Residual Currents and Flux Estimates in a Partially-mixed Estuary

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An intensive sampling programme of physical parameters has been carried out at three representative areas to examine the variable circulation and mixing conditions in Southampton Water and the Test Estuary (Hampshire, southern England). Field data of current velocity and salinity were collected under neap and spring tidal cycles, and under winter and summer circulations, to account for the fortnight and seasonal effects in the estuary. Both were collected to a very high vertical resolution using ADCP and CTD instrumentation. The instantaneous and residual fields of the above measured variables were examined, together with a thorough consideration of the mechanisms responsible for residual transport. Variation of the instantaneous field during different stages of the tidal cycle is shown, corresponding mainly to changes in the intensity of turbulence introduced in the water column. Tidal effects appear to be much more important than river flow discharge and wind shear stress for the longitudinal–vertical distribution of physical variables.

Analysis of residual longitudinal and lateral currents and fluxes of water and salt, however, reveals the significance of non-tidal effects in the mean tidal transport. The Stokes drift mechanism appears mostly positive in direction, thus pushing water and salt upstream and increasing its magnitude under spring tidal conditions. Lateral Eulerian transports are of the same order of magnitude as the longitudinal ones, but with smaller values, especially under spring tidal amplitudes, when the flow coincides better with the longitudinal axis of the estuary. Vertical eddy diffusivity and viscosity coefficients were examined showing near zero values at the surface and bottom layers and maximum values at the mid-depth region, where most intense mixing occurs.

\textbf{Keywords:} estuarine circulation; anchor stations; salt balance; water balance; residual currents; Southampton Water

\section*{Introduction}

Southampton Water (Hampshire, southern England) is part of a complex and highly populated estuarine system, The Solent, within which processes of water transport and mixing, as well as interactions between different parts of the system, are tidally dominated. In a plan view the estuary is almost rectangular with a length of about 10 km and a mean width of 2 km at high waters. The central part of the estuary consists of a well-defined, narrow, 13 m deep dredged water channel, which runs parallel to the estuarine axis in a northwest-southeast direction (Figure 1). It is bordered by broad shallow intertidal mudflats with shingle and sand on the eastern side and a salt marsh to the west. The largest source of freshwater input in Southampton Water is introduced at the head of the estuary by the rivers Test and Itchen, having a mean annual discharge of 8.81 and 3.26 m$^3$ s$^{-1}$, respectively. A much smaller source from the river Hamble (0.28 m$^3$ s$^{-1}$) enters the estuary at its eastern bank.

Previous investigations in Southampton Water indicate that partially-mixed conditions prevail in the estuary (Dyer, 1973; Westwood, 1982). Dyer (1973) observed important cross-sectional variations of salinity, which result in a circulation opposite to what could have been expected under the influence of Coriolis force. Local topographic irregularities, especially on either sides of the main channel, may be responsible for these patterns. Tidal currents in Southampton Water have been studied extensively by Dyer (1973), Blain (1980), Westwood (1982), Sylaios (1994) and Sylaios and Boxall (1996). As expected, maximum currents were recorded during ebb, due to the shorter duration of this tidal phase, as compared to flood. A tidal range of 2-4 m implies strong currents accounting for the high transport rates through the narrow cross-section of the main channel during a tidal period. The purpose of this work is to present measurements of velocity and salinity made at three representative areas in the main channel of Southampton Water and Test Estuary, to examine the instantaneous and...
residual field of the above measured variables and to consider thoroughly the mechanisms responsible for residual transport of water and salt in the estuary.

**Materials and methods**

**Sampling locations and parameters measured**

Three stations (A, B and C) located on the main channel of Southampton Water and the Test Estuary were chosen as indicative of the different water circulation and mixing characteristics in the area of interest (Figure 1). Station A is placed at the lower part of the estuary, near its seaward boundary (Calshot Buoy), where almost well-mixed conditions exist. Station B is positioned at the upper part of Southampton Water, midway along the main channel (NW Netley Buoy) having a highly variable water column. Station C is located near the tidal limit of the Test Estuary (Cracknore Buoy) receiving a large rainfall-dependent river input.

Profiles of velocity and salinity were collected at these stations during 12 different tidal cycles, for sampling periods of 13 h. A sampling rate of 15 min was appropriate to resolve the circulation and mixing processes of interest. A ship-mounted acoustic doppler current profiler (ADCP 1200 kHz, RD Instruments) was used to measure horizontal and vertical velocity components at 1 m intervals throughout the water column (horizontal speed accuracy=0.5%; horizontal velocity resolution=0.125 cm s$^{-1}$), using ensembles of 60 pings at 1 s ping interval. First and last depth cells of 0.5 m at the surface and bottom of the water column were contaminated from side-lobes effect, and, thus, results at those depth intervals needed correction for flux estimation using simple extrapolation techniques. Salinity was recorded using a Neil Brown Smart CTD (accuracy=0.01; resolution=0.001), at a vertical resolution of 0.1 m.

**Data manipulation and basic definitions**

Physical data obtained in the survey were manipulated according to Kjerfve (1975), since depth non-dimensionalization is essential for estuaries of tidal range to mean depth ratio ($\epsilon$) higher than 0.3. Southampton Water main channel with a mean depth of 10.78 m (below mean tide level) and a tidal range at spring tides of 4.08 m, has an $\epsilon$-value of 0.38. Hence, the instantaneous depth-average of a variable was obtained by averaging its values over $\sigma$, where $\sigma=z/d(t)$. At all times $\sigma=0$ represents the surface and $\sigma=1$ the sea bottom. The net or tidal average value of a variable was obtained by averaging over a 12.5 h period, for the different standard fractional depths $\sigma_0$, $\sigma_0 \ldots \sigma_1$.

Following Uncles and Jordan (1979) and Uncles et al., (1986), the instantaneous rate of water transport per unit of width through a water column of depth, H, is given by $Q$ as:

$$Q = H \overline{U} = H (\overline{u}, \overline{v})$$  \hspace{1cm} (1)

where $u, v$ are components of $U$ in the x, y directions, respectively. The overbar denotes an average over depth.
Furthermore, the residual rate of water transport will be given by:

$$
\langle Q \rangle = \langle H \rangle \cdot [V_1 + V_2]
$$  \hspace{1cm} (2)

where $\langle n \rangle$ represents the tidal average of a variable, $V_1$ the depth-averaged Eulerian residual transport, or the non-tidal drift, which is expressed by Dyer (1974) as:

$$
V_1 = \langle U \rangle
$$  \hspace{1cm} (3)

and $V_2$ a measure of the residual rate of transport of water resulting from tidal pumping, which is expressed by Hunter (1972) and Tee (1976) as:

$$
V_2 = \langle H \Delta U \rangle / \langle H \rangle
$$  \hspace{1cm} (4)

where $\Delta$ represents the depth-average instantaneous deviations from the tidal mean. The $V_2$ current is referred to as the mass transport Stokes drift, to distinguish from the related Stokes drift.

From the above analysis, it occurs that when the depth-averaged longitudinal current velocity is multiplied by the instantaneous water depth of each profile at the survey stations of Southampton Water, a 'local line flux' is obtained, being the area of a vertical plane of a body of water passing through a certain location ($y$), per unit of time ($t$):

$$
Q(y,t) = \sigma(y,t)H(y,t)
$$  \hspace{1cm} (5)

where $Q$ is the local water flux per unit of channel width ($m^3 s^{-1} m^{-1}$) and $y$ the lateral position along the cross-section.

These longitudinal local water fluxes per metre width per station ($Q_y$), can then be integrated over the flood and ebb period, respectively, resulting in:

$$
V_{\text{flood}}(y) = \int_{t=0}^{T} Q(y,t) \, dt \quad \text{and} \quad V_{\text{ebb}}(y) = \int_{t=T}^{t=0} Q(y,t) \, dt
$$  \hspace{1cm} (6)

in which $V_{\text{flood}}(y)$ or $V_{\text{ebb}}(y)$ is the water prism during flood or ebb period per metre of width ($m^3 m^{-1}$) and $t$ the tidal period ($s$) and $t=0$, $t=T$ and $t=T$ represent the slack tidal times. The tidal prism per unit of width ($m^3 m^{-1}$) is, therefore, defined as $V_{\text{prism}} = V_{\text{flood}} - V_{\text{ebb}}$ (de Jonge, 1992).

Similarly, the instantaneous transport rate of salt per unit of width through a column of depth, $H$, is given as:

$$
Q_S = H \Delta U \Delta S
$$  \hspace{1cm} (7)

The residual transport rate of salt is in the form:

$$
\langle Q_S \rangle = \langle H \rangle [V_{S,1} + V_{S,2} + V_{S,3}]
$$  \hspace{1cm} (8)

where $V_{S,1}$ represents the depth-averaged residual flux of salt due to the residual transport of water, and is expressed as:

$$
V_{S,1} = \langle Q \rangle \langle S \rangle / \langle H \rangle
$$  \hspace{1cm} (9)

where $V_{S,2}$ represents the depth-averaged residual flux due to tidal pumping, resulting from the non-zero correlations between $Q$ and $S$, expressed as:

$$
V_{S,2} = \langle Q \rangle \langle S \rangle / \langle H \rangle
$$  \hspace{1cm} (10)

and $V_{S,3}$ represents the depth-averaged residual flux of salt due to the vertical shear between the tidal and residual currents, expressed as:

$$
V_{S,3} = \langle H \Delta V \Delta S \rangle / \langle H \rangle
$$  \hspace{1cm} (11)

where $V'$ and $S'$ represent the deviations of velocity and salinity from the depth-averaged value, respectively.

The parameterization of vertical fluxes of momentum and salt by the conventional manner in estuarine studies leads to the introduction of eddy viscosity and eddy diffusivity coefficients. Uncles and Jordan (1979) presented an expression for the above coefficients, assuming that the spatial variation of salinity was approximated by the spatial variation of depth-averaged salinity, an assumption valid for the well-mixed Severn Estuary, but not so good for Southampton Water and the Test Estuary:

$$
K_z^* = \langle H \rangle^2 \left[ \frac{\partial \langle S \rangle}{\partial \sigma} \right]_{\sigma=0}^{-1} \left[ \int_{\sigma=0}^{\sigma=1} \frac{\partial \langle S \rangle}{\partial \sigma} \, d\sigma + \langle \langle S \rangle \rangle \cdot \int_{\sigma=0}^{\sigma=1} \langle \langle V \rangle \rangle \, d\sigma \right]
$$  \hspace{1cm} (12)

and

$$
N_z^* = \left[ \frac{\partial \langle U \rangle}{\partial \sigma} \right]_{\sigma=0}^{-1} \left[ \langle H \rangle^2 \cdot \left( -f \right) \int_{\sigma=0}^{\sigma=1} \langle \langle V \rangle \rangle \, d\sigma + \frac{1}{2} \langle \langle H \rangle \rangle \cdot \sigma (\sigma-1) \frac{\partial \ln \langle \rho \rangle}{\partial \sigma} \right]
$$

$$
+ \int_{\sigma=0}^{\sigma=1} \frac{\partial \langle U \rangle}{\partial \sigma} \, d\sigma \right] + \sigma \left[ N_z^* \frac{\partial \langle U \rangle}{\partial \sigma} \right]_{\sigma=0}^{\sigma=1}
$$

$$
+ (1-\sigma) \left[ N_z^* \frac{\partial \langle U \rangle}{\partial \sigma} \right]_{\sigma=0}
$$

(13)
Typical formulas for the representation of the instantaneous values of the above mixing coefficients are given by Blumberg (1977) as functions of the local Richardson number. These formulations will be used in this work for comparative purposes but also for the determination of the range by which these coefficients vary within typical tidal cycles in Southampton Water.

**Results**

Instantaneous fields

Tides in Southampton Water are dominated by the semi-diurnal $M_2$ motion, on which the marked influence of overtides give rise to special features such as the ‘double high water’ and the ‘young flood stand’. Hence, the instantaneous fields of current and salinity exhibit complicated patterns directly affected by the rate of turbulent energy induced into the system by tidal dynamics.

One representative tidal period measured at Calshot is presented in Figure 2 under neap tidal conditions. The homogeneous ebbing of water throughout the column is noted (Figure 2(a)). Maximum ebb currents occur at the surface, approximately 2 h before low water, having a magnitude of 0-6 m s$^{-1}$. These currents last until surface elevation obtains its minimum value. After that there is a gradual decrease in the magnitude of the ebbing current, until 1 h after low water, when flow reversal starts. Maximum flood (0-4 m s$^{-1}$) occurs approximately 6 h after low water and is again confined to the surface. The ‘young flood stand’ feature lasts approximately 1 h and creates almost slack conditions within the water column, with small negative velocities at certain depths. The continuation of flood is not as severe, especially at the bottom and mid-depth levels, where negative velocities occur again during high-water slack. In general, vertical shear in the water column appears increased during low water, since the current gradually reduces its speed from surface to bottom.

Lateral currents, positioned perpendicular to the main channel axis were considered positive to the right facing upstream of the estuary (Figure 2(b)). Low water at Calshot induces negative velocities in ebb tidal flow, meaning that it is deflected right, towards the western bank of the channel, which agrees well with the influence of Coriolis force. Maximum lateral velocity during ebb occurs at the same time as the longitudinal with a value of 0-12 m s$^{-1}$. The introduction of flood reverses this circulation giving a maximum of 0-2 m s$^{-1}$ approximately 4 h after low water.

The salinity field under neap tides shows a definite stratification feature, with fresher water of 31-0 at the upper layer and saltier water of 33-5 at the bottom (Figure 2(c)). The freshwater layer increases in thickness towards low water, when the greatest salinity gradient is observed. This stratification effect is highly confined at the surface of the water column and has a duration of 2-5 h, a behaviour mostly related to tidal straining of the estuary. After low water, the highly turbulent tidal flood currents mix the whole water column, raising salinity to 33-0.

Instantaneous fluxes of water and salt

Figure 3 shows the water local line flux variability within four tidal cycles at stations A and B under winter and summer circulation conditions, as obtained by solving equation 5 and considering velocity components relative to the direction of the main channel. In all plots time $t=0$ h represents the occurrence of low water. The basic features appear to be similar in these cycles, with the inward water transport exactly at low water, the influence of ‘young flood stand’ with the reduction or even reversal of flood transport, and the presence of maximum inward flux during the second phase of the flood. First high water takes place about 7 h after low water, reducing the instantaneous flux into the near zero values. The main ebb commences approximately 9 h after low water, giving an outward water transport peak at the 11th hour of the cycle. The instantaneous depth-averaged local fluxes of salt for any of the examined tidal periods shows similar patterns as those found for the transport of water, mostly due to the fact that the variation of depth ($H$) within a tidal cycle is much more important than the change of depth-averaged salinity within the same cycle.

Residual currents and fluxes

Figure 4 shows the variation of tidally-averaged longitudinal and lateral velocity within the water column, for tidal cycles obtained at the three representative areas of the estuary. The net velocity profile at Cracknore shows a seaward movement throughout the water column, having a maximum value at the surface of 0-14 m s$^{-1}$. This behaviour is mostly associated with the freshwater flow influence on the instantaneous current. The Eulerian residual current reduces its magnitude with depth near the sea-bed, where it is almost zero. At Netley, the distinctive two-layer circulation, characteristic of a partially mixed estuary, is shown. The net longitudinal velocity profile at Calshot for the winter case, under a
southward moving wind, creates an opposite circulation from the one discussed above. Lateral currents show that the seaward flowing upper layer has a tendency to deviate towards the east side of the channel, especially in upper parts of the estuary.

The calculated values of the local water prisms obtained by solving Equation 6 during flood, ebb and the whole tidal period for Southampton Water, are given in Table 1. At this point, it must be noted that the determination of the error bounds for the water prisms and the fluxes following is difficult, owing to the smoothing and subsequent averaging of the basic data set; for this reason, the results for each tidal period are given, and consistency between values is taken to imply that the determinations are meaningful (Uncles & Jordan, 1976). Furthermore, results are significant when considered as representative for the part of the main dredged channel of the cross-section, and not for the total balances within the estuary. The figures in brackets designate a deficit in the residual volume of the estuary, due to a higher seaward transport. The last three columns give an indication on the magnitude of the different mechanisms responsible for residual transport. Figures in column 8 show the prevailed tidal range, in column 9 the weekly-averaged river discharge for the days prior to the date of survey and for the rivers upstream of each station, and in column 10 the average wind speed and the most frequent direction as measured using an in situ portable anemometer.

Residual fluxes of water have been calculated for the Eulerian residual current and the Stokes drift using Equations 3 and 4, respectively. Results obtained for longitudinal and lateral directions are presented in Table 2. The longitudinal Eulerian residual current always has values an order of magnitude higher than the Stokes drift effect. The relative contribution of individual salt-flux terms to the net transport varies considerably along the estuary and under different conditions responsible for residual transport (Table 3). In all cycles, the main mechanism responsible for salt flux is the advection of mean salinity by residual flow of water.

Mixing considerations

Values of $K_z^*$ can be derived from equation 12 using the results of $\langle S \rangle$ and $\langle U \rangle$ for the tidal cycles reported. The term $\partial \langle S \rangle / \partial y$ cannot be determined from observations. Dyer (1973) described the longitudinal
distribution of mean salinity in Southampton Water as:

\[ \langle S \rangle = 32.4 - 0.12x \]  \hspace{1cm} (14)  

where \( x \) is the distance from the mouth of the estuary, in kilometres. Thus, the term \( \partial(S)/\partial x \) is approximated by \(-0.12 \text{ km}^{-1}\). Furthermore, the term \( \partial(S)/\partial t \) which is included in Equation 12 has been shown by Uncles and Jordan (1979) to modify the computed values of \( K_z^* \) by less than 4%, and, thus, has been ignored.

Values of \( N_z^* \) are computed from Equation 13 using the results of \( \partial(U)/\partial x \), \( \partial(V)/\partial x \) and \( \langle S \rangle \). Since wind effects are not considered, it follows that \( N_z^*(\partial(U)/\partial x) = 0 \) at \( \sigma = 0 \). Measurements also do not determine sea-bottom stress, leading to the assumption that \( N_z^*(\partial(U)/\partial x) = 0 \) at \( \sigma = 1 \) (Bowden & Sharaf el Din, 1966). Finally, the inclusion of the term \( \partial(U)/\partial t \) modifies the computed \( N_z^* \) values by less than 1% and is ignored. Calculated values for the effective eddy viscosity and diffusivity coefficients together with the minimum and maximum coefficients as obtained from instantaneous velocity and salinity data are presented in Figure 5.

Discussion

Southampton Water is a partially-mixed estuary with a high tidal range and a significant freshwater discharge. Its circulation patterns vary in position along the main channel, depending on tidal amplitude, river discharge and meteorological conditions. The current system shows very complex patterns along the extensive tidal flats and the main channel. Although tidal flats experience flooding only for a few hours in a tidal cycle, providing the dynamics with their own distinct...
directional characteristics, it is the main channel of Southampton Water that is responsible for the overall transport capacity of the estuary. Here, large water masses are being moved back and forth influencing the distribution of transported substances and properties of water. Direct observations of the basic...
physical parameters using a time-series approach for complete tidal cycles illustrated the different circulation and mixing patterns that exist at three representative estuarine areas, under varying external boundary conditions. The variability at the instantaneous field of any variable during different stages of a tidal cycle, corresponds well with changes in the intensity of turbulence introduced in the water column.

Longitudinal Eulerian residual profiles appeared strongly dependent on the axial position. The upstream station (Cracknore) showed a net seaward flow throughout the water column, the mid-channel station (Netley) revealed a characteristic two-layered estuarine circulation due to the combined effect of rivers Test and Itchen, and the open boundary station (Calshot) was directly affected by meteorological forcings and local topographic irregularities. A deviation of the net current towards the east bank of the channel became obvious, especially in the upper parts of the estuary. This behaviour has also been observed by Dyer (1973, 1982) who reported lower cross-sectional salinities on the eastern side, leaving the oceanic water to enter the channel from the west. Such a circulation pattern is rare for rivers and estuaries, and must be accounted to the complicated bathymetry of the region and the presence of a local eddy producing early flooding effects on the eastern side. In particular for the Cracknore tidal cycle, it can be seen that the overall water column has an eastward deflection, probably due to the meandering structure of the river Test at that position.

Tidal prism values are representative of the deep channel conditions, where most of the cross-sectional flow occurs. The sign of local water prism volumes is mostly negative under neap tidal conditions and moderate to high river discharge. As tidal range increases at spring conditions, the volume of water entering the estuary during flood exceeds that of the ebb (Calshot, 8 November 1991). However, even under increased tidal range, the effect of high river flow and strong northerly winds can lead to an overall deficit in the estuary (Calshot, 12 February 1993). For Station B, similar conclusions can be drawn. The position of the station, however, makes the influence of river discharge much more considerable. Neap tidal conditions lead to negative values in the tidal prism water volume (Netley, 18 October 1991), especially under

<table>
<thead>
<tr>
<th>Tidal cycle</th>
<th>Longitudinal residual flux</th>
<th>Lateral residual flux</th>
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<tbody>
<tr>
<td></td>
<td>$V_1$ (m s$^{-1}$)</td>
<td>$V_2$ (m s$^{-1}$)</td>
</tr>
<tr>
<td>Calshot (8 March 1991)</td>
<td>0.015</td>
<td>0.004</td>
</tr>
<tr>
<td>Calshot (5 July 1991)</td>
<td>-0.027</td>
<td>0.009</td>
</tr>
<tr>
<td>Calshot (8 November 1991)</td>
<td>0.019</td>
<td>0.005</td>
</tr>
<tr>
<td>Calshot (12 February 1993)</td>
<td>-0.050</td>
<td>0.007</td>
</tr>
<tr>
<td>Netley (8 March 1991)</td>
<td>-0.041</td>
<td>0.007</td>
</tr>
<tr>
<td>Netley (1 July 1991)</td>
<td>0.036</td>
<td>0.013</td>
</tr>
<tr>
<td>Netley (7 August 1991)</td>
<td>-0.047</td>
<td>0.003</td>
</tr>
<tr>
<td>Netley (18 October 1991)</td>
<td>-0.012</td>
<td>0.001</td>
</tr>
<tr>
<td>Cracknore (26 September 1991)</td>
<td>-0.041</td>
<td>-0.009</td>
</tr>
<tr>
<td>Cracknore (1 October 1991)</td>
<td>-0.026</td>
<td>0.001</td>
</tr>
<tr>
<td>Cracknore (8 October 1991)</td>
<td>-0.039</td>
<td>-0.011</td>
</tr>
<tr>
<td>Cracknore (15 October 1991)</td>
<td>-0.032</td>
<td>-0.005</td>
</tr>
</tbody>
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The transition of the tidal prism values from negative to positive takes place for a tidal range between 2.7 m (Netley, 7 August 1991) and 3.1 m (Netley, 1 July 1991). For Station C, the tidal prism is always negative, as the effect of river flow is the most important factor for water residual transport. Values of the same order of magnitude have been calculated by de Jonge (1992) for the Ems Estuary. Results from that estuary show that water transport is relatively independent of external factors such as changes in tidal amplitude, river discharge and wind, thus representing more persistent phenomena (Zimmerman, 1976).

Lateral volumes in Southampton Water seem to become important and of the same order of magnitude as the longitudinal volumes, during weak neap tidal conditions. As the tidal amplitude increases for the spring tides, a considerable reduction in the lateral water transport is observed. This is mostly due to the larger area of exposed banks at low water leading to a considerable change in the shape of bank boundaries and thus influencing transverse flow to a greater extent than the longitudinal (de Jonge, 1992).

The analysis of the mechanisms responsible for residual water transport showed that the Stokes drift effect is apparently an order of magnitude smaller than the dominant advective mechanism. This can be explained by considering that although Stokes drift has been demonstrated to be an important mechanism in the generation of long-term residual transport in straits (Dyer & King, 1975) and channels (Hunter, 1972), its value remains small within each individual tidal period (Dyke, 1980). Similar results have also been reported by Uncles et al., 1986 for the Tamar Estuary. The longitudinal Eulerian residual current obtains its maximum seaward value under conditions of increased river flow (Calshot, 12 February 1993; Netley, 8 March 1991). The Stokes drift mechanism is mostly positive, thus pushing water upstream. It obtains its maximum value under spring tidal conditions (Calshot, 12 February 1993; Netley, 1 July 1991; Cracknore, 26 September 1991) at all stations. The lateral component of the Eulerian residual current obtains its maximum value under neap tidal conditions, whereas the spring tidal status results in restricting the flow towards the longitudinal direction.

<table>
<thead>
<tr>
<th>Tidal cycle</th>
<th>Longitudinal residual flux</th>
<th>Lateral residual flux</th>
</tr>
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<tbody>
<tr>
<td></td>
<td>$V_{S,1}$ ($10^{-3}$ kg m$^{-2}$ s$^{-1}$)</td>
<td>$V_{S,2}$ ($10^{-3}$ kg m$^{-2}$ s$^{-1}$)</td>
</tr>
<tr>
<td>Calshot (8 March 1991)</td>
<td>0.601</td>
<td>0.052</td>
</tr>
<tr>
<td>Calshot (5 July 1991)</td>
<td>-0.572</td>
<td>0.009</td>
</tr>
<tr>
<td>Calshot (8 November 1991)</td>
<td>0.467</td>
<td>0.006</td>
</tr>
<tr>
<td>Calshot (12 February 1993)</td>
<td>-1.715</td>
<td>0.053</td>
</tr>
<tr>
<td>Netley (8 March 1991)</td>
<td>-1.060</td>
<td>0.028</td>
</tr>
<tr>
<td>Netley (1 July 1991)</td>
<td>1.581</td>
<td>0.020</td>
</tr>
<tr>
<td>Netley (7 August 1991)</td>
<td>-1.437</td>
<td>0.000</td>
</tr>
<tr>
<td>Netley (18 October 1991)</td>
<td>-0.365</td>
<td>0.002</td>
</tr>
<tr>
<td>Cracknore (26 September 1991)</td>
<td>-1.588</td>
<td>0.013</td>
</tr>
<tr>
<td>Cracknore (1 October 1991)</td>
<td>-0.734</td>
<td>0.003</td>
</tr>
<tr>
<td>Cracknore (8 October 1991)</td>
<td>-1.379</td>
<td>0.011</td>
</tr>
<tr>
<td>Cracknore (15 October 1991)</td>
<td>-0.823</td>
<td>0.004</td>
</tr>
</tbody>
</table>
Stokes drift perpendicular to main axis is of the order of 0.001 m s\(^{-1}\), being mostly positive under spring tidal conditions. As a result, the lateral water flux is greatly affected by the magnitude of the non-tidal drift under neap tides and winter circulations, and by both the non-tidal and the Stokes drift under spring tide conditions.

The main mechanism of salt flux is the advection of mean salinity by residual water flow. This indicates that cross-sectional shear must be an important mechanism for the axial transport of salt in Southampton Water. However, a quantitative study of this contribution to the overall salt balance would require simultaneous observations over the whole cross-section. The magnitude of residual salt flux due to tidal pumping, indicates that this process is of particular importance for estuarine salt balance. In general, tidal pumping can become an important mechanism for the transport of salt in mesotidal and macrotidal partially-mixed and well-mixed estuaries (Uncles et al., 1985). Furthermore, tidal pumping, in conjunction with the resuspension of bed sediments appears to be important for the transport of suspended sediments in these estuaries (Uncles et al., 1985). The axial dispersion of salt due to vertical shear appears to be of the same magnitude as tidal pumping, with increased contribution as salt intrusion becomes more intense under spring tides.

Fluxes due to transverse variations in velocity, salinity and depth are of the same magnitude but have smaller values than the longitudinal ones. Similar results have also been obtained by Lewis and Lewis (1983), when calculating the salt fluxes of Tees Estuary, another narrow, macrotidal partially mixed system, basically due to the effect of wide and narrow shallow parts on each side of the main channel.

Effective coefficients are generally reduced in magnitude compared to the maximum instantaneous, since individual mixing incidents occurring in the water column during a tidal cycle are being smoothed by averaging. It is shown that the longitudinal variation of these coefficients corresponds to changes in mixing intensity. At the lower parts of the estuary (Calshot), a range of 0.2-15 cm\(^2\) s\(^{-1}\), with an effective value of 5 cm\(^2\) s\(^{-1}\), can be considered as

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**Figure 5.** Profiles of instantaneous minimum (□—□) and maximum (○—○) vertical salt and momentum eddy diffusion coefficients, \(K_Z, N_Z\) and effective \(K_Z^*, N_Z^*\) (△—△), for: (a) Calshot; (b) Netley; and (c) Cracknore.
Conclusions

This paper forms a basic and important preliminary work in understanding the physical characteristics of circulation and mixing of the main dredged channel of Southampton Water and the Test Estuary. Tidal effects appeared to be the dominant factor for the longitudinal and vertical distribution of physical variables, in relation to river flow and wind shear stress. However, analysis of the net longitudinal and lateral currents and fluxes of water and salt revealed the importance of these non-tidal effects in the mean tidal transport. The Stokes drift effect is an order of magnitude smaller than the previously mentioned advective mechanisms. It appears mostly positive in direction, for the tidal periods examined in Southampton Water, thus pushing water and salt upstream and increasing its magnitude during spring tidal cycles. The lateral Eulerian transports of water and salt are of the same order of magnitude as the longitudinal ones, but with smaller values, especially under spring tidal amplitudes, when the flow coincides better with the longitudinal estuarine axis.

Stratification conditions in the water column of the estuary were examined in instantaneous and mean tidal terms. As expected, well-mixed conditions were apparent during most of the tidal cycle at the lower parts of Southampton Water, with stratification effects being enhanced moving in the upstream direction. Direct observations at these three representative areas, in terms of mixing patterns, were used to calculate the range within which these coefficients vary in time and space under different boundary forcings. It was noted that these coefficients obtain near-zero values at the surface and the bottom layers, but maximum values at the mid-depth region, where most intense mixing occurs, especially when the levels of turbulence from the tidal current remain high.

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