CHAPTER III

Mapping exchange and residence time

1. Introduction

This study examines the mechanisms and timescales of exchange between Willapa Bay, Washington, USA (Fig. 3.1) and the coastal ocean. Willapa is the largest in a series of shallow, coastal-plain estuaries that spans the U.S. Pacific Northwest coast from central Washington to northern California. Note that The Columbia River, immediately south of Willapa, is so different from its neighbors in morphology and river input (Hickey and Banas 2003) that we do not treat it as part of this series. These estuaries are forced by strong (2–3 m) tides, and have relatively deep and unchannelized intertidal zones (Emmett 2000): fully half of Willapa's area and volume are intertidal, and this appears to be typical (Hickey and Banas 2003). In addition, these estuaries are subject to changes in ocean water properties forced by wind-driven upwelling-downwelling transitions on both seasonal and event (2-10 d) timescales. The best analogies are the shallow, macrotidal estuaries of Britain and Northern Europe (Bowden and Gilligan 1971, Zimmerman 1976, Dronkers and van de Kreeke 1986, Simpson et al. 2001) and the upwelling-influenced estuaries of Spain and other eastern boundary systems (Álvarez-Salgado et al. 2000, Monteiro and Largier 1999).

River input into these estuaries occurs mostly in winter, outside the growing season (Hickey and Banas 2003). Accordingly, primary production in Willapa appears to be fueled less by riverine nutrients than by oceanic nutrients or the direct import of oceanic phytoplankton blooms (Roegner et al. 2002, Ruesink et al. 2003, Newton and Horner 2003). The dynamics of ocean-estuary exchange—its spatial pathways, its constancy or variability under changing forcing—are thus essential ecological concerns. These dynamics are the focus of the modeling work described in this paper.

Recent observational studies have provided a preliminary description of exchange processes in Willapa. Hickey et al. (2002) found that variations in ocean temperature and
salinity propagated upstream at approximately 10 cm s\(^{-1}\) even in summer, low-riverflow conditions, much faster than could be attributed to the river-driven gravitational circulation. Banas et al. (2004) found that even during winter storms, when riverflow is 200 times higher than in the late-summer dry season, the overall rate of exchange only increases by a factor of 3. Thus unlike otherwise similar systems like Tomales Bay, California, where a strong seasonal cycle in riverflow yields a strong seasonal cycle in flushing rate (Smith et al. 1991), exchange between Willapa and the ocean appears to be cushioned by a high "baseline" flushing rate independent of the river-driven circulation. Banas et al. (2004) suggested that the bulk of this baseline exchange is caused by lateral tidal stirring.

The observational analyses cited above relied on multi-year velocity and salinity time series in a few point locations, and thus addressed the temporal variability of Willapa's circulation more than its spatial complexity. Even in very simple estuarine geometries, understanding the horizontal structure of the net circulation is an active research problem (e.g., Valle-Levinson et al. 2003). Willapa's geometry, moreover, is not simple: the bay is a network of branching channels formed when a sand spit from the Columbia River (the Long Beach Peninsula) trapped a number of small rivers behind one connection to the ocean (Fig. 3.1; Emmett 2000). The purpose of this modeling study is to provide detailed maps of exchange and residence time in this complex geometry over a typical seasonal cycle.

In many numerical studies of estuaries, flushing rate and residence time are by necessity pure model predictions that can only be validated circumstantially. Here we use independent, observational estimates of dispersion rates at four salinity time-series locations (Banas et al. 2004; Sec. 3b below) to test modeled dispersion rates and residence times directly. The model's role, then, is principally to fill in spatial detail around these previous empirical conclusions.

Section 2 of this paper describes the numerical model. In Sec. 3 we describe the model's representation of the tidal circulation and map out the net tidal transport pathways. In Sec. 4 we add riverflow and density effects to the model, and map the changes in residence time between low-flow (summer) and high-flow (winter) conditions.
These results rely on particle-tracking and nonconservative-tracer model methods that are likely to be useful in other complex estuaries.

2. The model

a. Model physics

The Willapa circulation model is an implementation of GETM (General Estuarine Transport Model), a finite-difference, primitive-equation model designed for shallow-water applications, cases like Willapa where flow over complex topography, mixing in strong and changing stratification, and flooding and drying of intertidal areas are all important. GETM has previously been used to model tidal dynamics in the East Frisian Wadden Sea, which is more than half intertidal (Stanev et al. 2003); baroclinic dynamics and the estuarine turbidity maximum in the Elbe (Burchard et al. 2004); and seasonal hydrographic patterns in the North Sea (Stips et al. 2004). GETM is open-source and still under development: the project home page can be found at http://www.bolding-burchard.com. Below we highlight only some key features of GETM: for a systematic description of model physics and numerics, see Burchard and Bolding (2002) and Burchard et al. (2003).

GETM solves the equations of motion on a curvilinear Arakawa-C grid (Arakawa and Lamb 1977). To resolve fast-moving surface gravity waves without limiting the timestep for the entire calculation, the solution is split into internal and external modes, with ten external (barotropic) timesteps taken within each internal timestep. The model contains a simple, stable scheme for handling the flooding and drying of intertidal areas, in which a depth-dependent multiplier is applied to the equations of motion to smoothly reduce the role of all terms except tendency, bottom friction, and barotropic pressure gradient as the water depth decreases to a few cm. In addition, to prevent unphysical pressure gradients from developing, a correction to sea level is made where the cell-to-cell change in elevation is comparable to the water depth (Burchard et al. 2004). This
scheme allows the model to accommodate strong tracer gradients in intertidal areas.

GETM allows a wide choice of high-order advection schemes for velocity and tracers. High-order schemes in general are far less diffusive than a basic first-order upwind scheme, and therefore better suited to resolving strong vertical and horizontal gradients (Stips et al. 2004) and flows in relatively quiescent areas like shallow banks (Gross et al. 1999). Both Gross et al. (1999) and Stips et al. (2004), in their evaluations of a number of advection schemes in coastal and estuarine simulations, ultimately recommend the Superbee (Roe 1985) and Quickest (Leonard 1979, 1991) schemes. Both of these are total-variation-diminishing (TVD) and therefore free from numerical oscillations and guaranteed to be stable. The two schemes perform comparably well in both studies (Gross et al. 1999, Stips et al. 2004); in this study we use Superbee, which is less computationally expensive.

GETM also allows a choice of turbulence closures by coupling to the one-dimensional model GOTM (General Ocean Turbulence Model: Burchard et al. 1999). Here we use a standard $k$-$\varepsilon$ scheme (e.g., Burchard and Bolding 2001) with the stability functions suggested by Canuto et al. (2001). This scheme is among those that Umlauf and Burchard (2003) have shown can be represented as a special case of a single two-equation, "generic length scale" formulation with variable coefficients. Warner et al. (2005) compared a number of these two-equation special cases ($k$-$kl$, $k$-$wo$, $gen$, and the $k$-$\varepsilon$ scheme we use) and found the differences in their ability to reproduce estuarine salinity fields to be relatively minor, although all performed better than the original Mellor-Yamada level 2.5 scheme (Mellor and Yamada 1982). For a general review of two-equation closures, see Burchard (2002).

b. Implementation for Willapa Bay

This study uses a model grid with 175-by-82 cells in the horizontal and 12 sigma levels in the vertical. The bay itself is covered with uniform horizontal squares 255.5 m on a side; beyond the bay mouth lies an idealized rectangular "ocean" in which the grid size expands gradually to 6 km (Fig. 3.1). This domain was designed to resolve Willapa's
internal dynamics only, not the dynamics of the adjacent coastal ocean. The model "ocean" (Fig. 3.1) is a simple semi-enclosed reservoir, open only to the west, with depth limited to 30 m, since realistic depths would limit the timestep for the whole model. Additionally, the model has no alongshore currents, which on the real Washington coast are tens of cm s\(^{-1}\) and highly variable (Hickey 1989). Removing any of these simplifications would provide a false sense of realism unless they were all removed, and even then model validation would be tenuous given the scarcity of high-resolution observations on the adjacent shelf. Coupling Willapa's internal dynamics to shelf processes thus remains a future goal.

The bathymetric grid within the estuary is interpolated from a finite-element model grid developed by the U. S. Army Corps of Engineers Seattle District, who resurveyed most of the subtidal area of the bay in 1998 (Kraus 2000). Willapa's primary channels and many of its secondary channels are wide enough (500–3000 m) to be well-resolved by our 250 m model grid, although in many locations real channel edges are steeper than our grid allows. Intertidal bathymetry in our model, like that in the published NOAA charts for Willapa, is a composite of surveys over many decades. In 2002 the NOAA Coastal Services Center conducted a high-resolution LIDAR (Light Detection and Ranging) survey of the shoals of Willapa Bay (http://www.csc.noaa.gov/lidar), but no seamless, georeferenced bathymetry based on this dataset has yet appeared. Willapa's morphology is, in any case, a moving target—channels can migrate 100 m or more in a single year (e.g., Hands and Shepans 1999) —and therefore as much as possible the analysis below is based on spatially integrative validation methods that do not rely on point-by-point comparisons.

All model scenarios described in this paper are summarized in Tables 3.1 and 3.2. A first set of runs, used for model validation (runs A-C), simulates particular time periods. Another set (E-G) uses idealized tidal and river forcing to represent a typical seasonal cycle, from winter storms (riverflow ~ 1000 m\(^3\) s\(^{-1}\)) to spring (flow ~ 100 m\(^3\) s\(^{-1}\)) to late summer (flow ~ 0). Run D is a version of the idealized late-summer scenario used both for validation and for examining tidal transport pathways in detail (Sec. 3). A final set of runs (H-I) models the propagation of individual, event-scale pulses of ocean water
into the estuary.

The tidal, river, and ocean-water-property forcings applied in each run are specified in Tables 3.1 and 3.2. The tide in the model was generated by imposing a sea-level time series on the open ocean boundary (Fig. 3.1). Ocean water properties, when included, were applied on the open boundary as well. In runs where river input was applied, it was divided among the bay’s three largest rivers according to watershed area: 47% in the North River, 35% in the Willapa, and 18% in the Naselle (Fig. 3.1). In several runs where riverflow was not applied (A, B, D, E, H: Tables 3.1 and 3.2), we eliminated all density effects—i.e., all density variations in both space and time—in order to isolate the role of tidal dynamics. This was done by switching off the call to the equation-of-state calculation in the code, so that model tracers had no effect on the circulation, acting like dye rather than like salinity. In addition to "dye" and "salt," a third type of tracer—"age," a nonconservative, passive tracer representing average time spent in the estuary—was included in some runs, as explained in detail in Sec. 4.

The tide-only, density-off runs (A, B, D, E, H) were spun up from rest over 4 tidal cycles (2.1 days). This is a short adjustment period, but sufficient that when a pure M2 tide was applied, the variation in tidal currents among subsequent cycles (which in a fully spun-up model would be be zero) was only 0.02%. Runs that included density (C, F, G, I) were spun up for 100 days—approximately twice the longest residence times we calculate in Sec. 4—with initial tracer fields held constant before tracers were released and allowed to evolve. These spin-up periods are not included in the results below: in all cases the "initial" state of a run refers to the end of the spin-up period.

3. Tidal exchange

We begin by evaluating the simplest dynamical situation, in which the oceanic tide is the only forcing: river input is zero, as if in late summer, and no density variations in either space or time are included. We will begin by verifying the overall strength of modeled tidal currents against data (Sec. 3a), and then move on to the much weaker but dynamically more important subtidal (i.e., tidal-cycle-average or "residual") circulation.
a. Validation of tidal currents

For validation of the tidal model (runs A and B: Table 3.1), we aimed to reproduce the tides during particular time periods in which velocity data were collected. No tuning of model parameters was done: the bottom roughness \( z_o \), sometimes treated as a free parameter, was set to a standard value of 1 mm. NOAA tidal-height observations from Toke Point (station 9440910; Fig. 3.1) were used as the oceanic boundary condition with no adjustment, except for the shift of timebase necessary to match the observed tidal phase at Toke Point. This is a simple and convenient method, but a source of error: as this tidal signal propagates into the bay it distorts slightly and produces a tidal-height discrepancy at Toke Point of 5-10%.

Figure 3.2 shows the depth-averaged amplitude of the M2 (semdiurnal) tidal constituent for the period Oct 15 – Nov 11, 1998 at four locations, stations at which the Army Corps of Engineers Seattle District collected acoustic doppler current meter (ADCP) data during this period (run A: Table 3.1). The M2 constituent was extracted from data and model time series using the harmonic-analysis package \textit{t_tide} (Pawlowicz et al. 2002). Model velocity amplitudes agree with observations to 3-20%, an error not much larger, encouragingly, than the 5-10% error caused by the imprecision of the boundary forcing. The model does not, however, reproduce the direction of M2 currents at the three stations near the mouth: the real flow follows bends in the channel that the modeled flow does not, presumably because the model bathymetry is smoother and channel edges less abrupt than in the real bay.

This sort of point-by-point bathymetry-driven bias may or may not translate into bias in overall tidal transport, however. To test the tidal model more integrally, on a larger spatial scale, the model was run for the week ending on May 5, 2000, a day on which small-boat transects with a 300 kHz ADCP were repeated over a cross-section of the main channel for a full flood tide (run B: Table 3.1). Instantaneous transport through this cross-section (expressed as mean velocity) is shown in Fig. 3.2 for both the observed and modeled flow: each black dot represents an ADCP transect. Instantaneous mean
velocities in model and data agree within 10% for each of the ten transects. This suggests that local biases like the directional errors near the mouth described above (Fig. 3.2) are indeed local, and that the total tidal transport in the model is well-represented.

b. Validation of tidal dispersion: the horizontal diffusivity $K$

The tidal validation above mainly tests the model’s reproduction of the oscillatory part of the tide, which, although the largest velocity signal in the estuary ($\sim 1 \text{ m s}^{-1}$: Fig. 3.2) by definition does not contribute to net dispersion or exchange: it simply advects water back and forth. The “rectified” or “residual” tidal currents actually responsible for dispersion in estuaries tend to be only $\sim 10\%$ of the tidal amplitude (Zimmerman 1986), comparable to model-data error in our case as in many cases. Horizontal tidal dispersion in the model must thus be evaluated by other means.

Banas et al. (2004) calculated a “total effective horizontal diffusivity” ($K$) from salinity time series at each of four monitoring stations along the main channel, and used the resulting along-channel profile to characterize the subtidal circulation. In this section we will use the same parameter and compare results directly with those from observations. This is a stringent test of a tidal model, one not often performed for lack of a method for estimating $K$ from data.

A brief explanation of $K$ in terms of the estuarine salt budget follows: a more detailed derivation can be found in Banas et al. (2004). In general, we can express the conservation of salt (or some other tracer) on subtidal timescales as

$$\frac{\partial}{\partial t} \int_{-\infty}^{x} \langle a \bar{s} \rangle dx = \langle a \bar{u} \bar{s} \rangle$$  (3.1)

where $x$ is distance along the main axis of the estuary ($x = -\infty$ denotes "far upstream"), $t$ time, $a$ cross-sectional area, $u$ velocity perpendicular to the cross section, and $s$ salinity. Triangular brackets denote the average over a tidal cycle, and an overbar denotes the cross-sectional average: for any variable $v$, $\bar{v} = a^{-1} \int v da$. Equation (3.1) simply states that the rate of change of salt storage upstream of $x$ equals the total salt flux across a section at $x$. The salt flux $\langle a \bar{u} \bar{s} \rangle$ contains all processes moving salt in and out of the
estuary, including the river-driven circulation, tidal stirring, wind-driven exchange, and interactions among them. A common approach (e.g., Ridderinkhof and Zimmerman 1990, Monismith et al. 2002, Austin 2004) is to assume that these processes combine to form a pattern of advection too complicated to describe as anything but diffusion. Following Banas et al. (2004) we can write this as

\[ \frac{\partial}{\partial t} \int_{-\infty}^{x} \langle a\tilde{s}\rangle dx + Q\langle \tilde{s}\rangle = \langle a\rangle K \frac{\partial \langle \tilde{s}\rangle}{\partial x} \]  

(3.2)

where \( Q \) is riverflow — \( Q < \tilde{s} > \) is the seaward flushing of salt by the mean flow — and \( K \) is a horizontal diffusivity parameterizing all processes moving salt upstream. If we confine our attention to low-riverflow conditions, in which \( Q \approx 0 \), we can rearrange (3.2) into a formula for calculating \( K \) from the salinity field:

\[ K = \langle a \rangle^{-1} \left( \frac{\partial \langle \tilde{s}\rangle}{\partial x} \right)^{-1} \frac{\partial}{\partial t} \int_{-\infty}^{x} \langle a\tilde{s}\rangle dx \]  

(3.3)

We use salinity here only because it is the most common tracer used to indicate seawater content; to the extent that \( K \) really does represent a large-scale diffusive process, any conserved tracer can be substituted.

To calculate \( K \) from the tidal model, we constructed the following idealized scenario (run D: Table 3.1). We impose a simple, uniform semidiurnal signal whose amplitude (1.2 m) was chosen to match the long-term-average tidal-height variance at Toke Point. An initial field of "dye" with a uniform gradient in the along-channel direction is released into the spun-up tidal circulation (Fig. 3.3a). The initial dye concentration is high in the ocean and decreases up-estuary as salinity would, although this dye has no density and thus does not affect the circulation. The rearrangement of this dye field by a single tidal cycle is shown in Figs. 3.3a-c, and its state after 12 cycles (6.2 days) is shown in Fig. 3.3d. The large tongue of high dye concentration in Fig. 3.3d indicates water that has entered the domain across the open ocean boundary, where a constant concentration for incoming water was imposed.

An average value of \( K \) was calculated using (3.3) for this tracer field over the 12 tidal cycles shown, at a series of cross-sections along the longest axis of the bay (Fig. 3.3e) from the mouth to Shoalwater Bay in the south. Empirical values of \( K \) are given for
comparison at four locations. Three are values calculated by Banas et al. (2004) from low-riverflow-period salinity data (Fig. 3.3f gives an example). The fourth value, previously unreported, was calculated from a four-week (Aug - Sep 2002) deployment of a YSI temperature-salinity sensor at the edge of the main channel in Shoalwater Bay. The rate of change of salt storage at this station was calculated from the trend in salinity rather than event-scale fluctuations as in Fig. 3.3f, but otherwise followed the method described by Banas et al. (2004).

The four empirical $K$ values vary by more than an order of magnitude from mouth to head. Crucially, the model captures both high and low values accurately (Fig. 3.3e). This spatial gradient in diffusivity can be seen qualitatively in Figs. 3.3a-d: the dye field in the southern part of the bay (low $K$) changes only gradually, whereas a single tidal cycle (Figs. 3.3a,c) is enough to dramatically displace, stretch, and fold the dye field near the mouth (high $K$). Note that the only station where model and data values of $K$ differ significantly, Oysterville, lies at the junction of two channels (Fig. 3.1), where the geometric assumptions behind the Banas et al. (2004) calculation may be violated: either the model or the "empirical" estimate may be the source of error at this station.

c. In pursuit of salt-flux mechanisms

A map of $K$ as in Fig. 3.3e is a convenient summary of the subtidal circulation, but obscures the actual mechanisms that drive it. What is different about the tidal flow or the bathymetry in the seaward and landward reaches of the estuary, that the net dispersion rate should vary so much between them? What aspect of the bathymetry—channel width, channel curvature, channel junctions, bank-to-channel depth variations—is most responsible for generating the tidal asymmetries that in turn generate $K$? A standard approach to these questions (Fischer 1976, Hughes and Rattray 1980, Lewis and Lewis 1983) is a Reynolds-like decomposition of the total salt flux $<a\bar{s}>$ (see equation (3.1)). That is, we can decompose $s$ into mean and varying parts in space and time:

$$s = <\bar{s}> + <s'> + \bar{s}_1 + s'_1$$

(3.4)

where triangular brackets and an overbar denote tidal-cycle and cross-sectional means as
before, a subscript 1 denotes variations over a tidal cycle, and a tick denotes variations over a cross-section. We can also, following Dronkers and van de Kreeke (1986), write the mass flux through a small cross-sectional area element as

$$dq = u \, da = <dq> + dq_1$$

(3.5)

so that the total tidally averaged salt flux through a cross-section becomes

$$<a\bar{s}t> = \int s <dq>$$

$$= <\bar{s}> <q> + \int <s'> <dq> + <\bar{s}_1, q_1> + \int <s_1', dq_1>$$

(3.6)

If the volume of the estuary is constant and river input $Q = 0$, $<q> = 0$, leaving three terms on the right-hand side. Term $(i)$ represents steady shear dispersion by the tidal-residual eddy field $<u>$: Zimmerman (1986) refers to this mechanism as "Lagrangian chaos," since the velocity field itself is steady and deterministic but randomized by bathymetry to the point where tracers and particles disperse chaotically. Term $(ii)$, which describes correlations over the tidal cycle between $s$ and $q$, can be thought of as "tidal trapping" (Okubo 1973, Fischer 1976), the result of a phase lag that arises when salt is diverted into side embayments and reenters the main flow at a different point in the tidal cycle. The cross-term $(iii)$ involves variations in both space and time: unsteady shear dispersion, or spatially non-uniform trapping.

The relative strength of $(i)$, $(ii)$, and $(iii)$ are shown in Fig. 3.4 for each of the cross-sections where $K$ was calculated in the previous section. These salt fluxes have been converted into diffusivity units (m$^2$ s$^{-1}$) by combining equations (3.3) and (3.6), i.e., by normalizing by $<a><\partial \bar{s}/\partial x>$. Total $K$ (dashed line) decreases smoothly from the mouth upstream, but the three constituent diffusivities (bars) rise and fall in a much more complicated pattern. Each of $(i)$, $(ii)$, and $(iii)$ is in some location the dominant term; in fact, the cross-term $(iii)$ (gray bars), the term hardest to interpret, is dominant over much of the upper estuary. This implies that much of the net exchange of ocean water over the sections examined is accomplished not by general, easily parameterizable features of the flow but by processes localized in both space and time. Furthermore, all three terms are in some location negative, which could mean that a decomposition based on cross-sectional averages is not physically meaningful in bathymetry this complicated.
Alternatively, these negative constituent diffusivities could mean that certain tidal-residual processes in the bay are actually anti-diffusive in a coarse-grained sense, tending to sharpen fronts rather than smooth them away. Note that the initially smooth tracer field shown in Fig. 3.3a does in fact become more inhomogeneous as tidal stirring acts on it (Fig. 3.3d).

It appears, then, first, that the cross-sectionally integrated diffusivities shown in Figs. 3.3e and 3.4 are averages over a great deal of small-scale complexity and variety; and second, that the traditional method of decomposition, based on fixed-location tidal-cycle and cross-sectional averages, is not an effective way to tease apart this complexity. In the next section we will describe an alternative, flow-following approach that gives clearer results.

d. A Lagrangian map of exchange

We begin by releasing an imaginary particle in the center of each model grid cell, and track the trajectory of these particles as the tidal flow calculated in run D (Fig. 3.3) advects them seaward and landward. Three-dimensional particle tracking is not straightforward in sigma coordinates and steep bathymetry; to keep the problem tractable we follow particles in two dimensions only, using depth-average currents. This simplification would not be dynamically sensible in many scenarios, but in our tide-only case it appears to be fair: when terms (i) and (iii) in equation (3.6) are decomposed into vertical-mean and -varying parts (not shown), horizontal shear dispersion dominates over vertical dispersion by an order of magnitude. Nevertheless, the two-dimensional particle-tracking approximation introduces 10-20% volume-conservation errors, which should be taken as a lower limit on the uncertainty in this analysis.

The only information we retain from these particle tracks is the index of the grid cell where each particle is launched and the cell that contains it one tidal cycle later. The rearrangement done by the subtidal circulation is thus encapsulated in a single two-column table, a function that maps the list of grid-cell indices back onto itself: in the language of nonlinear dynamics, a "return map" or Poincaré map (Beerens et al. 1994).
One application of this Lagrangian return map is shown in Fig. 3.5. Here the only particles drawn are those which cross over a given section (green lines) within one tidal cycle. Lighter shades indicate initial position and darker points final position one tidal cycle later; particles are colored red if they are crossing up-estuary and blue if they are crossing seaward, as illustrated in the schematic in Fig. 3.5a. This illustration of the crossing regions for each section of interest (Fig. 3.5b-g) highlights not only the total amount of exchange at each section (which could also be inferred from the diffusivity $K$) but specifically what water is being exchanged for what other water in a typical tidal cycle. The details of these results depend on the tidal phase at which particles are released, but the qualitative patterns discussed below are not sensitive to it. Results are shown in Fig. 3.5 for particles released at high slack, the release time that minimized volume-conservation errors.

Note that dividing the volume of a given section's crossing regions by the intertidal volume upstream gives the "tidal exchange ratio," or fraction of each tidal prism that does not return on the following tide. The crossing regions shown in Fig. 3.5b-g correspond to tidal exchange ratios of 0.15 - 0.3, consistent with the range of typical values reported by Dyer (1973) and the estimate for Willapa from data in Banas et al. (2004).

Coherent structures in the net circulation emerge. Exchange near the mouth (Figs. 3.5b,c) consists mainly of a simple lateral exchange, into the estuary over the southern part of the mouth and seaward in the deeper main channel to the north. This pattern, in which the subtidal flow tends to the right both entering and leaving the estuary is consistent with the expectation for rotation-driven exchange (Valle-Levinson et al. 2003), but it persists even when the Coriolis acceleration is turned off in the model (Fig. 3.5h). An alternative explanation may be tidal Stokes drift, which tends to produce an up-estuary Lagrangian transport through the shallower part of a cross-section, balanced by a seaward return flow in the deeper main channel (Li and O'Donnell 1997), as seen here. The lateral exchange pattern persists for 15-20 km upstream from the mouth (Fig. 3.5d-f), but its relationship with the depth of the underlying bathymetry reverses: more than a few km from the mouth (Fig. 3.5d,e) flow over the banks is primarily seaward and flow in the
main channel up-estuary (bathymetry shown in Fig. 3.1). Analytical studies and observations of lateral transport gradients in other estuaries have found both the pattern at Willapa's mouth (channel flow seaward: Friedrichs and Hamrick 1996, Li and O'Donnell 1997), the pattern in Willapa's interior (channel flow landward: Valle-Levinson and O'Donnell 1996, Winant and Gutiérrez de Velasco 2003), and indeed a reversal of the pattern along the estuary (Li 1996), as we observe.

In the interior of the estuary (Fig. 3.5d-g) there are places where the Lagrangian residual circulation appears more diffusive than advective, in the sense that holes and discontinuities appear in the crossing regions, and in places the incoming and outgoing water masses appear to move through each other rather than past each other (Fig. 3.5g). Such complexity is especially visible in the exchanges around channel junctions highlighted in Fig. 3.6. These junction exchanges combine coherent, advective motion (water enters Willapa Channel and then leaves again, as on a counterclockwise conveyor belt: Fig. 3.6b) with discontinuities and dispersion on scales from the grid resolution up to 10 km.

Thus in contrast to the $K$ formulation (Sec. 3b) in which all exchange flows are treated mathematically as diffusion, and also in contrast to the Eulerian decomposition (Sec. 3c) in which only simple advective patterns are interpretable, this crossing-region method lets us distinguish in detail between advection-like and diffusion-like processes. It also lets us visualize hybrid, unnameable transport processes like those around channel junctions. Note that these hybrid junction processes may be important to exchange overall: the highest diffusivities in the southern half of the estuary (Fig. 3.3e) lie at the junction at the mouth of Shoalwater Bay (Fig. 3.6a).

4. Density-driven exchange

a. Baroclinic validation: the salinity field

So far our analysis has considered negligible-riverflow, tide-dominated conditions only. In this section we add rivers and ocean density to the model, and evaluate their
effect on flushing rate relative to the tidal baseline. We must begin, however, by verifying the model's baroclinic behavior in general: we will do this by validating the model salinity field against observations across a range of river and ocean forcing.

For validation, the model was run with time-varying sea-level, riverflow, and ocean-salinity forcing for the 11-month period July 1999 – June 2000 (run C: Table 3.1). During this period, as described by Banas et al. (2004), point salinity time series were collected at four main-channel stations: W3 near the mouth at 8 m depth, and Bay Center, Oysterville, and Naselle farther up-estuary at 1 m depth (Fig. 3.7, insets). For model forcing, sea level from Toke Point was imposed at the ocean boundary as in runs A and B (Sec. 3a); both tidal and subtidal components were included. River input into the North, Willapa, and Naselle Rivers was calculated from USGS gauge data as described in Banas et al. (2004). Finally, for lack of a better option for representing changing ocean salinity, tidally-averaged salinity from W3 was imposed at the ocean boundary as well.

This last approximation is the strongest limit on our ability to simulate particular time periods. Over this 11-month run, the distortion as the imposed W3 salinity time series propagates from the open boundary back to W3 is 27% of the signal. As a result, it does not seem meaningful to compare model results with data instant by instant, but rather to treat model and data as if they represented similar but distinct time periods. A stricter baroclinic model validation must wait until better offshore data or a high-resolution, large-scale coastal model is available to provide better salinity boundary conditions.

The most integrative test of the model's baroclinic circulation is the length of the salt intrusion as a function of riverflow. A relative measure of salt intrusion length—salinity at Bay Center, Oysterville, and Naselle as a fraction of salinity at the mouth (station W3)—is shown for both observations and model results in Fig. 3.7. This relative salinity is displayed as a function of bay-total riverflow Q: the gap around Q = 30 m$^3$ s$^{-1}$ corresponds to a six-week period in autumn 1999 when W3 salinity was unavailable (a linear trend was substituted as the model boundary condition during this period). At Oysterville and Naselle, the slope between relative salinity and riverflow is indistinguishable between model and data at the 95% confidence level. Furthermore, the
model reproduces, qualitatively, the scatter—the event-to-event variance—in relative salinity at a given riverflow level. At Bay Center, however, the dependence of salinity on riverflow in the model is too weak. This is a sign either that our simplified treatment of the coastal ocean distorts the salinity field near the mouth, or that the model is not correctly distributing freshwater from the North and Willapa Rivers around the Stanley Channel-Willapa Channel junction (Fig. 3.1). In either case, the bias in the salinity field appears to be a local effect that does not reach as far south as Oysterville.

Furthermore, the model appears to match, qualitatively, observations of stratification at Bay Center from a more recent (previously unreported) experiment. Two SeaBird conductivity-temperature (CT) sensors were deployed at Bay Center from Oct 2003 to Apr 2004, one at 1 m depth as above, and the other 0.5 m off the bottom. W3 was not occupied during this deployment, and so for lack of a salinity boundary condition we cannot simulate this time period in the model directly. The 1999-2000 model run described above (run C) does, however, reproduce the most striking feature of the 2003-04 observations (Fig. 3.8a,b): at riverflows above the long-term median (~ 100 m³ s⁻¹), the water column stratifies 1-5 psu on each ebb and then destratifies again on flood. Similar patterns of strain-induced periodic stratification (SIPS) have been frequently observed in macrotidal estuaries (Simpson et al. 1990, Stacey et al. 2001), although they have not been documented previously in Willapa. Notably for model validation, in both model and data, the only flow levels at which the flood tide is unable to completely re-homogenize the water column for more than 1-2 cycles at a time are those close to the annual maximum, > 1000 m³ s⁻¹ (in observations, early Feb 2004; in model, late Dec 1999; arrows, Fig. 3.8a,b). This model-data comparison is shown more systematically in Fig. 3.8c,d. At low-to-moderate riverflows, the maximum stratification attained during each tidal cycle ranges from 0-3 psu in both data and model, as the bay responds to various combinations of river and ocean water-property forcing. At high flows, the model underestimates the highest stratifications at Bay Center, but reproduces the fact that during at least some of these high-flow events, tidal-cycle-minimum stratification does not drop below the observational limit of 0.05 psu, as it does during almost all other conditions (Fig. 3.8d).
Until better boundary-condition data are available to make the baroclinic validation more precise, it would be over-trusting to use the model to map Willapa's river-driven exchange circulation in as much detail as we mapped the tidal circulation in Sec. 3d. Still, we have found that the modeled vertical circulation at mid-estuary (as indexed by Bay Center stratification) is, at least, not grossly wrong; and in fact that the slope of relative salinity at Oysterville and Naselle vs. riverflow—i.e., the distance that water entering the mouth reaches into the southern estuary under complex, changing forcing—matches the slope from observations within confidence limits (Fig. 3.7). In the next section we describe a numerical method by which river- and ocean-density-forced changes in circulation can be translated into summary maps of bay-wide residence time.

b. A method for mapping residence time

The method we use to determine residence time—the nonconservative "age" tracer mentioned in Sec. 2—is similar to the ventilation-time method (time since a water parcel has been exposed to the atmosphere) used in global circulation modeling (Thiele and Sarmiento 1990, England 1995). For a rigorous, general derivation of this and a variety of other age methods in ocean modeling, see the thorough review by Deleersnijder et al. (2001); we will explain this age-tracer more briefly here.

Consider a passive tracer whose concentration is initially zero everywhere, and that is always zero in the estuary's river and ocean source waters, but that is made to grow at a constant, specified rate in the estuary's interior. Between the open ocean boundary and the mouth of the estuary, the tracer simply advects and mixes without growing. As time passes, if forcing is held constant, the tracer concentration will approach an equilibrium in which the steady creation of tracer within the estuary is balanced by flushing out to sea (which replaces high-concentration estuarine water with zero-concentration ocean water). At this equilibrium, the concentration in each grid cell is directly proportional to the average length of time the water in that grid cell has spent growing, i.e., spent within the estuary. Thus we can think of this tracer as average water age.
As an illustration, consider the case of a riverless, linear channel: in the cross-sectional average, the conservation equation for age can be written

$$\frac{\partial c}{\partial t} = \frac{1}{a} \frac{\partial}{\partial x} \left( aK \frac{\partial c}{\partial x} \right) + \gamma$$  \hspace{1cm} (3.7)

where $c$ is age-tracer concentration, $a$ and $K$ are, as before, cross-sectional area and a total effective diffusivity, and $\gamma$ is the imposed tracer growth rate. We can pick the units of $c$ so that $\gamma = 1$. For given profiles of $a(x)$ and $K(x)$, Equation (3.7) has a steady solution, given by a double integration of

$$-\frac{1}{a} \frac{\partial}{\partial x} \left( aK \frac{\partial c}{\partial x} \right) = 1$$  \hspace{1cm} (3.8)

with the boundary conditions $c = 0$ at the ocean and $(aK)(\partial c/\partial x) = 0$ (no tracer flux) far upstream. Equation (3.8) describes a state in which the along-channel tracer gradient $\partial c/\partial x$ has adjusted so that the diffusive loss of tracer (left hand side) balances the steady, imposed tracer growth rate (right hand side) at every location. If $a$ and $K$ decrease exponentially toward the head with e-folding lengths $l_a$, $l_K$, then at equilibrium, age increases approximately exponentially from mouth to head: $c(x) \propto l_a l_K K(x)^{-1}$. In this simple case, as one might expect, age $c$ (an estuarine residence time) is inversely proportional to $K$ (a measure of the estuarine exchange rate). Equation (3.8) implies that this is likely to be true in general, at least in a scaling sense, a fact we can use (Sec. 4d below) to check the reasonableness of the residence times for Willapa that the age-tracer method yields.

We implemented this age-tracer in GETM by first removing references to temperature in the equation-of-state calculation (so that it can be used as a generic passive tracer) and second adding code to the temperature-advection subroutine to increment the tracer one unit each timestep at grid points within the estuary. This tracer thus is transported and mixed using GETM’s full advection and turbulence schemes, without the need for postprocessing. Crokelet and Ebbsемсяyer (1991) used a similar method to calculate residence time in a simplified box model of Puget Sound. We have not seen this method used in a high-resolution estuary model, although it requires only a few lines of code and presumably can be added straightforwardly to any primitive-
equation model. Note that unlike the simple illustrative case described in (3.7) and (3.8), the full age-tracer calculation yields three-dimensional maps of residence time at the same resolution as the model grid.

Tidally-averaged maps of age for three model scenarios representing an idealized seasonal cycle (runs E-G: Table 3.2) are shown in Fig. 3.9. The first of these (run E) contains a tidal circulation only, without density effects or river input, representing the sustained upwelling-favorable, low-riverflow conditions typical in late summer (Hickey et al. 2002, Banas et al. 2004). This run is identical to run D, discussed in Sec. 3b-d, except that the age tracer has been added. The next run (F) adds steady river input of 100 m$^3$ s$^{-1}$ and a constant ocean salinity of 30 psu: these forcings represent the fair-weather (upwelling-favorable) conditions typical in spring and early summer (Banas et al. 2004). In the third run (G), river input is 1000 m$^3$ s$^{-1}$ and ocean salinity 20 psu, representing sustained southward-wind, foul-weather, downwelling-favorable conditions and intrusion of the Columbia River Plume (Hickey and Banas 2003). In the following subsections we will examine the tidal residence-time pattern (run E) and then compare with the river-forced cases (F,G).

c. Age distribution under tidal forcing alone (run E)

The spatial distribution of ages in Fig. 3.9a encapsulates much of the information in the dye-intrusion (Fig. 3.3d), diffusivity (Fig. 3.3e), and crossing-region (Fig. 3.5) views of the tidal circulation discussed above. A tongue of low-age water entering through the southern side of the mouth corresponds to high, oceanic dye concentrations in Fig. 3.3d and the up-estuary-directed half of the lateral exchange flow shown in Fig. 3.5b-f. The compensating exit flow on the north side of the mouth (Figs. 3.3d, 3.5b,c) is also visible, above the 3-day age contour. Where the Lagrangian lateral-exchange flow breaks up in the middle of the estuary (Fig. 3.5g), $K$ has a local minimum (Fig. 3.3e), and there is a strong gradient, almost a front, in age (Fig. 3.9a). At the other $K$ minima as well—the tips of the Willapa Channel and Shoalwater Bay—the age gradient is intensified, and wherever $K$ is large—at the mouth, for example, the age gradient is
flatter. The age map thus contains two scales of circulation information. The age itself indicates the flushing-and-resupply rate for a given region on the whole-estuary scale—Shoalwater Bay, for example, is ~ 50 days "distant" from the mouth in a circulational sense (Fig. 3.9a)—while the gradient in age suggests dispersiveness or quiescence on the local scale.

There is an intense transverse gradient in age (up to 9 d km\(^{-1}\)) across the shallow banks at mid-estuary where the sections in Fig. 3.9d-f are taken. The simplest interpretation is that this gradient, just like the strong transverse temperature and salinity variations often seen in the same location (up to 5 °C, 4 psu: Hickey and Banas 2003), is caused by differential tidal advection (Huzzey and Brubaker 1988, O'Donnell 1993, Hickey and Banas 2003). In this process the channel-to-bank difference in the strength of flood-tide currents strains tracer fields, creating lateral gradients out of along-channel gradients. In other words, just as in summer water on Willapa's shallowest banks tends to be noticeably warmer and fresher—more like water 5-10 km up-estuary—than water just 1-2 km away in the main channel, so may that bank water be 2-3 weeks older, simply through rearrangement by the flood tide.

d. Age distribution under river forcing (runs F, G)

As river input increases (Fig. 3.9b,c), along-estuary variation in age decreases. (Note that the mid-estuary bank-to-channel gradient discussed above decreases proportionally, consistent with the differential-advection mechanism). The northern channel, where 81% of riverflow enters (Sec. 2b), is especially homogenized and quickly flushed: there is now a source of zero-age water at both ends, and even at moderate riverflows (Fig. 3.9b), average age is < 6 days everywhere between the mouth and the Willapa and North rivers. Shoalwater Bay, in contrast, which has no river input of its own, is largely bypassed by the river-driven increase in flushing rate. Even at storm-level riverflows (Fig. 3.9c), water at the southern end of Shoalwater Bay is 3 weeks older than water in the central bay.

In the absence of riverflow (Fig. 3.9d) the age distribution is vertically uniform,
but at high flows (Fig. 3.9f) water at depth in the main channel, far from the river mouths, is 1-2 d younger than surface water. This is a sign that a vertical gravitational circulation has developed, following the classical pattern (inward at depth, outward at the surface) first described by Pritchard (1952). Notably, the exchange pattern at flow levels near the long-term mean (Fig. 3.9b,e) more closely resembles the tide-driven pattern (Fig. 3.9a,d) than it does this high-flow, vertical exchange pattern (Fig. 3.9c,f). This is consistent with the results of Banas et al. (2004), who found that at moderate flows (100 m³ s⁻¹) the total effective diffusivity at Bay Center increases only 40% above its low-riverflow, tidal baseline value, while at very high flows (1000 m³ s⁻¹) the total diffusivity is 3.1 times the baseline value. For comparison, in these model results average age in the central estuary (from Bay Center to Oysterville: see Fig. 3.3e) is 30% higher at moderate flows (100 m³ s⁻¹) than under tidal forcing only and 2.4 times higher at high flows (1000 m³ s⁻¹) than under tidal forcing (Fig. 3.9a-c).

e. Intrusion of ocean water on the event scale (runs H, I)

The maps of average age discussed above suggest that at riverflow levels from the long-term median down—i.e., most of spring and summer—river- and ocean-forced density effects are secondary to the tides in setting Willapa Bay’s residence time, except in the relatively small volumes near the river sources themselves. It is possible, however, that on the event (2-10 d) scale, non-tidal influences have intermittent but sometimes much greater dynamical importance. Studies of estuarine adjustment (Kranenburg 1986, MacCready 1999) have shown that an estuary may respond to a pulse of river input very differently than to sustained river input. In addition, past work on ocean-estuary exchange on upwelling coasts (Duxbury 1979, Monteiro and Largier 1999, Hickey et al. 2002) suggests that the time-dependent response to changing ocean salinity may be an important modulation of the gravitational circulation.

Nevertheless, results from a final set of model runs (H-I: Table 3.2) suggest that the tide-driven residual circulation by itself is a good first-order approximation for a wide range of spring and summer conditions in Willapa, even on event timescales. Results
from these runs, which simulate the propagation of an individual 2 d pulse of ocean water into the bay, are shown in Fig. 3.11. The first of these runs (H) is identical to the idealized-tide run D described in Sec. 3, except that now the initial dye concentration is zero everywhere, and after the tidal flow is spun up dye is injected along the open ocean boundary for 2 d and then shut off. The resulting dye patch—which might be thought of as a marker for a mass of nutrient- and biomass-rich or -poor water produced by an individual upwelling or downwelling event—propagates into the estuary as shown in Fig. 3.10a. In the second of these runs (I: Fig. 3.10b), a constant background riverflow of 100 m$^3$ s$^{-1}$ has been added as in run F (Fig. 3.9b), and in addition, ocean salinity smoothly increases from 28 to 32 psu and back again during the 2 d that dye is injected. This combination of forcings represents a rapid spring upwelling event (Hickey and Banas 2003). With river and ocean-density forcing added, 28% more dye (i.e., new ocean water) enters initially (Fig. 3.10b, day 4) and the half-life of the dye in the estuary is 10% longer. When the duration of the dye release/upwelling event is extended to 8 d (not shown), the river- and ocean-forced modulation of the intrusion is even weaker: 10% more dye enters, and has a half-life only 5% longer, than in the tide-only case.

5. Discussion

In the pulse-propagation experiment just described, not only the horizontal spreading of an isolated ocean water mass, but also its steady migration up-estuary, is primarily controlled by tidal stirring. This may seem counterintuitive, given that tidal stirring is so often treated mathematically, as the label "stirring" implies, as a simple Fickian diffusion process (Hansen and Rattray 1965, MacCready 2004, Sec. 3b above). Indeed, throughout our analysis we have been, inevitably, shuttling between the language of "diffusion," "dispersion," and "stirring" on the one hand and "advection," "transport pathways," and "conveyor belts" on the other. In this section, we will try to reconcile these ideas, and show what we can infer about the actual mechanisms of tidal exchange in Willapa when all the components of our analysis are taken together.
a. Tidal flushing and tidal retention

A schematic of tidal exchange is shown in Fig. 3.11. Arrows indicate the Lagrangian transport pathways most visible in the crossing-region analysis from Sec. 3d. The most coherent feature is the strong lateral exchange flow in the seaward reach. The inflow curves around the southern and western sides of the entrance, crossing in and out of the curving main channel as shown (Figs. 3.5b-g, 3.11). The compensating outflow, through the deep channel to the north of the entrance, draws water from a broad area on both sides of the Stanley-Willapa junction, although some water from the central estuary disperses into the Willapa Channel (Fig. 3.6b) before entering this outflow, in a tangled, "diffusive" process. A similar interleaving of incoming and outgoing water masses occurs at the junction at the Shoalwater Bay entrance (Figs. 3.6a, 3.11).

Thus as we commented above (Sec. 3d), the along-channel profile of effective diffusivity $K$ (Fig. 3.3e) describes not just diffusion but also coherent, advective flow structures. Furthermore, on large scales these flows sometimes actually appear anti-diffusive, creating fronts and discontinuities instead of smoothing them away (Fig. 3.3d). The key to this phenomenon, in mathematical terms, is that a gradient in diffusivity acts on a tracer as a kind of "psuedoadvection." That is, the general tracer-diffusion equation

$$\frac{\partial s}{\partial t} = \frac{\partial}{\partial x}\left(K \frac{\partial s}{\partial x}\right)$$

(3.9)

(where $s$ is tracer concentration) can be rewritten

$$\frac{\partial s}{\partial t} + \left(-\frac{dK}{dx}\right)\left(\frac{\partial s}{\partial x}\right) = K \frac{\partial^2 s}{\partial x^2}$$

(3.10)

as if $K$ were locally constant and a background velocity $-dK/dx$ were advecting the tracer in the direction of decreasing diffusivity. This idea helps explain why any particular water mass tends to collect or stop advancing where $K$ is low: southern Shoalwater Bay, the head of the northern channel (Figs. 3.9a, 3.10a), and the mid-estuary minimum in $K$ (Figs. 3.3e, 3.5g; dotted line in Fig. 3.11). A local minimum in $K$—i.e. a region where the interaction of tides and bathymetry happens to generate few residual currents—acts as a semipermeable barrier to tidal exchange. A minimum in $K$ is a gap in subtidal transport, a
valley that can be crossed only slowly, and thus a place where relatively sharp breaks in water properties may form (Figs. 3.3d, 3.9a).

The potential significance of such a "dispersion gap" (and additional support for our model results) can be seen in a fact well-known in the Willapa Bay oyster-culture community, although not well documented in the scientific literature (Hedgpeth and Obrebski 1981; J. Ruesink and A. Trimble, pers. comm.) Natural settlement of oyster larvae, which requires retention in the water column for three weeks after spawning, occurs in the southern reaches of Willapa Bay but not in the northern. The dividing line lies north of the northern tip of Long Island, close to where we find the $K$ minimum and sharp gradient in age discussed above (Fig. 3.11). Just as new ocean water advances quickly up to this point in the bay but continues past it only weakly (Fig. 3.3d), so can water up-estuary of this point (carrying recently spawned oyster larvae, say) be trapped behind it. The online drifter-dispersion visualization described in Sec. 3e shows this directly. Our model thus reproduces a striking feature of Willapa's circulation: despite the fact that half the volume of the bay enters and leaves with every tide (Sec. 1), a typical water parcel in the upper third of the estuary is retained for several weeks or even months.

b. The importance of along-channel complexity

In the Lagrangian view, the high diffusivities and rapid exchange seaward of the mid-estuary dispersion gap appear to be the result of a single, not especially complex lateral-shear flow structure (Figs. 3.5, 3.11). The Eulerian view of the same area, however, is decidedly lacking in simple shear structures (Fig. 3.4): in fact, near the mouth the salt-flux constituent that dominates in equation (3.6) is term $(ii)$, the only term that does not involve lateral shear at all.

Dronkers and van de Kreeke (1986) present an interpretation of the salt-flux constituents in equation (3.6) that helps explain this discrepancy. They refer to term $(i)$ as the "local salt flux," since it involves only the structure of $u$ and $s$ within a cross-section. They refer to $(ii)$ and $(iii)$ together as the "nonlocal salt flux," since the temporal
variations in these terms \((s_1, dq_1)\) are really signs of spatial variation outside the cross-section being analyzed: i.e., variations over a tidal excursion in the along-channel direction. They then demonstrate that the nonlocal salt flux is approximately equal to the difference between the local flux through a stationary section and the local flux through a section moving with the flow. In other words, where terms \((ii)\) and \((iii)\) are large (in Willapa Bay, nearly everywhere: Fig. 3.4), it is definitional that the Eulerian circulation pattern does not reflect the underlying Lagrangian circulation, as we have found.

Furthermore, the fact that the nonlocal terms are non-negligible or dominant over most of the estuary (Fig. 3.4) can be taken to mean that the Lagrangian transport pathways we have described are driven at least as much by along-channel (nonlocal) bathymetric complexity as by transverse complexity. It is not surprising, then, that as discussed in Sec. 3d, Willapa's exchange circulation is not readily schematized as a lateral, channel-flow-versus-bank-flow pattern, despite the ubiquity of such patterns in the estuarine literature (Li and O'Donnell 1997, other references above). There can be no typical lateral pattern if the local salt flux \((i)\) is weak and changeable; in Willapa term \((i)\) changes sign and weakens over the 20 km in from the mouth (Fig. 3.4), and indeed the landward and seaward halves of the Lagrangian exchange flow migrate between bank and channel and dissipate over the same distance (Figs. 3.5, 3.11). If the circulation in estuaries like Willapa is to be schematized, we propose that one start not with lateral depth variation but along-estuary complexity: channel curvature, branching and junctions, and sets of ebb-dominant/flood-dominant channel pairs. To date, these common features of coastal-plain estuaries have received more attention in the morphodynamics literature (e.g., van Veen 1950, Hibma et al. 2003) than in the hydrodynamics literature.

6. Conclusions

This modeling study and a previous observational analysis (Banas et al. 2004) paint a consistent picture of the processes that control flushing in Willapa Bay. Under high-flow, winter conditions, the density-driven and tide-driven circulations are both
first-order contributors: neither can be neglected (Fig. 3.9c). During spring and summer, however, when river input is generally below the long-term average, tidal effects dominate, and baroclinic effects are discernable (Fig. 3.9e) but a second-order correction. This appears to be true on both seasonal timescales (Fig. 3.9) and the event scale (Fig. 3.10).

We did not, however, test a full suite of event-scale scenarios involving, for example, bursts of storm-level, northward winds accompanied by sharp pulses of riverflow and drops in ocean salinity (a set of forcings that tends to occur in concert on the Washington coast in winter and early spring: Hickey and Banas 2003, Banas et al. 2004). Examining a range of such scenarios in more detail may be important to understanding phenomena that unfold not over days or weeks but over individual tidal cycles: patterns of larval settlement in the seaward reach of the estuary, for example (Eggleston et al. 1998, Roegner et al. 2003). Before such processes near the mouth can be examined reliably in detail, however, better offshore data is needed for model forcing and validation.

Tidal dispersion has often been represented in the theory of estuarine circulation in particularly stripped-down, generic terms. In Willapa the tidal circulation is not only dynamically central but extremely intricate and variegated, diffusion-like in some places and conveyor-belt-like in others (Sec. 5). Our description of tidal dispersion in Willapa relies on two Lagrangian methods: the residual-circulation return map, built through particle tracking (Sec. 3d), and the steady-state age distribution, calculated using a nonconservative tracer (Sec. 4b). The age distribution, in particular, provides the most comprehensive picture of the flushing of the estuary of all the methods used in this study. It does not seem coincidental that it is also the only method that does not impose an artificial geometry (a preferred "along-channel" direction, straight-line cross-sections) on the flow. We recommend particle and tracer methods along these lines, and attention to along-estuary geometric intricacies—exactly those channel bends and flares and junctions that have been hardest to treat analytically—in the analysis of flushing and residence time in other complex systems.
Table 3.1. Overview of model runs and datasets used for model validation.

<table>
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<tr>
<th>run ID</th>
<th>validation runs</th>
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<th>B</th>
<th>C</th>
<th>D</th>
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<td></td>
<td></td>
<td>tidal velocity 1</td>
<td>tidal velocity 2</td>
<td>realistic year</td>
<td>mean tides</td>
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<td></td>
<td>(Sec. 3a, Fig. 3.2)</td>
<td>(Sec. 3a, Fig. 3.2)</td>
<td>(Sec. 4a, Figs. 3.7-3.8)</td>
<td>(Sec. 3b, Fig. 3.3)</td>
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<tr>
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<td>density?</td>
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<td>off</td>
<td>on</td>
<td>off</td>
<td></td>
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</tbody>
</table>

**model forcing**

| sea level | observations from Toke Point (NOAA station 9440910) | observations from Toke Point | observations from Toke Point | pure M2 tide (±1.2 m) |
| river input | — | — | observations from USGS gauges 12013500, 12010000, adjusted for watershed area | — |
| ocean tracers | — | — | salinity at bay mouth (station W3: Banas et al. 2004) | constant dye concentration (initial dye field in estuary: Fig. 3.3a) |

**model validation**

| criterion | M2 constituent of depth-averaged tidal velocity | instantaneous, cross-sectionally-averaged tidal velocity | salt intrusion length | stratification at Bay Center | effective horizontal diffusivity $K$ |
| data source | US Army Corps of Engineers Seattle District: Kraus 2000 | (previously unreported) | Banas et al. 2004 (previously unreported; courtesy of J. Newton/ Wash. Dept. of Ecology) | Banas et al. 2004 (W3, Bay Center, Oysterville; Shoalwater data previously unreported, courtesy of J. Ruesink and A. Trimble) |
Table 3.2. Overview of idealized model scenarios used to examine residence time and exchange patterns.

<table>
<thead>
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<th>idealized, pulsed ocean forcing</th>
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<td></td>
<td>E</td>
<td>F</td>
</tr>
<tr>
<td></td>
<td>LATE SUMMER</td>
<td>SPRING</td>
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<td></td>
<td>(Sec. 4b, Fig. 3.9)</td>
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<tr>
<td>simulates</td>
<td>sustained low riverflow + upwelling</td>
<td>sustained moderate riverflow + upwelling</td>
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<td>tracers</td>
<td>age</td>
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<td>density?</td>
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<td>sea level</td>
<td>—</td>
<td>all runs: pure M2 tide (+1.2 m)</td>
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<td>river input</td>
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<tr>
<td></td>
<td>salinity = 30 psu</td>
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(also repeated with 8 d pulses: Sec. 4e)
Figure 3.11. Schematic of the tide-driven residual circulation, based on Figs. 3.3, 3.5, and 3.9a. Arrows indicate the direction and magnitude of the residual flows. The model grid expands offshore, and the closed boundary is shown.

Figure 3.1. Map of Willapa Bay. The offshore boundaries of the model domain and samples of the model grid are also shown.
Figure 3.2. Results from runs A and B (Table 3.1, Sec. 3a) compared with observations. Depth-averaged M2 tidal velocities from model and data are shown at four locations (three near the mouth, one near 46°30'). Modeled and observed tidal height and the north-south component of instantaneous, cross-sectional-mean tidal velocity are shown for one flood tide on May 5, 2000.
Figure 3.3. (a-d) Evolution of an initially uniform tracer gradient under idealized tidal flow (run D). (e) Total effective horizontal diffusivity $K$, calculated for this run from tracer fluxes across 12 sections (green lines); also, empirical values of $K$ calculated from salinity time series at four stations (blue dots) using the method of Banas et al. (2004). (f) An example of the empirical $K$ calculation (adapted from Banas et al. 2004).
Figure 3.4. Contributions to the total diffusivity $K$ associated with terms (i) (steady shear), (ii) (trapping), and (iii) (a cross-term) in the salt-flux decomposition (3.6). Fluxes are calculated from model run D at the 12 cross-sections shown in the inset and Fig. 3.3e.
Figure 5.5. (a) Schematic of the method for constructing Lagrangian "crossing regions" from numerical particle tracking as described in Sec. 3c. Initial particle positions are shown in light shades, positions one tidal cycle later as dark points; for a given cross-section (green line), particles crossing seaward are shown in blue, particles crossing landward in red; particles that begin and end on the same side of the cross-section are not plotted. (b-g) Crossing regions for six sections under idealized tidal flow (run D: Table 3.1). (h) Crossing regions for the section shown in (c) when run D is repeated with the Coriolis acceleration turned off.
Figure 3.6. Crossing regions, like those in Fig. 3.5b-g, for two sections at channel junctions.
Figure 3.7. Salinity at three stations (Bay Center, Oysterville, Naselle) relative to salinity at the mouth (W3), as a function of bay-total riverflow. Results are given for both model run C and observations Jul 1999 – Jun 2000 (see Table 3.1).
Figure 3.11. Schematic of the tide-driven residual circulation, based on Figs. 3.3, 3.5, and 3.9a. Arrows indicate the direction of flow.

Figure 3.8. Time series of stratification at Bay Center and bay-total riverflow, from (a) observations in 2003-04 (see Table 3.1) and (b) model run C. Sustained stratification events are marked by arrows. (c,d) Maximum and minimum stratification during each tidal cycle shown in (a) and (b), displayed as a function of riverflow.
Figure 3.9. Maps of water age (average time that the water in each grid cell has spent within the estuary) under three constant-forcing scenarios (runs E-G: Table 3.2). Riverflow $Q$ and ocean salinity $s_{oc}$ are given for each case where they are applied. (a-c) Depth-averaged age; arrows mark river inputs. (d-f) Vertical sections of age at mid-estuary.
Figure 3.10. Evolution of pulses of ocean water marked by a two-day dye release in the model. Each frame shows depth-averaged dye concentration relative to the initial source concentration. Two scenarios are shown: propagation under idealized tides only, without density effects (run H), and propagation when riverflow of 100 m$^3$ s$^{-1}$ is added and ocean salinity temporarily increases from 28 to 32 psu during the dye release (run I; see Table 3.2). The first case represents typical late-summer conditions, the second a typical spring upwelling event.
Figure 3.11. Schematic of the tide-driven residual circulation, based on Figs. 3.3, 3.5, and 3.9a. Arrows indicate Lagrangian transport pathways; note that point-by-point, Eulerian residual velocities are generally substantially different. A dotted line marks the water-mass-isolating “dispersion gap” discussed in Sec. 3.5a.