Port of Emden and its approach channel — a discrepancy they attributed to the hyperturbid conditions in the lower Ems River. Both mechanisms may cause the net baroclinic forcing which generates the two-layer estuarine circulation. The seaward residual flow observed during the 2024 survey without fluid mud is consistent with a configuration in which the typical up-estuary increase in SSC possibly outweighs the background salinity gradient in the opposite direction.

For comparison, historical surface SSC and salinity records at the Emden and Pogum tidal stations (about 5 km apart) over the same survey period yield a near-surface sediment concentration gradient of $7.5 \times 10^{-5} \text{ kg m}^{-4}$ and a background salinity gradient equivalent to $4.15 \times 10^{-4} \text{ kg m}^{-4}$, both smaller in magnitude than the near-bed sediment gradient (Fig. 7b). These surface-based estimates reflect a coarser spatial scale than the buoy-based near-bed gradient, so this comparison is indicative of relative order of magnitude rather than a strictly scale-matched estimate.

4 Discussion

4.1 Fluid mud and river discharge

The thickness of fluid mud differed markedly among the three surveys (Fig. 2): fluid mud was thick ($>2 \text{ m}$) and persistent at $Q = 200 \text{ m}^3 \text{ s}^{-1}$, thin ($\sim 1 \text{ m}$) and intermittent at $Q = 125 \text{ m}^3 \text{ s}^{-1}$, and absent at $Q = 45 \text{ m}^3 \text{ s}^{-1}$. In 2022, along-channel backscatter profiles revealed a thick fluid mud layer throughout the $\sim 2 \text{ km}$ surveyed reach (Fig. S1 in the Supplement), indi-

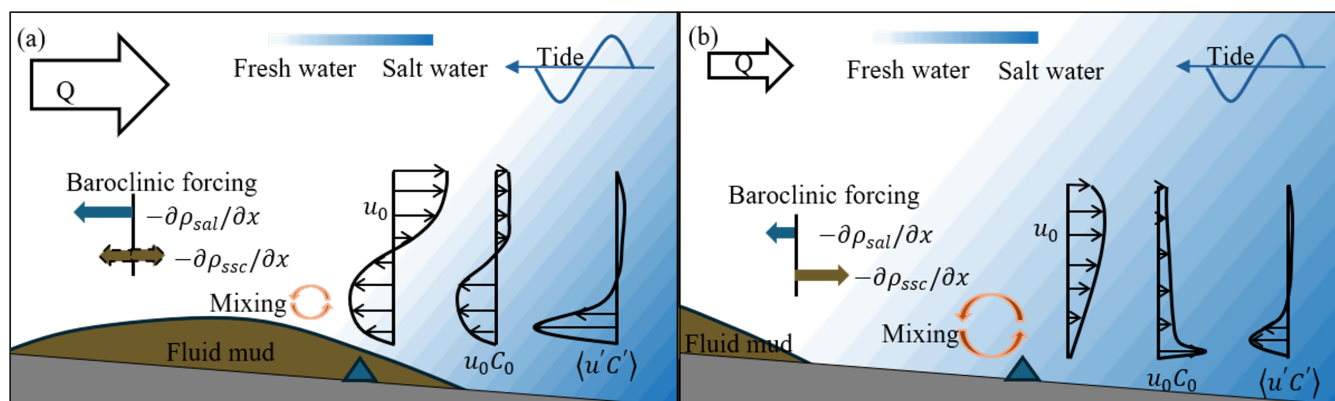


Figure 8. Schematic of the variations in river discharge, baroclinic forcing, residual flow and sediment flux due to advection and tidal pumping. The blue shading represents the salinity intrusion, the triangle marks the observation site, and the size of the mixing symbol indicates the intensity of vertical mixing. (a) High discharge: baroclinic forcing by salinity and sediment produces a two-layer residual flow. (b) Low discharge: the sediment-induced density gradient opposes the salinity-induced gradient, generating a predominantly seaward residual flow.

cating that at high discharge the fluid mud pool as a whole, rather than a local patch, extended over the study area. We therefore interpret the contrast among the surveys as a discharge-driven shift of the fluid mud pool, which extends seaward over the study
 385 area at high discharge (Fig. 8a) and retreats landward of it at low discharge (Fig. 8b). This interpretation is based on surveys at different times rather than on continuous observation of the migration, and differences in other conditions among the surveys, such as wind and water temperature, may also have contributed to the contrast through their effects on hydrodynamics and sediment rheology.

Fluid mud forms where processes that enhance the downward sediment flux coincide with trapping in a convergence zone
 390 (Kineke et al., 1996), so its location follows that of the convergence zone. Increasing river discharge acts on both the supply and the trapping of sediment. In the upper estuary, higher river flow increases bed shear stress and transport capacity, flushing sediment seaward (Guo et al., 2014; Jalón-Rojas et al., 2015; Yan et al., 2021). At the same time, it counteracts salt intrusion, thereby pushing the zone of salinity-driven stratification and gravitational circulation seaward, where landward near-bed transport converges with the seaward river flow and fine sediment is retained (Zhu et al., 2021; Siemes et al., 2026). Within this
 395 displaced convergence zone, high river flow also damps flood velocities, promotes the aggregation of fine sediment and reduces resuspension, so that sediment is trapped rather than flushed further seaward (Horrevoets et al., 2004; Winterwerp et al., 2017; Winterwerp, 2011). As discharge decreases, the convergence zone and the fluid mud pool retreat landward. Analogous to the discharge-driven migration of the estuarine turbidity maximum (ETM) (Grasso et al., 2018; Camargo et al., 2026), the location of fluid mud thus reflects a discharge-dependent balance between flushing and trapping.



400 4.2 Feedback between fluid mud, hydrodynamics and sediment import

Suspended sediment is commonly treated as a passive tracer transported by a prescribed flow field (Rouse, 1937). Our observations add to the growing evidence that high suspended sediment concentrations and fluid mud, in turn, modify estuarine hydrodynamics (Li et al., 2023; Egan et al., 2020; Gabioux et al., 2005). Fluid mud affects both the horizontal density gradient that drives the residual flow and the vertical density structure that controls how the flow responds to this forcing (Fig. 8).
 405 Where fluid mud is present, it constitutes a source of sediment for resuspension, and the resulting near-bed SSCs introduce a substantial density anomaly (Geyer et al., 2004; Kineke et al., 1996). This anomaly is concentrated in the near-bed layer and varies along the channel with the thickness and concentration of fluid mud, generating a strong near-bed horizontal density gradient (Sect. 3.5). Unlike the salinity gradient, which acts over the entire water column, the sediment-induced gradient therefore produces a baroclinic forcing concentrated near the bed, substantially altering the horizontal pressure gradient that drives
 410 the near-bed residual flow (Donker and de Swart, 2013; Talke et al., 2009). The direction of this forcing depends on where the fluid mud pool lies relative to the study area. At low discharge, the pool lies upstream, so near-bed density decreases seaward and the sediment-induced forcing is directed seaward, opposite to the salinity-induced forcing (Fig. 8b).

The residual flow in 2024 was seaward over the entire water column and strongest near the surface and the bed, suggesting that the seaward forcing by an SSC gradient exceeded the salinity-induced forcing near the bed. Its magnitude was not mea-
 415 sured, however, and other processes, such as the combination of the Stokes transport and compensating return flow, may also play a role (Garel and Ferreira, 2013). As the pool shifts seaward and extends over the study area, the seaward sediment-induced forcing there weakens and may even reverse, depending on the position within the pool and on local bathymetry (Sect. 3.5). Once it no longer exceeds the salinity-induced forcing, the net near-bed forcing becomes landward, and a two-layer residual circulation develops, as observed when fluid mud was present (Fig. 8a).

420 Fluid mud also modifies the vertical structure of the flow. During flood in 2025, Ri_{SSC} near the bed exceeded the critical value of 0.25, whereas in 2024 it did so only sporadically. Sediment-induced stratification alone was thus sufficient to suppress near-bed turbulence when fluid mud was present. When salinity was included, Ri_g exceeded 0.25 over much of the water column in 2025, but only near the surface in 2024. Such stratification damps turbulence and reduces vertical momentum exchange across the lutocline, which can effectively reduce the bottom drag experienced by the overlying flow (Wu et al.,
 425 2023, 2022; Winterwerp et al., 2009; Moodie et al., 2022). Although bottom stress and turbulence were not measured, the velocity structure is consistent with this mechanism. In 2025, a mid-depth velocity maximum up to 0.9 m s^{-1} developed during flood deceleration, whereas no such local maximum occurred in 2024. While the cause of the mid-depth velocity maximum remains unclear, its descent with sediment settling and its absence in 2024 (Fig. 4c–f) are consistent with sediment-induced stratification reducing momentum exchange with the bed. At the subtidal timescale, reduced drag allows a given mean pressure
 430 gradient to drive a stronger near-bed residual flow (Egan et al., 2020). Once the net near-bed forcing is landward, this reinforces the two-layer circulation observed when fluid mud was present (Fig. 8a).

Near-bed transport dominated the net sediment import in both years, and with fluid mud present this import was about an order of magnitude larger (Table 2). Fluid mud contributed to this increase in two ways. The first is direct: as a sediment



source, fluid mud raises near-bed SSC and thereby amplifies tidal pumping, which was the largest near-bed flux component. Tidal pumping arises from the asymmetry between flood and early ebb near the bed. During the long HWS observed in both years, flocs may grow and settle, forming dense near-bed suspensions that can develop into fluid mud (Wu et al., 2023; Guo et al., 2017). Because the surface flow turns to ebb before the near-bed flow (Fig. 4e,f), the near-bed ebb begins late, and the settled sediment must first be resuspended before it can be transported seaward. Near-bed concentrations were therefore lower during early ebb than during flood (Fig. 4c,d), yielding a net landward tidal pumping flux. With fluid mud present, far more sediment takes part in this asymmetric exchange, strengthening tidal pumping (Fig. 8a versus b). The second pathway is indirect, through the hydrodynamic changes discussed above: the near-bed residual flow reversed from seaward to landward when fluid mud was present, so that near-bed residual advection changed from exporting to importing sediment.

The enhanced near-bed import, in turn, can supply sediment to the fluid mud pool, closing a positive feedback in which fluid mud promotes the import that sustains it. Because the pool shifts with discharge, the two panels of Fig. 8 can also be read as two positions along the estuary relative to the pool. Seaward of the convergence zone, where fluid mud is absent, tidal pumping imports sediment towards the zone, where it may be trapped (Fig. 8b). Within the convergence zone, fluid mud both supplies sediment for tidal pumping and modifies the local hydrodynamics, together transporting sediment further landward (Fig. 8a). This feedback may thus contribute to the persistent hyperturbidity observed in the Ems Estuary and the lower Ems River (Winterwerp, 2011; Talke et al., 2009). Although our observations do not resolve effects at the scale of the tidal wave, sediment-induced drag reduction over fluid mud can alter tidal propagation (Gabioux et al., 2005), and similar feedbacks between drag reduction, tidal amplification and sediment import have been proposed to explain the shift of the lower Ems River to hyperturbid conditions (Van Maren et al., 2015b; Winterwerp and Wang, 2013; Winterwerp et al., 2013; Dijkstra et al., 2019). The local feedback observed here may thus be one element of this larger-scale response. Stratification, estuarine circulation and tidal asymmetry are recognized as the main mechanisms maintaining high turbidity in other estuaries (Burchard et al., 2018; Van Maren et al., 2023; Jay and Musiak, 1994), such as the Yangtze Estuary (Zhu et al., 2022; Lin et al., 2025), the Seine Estuary (Grasso and Le Hir, 2019) and the Hudson River estuary (Geyer et al., 2001). Our results suggest that where fluid mud forms, it does not merely respond to these mechanisms but also modifies them, which may reinforce sediment trapping in such systems as well.

4.3 Limitations of the sediment flux estimates

Our results demonstrate that the near-bed layer ($z/H < 0.3$) dominates the net landward sediment flux, so the reliability of flux estimates depends on how well the profiles of velocity and SSC are resolved. In hyperturbid estuaries, however, sediment flux measurements can hardly achieve high temporal and high spatial resolution at the same time. Whereas velocity profiles can be obtained with relatively high temporal and spatial resolution by ADCPs, the estimation of SSC is limited by its strong spatial heterogeneity, particularly the steep near-bed gradients. Bottom-mounted instruments, such as acoustic backscatter systems, resolve near-bed SSC at high resolution but only at fixed points (Thorne and Hanes, 2002; Becker et al., 2018). In this study, bottom stress and turbulence were not measured, so the drag reduction discussed in Sect. 4.2 remains inferred.



The observations were collected during neap tides, when vertical stratification is relatively stable and fluid mud layers persist (Wu et al., 2022). These conditions are advantageous for reconstructing monotonic SSC profiles and estimating sediment fluxes, but the transport estimates represent only part of the spring–neap variability. During spring tides, enhanced tidal mixing is expected to weaken stratification and may entrain fluid mud into suspension, weakening the feedback and reducing circulation-driven residual transport. At the same time, higher bed shear stress and SSC may enhance resuspension and tidal pumping (Ge et al., 2018; Bailey et al., 2025), partly compensating for the weaker residual transport. The net effect on sediment import over a spring–neap cycle therefore remains uncertain.

Future observations would benefit from synchronous velocity and SSC measurements in hyperturbid environments, direct measurements of turbulence and bottom stress to test sediment-induced drag reduction, simultaneous along-channel measurements of near-bed salinity and SSC to compare their density gradients directly, and surveys spanning complete spring–neap cycles and a wider range of discharges to capture the migration of fluid mud. Such observations would help to quantify near-bed transport in hyperturbid estuaries, where neglecting this layer may lead to substantial underestimation of sediment transport and obscure the mechanisms responsible for landward sediment trapping.

4.4 Lateral asymmetry in sediment transport and its implications for modeling

Our observations reveal pronounced lateral asymmetry in residual advection, SSC distribution and tidal pumping, particularly in the 2025 survey, when fluid mud was present. This cross-channel variability suggests that depth-averaged or single-vertical representations of sediment transport may misrepresent the true along- and cross-channel structure of landward sediment flux. This is because the lateral variability strongly influences the stratification, momentum balance and longitudinal mass balance (Geyer et al., 2020; Lerczak and Geyer, 2004). Capturing this behavior likely requires high-resolution three-dimensional modeling approaches capable of resolving both lutocline dynamics and lateral circulation structures (Chen et al., 2020; Zhou et al., 2021).

5 Conclusions

Observations in the Emden Fairway of the hyperturbid Ems Estuary show that the strength of landward sediment transport is strongly controlled by the presence of a fluid mud pool, whose extent changes with river discharge. By combining near-bed-resolved SSC profiles with velocity measurements under three river discharge conditions, we quantified how sediment flux is partitioned between the near-bed and upper layers and between residual advection and tidal pumping, and how this partitioning changes with the presence of fluid mud. Fluid mud was thick and persistent at high discharge, thin and intermittent at intermediate discharge, and absent at low discharge. These observations are consistent with a seaward shift of the fluid mud pool with increasing discharge, reflecting a discharge-dependent balance between flushing and trapping. Regardless of the presence of fluid mud, the near-bed layer ($z/H < 0.3$) carried all of the net landward sediment transport; with fluid mud present, however, this transport was about an order of magnitude larger, and the near-bed layer carried a larger share of the gross transport. Without fluid mud, the residual flow was seaward over the entire water column, and net import relied on tidal



pumping, which arises from the asymmetry in near-bed resuspension between flood and early ebb. With fluid mud present, the near-bed residual flow turned landward and elevated near-bed SSC amplified tidal pumping, so that residual circulation and tidal pumping acted together to drive sediment import. The resulting near-bed import in turn supplies the fluid mud pool, forming a positive feedback that may contribute to the persistent hyperturbidity of the Ems Estuary and the lower Ems River. These results show that resolving the near-bed layer is essential for quantifying sediment transport in hyperturbid estuaries, which in turn underpins understanding of their morphological evolution and the siltation of navigation channels.

Code and data availability. A MATLAB toolbox used to read and pre-process the raw ADCP data (adcpools) is publicly available at: <https://github.com/bartverm/adcpools.git>. Water level and river discharge data were obtained from the Federal Maritime and Hydrographic Agency (BSH; <https://wasserstand-nordsee.bsh.de>). Surface SSC and salinity records at the Emden and Pogum stations were obtained on request from the Lower Saxony Water Management, Coastal Defence and Nature Conservation Agency (NLWKN). All measurement data and analysis scripts supporting this study are archived and available at 4TU.ResearchData: <https://doi.org/10.4121/6c4b1c4c-b71a-4159-b66b-286dbdae6361>.

Appendix A: Change point detection

Acoustic inversion of backscatter to SSC assumes that the acoustic properties of the suspended sediment (e.g., particle size, shape and composition) are uniform within the ensonified volume (Thorne and Hanes, 2002; Sassi et al., 2012). In hyperturbid environments, this assumption is frequently violated along the acoustic path (Venditti et al., 2016; Thorne et al., 2021). Successful inversions are typically associated with corrected backscatter profiles that vary linearly with range (Guerrero and Di Federico, 2018; Haught et al., 2017). We therefore used a linearity-based change point detection on the $S'_{v,\alpha_s=0}$ profiles to identify, *a priori*, profiles in which this condition is violated (Bao et al., 2026).

Change points were identified by fitting piecewise linear models with MATLAB's `findchangepts` function, which minimizes the total squared error subject to a penalty (MinThreshold) against overfitting, following the optimal partitioning framework of Killick et al. (2012). Profiles were classified as Type 1 (no change point) or Type 2 (one or more change points).

The classification was validated with frame-mounted ADCP profiles and near-bed water samples from the 2022, 2024 and 2025 surveys (Bao et al., 2026), using the samples closest to the ADCP and to the bed for calibration and the two middle samples for validation. Inverted SSC agreed well with the water samples for Type 1 profiles ($R_{Pearson} = 0.98$, up to $\sim 20 \text{ g L}^{-1}$), but poorly for Type 2 profiles ($R_{Pearson} = 0.28$), with a systematic underestimation at high concentrations. The classification was insensitive to MinThreshold values between 10 and 30; a value of 20 was used for these short profiles, whereas a value of 400 was used for the full-depth vessel-based profiles (Sect. 2.2.1).



Appendix B: Acoustic inversion

The volume scattering strength S_v of sediment suspensions at a range R from an ADCP transducer is equal to the sum of the echo intensity directly measured by the ADCP (EI, dB), attenuation by water and sediment in the two-way path along the range
 530 to the sediment suspension (dB), signal loss from spherical spreading (dB) and an instrument specific constant (G , dB):

$$S_v = EI + 2\alpha_w R + 2 \int_0^R \alpha_s(r) dr + 10 \log_{10} \left(\frac{T_T R^2 \psi^2}{L P_T} \right) + G \quad (\text{B1})$$

where S_v is in dB, α_w represents sound attenuation due to absorption by water (dB/m), α_s is attenuation by suspended particles (dB/m). The fourth term on the right hand side is the signal loss from spherical spreading (dB), where ψ is a function that accounts for the departure from spherical spreading within the near field of the transducer, T_T is the transducer temperature
 535 ($^{\circ}\text{C}$), L is the transmit pulse length (m), P_T is the transmit power (W). The instrument specific constant is sometimes provided by the manufacturer.

Focusing on the unknown sediment attenuation, when G is unknown, the sonar equation for S_v reads:

$$\begin{aligned} S_v(R) &= S'_{v,\alpha_s=0}(R) + 2 \int_0^R \alpha_s(r) dr + G \\ &= S'_{v,\alpha_s=0}(R) + 2 \int_0^R \xi_{total}(r) M_s(r) dr + G \\ &= 10 \log_{10}(k_s^2(R) M_s(R)) \end{aligned} \quad (\text{B2})$$

Herein, the backscatter term $S'_{v,\alpha_s=0}$ is the sum of EI, attenuation by water and spherical spreading, which are all readily
 540 calculated. We capture the combination of scattering and viscous absorption in ξ_{total} (dB m² / kg). α_s is written as the product of sediment attenuation ξ_{total} and mass concentration, and k_s^2 describes the scattering properties of the suspended sediment (dB/(m kg)). Equation (B2) shows which factors impact the inversion relationship. The mutual dependence of M_s , S_v and distance-integrated acoustic attenuation complicates ADCP backscatter calibrations. To estimate M_s , profiles of S_v have to be corrected for sound attenuation, which in turn depends on M_s (Thorne and Hanes, 2002).

545 Under the quasi-homogeneity assumption where k_s^2 and ξ_{total} remain constant over the ensonified volume, the depth-averaged explicit calibration method is adopted following Sassi et al. (2012) and Thorne and Hanes (2002):

$$M_s(R) = \frac{\beta(R)}{K(R_{ref}) - \frac{\ln 10}{5} \xi_{total} \int_{R_{ref}}^R \beta(r) dr} \quad (\text{B3})$$

where

$$\beta = 10^{\frac{S'_{v,\alpha_s=0}}{10}}, K(R_{ref}) = \frac{\beta(R_{ref})}{M_{ref}} \quad (\text{B4})$$

550 Equation (B3) yields the concentration profile over the entire range R , provided a reference $K(R_{ref})$ at a distance R_{ref} from the transducer and attenuation parameter ξ_{total} .



A calibration procedure is needed to translate ADCP backscatter to concentration profiles. The backscatter profile is 2-min averaged volume backscattering strength from the vertical beam as described above. By collecting water samples at a single range R_{ref} , Holdaway et al. (1999) developed a calibration in which $K(R_{ref})$ was obtained as the average over all measurements corresponding to the times when bottle samples were collected. This approach is well-suited for rigid deployments where the ADCP is deployed at a known distance above the bottom. In boat-mounted deployments, water samples are never collected exactly within the ADCP target volume, so R_{ref} is not constant. By assuming relatively little spatial-temporal variation of suspension heterogeneity at the reference range R_{ref} , we expect a linear relationship between $10\log_{10}(K(R_{ref}))$ and $S'_{v,\alpha=0}(\tilde{R}_{ref})$,

$$10\log_{10}(K(R_{ref})) = aS'_{v,\alpha=0}(\tilde{R}_{ref}) + b \quad (B5)$$

where a and b are linear regression coefficients, $\tilde{R}_{ref}$ is chosen to coincide with the mean reference range.

In this study, the mid-depth water samples are used to determine $K(R_{ref})$ because they are normally located in the well-mixed part of the water column (Fig. 4g and h). In addition, the surface water samples cannot match with backscatter data and the ranges of near-bed water samples may cause substantial sediment attenuation. The reference samples provide pairs of M_{ref} and $\beta(R_{ref})$ for which $K(R_{ref})$ is computed. The attenuation parameter is determined empirically by collecting another water sample at a known distance from the ADCP, which reads

$$\xi_{total} = \frac{5}{\ln 10} \frac{K(R_{ref}) - \frac{\beta(R_2)}{M_s(R_2)}}{\int_{R_{ref}}^{R_2} \beta(r) dr} \quad (B6)$$

R_2 is chosen as the lowest water sample above the second changepoint, to better represent the depth-averaged attenuation within the inverted segment. Additional samples collected in between were used for validation of the inversion performance.

Appendix C: Salinity Distribution

Salinity was measured by CTD casts at the buoys shown in Fig. 1d during the 2024 and 2025 surveys. The CTD was mounted on the mobile frame where a pressure sensor was deployed.

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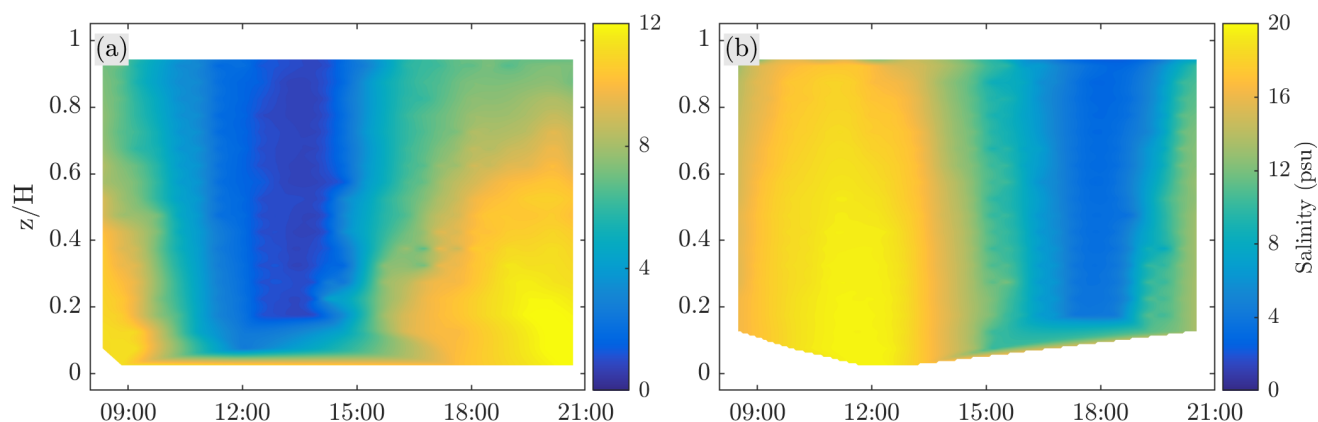


Figure C1. Salinity distribution at North Buoy during the 2025 (a) and 2024 (b) surveys.

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