The Classification of Estuarine Systems

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Abstract

Exchange flows through straits connecting oceans and basins, such as estuaries, have important implications for water quality. Therefore understanding the basic hydrodynamics in the system is important. This study establishes an estuarine classification system based on the dominant hydraulic forcing. The extent of dominance of the buoyancy, viscous, advective, diffusive and tidal forcing is considered on a seasonal basis, using the non-dimensional Grashof and tidal parameters, $Gr_t$ and $Et$. This study also provides a comparison of exchange flow calculations based on two simple models. The first model uses field data to calculate the residence time in the basin and the second model uses external forcing parameters.

The hydrodynamic classification system was successfully applied to four major West Australian estuaries, Wilson Inlet, the Swan River, Shark Bay and the Peel-Harvey Estuary. The dynamic regimes of the estuaries alter on a seasonal basis and there is significant variation between each system. The strength of the barotropic forcing is linked to the magnitude of the topographic constriction at the entrance of the basin and the seasonal basin properties. The relative dominance of buoyant and diffusive forcing also shows seasonal variation, the strength of internal mixing in the strait can alter within an estuarine system between hydraulically limited exchange flow to viscous, advective, diffusive forced flow. The exchange flow models were applied to the estuaries, the results show that some inaccuracy occurs in the models, yet the residence time applications show reasonable values, that are useable in terms of understanding the exchange flow magnitude and changes in an estuary.
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1.0 Introduction

Water quality is an area of increasing importance as awareness of the impacts of anthropogenic activity and resulting environmental degradation grows. Exchange flow through straits connecting enclosed basins and the ocean determine the residence time of water in the basin, which strongly influences water quality and the response of the ecosystem to external pressures. The motivation for this study is to consider the application of recent developments to field sites and examine the parameters governing exchange flow. Currently there are two limiting theories, hydraulic and diffusive limited flow. Hydraulic theory assumes that no internal mixing occurs in the channel, diffusive limited flow occurs at the opposite end of the spectrum whereby strong internal mixing occurs in the channel and the exchange flow is heavily dampened by the effects of mixing. This study considers whether exchange flow can be reasonably accurately estimated using simple external parameters, compared to field measurements. As in circumstances when obtaining field data is unviable or unavailable knowledge the use of external parameters to determine the exchange flow reasonably accurately is important. Using two simple models, one that uses field data and incorporates the extent of internal mixing, and the second, which operates using external parameters and hydraulic theory assumptions, the sensitivity of the exchange flow to the parameters is determined.

This study also aims to provide an estuarine classification system based on the dominant hydrodynamic forcing of exchange flow, based on the parameters $G_{rt}$, $A$ and $E_t$. $G_{rt}$ is the non-dimensional Grashof number, which is a ratio of the relative effects of viscous diffusive forcing to buoyant hydraulic forcing in the system. $A$ is the aspect ratio, it represents the topographic control of the exchange flow. The parameter $E_t$ is a non-dimensional tidal parameter that defines the strength of buoyancy forces compared to tidal forces. Most estuarine systems flow regimes alter seasonally, based on the relative magnitudes of the buoyancy, tidal and turbulent viscous forcing – therefore using the parameters $G_{rt}A^2$ and $E_t$ estuarine systems can be seasonally defined as having flow regimes closely linked to tidal, hydraulically or diffusively forced exchange flows. The hydrodynamic classification and the two models are important in identifying the major processes and time scales of the estuary or basin, from which the focus of more intensive and complex modelling can be determined.
2.0 Literature Review

2.1 Exchange Flows

Exchange flows are single, or more commonly bi-directional flow through a strait connecting two water bodies of differing fluid properties. Due to tidal influences the exchange flow can alter between single layered flow and bi-directional flow, introducing an unsteady element into the flow. Buoyancy fluxes in the connected reservoirs often create density differences between the two reservoirs or basins and thus provide the gravitational forcing for the flow, while topographic constrictions are a controlling parameter on the effectiveness of the flow (Armi and Farmer, 1986, Garret et al 1990, Bray et al 1995). The buoyancy fluxes in the basins can be caused by evaporation, precipitation, freshwater inflow and freezing, creating the density difference. The basin dynamics also impact on the strength of the exchange flow, as mixing influences the density gradients and therefore local flow forcing in the strait (Ivey, 2003).

Two-way exchange flows are found in many geographical situations and on many scales. Ranging from exchange between ocean bodies, such as between the Mediterranean Sea and the Atlantic Ocean through the strait of Gibraltar – to small scale exchange between rivers and the ocean, such as the Swan River and Wilson Inlet, where fresher inlet/river water exits into the ocean, being replaced by a denser salty ocean water that flows into the fresh water body as a bottom layer. Exchange flows are not only found in surface systems, but also are important in deep ocean situations; such as the abyssal exchange flow between the Antarctic bottom water and Pacific waters (Whitehead, 1998). In fresh water systems water quality is particularly partially dependant on the residence time of the water in the system and exchange flows determine this residence time. As such understanding of the residence time is a key aspect of management of water bodies.

2.2 Equations Governing exchange flows

Exchange flows were initially studied as an internal hydraulic problem, Prandtl (1952) considered single layered flows beneath or above a passive layer. This work was continued by Turner (1973) and Stommel and Farmer (1953) who examined two-layered flow where both layers interact to establish exchange flow control. Armi (1986) and Armi and Farmer (1986) provided the basis for dynamic theory of exchange flows, through their use of steady state internal hydraulic theory, defining the flow through the Froude number plane and considering the topographic control of the flow. The internal hydraulic theory was corrected for weakly

2.2.1 Basic fluid flow equations

The Navier Stokes (conservation of momentum) and continuity (conservation of mass) equations describe the motion of a stratified fluid. Using an inertial time frame Baines (1995) defined these equations as:

**Navier Stokes:**
\[
\frac{Du}{Dt} = -g\hat{z} - \frac{1}{\rho} \nabla p + \nu \nabla^2 \mathbf{u}
\]  
2.2.1

**Continuity:**
\[
\frac{1}{\rho} \frac{D\rho}{Dt} + \nabla \cdot \mathbf{u} = 0
\]  
2.2.2

Where \( \mathbf{u} = (u,v,w) \) is the fluid velocity in three dimensions, \( \rho \) is the fluid density, \( p \) is pressure, \( g \) is the gravitational force, \( \hat{z} (0,0,1) \) is the unit vector vertically upwards, \( D/Dt \) is the Langranian derivative operator with respect to time and \( \nabla \) is the Laplace derivative operator \((d/dx, d/dy, d/dz)\). The Navier Stokes equation states that the acceleration of a fluid element is equal to the balance of gravity, pressure and friction forces.

A fluid is incompressible if the density of every particle remains constant when it moves, even when subject to variations in pressure. The conservation of mass equation is a statement whereby the volume of fluid entering a control volume equals the volume of fluid leaving the control volume, for an incompressible fluid, as the temporal density gradient term in equation 2.2.2 is equal to zero reducing the equation to:

\[
\nabla \cdot \mathbf{u} = 0
\]  
2.2.3

Internal hydraulic theory considers hydrostatic pressure systems, therefore pressure and density fields are related by equation:

\[
\frac{dp}{dz} = -\rho g
\]  
2.2.4

The Bernoulli equation describes the total energy density per unit volume of a fluid parcel travelling on a streamline, this is the integrated equation of motion for an inviscid
homogeneous fluid is the Bernoulli equation (Eagleson, 1966). The energy associated with open channel fluid motion can be considered the sum of its elevation, pressure and velocity heads.

**Bernoulli Equation:**

\[ E = H = z + y + \frac{v^2}{2g} \]  

Where \( z \) is the elevation of the streamline above an arbitrary datum, \( y \) is the flow depth or pressure head, \( v \) is the fluid velocity along the streamline and \( \frac{v^2}{2g} \) is the velocity head.

### 2.2.2 Steady State Two-Layer Hydraulic Theory

Internal hydraulic theory extends the ideas from single layer hydraulics and can be used to determine flow rates between reservoirs of differing density connected by a channel of slowly varying height and width. When considering two-way exchange flow through a contraction \( (h_s(x)) \) or over a sill \( (b(x)) \), with a free surface the following convention is adopted. It is assumed there are two distinct layers with constant velocity and density and specific layer depths. The top, less dense layer is described with the subscript 1 and moves from left to right: the density is \( \rho_i \) and the layer moves with velocity \( u_i \) and the layer thickness equal to \( y_i \).

The schematic is illustrated below in Figure 2.2.1. Internal hydraulic theory assumes the fluid is inviscid, incompressible and flows through a strait with gradually changing height and width. The variations in the channel width are small enough that vertical velocity, friction and mixing effects can be ignored (Lawerence, 1990).

![Figure 2.2.1](image-url)

*Figure 2.2.1 (a) the schematic shows the side view of two-way exchange flow over a sill for sill depth \( h_s \) and layer depth \( y_i \). (b) The plan view for flow through a contraction.*
Armi (1986) summarised the steady momentum and continuity equations presenting them in the following form:

\[ C v_x = D f_x \quad (2.2.6) \]

Where \( C \) and \( D \) are the following matrices and \( v_x \) and \( f_x \) are the following vectors:

\[
\begin{bmatrix}
  u_1 & 0 & g & g \\
  0 & u_2 & r g & g \\
  y_1 & 0 & u_1 & 0 \\
  0 & y_2 & 0 & u_2 \\
\end{bmatrix}, \quad
\begin{bmatrix}
  -g & 0 \\
  -g & 0 \\
  0 & q_i \\
  0 & q_2 \\
\end{bmatrix}, \quad
\begin{bmatrix}
  u_1 \\
  u_2 \\
  y_1 \\
  y_2 \\
\end{bmatrix}, \quad
\begin{bmatrix}
  v_x \\
  h \\
\end{bmatrix}
\]

Where \( r = \frac{-1}{\_ \_} \). Subscript 1 refers to the top layer, subscript 2 refers to the bottom layer. \( y_i \) refers to the depth of the respective layers and \( u_i \) refers to the velocity, \( q_i \) to the flow rate. \( g \) is the gravitational acceleration, \( h \) the depth of the water column at the topographic constriction and \( b \) is the minimum width at the constriction. As a consequence of the assumption of hydrostatic pressure, equation (2.2.6) has only long wave solutions. The celerities of the long waves are specified by the eigenvalue equation (Lawerence, 1990):

\[ \text{Det}(C - I) = 0. \quad (2.2.8) \]

Where \( I \) is the identity matrix. The characteristic velocities are given by

\[ \lambda = \pm \frac{u_{\text{con}}}{c} \quad (2.2.9) \]

The ratio of the fluid velocity \( (u_{\text{con}}) \) to the phase speed \( (c) \), is traditionally known as the Froude number. Where \( u_{\text{con}} = \frac{u}{h} \) the velocity of the water, \( h \) is the height of the water, \( g \) is the gravitational force and \( c = (gh)^{0.5} \)

\[ F = \frac{u}{(gh)^{0.5}} \quad (2.2.10) \]

### 2.2.3 Froude Number Plane and Hydraulic Controls

Hydraulic control is the term used to describe the transition of flow from supercritical to subcritical, in a slowly varying channel critical flow is always established at the point of hydraulic controls (Lawerence, 1990). The Froude number is used to classify the conditions of the flow. In open channel flow where the Froude number is equal to unity the flow is unique and critical. For a two-layered flow Armi (1986) extended this idea and parameterised flow conditions in terms of the internal Froude numbers of each layer, the possible flow solutions are then shown using curves in the Froude-number plane. Equation (2.2.6) is quasi-linear, critical flow occurs if the \( \text{det}(C) = 0 \), and the dependant flow variables \( v_1, v_2, y_1 \) and \( y_2 \) can be solved as functions of the topographic variables \( b \) and \( h_s \). Maximal two-way exchange
requires a topographic and a virtual control point, the topographic control point occurs at the narrowest section of the channel. For a two-layer flow with a free surface the composite Froude number $G^2$ is defined by (Armi, 1986):

$$G^2 = F^2_1 + F^2_2 - (1-r) F^2_1 F^2_2$$  \hspace{1cm} (2.2.11)

Where $F^2_1$ is the internal Froude numbers of the individual layers from equation 2.2.9, $g' = (1-r)g$ is the reduced gravitational acceleration and $r = \frac{\rho_1}{\rho_2}$

The composite Froude number equals unity whenever the flow is critical. This can be simplified to equation 2.11 whenever the external Froude numbers and the non-dimensional density differences are small, $(1-r)<<1$.

$$G^2 = F^2_1 + F^2_2$$  \hspace{1cm} (2.2.12)

This critical condition collapses to a straight line, separating internally supercritical from subcritical flow (Armi, 1986). The flow rate is specified as $q_i = u_i y_i$ for each layer (Armi, 1986). Adopting Armi’s (1986) convention the non-dimensional volume flow rates $q_i$ are:

$$q'_i = \frac{q_i}{g^{0.5} b_o (y_1 + y_2)_0^{1.5}}$$  \hspace{1cm} (2.2.13)

Where $(y_1 + y_2)_0$ is the depth of layer 1 and 2 and $b$ is the channel breadth. The non-dimensional width, height and layer depths can be defined by:

$$b' = \frac{b}{b_o} \quad h' = \frac{h}{(y_1 + y_2)_0} \quad y'_i = \frac{y_i}{(y_1 + y_2)_0}$$  \hspace{1cm} (2.2.14)

Where $b'$ equals zero at the minimum width at the contraction. Armi (1986) and Lawerence (1990) fixed the volume flow ratio $q_r$ along the channel, therefore the location of all possible Froude number pairs for any volume flow rate, width and total depth of flow can be calculated using the continuity equation and assuming a level-free surface:

$$q_1^2 F_1^{-2} + q_2^2 F_2^{-2} = \frac{q'_2}{b'(1-h')^{1.5}}$$  \hspace{1cm} (2.2.15)

Where $q_r = q_1/q_2$, the ratio of flow rates in each layer. Equation 2.2.15 defines the possible Froude number pairs for any fixed ratio $q_r$ and $[q_r'/b'(1-h')^{3/2}]$. For a specified ratio of volume flow rates the steady state maximal two-layer exchange flow through a contraction can be found from the Froude-number plane, using the dimensionless Bernoulli equations and the continuity equation. Below we consider a number of flow configurations.
2.2.4 Flow through a contraction

Armi (1986) comprehensively studied and derived flow equations for moderate, steady exchange flow through a contraction. Flow through a contraction only encounters width variations, both the bottom and top surfaces remain horizontal as described by the following Bernoulli equations:

\[ H_1 = \frac{1}{2} \rho_1 u_1^2 + \rho_1 g(y_1 + y_2) + p \]  \hspace{1cm} 2.2.16

\[ H_2 = \frac{1}{2} \rho_2 u_2^2 + \rho_2 g y_1 + \rho_1 g y_1 + \rho_2 g y_2 + p \]  \hspace{1cm} 2.2.17

Where \( p \) is the pressure at the free surface, by definition equal to zero. The dominant hydrostatic pressure effects, not associated with the internal hydraulics can be removed through subtracting equation 2.2.17 from 2.2.16. Resulting in a conservative energy difference, provided there are no hydraulic jumps (Armi 1986). Non-dimensionalising by \( g' \rho_2 (y_1 + y_2) \) gives:

\[ \frac{H_2 - H_1}{g' \rho_2 (y_1 + y_2)} = \Delta H' = \frac{F_2^{-2} (1 + \frac{1}{2} F_2^2) - \frac{1}{2} q_r^{-2} F_1^{-2} F_2^{-2}}{q_r^{-2} F_1^{-2} + F_2^{-2}} \]  \hspace{1cm} 2.2.18

Farmer and Armi (1986) obtained solutions to the dimensionless Bernoulli equation 2.2.16, for several values of \( \Delta H' \) together with the continuity equation, below 2.2.19, for several volume flow rates per unit width.

\[ \frac{q_r^{1/2}}{b'} = q_r^{-1} [1 + q_r^{-1/2}]^{-2} \]  \hspace{1cm} 2.2.19

The following diagram, figure 2.2.2 illustrates solution curves for \( q_r=1 \), where the solid line is the solution to Bernoulli equation, the dashed lines are volume flow rate per unit width and the shade represents critical flow. The x-axis represents the internal Froude number of layer 1, the y-axis the internal Froude number of layer 2. This diagram illustrates the exchange flow that occurs for a particular layer depth and Froude number.
Figure 2.2.2 (a) Solution curves for the volume flow rate $q_r=1$, the solid lines are the solutions to the Bernoulli equation, dashed lines or fine lines are volume flow rate per unit width and the shade and straight line represent critical flow.

The theoretical flow explored by Armi and Farmer (1986) considered the inflowing layer to spread out in each reservoir and have a negligible thickness, therefore $F_1^2 = F_2^2 \to \infty$ in the reservoirs. Conditions for maximal exchange flow must satisfy the energy difference equation 2.2.15, for $q_r=1$, at the narrowest section $\frac{q'_2}{b'} = \frac{q'_1}{b'} = \frac{1}{4}$ and $F_1^2 = F_2^2 = 0.5$, as evident in
figure 2.2.2. The flow is just critical and the two layers are of equal thickness, and for maximal two-layer exchange supercritical flow must occur on either side of the control (Armi, 1986). For volume flow rates $q_2'/b'<0.25$ the flow rate locus intersects the critical condition $G^2=1$ twice. Therefore there are two possible controlled solutions through a contraction for each choice of flow rate. If $q_2'/b>0.25$ there are no intersections between the critical flow line and the line of possible internal Froude number pairs; therefore all flow is internally supercritical and the control must occur upstream at a wider section of the contraction, where $db/dx≠0$; the control is then termed a virtual control (Wood, 1968, Lawerence 1990). The importance of this result is that if the flow is controlled the volume flux through the contraction can then be calculated from a few parameters. Considering this two-layered contraction problem at the narrowest point of a contraction the two layers are of equal thickness and $q_r=1$, the maximal volume and mass flux in each layer are given by (Armi and Farmer, 1986, Lawerence, 1990, Hogg et al, 2001):

$$Q_h = \frac{1}{4} g^{\frac{1}{2}} bh^{\frac{3}{2}} \quad M_h = \frac{1}{4} \Delta \rho g^{\frac{1}{2}} h^{\frac{3}{2}} \quad 2.2.20$$

### 2.1.5 Flow over a sill

Armi (1986) assumed flow over a sill had constant width and the variation was only considered in the bottom height ($h$). The Bernoulli equations for each layer are:

$$H_1 = \frac{1}{2} \rho_1 u_1^2 + \rho_1 g(y_1 + y_2 + h) + p \quad 2.2.21$$

$$H_2 = \frac{1}{2} \rho_2 u_2^2 + \rho_2 g(y_1 + h) + p \quad 2.2.22$$

Again the dominant hydrostatic pressure force can be removed through subtraction of equation 2.2.10 from 2.2.21. Assuming the non-dimensional density difference is small and the surface condition is free and level, the non-dimensional energy difference is then given as:

$$\frac{H_1 - H_2}{g' \rho_2(y_1 + y_2)_0} + 1 = y'_1 (1 + \frac{1}{2} F_1^2) - \frac{1}{2} y'_2 F_2^2 \quad 2.2.23$$

Assuming that $q_i$ is independent of the position along the channel the continuity equation in Froude number space can be expressed as:

$$q_i^2 F_1^{-2} + F_2^{-2} = \frac{q_i^2}{(1-h)^{1.5}} \quad 2.2.24$$

Farmer and Armi (1986) expressed the solutions for a particular $q_r$ contours of $q_2^2/(1-h)^{3/2}$ as functions of $F_1^2$ and $F_2^2$ along with the dimensionless Bernoulli equation to compute maximum exchange flows. Flow, which is internally subcritical, can only have the bottom
layer accelerated during flow over a sill. The critical flow condition occurs only at the highest point on the sill. High flow rates or high bumps, represented by the parameter \( q'/(1-h)^{3/2} \) being sufficiently large then the flow is everywhere internally supercritical. For two layer flow at constant width \( db/dx=0 \) the flow is critical at

\[
G^2 = F_1^2 + F_2^2 - (1 - r)F_1^2F_2^2 = 1
\]

The free surface is controlled at the crest where \( G^2=1 \), at the location of the control of the interface upstream of the crest regularity conditions where \( dh/dx \neq 0 \) can be satisfied only if \( F_1^2 = 1, F_2^2 = 0 \). Armi and Farmer (1986) found that sub-maximal exchanges occur over a sill and through a constriction and for that case supercritical flow occurs on only one side of the control.

2.1.6 Flow through a combination of a sill and contraction

In many straits there is either a coincident sill and contraction or separated sills and contractions. The two way hydraulically controlled exchanges over a sill differs significantly from the problem of flow through a contraction. The control at the sill crest acts primarily through the deeper layer into which it protrudes and only indirectly controls the surface layer. Farmer and Armi (1986) described flow solutions to the problem of a displaced sill and contraction, quantifying the exchange flux for the topography of the Gibraltar strait. However no attempts have been made to quantify the exchange through a coincident sill and contraction as the regularity condition becomes complicated to solve. Farmer and Armi (1986) identified possible topographic controls for a strait with displaced sill and contraction.

1. If the contraction is located to the left of the sill it will not influence the two-way exchange at the sill. This is for the idealised situation of a deep reservoir on either side of the sill – where the contraction only influences the surface layer.
2. If the contraction is to the left of the sill, in a strait not much deeper than the sill the control at the sill crest can be flooded- therefore the flow is solely influenced by the contraction.

2.1.7 The Effect of Basin Dynamics

Finnigan and Ivey (1999,2000) proposed a simple way of connecting the hydraulically controlled flow through a strait to the large-scale dynamics of the basins at each end of the strait. Basin scale dynamics and exchange flow are closely related- once steady flow conditions are established the exchange flow is controlled by the forcing, geometric
parameters and mixing conditions within the basin, which in term are dependant on the exchange flow (Finnigan & Ivey, 1999). Initially there is an unsteady response to the basin forcing, but with continued forcing the system reaches steady state with a surface current and deep return layer. The extent of mixing in the basin is reflected through the buoyancy difference, which is established between the layers at the open end of the basin (Finnigan & Ivey, 1999). The exchange over a sill is increased by increased interior mixing, which decreases the buoyancy difference between layers at a sill. Finnigan and Ivey (1999) used the following method in analysing the connection between basin dynamics and exchange flow. The mean buoyancy in a basin is calculated as:

\[ b = \frac{\rho_o - \rho}{\rho_o}g \tag{2.2.26} \]

Finnigan and Ivey (2000) computed the surface buoyancy flux as:

\[ B_o = \frac{\alpha g H}{\rho C_p A} \tag{2.2.27} \]

Where \( A \) is the surface area, \( \alpha \) is the thermal expansion coefficient, \( C_p \) is the specific heat, \( \rho \) is the density and \( H \) is the averaged heat flux, and \( B_o \) is the buoyancy flux. Finnigan and Ivey (1999) described three forcing regimes for buoyancy forced flow over a sill, regime one occurs a short time after the forcing is initiated and the basin mixes over the entire depth by vertical turbulent convection. No buoyancy is exchanged across the sill, meaning there is no exchange flow. Regime two is also unsteady, initialised when a lateral exchange is established. The lateral surface current moves into the basin, loses its buoyancy and then departs from the surface, eventually flowing back out of the basin. The flow then reaches a steady state, described by regime three, the buoyancy in the underflow is determined by the buoyancy deficit acquired in the surface current as it traverses the length of the basin.
Figure 2.2.3 The effect of the buoyancy flux on the reservoir for each regime.

Since the entire flow is considered driven by the forcing in the basin the resulting velocity in each layer across the sill crest and buoyancy difference can be expressed as:

$$u_1 = \beta (B_o L)^\frac{1}{3} \frac{h_1}{h_2}$$

$$b = \gamma \left( \frac{B_o L}{h_1} \right)^2$$

Where $\beta$ and $\gamma$ are undetermined coefficients dependent upon the turbulence field within the basin as well as $B_o$, $L$ and $h_i$; which are the length ($L$) of the basin and height ($h_i$) of each layer (Finnigan and Ivey, 1999). Combining these two equations and considering the flow parameters are related by the critical condition at the sill rest leads to;
\[
\frac{\beta}{\gamma} = \frac{h_1^3}{h_2^3 + h_1^3}
\]  

This can be rearranged when considering \( h = h_1 + h_2 \) at the sill and the buoyancy conservation condition is \( B_0L = u_1h_1b \). Resulting in a statement, whereby the buoyancy flux and the undetermined coefficients are functions of the interface level at the sill crest, which are generally unknown (Finnigan and Ivey, 1999):

\[
B_0 = \frac{(h - h_1)^3}{(h - h_1)^3 + h_1^3} = \gamma^{-1}
\]  

Fluid in the upper layer (subscript 1) is stirred by convective turbulence, the middle layer is essentially (Finnigan and Ivey, 2000). The buoyancy scaling; \( g' \sim (B_0L)^{2/3}/h \), infers that \( g' \) at the sill increases with \( B_0 \). Turbulent convection due to the destabilizing effect of the surface buoyancy flux sustains a mixed region in the inflowing upper layer. At the sill Finnigan and Ivey (2000) rearranged the volume flux in each layer at the sill, using \( q = u_1h_1 = u_2h_2 \) and the buoyancy conservation condition:

\[
q = \left[ B_0L \frac{1}{h_1} + \frac{1}{(h - h_1)^{-1}} \right]^{1/3}
\]  

This solution is not closed, as \( H_1 \) is generally unknown. Using the relationship between the mixed layer depth (refer to equation 2.3.11) and flow rate determined by Hogg et al (2001), and considering a system of a semi-enclosed basin and strait system at steady state, where the strait is a simple contraction connected to an enclosed basin of dimensions \( B \) and \( L \). The dimensional flux can be written as:

\[
Q = 0.4 \left( 1 - \frac{1}{2} \left( \frac{\delta}{h} \right)^2 \right) \left( B_0b^2BL \right)^{1/3}
\]  

Which for the hydraulic limit simplifies to the following equation as the mixed layer equals zero.

\[
Q = 0.4h \left( B_0b^2BL \right)^{1/3}
\]  

### 2.3 Exchange flows with significant mixing effects

Winters and Seim (2000) investigated large-scale exchange flows and shear instabilities that form at the interface, causing turbulence. They found that inter-layer transport lead to flows with lower Froude numbers, larger transports and wider regions of subcritical flow in the contraction. The flow was best represented as a three-layer decomposition, to demonstrate the effects of mixing and entrainment. The effects of friction act to shift the site of the
topographic control downstream and the virtual control upstream (Pratt, 1986). The dominant effect of interfacial friction for maximal exchange simulations increases transport through this shift in control points, rather than from accelerated flow speeds (Winters and Seim, 2000). By altering the bottom boundary condition from free to non-slip Winters and Seim (2000) found that instabilities were created, associated with propagating bores, for steady state flow the flow resembled sub-maximal exchange flows described by Armi and Farmer (1986).

In hydraulic theory the pressure is considered hydrostatic and the channels have constant width and depth. The horizontal velocity is considered to be uniform with depth within each layer, this is only valid when the streamline curvature is small. Zhu and Lawerence (1998) modelled the curvature effects on a two-layered flow over a two-dimensional smooth sill in a channel of constant width and multi-layered flow in shallow water, using the Approach-controlled flow. The point of control is located at the foot of the obstacle- between the point of control and the obstacle the flow is similar to that predicted by hydraulic theory. Downstream from the crest the interface level drops and the flow changes from supercritical, where the upper layer changes from thin to thick (Zhu and Lawerence, 1998). Approach controlled flows are characterized by this change, called a supercritical leap (Lawerence, 1993). The supercritical leap involves streamline curvature and has subsequent non-hydrostatic effects that are not accounted for in hydraulic theory.

2.3.1 Turbulent Dominated Exchange Flow

Internal hydraulic theory excludes entrainment and mixing and their effect on the flux between two reservoirs. Winters and Seim (2000) showed that an interfacial layer of intermediate density formed by vertical entrainment and mixing can play an important role in horizontal exchange and carries mixed fluid away from a contraction in both directions. They found that internal mixing rates are greatest where the density gradient and eddy diffusivity are large. The density conservation equation:

\[ \frac{d\rho}{dt} + \mathbf{u} \cdot \nabla \rho = \nabla \cdot K \nabla \rho \]  \hspace{1cm} \text{2.3.1} \\

where diffusivity is \( K \), indicating that when the flow is steady streamlines will cross isopycnals where the diffusivity and density gradients are locally enhanced. The three layer analysis Winters and Seim (2000) undertook emphasized the inclusion of viscosity and diffusion leading to a change in the character of the solution. Bray et al (1995) emphasized the importance of the interface in carrying horizontal transport in their analysis of observations from the Strait of Gibraltar. Winters and Seim found that mixing in the vicinity
of hydraulic controls is of importance for the overall circulation and exchange between basins connected by straits or channels and suggested the overall flux can be easily reduced by 15% or more below the hydraulic predictions.

2.3.2 Viscous Advective Diffusive Limit

In flows with strong mixing between flowing layers, the horizontal velocity may be controlled by the turbulent eddy viscosity and diffusivity rather than the speed of internal waves (Hogg et al 2001). Therefore the role of the contraction is altered- experimentally it is found that the estimate of the flux can be reasonably made using a straight edged channel, with width equal to the minimum width of the contraction.

Officer (1976) and Hogg et al (2001) determined flux through a contraction in the VAD limit assuming the aspect ratio is small and maintaining density contrast at the vertical end walls. The aspect ratio is the ratio of the height in the channel to the length of the mixed region. Therefore for the VAD limit the velocities can be considered horizontal and constant in the x direction and the pressure is assumed hydrostatic.

The equations of motion for steady flow are then (Hogg et al 2001):

\[
\frac{dP}{dx} = \rho K_v \frac{d^2 u}{dz^2} \tag{2.3.2}
\]

Where \( P = \rho g (\zeta + z) \) . P is the pressure, \( \zeta \) is the surface elevation, \( K_v \) is the turbulent eddy viscosity. All solid boundaries are considered free slip and zero flux and the barotropic flow rate is assumed to be zero. Cormack et al (1974) found that \( \rho \) and \( \zeta \) are linear in \( x \) from the leading order asymptotic solution- therefore pressure can be eliminated from equation 2.26, giving:

\[
u = \frac{G_{st} K_v}{24L} (-4h^3 + 6h^2 -1) \tag{2.3.3}
\]

Where \( h' \) is the dimensionless height, \( h' = h/H \), \( L \) is the length of the channel (defined as, \( x=L/2 \) to \(-L/2\)) The Grashof number is a non-dimensional number that defines the transition between the diffusive and hydraulic limits of the flux by comparing the effect of buoyancy forces to the turbulent viscous forces:

\[
G_{st} = \frac{g' h^3}{K_v^2} \tag{2.3.4}
\]

The dimensional volume flux in either direction is then:
The mass flux can be written as

\[ M = K_v Ab \Delta \rho + \alpha G_{\gamma r} A^3 b K_v \Delta \rho \]  \hspace{1cm} 2.3.6

Where \( b \) is the width of the channel (equal to the minimum width of the contraction) and \( h \) is the minimum depth of the channel. The net mass flux is due to the horizontal diffusion and the advection of density. The VAD solution has been verified through laboratory experiments by Imberger (1976) and applied to some field data by Burling et al (1999).

2.3.3 Flow Ranges and the Grashof Number

Internal hydraulic theory and viscous advective diffusive theory describe the two limits characterising flux between two reservoirs with gravitational forcing. Internal hydraulic theory is valid in the absence of mixing, viscosity and friction (Wood, 1970, Armi, 1986, Armi and Farmer 1986, Lawrence, 1990, Dalziel, 1991) and the viscous advective diffusive solution (VAD) is valid when the exchange flow is dominated by turbulent mixing (Cormack et al 1974, Officer 1976). Between these two limits is a range of flows not accurately represented by either limit and Hogg et al (2001) used a numerical model and presented a description of the entire range of flows, using the naturally occurring non-dimensional Grashof parameter (equation 2.3.4). The flux of volume and density depends upon the Grashof number, the aspect ratio and the Prandtl number. The turbulent Prandtl number is assumed to be equal to one in the study completed by Hogg et al (2001), so that the Grashof number may equally well represent the turbulent diffusion on buoyancy forces.

\[ P_{\gamma T} = \frac{K_v}{K_p} \]  \hspace{1cm} 2.3.7

Where \( K_v \) is the turbulent eddy viscosity and \( K_p \) is the eddy diffusivity constant. The reduced gravity force is also important, \( g' = \frac{g \Delta \rho}{\rho_o} \). The hydraulic and VAD exchange flow solutions were non-dimensionalised by Hogg et al (2001) using the hydraulically limited equation below.

\[ Q_h = \frac{1}{4} bg' \gamma h^3 \]  \hspace{1cm} 2.3.8

Where \( h \) is the depth of the channel and \( b \) is the width of the channel. Therefore when there is no internal mixing and the exchange flow is hydraulically limited the exchange flow rate is equal to 1, as shown in equation 2.3.9.
The dimensional volume flux for the VAD solution in either direction is equation 2.3.5. Hogg et al (2001) non-dimensionalised the VAD solution with the hydraulically limited exchange flow to give,

\[ q_v = \frac{5(G_r A^2)^{\frac{1}{2}}}{96} \]  

2.3.10

Hogg et al (2001) determined the dependence of flux upon \( G_r A^2 \) within the two limits through experimentation. Where by the intermediate layer thickness only dependant on the \( G_r A^2 \) parameter:

\[ \frac{\delta}{h} = 3.4 \left( G_r A^2 \right)^{\frac{1}{4}} \]  

2.3.11

Hogg assumed that the velocity and density gradients in the interfacial layer are linear, using an integration of the idealised finite interfacial layer an estimate of flux as a function of \( G_r A^2 \) in the intermediate regime was created.

\[ q = \frac{(1 - 3.4(G_r A^2) - 0.25)}{2} \]  

2.3.12

The results from this equation are consistent with previous work- whereby an increase in mixing will decrease the total flux, Winters and Seim (2000) found that entrainment between the two layers causes recirculation within the channel, thereby reducing the total volume flux. Hogg et al (2001) concluded that when \( G_r A^2 > 10^5 \) the mass and volume flux are within 10% of the hydraulic prediction and when \( G_r A^2 < 40 \) the VAD solution is a good prediction. Between these two limiting values the intermediate regime described by equation 2.3.12 is representative of the flow. Hogg et al (2001) presented a unifying description of a large range of flows from these results, using the Grashof number to define the non-dimensional exchange flows as either represented by the hydraulic limit, the VAD limit or an intermediate regime.

\[ G_r A^2 < 40 \quad q = 0.052(G_r A^2)^5 \]

\[ 40 < G_r A^2 < 10^5 \quad q = 1 - 1.7(G_r A^2)^{-0.25} \]

\[ G_r A^2 > 10^5 \quad q = 1 \]  

2.3.13

Figure 2.3.1 illustrates the dampening effect that internal mixing has on the exchange flow. \( G_r A^2 \) is plotted on the x-axis, on the y-axis the mass flux that is non-dimensionalised by the hydraulic exchange flow. Gibraltar lies close to the hydraulic limit, there is little internal mixing, and the non-dimensional mass flux is almost equal to 1. Bosphorous and Burlington
are both within the transitional region, therefore exchange flow is dampened by some internal mixing, but the exchange flow is not adequately represented by either limit. The VAD limit occurs when \( G_{\tau}A^2 \) is less than 100, there is very strong internal mixing. The exchange flow is greatly decreased, with non-dimensional mass flux is less than 20% of the maximum exchange flux possible.

Figure 2.3.1 Plot of \( G_{\tau}A^2 \) against the non-dimensional mass flux for some field applications (Hogg et al, 2001).

2.3.4 Exchange flows with tidal forcing

Neither hydraulic theory nor the VAD solution incorporates the effects of tidal flow into their estimates of exchange rates. Tidal forces can significantly alter the dynamics of exchange flows through a constriction and therefore alter the flux (Phu, 2001). At any geographical
location the types of tide observed depend on the complex interaction between principal tidal components, topography, local effects such as topography and meteorological conditions to produce other tidal harmonics. Dietrich et al (1980) defined the behaviour of tides as a progressive wave. The tidal amplitude is the maximum vertical fluctuation in water levels over a tidal cycle, the flood tide is the rising tide and the falling tide is termed the ebb tide.

Tidal currents are identified as being a major process inducing mixing in estuaries, bays and straits (Fisher et al, 1979). The vertical oscillations of the tide are accompanied by tidal currents, which are horizontal water motions. Tidal currents are a complex three-dimensional process, by which water particles move in an elliptical path. In shallow seas this causes a frictional drag along the seabed. When tidal currents are strong this produces a vertical current shear, which generates turbulent and causes vertical mixing in the lower water layers (Brown et al, 1989).

Superimposed over this horizontal tidal current is a steady circulation, referred to as the residual current. Fisher et al (1979) defined the residual current as the velocity field obtained by averaging the velocity over the tidal cycle. The Coriolis force in large reservoirs can cause the residual circulation. More commonly it is due to the tidal flows interaction with the basin topography. The residual current is a mechanism through which tidal pumping occurs (Fisher et al, 1979). Tidal pumping is a mechanism by which mixing and flushing occurs in enclosed basins, illustrated in Figure 2.3.2 (Stommel and Farmer, 1952). The diagram shows the mechanisms of flood and ebb tidal flow. The flood flow through a contraction into the basin acts as a confined jet, the water moves a great length into the estuary and some of the water is entrapped in the bay as a result of the basin bathymetry. The ebb flow draws fluid from the entire mouth area as an outward sink flow. Therefore the water returning out into the ocean is not the same as the water entering; there is a net flushing of the basin (Fisher et al, 1979).
Armi and Farmer (1986) addressed the issue of tidal influence by imposing a net barotropic component under quasi-steady conditions on the steady state two-layer exchange in internal hydraulic theory. Their results both showed there is a great dependence of the flux on the tidal forcing strength – tidal forcing leads to an increase in exchange flux from the steady state hydraulic dominated case. Figure 2.2.3 illustrates these results, varying the strength of the barotropic component from moderate to intermediate to strong showing the effects on the layer depths and exchange fluxes (Armi and Farmer, 1986). If the tidal forcing is strong enough the two-layer exchange can be reverted to one-way exchange, with a motionless layer or both layers flowing in the same direction. Armi and Farmer (1986) further found that if the time taken for a long wave to propagate through the sub critical section between the topographic and virtual controls is short compared to the tidal period then the tidal flow has no effect on the exchange flow. If this is not true then tidal oscillation introduces a time dependant factor to exchange flows, introducing unsteadiness to the flow that is not considered in steady state hydraulic theory.
Figure 2.3.3 The effect of barotropic forcing on exchange layer depth (Armi and Farmer, 1986).

Exchange volume flux due to tidal forcing is a velocity parameter through the cross sectional area of the strait, where the velocity is induced by the rising and falling water levels caused by tidal effects (Phu, 2001). These effects are tidal currents, which are periodic and described by the following equation:

\[ u = u_t \sin\left(\frac{2\Pi t}{T}\right) \]  \hspace{1cm} 2.3.14

Where \( u \) is the tidal velocity at any time, \( t \), \( u_t \) is the maximum tidal velocity and \( T \) is the tidal period. The tidal current can further be characterised by the amplitude of the tidal current over the tidal period. Phu (2001) developed an expression describing the effect of tides on two layer exchange flows and quantifying the fluxes using simple flow parameters. \( E_t \) is a non-dimensional parameter expressing the relative magnitudes of baroclinic flux over barotropic flux (the relative strengths of buoyancy to tidal forces driving the observed exchange flow). The barotropic flux is estimated using the following equation:

\[ Q_t = \frac{\text{vol}_{\text{prism}}}{T} \approx \frac{1}{4} \frac{L}{T} \frac{aw}{4} = \frac{awL}{T} \]  \hspace{1cm} 2.3.15
The volume of the tidal prism is the volume of the estuary between low and high water, which is equal to the tidal range times the surface area. This method works on the assumption that the ocean water is mixed with estuary water and the water exiting over one tidal period does not return on the next. Where L was taken as the length of the experimental tank in work conducted by Phu (2001). This length scale is defined as the length of the tidal influence and in geophysical applications is taken as the length of the tidal excursion. w is the width of the channel and a is the tidal amplitude. The parameter $E_t$ can be calculated using equation (2.3.16).

$$E_t = \frac{0.25(g'h)^{0.5}hb}{awL}$$ \hspace{1cm} 2.3.16

Where $h$ is the height at the contraction, and $hb$ is the cross-sectional area of the contraction. $w$ is the width of the contraction. Phu (2001) derived the following classification system using the $E_t$ parameter to describe the prevalent forcing of the system.

**Tidal Regime**, tidal forces are dominant and flow is primarily a result of the mechanism of tidal pumping. $E_t<1$. The exchange flux can be approximated by the tidal forcing; $q_t = \frac{awL}{T}$.

**Transitional Regime**, the tidal and buoyancy forces are equally important $1<E_t<5$

**Steady State Hydraulic Regime**, the buoyancy forces dominant the flow is characterised by baroclinic forcing. $E_t>5$. The exchange flow can be approximated using steady state internal hydraulic theory; $Q_t = 0.25(gh)^{0.5}hb$.

**2.4 Current Modelling Programs**

Current estuarine modelling programs tend to use complex biological modelling and simple hydrodynamic forcing, generalised in terms of the location or ‘type’ of estuarine system. This highlights the need for a simple classification system based on the hydrodynamics of an estuary. This classification system accurately specifies the dominant forcing of the exchange flow rates, this determines the residence times of water in the basin. A very important parameter determining water quality, which is very useful for biological modelling. Therefore the use of simple models, which estimate the exchange flow, based on small amounts of field data, or if this is unavailable from external parameters is useful in terms of these modelling programs. Together with the classification of the hydrodynamics an overview of the system can be developed.
2.4.1 SERM model

The SERM model is a coupled biological and physical estuarine response model developed by the CSIRO (2001). It consists of a simple circulation model that predicts physical exchange in three different estuary types, based on simple box geometry and forcing data (river flow and tidal amplitude). The biogeochemical/ecological model is much more complex and has been developed through previous coastal studies, yet relies on the exchange flows predicted by the simple hydrodynamic model. SERM is a first attempt by the CSIRO to develop a broad, generic model for Australian estuaries. The modelling focuses on the relationship between the natural and anthropogenic pressures and the estuarine state – aiming to guide thinking about management responses through understanding the way estuaries respond to different pressures.

Some of the pressures used in the model are complex, there are many indicators related to the chemical forcing – such as nutrient and sediment loads, nutrients and dissolved oxygen, light attenuation and primary producers. For each parameter combination the model was run over a 10-year period, reaching a repeating seasonal cycle. The initial conditions represent a mesotrophic estuary with a viable benthic community, which may be inappropriate for highly degraded systems.

The SERM hydrodynamic model is very simple, providing a total of 6 estuarine types. The geometry and circulation are characterized by the estuarine type and the depth. The estuarine type specifies the nature of the circulation – lagoon, tidal or salt wedge, and the horizontal size and shape of the estuary. The depth is the average depth of the estuary. The river forcing is characterized by the climate zone – seasonal variation in flows, loads and temperature, the freshwater replacement time specifies the mean annual river flow in terms of the turnover time of the estuary volume.

The Lagoon hydrodynamic model represents the estuarine type as a single, well-mixed box, with fresh water replacement time and exchange with the ocean. In reality the rate at which a lagoon exchanges with the ocean depends on the geometry, depth and circulation near the mouth – in the model the size and shape of the estuary are of no consequence as flows and loads are specified relative to the volume and area. The fresh water replacement time specifies the time taken for the annual mean river flow to deliver a volume of fresh water equal to the volume of the estuary. The flushing time in the model is calculated by specifying the time for exchange processes at the mouth of the estuary to mix a volume of oceanic water equal to the
volume of the lagoon. The flushing time and fresh water replacement time are combined in
the model total exchanges between the estuary and ocean. If the freshwater replacement time
is much smaller than the flushing time the estuary will be fresh, if the flushing time is greater
the estuary will be close to marine salinity. The plan view is demonstrated below:

Figure 2.4.1 Lagoon hydrodynamic schematic

Tidal estuaries are modeled as vertically well mixed channels – fresh water replacement time
and diffusive exchange due to tides are modeled in 5 adjacent boxes. Extreme simplifications
have been made, including having the depth be uniform through out the estuary. The tidal
range and phase is uniform through out the estuary. The diffusive exchange through adjacent
boxes is proportional to the tidal flow volume passing through the interface between the
boxes. The total exchange between the two adjacent boxes is the sum of the two-way tidal
diffusive exchange plus the river forced downstream transport.
The salt wedge estuary is represented as a stratified channel split into 5 boxes and two layers. Circulation and exchange occur due to fresh water replacement time, horizontal diffusive exchange is due to tidal motions and vertical entrainment and mixing are modeled. The depth of the estuary is assumed to be uniform throughout the whole estuary, as is the tidal range and the height of both layers. Horizontal diffusive exchange is calculated from tidal motions, entrainment is represented by a uniform upwards velocity. The lower salinity is considered constant, equal to the ocean salinity. It is assumed that there is no internal mixing between the layers.
The simplifications made in the hydrodynamic model greatly affect the nature of the application of the SERM model. It is useful for general understanding of how estuaries are affected by different pressures – the biological modeling is adequate. Yet extreme caution is necessary when using the SERM model in analysis of individual estuaries – the geometry used is simple, the influence of shape and channel dimensions, depth and seasonal flow rates are not accounted for. The actual pressures on an individual estuary cannot be assumed from the SERM model.

The SERM project completed a case study on Wilson Inlet in Denmark. The study classified Wilson Inlet as a lagoon system, with an average river inflow of 3 m\(^3\)s\(^{-1}\) giving a fresh water replacement time of 330 days. The depth was characterized as shallow, from 1-3m. The flushing time was estimated as 200 days based on the long-term salinity balance, the model estimates the exchange flow between 100 and 500 days, without specification of seasonal changes.

2.4.2 Ozestuaries

Ozestuaries is a database of Australian estuaries created from an Australian Government Geoscience Australia and CRC for Coastal Zone Estuary and Waterway management initiative. The database provides a simple hydrodynamic classification based on 7 types of Australian estuaries and coastal waterways. For each classification system the structure, evolution, geomorphology is presented along with flow diagrams of the hydrology, sediment and nutrient dynamics. Each estuary is specified by its location, a condition assessment, the dominant forcing in the system, the basin and channel size, biological components and the tidal parameters. The condition assessment is based on the extent of modification in the catchment area and along the riverbanks, the water quality, amount of pressure on the system, habitat condition, fish condition and biological diversity. The 7 types are based on the relative influence of wave, tide and river power: wave dominated deltas, wave dominated estuaries, tide dominated deltas, tidal dominated estuaries, tidal flats/creek, coastal lagoons and strand plains.

The hydrodynamic classification occurs based on the location of the system. The South is considered wave dominated, the north is tidally dominated – this is based on the general distribution of the wave and tide forces on shelf environments. The climate and seasonality is also taken into account in this modelling- in strongly seasonal areas the conceptual model may vary between two classifications. Shark Bay is considered to always be a positive
estuary, whilst Wilson Inlet, the Peel-Harvey Estuary and the Swan River are positive hydrodynamic systems in winter and negative in summer. Shark Bay is classified, as a tidally dominated system, the Swan River, Wilson Inlet and the Peel-Harvey Estuary are all wave dominated. The classification of the three systems considers their biological state to be dependent on the wave energy; the systems have high sediment trapping efficiency, naturally low turbidity, partially mixed circulation and high risk of sedimentation. The Swan River and Peel-Harvey Estuary is considered to be extensively modified due to changes of land use, Wilson Inlet is classified as modified due to changes of catchment cover and Shark Bay is considered unmodified.

The wave dominated estuaries classification is considered to be a coastal bedrock embayment, partially filled with sediment and wave energy is the dominant forcing. They occur on exposed coastlines with small tidal influence, and have a constricted entrance. The narrow entrance restricts exchange flow in the model. The tidally dominated estuary is very similar, a coastal bedrock embayment partially filled with sediment, with tidal energy as the dominant forcing in the system. They occur on low gradient coastlines characterised by meso to micro tidal ranges. The entrance is very similar to the lagoon specified in the SERM model, a landward tapering funnel shaped valley that is bound by intertidal sedimentary environments. Embayment is bed rocked lined coastal indentations, without much sediment infill, they have wide round bays and unconstricted exchange with the ocean. The influence of wave and tidal power differs depending on the regional conditions. Lagoons are classified as small, shallow basins with low fresh water inputs. They experience strong tidal currents and as a result of this, their entrances are often intermittently closed due to sediment transport, therefore there is low wave energy within the lagoon.

The hydrodynamic modelling of this classification is very simplistic; the classification of estuaries occurs on the basis of their location, assuming generalised conditions are applicable. The simplification of the hydrodynamics implies that the Swan River estuary, the Peel-Harvey estuary and Wilson Inlet have similar processes, when these estuaries are in practise different. This classification system does allow for seasonal variation in the hydrodynamic mechanisms, which the SERM model does not take into account. The biological and sedimentary modelling is detailed and complex, yet relies on simple generalised hydrodynamic classification to determine flushing times- which is an important parameter determining water quality.
2.5 Field Applications

The sites chosen for the application of the internal and external model and the classification system are relatively diverse systems. The sites are important estuarine systems in Western Australia, Wilson Inlet, the Swan River estuary, the Peel-Harvey estuary and Shark Bay. The magnitude and type of the topographic constriction vary between the sites, as does the type and magnitude of the exchange flow forcing. The systems are all topical within Western Australia. Numerous studies have been completed on Wilson Inlet, the Swan River and the Peel-Harvey estuary, as increasing anthropogenic pressure has resulted in the deterioration of water quality and subsequent algal blooms and other issues. Shark Bay is of interest as the exchange flow between Hamelin Pool and Hopeless Reach is restricted and as such Hamelin Pool is hyper saline. There is increasing anthropogenic pressure on this system, due to increasing tourism in the region- therefore knowledge of the exchange flow and hydrodynamics of the basin is useful in terms of understanding how much stress the system can undergo before deterioration occurs.

2.5.1 Wilson Inlet

Wilson Inlet is situated on the Southwest coast of Western Australia. It is a relatively small estuary, 48 km$^2$ in area with an average depth of 1.78m. The estuary opens to the sea towards the western end of the bay, the mouth is sheltered by rocky headlands from the prevailing winter south-westerly winds and swells (Hodgkin and Clark 1988).

The water in Wilson Inlet is always brackish, salinity ranging from 12-20 ppt (Lukatelich et al 1984); this density difference is the driving force for the two-way exchange flow. The two-way exchange flow occurs as less dense inlet water flows out to the ocean as a top layer being replaced by dense ocean water, flowing into the inlet as a bottom layer. The exchange flow occurs through a narrow channel, a sand bar blocks the channel mouth for approximately seven months of the year. Currently it is artificially breached when the water level in the inlet reaches 1.1m above MSL (Hodgkin and Clark, 1988). This breaching creates high flow rates, which scour a channel of maximum dimensions 350m long, 2m deep and 100m wide (Ranasinghe 1995). The measured exchange rates differ greatly depending on the rainfall and river input. Previous studies on Wilson Inlet have found that the flow rate is highly dependant on river inflow (Ranasinghe, 1995). The river inflow is confined to the months of June-October and reduced inflow reduces exchange with the ocean increasing the size of the sand bar (Hodgkin and Clark, 1988).
Figure 2.5.1 Wilson Inlet, Denmark, Western Australia

2.5.2 Swan River

The Swan River is a South-West Australian estuary that runs through the city of Perth (Kurup et al 2000). The river has a free connection with the ocean that is permanently open. The system is of current interest due to increased nutrient loading causing stress on the system from groundwater, runoff and river inflows (Stephens and Imberger, 1996). By the early 1990’s algal blooms and fish deaths in the Swan River brought public attention to the deterioration of the system (Swan River Trust, 2001). Density stratification is prolonged, intense and seasonal, due to the summer drought conditions and the confinement of rainfall to the winter months (Spencer, 1955). The hydrodynamics of the system have been studied in various ways (Spencer, 1955; Stephens and Imberger, 1996; Ranasinghe and Pattiaratchi, 1999; Kurup et al, 2000; Hamilton et al, 2001; The Swan-Canning Cleanup Program, 2001); the exchange flow is controlled through two topographic constrictions. The exchange flow with the ocean is restricted by the presence of a sill approximately 3-5 km upstream, the propagation of the salt wedge shows seasonal variation and can extend almost the full extent of the river.
2.5.3 Peel Harvey

The Peel-Harvey estuary is situated in Mandurah, the region has had high residential and commercial growth over the past 25 years, which has put pressure on the estuarine system. Eutrophication is a major problem in the Peel-Harvey, causing large Nodularia blooms. In an effort to increase flushing rate of the system the Dawesville channel was constructed to connect the estuary to the Indian Ocean in 1995. This channel is 130-200m wide, 2 555m long and 4.5-6.5m deep (Ranasinghe, 2000). It is estimated that the flushing rate has increased by 50% since the opening was created (Waters and Rivers Commission, 2003). The original Mandurah channel has dimension of width 50m, length 4km and depth approximately 1m (Hearn et al, 1994).
Shark Bay is a large coastal embayment located on the central west coast of Australia. The system operates as an inverse estuary (Logan and Cebulski 1970). The large evaporation in the system, ten times the rainfall has made the embayment hyper saline in areas (Burling et al 1999). The bay is comprised of two major reaches; the exchange between them is severely hindered by the presence of the Faure sill. The exchange flow between Hamelin Pool and Hopeless Reach predominantly takes place through a channel 2 km wide, 6m deep and 20 km long (Burling et al 1999). There is a longitudinal density gradient through Shark Bay; the furthest reach of Hamelin Pool has the highest density. The density difference between Hamelin Pool and Hopeless Reach drives the exchange flow, with the higher density water from Hamelin Pool moving as a bottom layer into Hopeless reach and less dense water from Hopeless reach moving as a top layer into the Pool (Burling et al 1999). The exchange flow between these two reaches has been studied as a diffusively dominated system, well mixed vertically and with a linear gradient in salinity (Burling et al 1999). The system has classified as steady state by Logan and Cebulski (1970) over a 13 yr study period – with little seasonal or annual variation. But more recent field experiments have shown that the salinity structure can be seasonally variant. The range of salinities is unchanged but a strong two-layer structure observed to the east of Cape Peron during winter represents a dramatically different flow regime than that in summer.
Figure 2.5.4 Shark Bay, Western Australia. Note the Faure sill by which flow is limited between Heralds Loop and Hopeless Reach.
3.0 Methods

Exchange flows between four estuarine systems and the ocean were calculated using two models. The models are compared; one model, referred to as the internal model, uses field data to calculate the exchange flow rate and flushing time of the basin. The second model uses external parameters and provides an estimate of the exchange flow rate based on steady state hydraulic assumptions. When field data is unavailable or unviable to obtain, modelling based on external parameters is necessary and the reliability of the calculations is a matter of interest. Both models together with the classification of the dominant hydrodynamics of the basin provide a picture of what is occurring in the system, from which further modelling of greater complexity can be used to provide more accurate estimates of processes in the estuary.

The first model is referred to as the internal model, because field data is required for the calculation of the flushing time in the basin. This model uses the salinity and temperature of the ocean and the inlet to determine the exchange flow forcing, it takes into account the effect of mixing between the flowing layers on the exchange flow. Creating estimates of the Grashof parameter and the mixed layer depth the flow rate is calculated using the classification system defined by Hogg et al (2001). The strength of the tidal influence is calculated using the method devised by Phu (2001). These factors are combined to create a classification based on the dominant hydrodynamics of the system. The external model is named according to its reliance on externally available parameters, the buoyancy flux out of the system. This model assumes there is no internal mixing in the channel and the estuary has been at steady state with the ocean for a long time, therefore the estuary and ocean have the same density.

The models and the classification of dominant hydrodynamics are applied to the Swan River estuary, Wilson Inlet, the Peel-Harvey estuary and Shark Bay are The Swan River estuary, Wilson Inlet and the Peel-Harvey Estuary are seasonally variable systems, therefore the exchange flow and parameter sensitivity is calculated on a monthly basis. Where as Shark Bay shows little seasonal variation in the density range, there is little precipitation and no river inflow, therefore the models are calculated on a yearly basis, as the classification of the hydrodynamics of the system.
3.1 Internal Model

The exchange flow can be calculated using an iterative process, when the density difference between the ocean and inlet are known. Initially the system is assumed to be viscous advective diffusively limited and the mixed layer ratio is assumed equal to 1.

\[
\frac{\delta}{h} = 1
\]

Where \( \delta \) is the mixed layer depth and \( h \) is the depth of the channel.

The Grashof number can be estimated using this mixed layer value from Hogg et al (2001).

\[
G_nA^2 = \left( \frac{\delta}{\frac{h}{3.4}} \right)^{\frac{1}{4}}
\]

3.1.1

From this value of \( G_nA^2 \) a non-dimensional flow rate can be calculated. Depending on the value of \( G_nA^2 \) the equation used to calculate this non-dimensional flow rate would change according to equations 3.2 of the literature review.

Initially the non-dimensional flow rate will be calculated using (Hogg et al, 2001)

\[
q = \frac{5}{96} \left( G_nA^2 \right)^{\frac{1}{4}}
\]

3.1.2

The velocity of the flow through the channel can be estimated using \( u = q/(bh) \); where \( b \) is the width of the channel and \( h \) is the height of the channel. A rough estimate of \( G_nA^2 \) can be calculated by assuming momentum eddy diffusivity parameter \( K_v \) can be estimated using the following equation (Fisher et al, 1979);

\[
K_v = 0.1u * h
\]

3.1.3

From this estimate of the momentum eddy diffusivity parameter a new estimate of \( G_nA^2 \) can be calculated (Hogg et al, 2001).

\[
G_nA^2 = \frac{g' h^3 A^2}{K_v^2}
\]

3.1.4

Where \( g' \) is the reduced gravitational force, \( g' = g(\rho_o - \rho_i) / \rho_o \), where \( \rho_o \) is the density of the ocean outside of the channel and \( \rho_i \) is the density of the inlet outside of the channel. \( A \) is the aspect ratio, \( h/L \) where \( h \) is the height of the channel and \( L \) is the length of the channel.

The density in the estuaries is calculated using salinity, temperature and water depth profiles available from the Waters and Rivers Commission. Then using the SEAWATER, Version 1.2 library of computational seawater routines for MATLAB the density of the inlet can be
calculated. The density of the ocean outside the inlet can be calculated on a monthly basis using the SEAWATER program and CTD data from the National Oceanographic Data Centre.

From this estimate of $G \rho A^2$ the mixed layer ratio can be recalculated using the initial equation. This process is repeated until the answers converge. The final result is a non-dimensional flow rate based on the estimate of the mixed layer. From this non-dimensional flow rate the actual flow rate can be calculated using the following equation (Hogg et al, 2001);

$$Q = q * \frac{1}{4} g^\frac{1}{2} bh^2 \left [ m^3/s \right ]$$  \hspace{1cm} 3.1.4

Using this flow rate the residence time can be estimated using the same method as specified in the hydraulically limited section.

3.2 External Model

This method calculates the exchange flow through a topographic constriction using externally available data. Conceptually, before any forcing is applied the inlet/estuarine density is assumed equal to the ocean density. The buoyancy forcing of evaporation, precipitation and river inflow subsequently causes density differences between the ocean and estuary creating exchange flow between the inlet and the ocean (Finnigan et al, 1999, 2000).

Firstly the monthly evaporation and precipitation data was collected over the period of 1 year. This is available from the Bureau of Meteorology. The evaporation and precipitation considered over the whole Swan River surface area were collected from the Perth Airport Meteorology Station (Bureau of Meteorology, 2001). Monthly evaporation and precipitation data for Wilson Inlet is from the Albany Meteorology Station. The evaporation and precipitation data is collected from the nearest station to the Peel-Harvey Estuary at Roelands and Wokalup. The Hamelin Pool shows little seasonal variation in density range, therefore the exchange flow was calculated on a yearly basis, using evaporation to precipitation ratio from Burling et al (1999).

The buoyancy flux in/out of the system is calculated as a net buoyancy flux; therefore the net evaporation/precipitation for the month is used. The total evaporation/precipitation is considered to act in a distributed manner over the whole surface area of the basin.

I.e. Precipitation $P=95$ mm/month
Evaporation $E = 15$ mm/month

The net precipitation rate $P = (15-95)/(1000\times4\times7\times24\times3600)=3.3\times10^{-8}$ m/s

The reduced gravitational force is given by (Armi, 1986);

$$g' = \frac{g(\rho_o - \rho_{fw})}{\rho_o}$$

3.2.1

g is gravity, $\rho_o$ is the density of the ocean, assumed to be constant at 1025 kg m$^{-3}$ and $\rho_{fw}$ is the density of freshwater, 1000 kg m$^{-3}$. Using equations 3.1.0 and 3.1.1 the equivalent net buoyancy flux over the entire surface for the month can be calculated using (Finnigan and Ivey, 1999).

$$B_o = Pg' \quad [m^2 s^{-3}]$$

3.2.2

Where P is the precipitation rate. The buoyancy flux is an inward flux over the surface of the basin. For months where the evaporation exceeds the precipitation the heat flux is calculated, using the method outlined by Fisher (1979).

$$\hat{H} = \rho EL \quad [W m^{-2}]$$

3.2.3

Where $\rho$ is the density of freshwater, $E$ is the evaporation rate, calculated in the same way that the precipitation rate is (above). L is the latent heat coefficient, dependant on temperature. For the case studies used in this thesis $L = 2.4\times10^6$ (JKg$^{-1}$).

The buoyancy flux due to evaporation is calculated using the following equation (Finnigan and Ivey, 2000).

$$B_o = \frac{\alpha g \hat{H}}{\rho C_p} \quad [m^2 s^{-3}]$$

3.2.4

Where $\alpha = 2\times10^{-4}$, g is gravity, $H$ is the heat flux calculated above, $\rho$ is the density of freshwater and $C_p = 4.181\times103$ (JKg$^{-1}$) (Fisher et al, 1979). The net buoyancy forcing of the system also takes into account the freshwater inflow into the system. River inflow forms a buoyancy plume that we model to form a layer of freshwater over the entire surface area.

The inflow rates for each river are available from the Waters and Rivers Commission. The point of discharge is not of importance for this model and a total inflow rate is calculated by simply adding each inflow rate. The effect of river flow is then distributed using the surface area of the entire inlet. Together with the reduced gravitational force is (equation (3.1.1)) the buoyancy flux due to river inflow is then calculated (Fisher, 1979).

$$B_o = \frac{Q g'}{A} \quad [m^2 s^{-3}]$$

3.2.5
The buoyancy fluxes caused by river inflow, precipitation and evaporation were then added, giving a total buoyancy flux for each month. The buoyancy flux out of the system caused by evaporation automatically mixes and occurs across the entire basin surface area. In the case of precipitation and river inflow, as the simple schematic below indicates, they cause a stabilising buoyancy flux. It is assumed in the model that some mechanism, such as wind or tide, mixes the estuarine waters and creates a density difference with the ocean. Therefore due to the river inflow and precipitation the estuary becomes less dense than the ocean, in the case of evaporation this buoyancy flux acts to increase the density of the inlet.

![Exchange flow mechanism due to buoyancy flux](Image)

Using these values of buoyancy flux for each month, the hydraulically limited exchange flow through a topographic constriction can be calculated (Ivey, 2003, Finnigan and Ivey, 2000).

$$Q = 0.4h(B_o b^2 BL)^{1/3} \text{[m}^3\text{/s] } \quad 3.2.6$$

Where \(h\) is the height at the contraction, \(B_o\) is the buoyancy flux, \(b\) is the minimum width at the contraction and \(BL\) is the surface area of the basin.

Using the exchange flow rates and the volume of water in the basin the residence time of the water can be calculated very simply.

Where \(T = \frac{V}{Q}\) [s]; \(V\) is the volume in \(m^3\) and \(Q\) is the flow rate in \(m^3/s\).

### 3.3 Tidal Effects
The $E_t$ parameter is a ratio baroclinic flux over barotropic flux. First the tidally dominated flow rate is found using the following equation (Phu, 2001);

\[ Q_t = \frac{abL}{T} \]  \hspace{1cm} 3.3.1

Where $a$ is the tidal amplitude [m], $b$ is the width of the channel [m], $L$ is the length of the tidal excursion [m] and $T$ is the tidal period [s]. The hydraulically limited flow rate is calculated using (Armi and Farmer, 1986);

\[ Q_h = \frac{1}{4}(g'h)^\frac{1}{2}hb \] \hspace{1cm} 3.3.2

Where $h$ is the height of the channel and $b$ is the width, and $g'$ is found using the monthly density values, as specified in section 2 (Phu, 2001).

\[ E_t = \frac{Q_h}{Q_t} \] \hspace{1cm} 3.3.3

Using this parameter an estimate of the dominant forces of the system can be illustrated by graphing the parameter $G_tA^2$ against $E_t$. 
4.0 Results

Results of the application of the internal and external model and the classification system to Wilson Inlet, the Swan River, the Peel-Harvey estuary and Shark Bay are presented in this section. The results for each application are presented separately, beginning with the exchange flow rate and flushing time results from the internal model, followed by results from the internal model and a comparison of the model results. Then the results of tidal exchange and the classification of the dominant hydrodynamics in the system are presented. The models are used in a predictive sense to calculate the exchange flow rate and flushing time for Wilson Inlet based on hypothetical dredging situations. Finally a regime plot of the dominant hydrodynamics for all for sites is presented with a short summary.

4.1 Wilson Inlet

4.1.1 Exchange Flow estimates using Internal model

The surface area of Wilson Inlet is approximated at $48 \times 10^6$ m$^2$, the volume at mean sea level is $85 \times 10^6$ m$^3$ (Hodgkin and Clark, 1988). A sand bar at the inlet contraction blocks Wilson Inlet for approximately 7 months of the year. The bar artificially breached when the water level in the inlet reaches 1.1 m above mean sea level, usually in July. This breaching causes a channel to be scoured of initial approximate dimensions; height is 2m, length 100m and width 100m (Ranasinghe and Pattiaratchi, 1996). The channel dimensions then decrease gradually until the inlet is closed to the ocean mid-December.

The actual density values in the inlet were measured in June, July, December and March as part of a study on Wilson Inlet completed by Ranasinghe (1995). Table 4.1.1 shows the reduced gravitational values for July, August and December ($g'$), the final estimates of velocity ($u$), final estimates of the turbulent eddy diffusivity parameter ($Kv$), final estimate of the non-dimensional Grashof number parameter ($GrtA^2$), the mixed layer ratio ($\delta/h$) and non-dimensional exchange flow rate ($q$). As the density difference between the ocean and the inlet decrease the reduced gravitational force decreases ($g'$). As this parameter decrease the turbulent grashof parameter decreases. This corresponds to an increase in the mixed layer ratio. The mixed layer ratio is the ratio of the vertical scale of the mixed layer to the depth of the water column. Therefore as the mixed layer depth ratio increases there is a subsequent decrease in the non-dimensional exchange flow rate. In December the water column in the
channel is almost entirely mixed indicating that the system is closer to the system predicted in the viscous, advective, diffusive limit, than the hydraulically limited system that is assumed in the external data exchange flow calculations. July and August have similar profiles, the mixed layer ratio is smaller and the inlet hydrodynamics are closely linked to the hydraulically limited system. The exchange flow rate and flushing time for the internal model are available in Table 8.1.1 in the appendix.

Table 4.1.1 Wilson Inlet estuary parameters for internal model

<table>
<thead>
<tr>
<th></th>
<th>2001</th>
<th>g’</th>
<th>u [ms⁻¹]</th>
<th>Kᵥ [m²s⁻¹]</th>
<th>GᵥA²</th>
<th>δ/h</th>
<th>q</th>
</tr>
</thead>
<tbody>
<tr>
<td>July</td>
<td>0.1398</td>
<td>4.4×10⁻³</td>
<td>8.8×10⁻⁵</td>
<td>56530</td>
<td>0.22</td>
<td>0.889</td>
<td></td>
</tr>
<tr>
<td>August</td>
<td>0.1565</td>
<td>4.4×10⁻³</td>
<td>8.9×10⁻⁵</td>
<td>62874</td>
<td>0.21</td>
<td>0.892</td>
<td></td>
</tr>
<tr>
<td>December</td>
<td>0.1174</td>
<td>2.6×10⁻²</td>
<td>2.6×10⁻⁴</td>
<td>1693</td>
<td>0.94</td>
<td>0.676</td>
<td></td>
</tr>
</tbody>
</table>

* g’ values from Ranasinghe and Pattiaratchi, 1995 and NODC data 1995.

4.1.2 Hypothetical dredging scenarios using the internal model

The internal model is used in a predictive sense to determine the exchange flow rate and flushing time for different dredging scenarios. The channel width and length can be altered greatly by dredging. Figure 4.1.1 shows monthly exchange flow rates for Wilson Inlet based on the same density differences in the inlet and ocean measured by Ranasinghe (1995) in March, July, August and December. The month of March is represented by 3 on the below figure, December is 12 for the year 1995/96. The channel is considered permanently open and the dimensions do not change within the Six scenarios are shown, the height of the channel remains constant at 2.5m. Increasing the width of the channel greatly increases the exchange flow. However increasing the length of the channel can hinder the exchange flow in the model- shown through the difference between the length equal to 200 and 2000m with the width of the channel equal to 500m.
4.1.1 Internal model monthly exchange flow rates for Wilson Inlet multiple dredging scenarios

Figure 4.1.1 illustrates the flushing times of the inlet based on the above exchange flow rates for each scenario. The flushing times calculated would occur if the conditions for the month were to continue. The flushing times greatly decrease when the width increases, the largest difference in flushing times occurs between width 200-300m and length 150-300m.

4.1.2 Internal model monthly residence times for Wilson Inlet with multiple dredging scenarios.

Figure 4.1.2

4.1.3 Exchange flow estimates using External model

The total river inflow is an addition of the Hay, Denmark and Sleeman Rivers, which respectively contribute 65%, 25% and 10% to the total annual flow (Ranasinghe and Pattiaratchi, 1996). The buoyancy flux was calculated on a monthly basis from precipitation...

Figure 4.1.3 illustrates the monthly buoyancy fluxes at Wilson Inlet estimated by the methods in section 3. The buoyancy flux is due to a combination of evaporation, precipitation and river inflow for the year 1995. There is variation in the buoyancy flux between each month, however the flux remains within the same order of magnitude for the entire year. The evaporation rate, heat flux, river inflow and resulting buoyancy fluxes are shown in Table 8.1.2 in the appendix.

![Monthly Buoyancy Flux for Wilson Inlet](image)

**Figure 4.1.3 Monthly net buoyancy fluxes for Wilson Inlet.**

Table 8.1.3 in the appendix contains the exchange flow rates calculated using the external model for the months of July, August and December. In July the bar has just been breached, this occurs when the water is 1m above mean sea level, therefore the channel dimensions and the volume in the inlet are both maximums (Hodgkin and Clark, 1985). In December the channel is greatly reduced and the height is 1m, length is 100m and width is 20m (Ranasinghe, 1996). The flushing time and exchange flow rate appear highly dependant on the channel width, there is a large increase in the flushing time for December, despite similar buoyancy forcing.

Figure 4.1.4 illustrates a comparison between the monthly exchange flow rates for Wilson Inlet calculated using the internal and external model. The comparison occurs for the months of July, August and December as the channel dimensions and density of the ocean and inlet are known for these months (Ranasinghe, 1996). The exchange flows fro July and August are very close and there is very little difference between the models predictions. The exchange
flow rate for December is greatly reduced, the internal and external model predict very close results.

Figure 4.1.4 Monthly exchange flow rates for Wilson Inlet based on internal and external model

Figure 4.1.5 illustrates a comparison of the monthly flushing times for July, August and December for Wilson Inlet based on the internal and external model. The flushing times are less than 100 days for July and August and very similar for both the external and internal models.

Figure 4.1.5 Monthly Flushing Times for Wilson Inlet based on internal and external models.

4.1.4 Hypothetical dredging scenarios using the external model

Figure 4.1.6 shows the monthly exchange flow rates for Wilson Inlet based on the buoyancy fluxes calculated above and using different channel dimensions. The inlet is assumed to be permanently open and to have constant dimensions, all residence time estimates have a
channel depth of 2.5m. The length (L) is measured in metres, as is the width (b). With increasing channel width the exchange flow increases largely. Based on the external model the length of the channel has no influence on the exchange rate. This is evident as the exchange flow for a channel of dimensions, length 2000m and width 400m has the same exchange flow as a channel of the same width but length equal to 200m.

Figure 4.1.6 External model monthly exchange flow rates for Wilson Inlet multiple dredging scenarios.

Figure 4.1.7 illustrates the corresponding flushing times for the exchange flow rates in figure 4.1.6. The residence time calculated for each channel dimension follows the same monthly trend, increasing with increasing channel width. The minimum flushing time corresponds to the greatest exchange flow rate, for the channel of the greatest width, which is equal to 500m.

Figure 4.1.7 External model monthly residence times for Wilson Inlet with multiple dredging scenarios.
4.1.5 Tidal Influence on exchange flow

Calculating the tidal influence on the exchange flow rate requires an estimate of the tidally forced exchange flow. The tidal amplitude is 1.1m, the tidal period is 43200 s and the length of the tidal excursion is 2000m when the channel is open. The tidally forced exchange flow rate alters with changing width of the channel, for July and August $Q_t = 5.09 \text{ m}^3\text{s}^{-1}$, in December it is smaller, $Q_t=0.05 \text{ m}^3\text{s}^{-1}$ as the tidal excursion is restricted.

Figure 4.1.8 illustrates the monthly variation of the influence of the parameters $G_tA^2$ and $E_t$. $G_tA^2$ is much greater for the months of July and August, it is close to the hydraulic limit, in December $G_tA^2$ is greatly reduced, closer to the diffusive limit. The ratio of buoyancy forcing to tidal forcing is much greater in December, the value of $E_t$ is large and there is practically no tidal excursion into the inlet. July and August have much smaller values of $E_t$, therefore the effect of tidal forcing is greater, although the values are just greater than 5.

![Hydrodynamic Classification of Wilson Inlet](image)

*Figure 4.1.8 The Hydrodynamic classification of Wilson Inlet, $E_t$ plotted against the $G_tA^2$ on a logarithmic scale.*

4.2.0 Swan River

4.2.1 Exchange Flow estimates using Internal model

The surface area of the Swan River estuary is approximated at $25.9 \times 10^6 \text{ m}^2$, the volume of the Swan River is $75.7 \times 10^7 \text{ m}^3$ (Spencer, 1955). The topographic constriction that acts an
exchange flow control is the Fremantle Sill, depth at the sill is 5m (Stephens and Imberger, 1996). The width is 200m and the length of the channel is 8 km (Hamilton, 2001).

Table 4.2.1 shows the flow parameters for the Swan River estuary for the year 2001. The mixed layer depth varies significantly throughout the year. There is significant internal mixing in the channel throughout the whole year, there is maximum mixing from April through to June, where the channel is completely mixed. Table 8.2.1 in the appendix shows the exchange flow rate and flushing time calculated using these parameters in the internal model.

**Table 4.2.1 Swan River Estuary Parameters from the Internal Model.**

<table>
<thead>
<tr>
<th></th>
<th>g’</th>
<th>u [ms⁻¹]</th>
<th>Kᵥ [m²s⁻¹]</th>
<th>GᵣA²</th>
<th>δ/h</th>
<th>q</th>
</tr>
</thead>
<tbody>
<tr>
<td>January</td>
<td>0.0831</td>
<td>7×10⁻⁴</td>
<td>3.8×10⁻⁵</td>
<td>2710</td>
<td>0.48</td>
<td>0.76</td>
</tr>
<tr>
<td>February</td>
<td>0.0038</td>
<td>5×10⁻⁴</td>
<td>2.8×10⁻⁴</td>
<td>232</td>
<td>0.87</td>
<td>0.56</td>
</tr>
<tr>
<td>March</td>
<td>0.0038</td>
<td>5×10⁻⁴</td>
<td>2.5×10⁻⁴</td>
<td>291</td>
<td>0.82</td>
<td>0.58</td>
</tr>
<tr>
<td>April</td>
<td>0.009</td>
<td>4×10⁻⁴</td>
<td>2.3×10⁻⁴</td>
<td>83.7</td>
<td>1.12</td>
<td>0.44</td>
</tr>
<tr>
<td>May</td>
<td>0.0019</td>
<td>5×10⁻⁴</td>
<td>2.5×10⁻⁴</td>
<td>140.6</td>
<td>0.98</td>
<td>0.50</td>
</tr>
<tr>
<td>June</td>
<td>0.0009</td>
<td>4×10⁻⁴</td>
<td>2.3×10⁻⁴</td>
<td>83.7</td>
<td>1.12</td>
<td>0.43</td>
</tr>
<tr>
<td>July</td>
<td>0.0086</td>
<td>6×10⁻⁴</td>
<td>3×10⁻⁵</td>
<td>428</td>
<td>0.74</td>
<td>0.62</td>
</tr>
<tr>
<td>August</td>
<td>0.0659</td>
<td>7×10⁻⁴</td>
<td>3.7×10⁻⁵</td>
<td>2266</td>
<td>0.49</td>
<td>0.75</td>
</tr>
<tr>
<td>September</td>
<td>0.0773</td>
<td>7×10⁻⁴</td>
<td>3.8×10⁻⁵</td>
<td>2591</td>
<td>0.47</td>
<td>0.76</td>
</tr>
<tr>
<td>October</td>
<td>0.0620</td>
<td>7×10⁻⁴</td>
<td>3.7×10⁻⁵</td>
<td>2151</td>
<td>0.49</td>
<td>0.75</td>
</tr>
<tr>
<td>November</td>
<td>0.0474</td>
<td>7×10⁻⁴</td>
<td>3.6×10⁻⁵</td>
<td>720</td>
<td>0.52</td>
<td>0.73</td>
</tr>
<tr>
<td>December</td>
<td>0.0229</td>
<td>6×10⁻⁴</td>
<td>3.4×10⁻⁵</td>
<td>933</td>
<td>0.61</td>
<td>0.69</td>
</tr>
</tbody>
</table>

* g’ values from CTD data from the Waters and Rivers Commission, 2001 and NODC data, 2001.

Figure 4.2.1 shows the monthly variation in the Grashof parameter GᵣA² in the Swan River. The system varies greatly over the yearly period but remains in the transitional region. There is a quite distinct seasonal pattern, the value of GᵣA² is much smaller for February through July and close to the VAD limit, increasing dramatically during the other months.
4.2.1 The monthly variation in $G_{rA^2}$ in the Swan River.

4.2.2 Exchange flow estimates using External model

The buoyancy flux was calculated on a monthly basis from precipitation and evaporation data from the Bureau of Meteorology and river inflow data from the Waters and Rivers Commission. The reduced gravitational force is $g' = 0.239$.

The total river inflow is assumed approximately equal to the inflow of the Avon and Ellen Brook Rivers, which together account for 93% of river inflow (Swan River Trust, 2001). Figure 4.2.2 illustrates the changing buoyancy flux for the Swan River estuary for each month. Month 1 shows the average buoyancy flux for January, 2001 based on the evaporation, precipitation and river inflow for the month. The buoyancy flux shows large variation between each month. A seasonal trend is evident, summer values of buoyancy are greater the flux decreases for winter. The month of August has maximum buoyancy flux due to a high stream flow event. Table 8.2.2 in the appendix contains the data that this buoyancy flux was calculated from and the buoyancy fluxes.
Figure 4.2.2 Monthly net buoyancy flux for the Swan River.

Table 8.2.3 in the appendix contains the calculated exchange flow rate and flushing time for the Swan River estuary based on the above buoyancy flux and calculated using the external model. Figure 4.2.3 illustrates the monthly exchange flow rates for the Swan River based on external and internal models. The values for each model are comparable and follow the same general trend. The exchange flow rates are greater during the summer months and then decrease during the winter months. The external model generally estimates greater flushing rates than the internal model.

Figure 4.2.3 The monthly exchange flow rates for the Swan River based on external and internal models.

Figure 4.2.4 illustrates the flushing time estimates based on the internal and external model. The predicted flushing times are similar for August through to March, there is very high
variation between the models for the other months. During these months of high variation, April through to July, the flushing time peaks.

![Monthly Flushing Time of the Swan River](image)

**Figure 4.2.4** The monthly residence times for the Swan River estuary based on the internal and external model.

### 4.2.3 Tidal Influence on exchange flow

The tidal period is 21 hours, the tidal amplitude is 0.8m and the length of the tidal excursion is 3 km (Hamilton, 2001). No monthly variation is taken into account, the tidally forced exchange flow rate is considered constant. $Q_t$ equals 5.71 m$^3$s$^{-1}$.

Figure 4.2.5 shows the variation in the tidal influence parameter $E_t$ with each month in the Swan River estuary for the year 2001. There is considerable variation on a monthly basis, and reasonable evidence of a seasonal trend. The relative importance of the tidal forcing is much greater in winter, decreasing for the summer months where the system is buoyancy forced.
Figure 4.2.5 Monthly Variation of $E_t$ in the Swan River Estuary.

Figure 4.2.6 illustrates the monthly variation of the parameters $G_{rt}A^2$ and $E_t$. For the months of January, August, September and October $G_{rt}A^2$ is high, decreasing slightly for the month of November and then further for December. The remaining months have the lowest values of $G_{rt}A^2$ for this system, ranging from 80-300. The low values of $G_{rt}A^2$ also correspond to the lower values of $E_t$ for this system.

Figure 4.2.6 The hydrodynamic classification of the Swan River on a monthly basis.

4.3.0 The Peel-Harvey Estuary

4.3.1 Exchange Flow estimates using Internal exchange flow

Most of the exchange flow with the ocean in the Peel-Harvey Estuary occurs through the Dawesville Channel. This channel has length 2555m, width 130-200m and height 4.5-6.5m. The Mandurah Channel has dimensions of length 4000m, width 50m and height 1m (Hearn et
al, 1994). The surface area of the estuary and inlet is equal to $1.33 \times 10^8 \text{ m}^2$, and the volume is $1.5 \times 10^8 \text{ m}^3$ (Ranasinghe et al, 2000).

Table 4.3.1 shows the flow parameters for the Peel-Harvey Estuary for the year 2001. The reduced gravitational force is taken near the Dawesville Channel, in the Harvey Estuary. The flushing time is practically determined by the exchange flow through the Dawesville channel, it is of a much greater magnitude than that through the Mandurah channel. The Mandurah channel is entirely internally mixed and the exchange flow is diffusively limited, $G_tA^2$ ranges between 0 and 10 throughout the year and the exchange flow ranges between 0 and $20 \text{ m}^3\text{s}^{-1}$. The exchange flow through the Dawesville channel also experiences strong internal mixing. The mixed layer depth ranges from 62% to fully mixed. The dimensions of the Dawesville channel are taken to be the minimum channel dimensions, to give an upper limit on the residence time in the inlet and estuary. The exchange flow rate and flushing times calculated from these values using the internal model are contained in the appendix, Table 8.3.1.

Table 4.3.1 Peel-Harvey estuary parameters based internal model.

<table>
<thead>
<tr>
<th></th>
<th>g’</th>
<th>u [m s$^{-1}$]</th>
<th>$K_v$ [m$^2$s$^{-1}$]</th>
<th>$G_tA^2$</th>
<th>$\delta/h$</th>
<th>q</th>
</tr>
</thead>
<tbody>
<tr>
<td>January</td>
<td>0.0551</td>
<td>$1 \times 10^{-3}$</td>
<td>$5.0 \times 10^{-5}$</td>
<td>546</td>
<td>0.70</td>
<td>0.64</td>
</tr>
<tr>
<td>February</td>
<td>0.0513</td>
<td>$1 \times 10^{-3}$</td>
<td>$4.9 \times 10^{-5}$</td>
<td>516</td>
<td>0.71</td>
<td>0.64</td>
</tr>
<tr>
<td>March</td>
<td>0.0974</td>
<td>$1 \times 10^{-3}$</td>
<td>$5.2 \times 10^{-5}$</td>
<td>865</td>
<td>0.62</td>
<td>0.68</td>
</tr>
<tr>
<td>April</td>
<td>0.00287</td>
<td>$9 \times 10^{-4}$</td>
<td>$4.2 \times 10^{-5}$</td>
<td>10</td>
<td>1.3</td>
<td>0.32</td>
</tr>
<tr>
<td>May</td>
<td>0.0229</td>
<td>$1 \times 10^{-3}$</td>
<td>$4.5 \times 10^{-5}$</td>
<td>276</td>
<td>0.83</td>
<td>0.58</td>
</tr>
<tr>
<td>June</td>
<td>0.0325</td>
<td>$1 \times 10^{-3}$</td>
<td>$4.7 \times 10^{-5}$</td>
<td>361</td>
<td>0.78</td>
<td>0.61</td>
</tr>
<tr>
<td>July</td>
<td>0.0934</td>
<td>$1 \times 10^{-3}$</td>
<td>$5.2 \times 10^{-5}$</td>
<td>836</td>
<td>0.63</td>
<td>0.68</td>
</tr>
<tr>
<td>August</td>
<td>0.0422</td>
<td>$1 \times 10^{-3}$</td>
<td>$4.8 \times 10^{-5}$</td>
<td>442</td>
<td>0.74</td>
<td>0.63</td>
</tr>
<tr>
<td>September</td>
<td>0.0162</td>
<td>$9 \times 10^{-4}$</td>
<td>$4.3 \times 10^{-5}$</td>
<td>212</td>
<td>0.89</td>
<td>0.55</td>
</tr>
<tr>
<td>October</td>
<td>0.0172</td>
<td>$9 \times 10^{-4}$</td>
<td>$4.4 \times 10^{-5}$</td>
<td>221</td>
<td>0.88</td>
<td>0.56</td>
</tr>
<tr>
<td>November</td>
<td>0.0162</td>
<td>$9 \times 10^{-4}$</td>
<td>$4.3 \times 10^{-5}$</td>
<td>211</td>
<td>0.89</td>
<td>0.55</td>
</tr>
<tr>
<td>December</td>
<td>0.0491</td>
<td>$1 \times 10^{-3}$</td>
<td>$4.9 \times 10^{-5}$</td>
<td>501</td>
<td>0.71</td>
<td>0.64</td>
</tr>
</tbody>
</table>

g’ values from CTD data from the Waters and Rivers Commission, 2001 and NODC data, 2001.
Table 4.4.3 in the appendix shows the exchange flow rates and flushing times of the Peel-Harvey estuary based on the internal model. Figure 4.3.1 shows the monthly variation in the Grashof parameter $G_rA^2$ in the Peel-Harvey estuary. There is no seasonal trend evident, the values are within the transitional region, with the exception of April where the exchange flow is within the VAD limit.

![Graph of monthly variation in $G_rA^2$ for the Peel-Harvey Estuary](image)

*Figure 4.3.1 The variation in $G_rA^2$ in the Peel-Harvey estuary on a monthly basis.*

### 4.3.2 Exchange flow estimates using External model

The buoyancy flux was calculated on a monthly basis from precipitation and evaporation data from the Bureau of Meteorology and river inflow data from the Waters and Rivers Commission. The nearest rainfall station to Mandurah is at Wokalup and the nearest evaporation centre is from Roelands, therefore the actual precipitation and evaporation experienced by the Peel-Harvey Inlet may vary slightly. The reduced gravitational force is $g'=0.239$.

Figure 4.3.2 illustrates the changing buoyancy flux for the Peel-Harvey estuary for each month. Month 1 shows the average buoyancy flux for January, 2001 based on the evaporation, precipitation and river inflow for the month. The buoyancy flux shows a seasonal trend, it is greater in summer, decreasing for the winter months. Table 8.3.2 in the appendix shows the data the river inflow, evaporation and precipitation values and the buoyancy values.
Figure 4.3.2 Monthly net buoyancy flux for the Peel-Harvey Estuary.

Table 8.3.3 in the appendix contains the exchange flow rate and flushing times based on the external model. Figure 4.3.3 illustrates the monthly exchange flow rates for the Peel-Harvey estuary based on external and internal data models. The internal model generally has a greater exchange flow rate, with the exception of April, which is the minimum exchange flow estimated by either model for any month. The exchange flow predicted by the models is very similar for September through to November. The difference between the model predictions is greatest in winter and is relatively high, up to 175 m$^3$ s$^{-1}$. The internal model shows much more monthly variation than the external model.

Figure 4.3.3 Monthly exchange flow rates for the Peel-Harvey estuary based on external and internal methods of calculation.
Figure 4.3.4 illustrates the flushing time estimates based on the internal and external model. The flushing time peaks for the internal model in April, this is much greater than the rest of the year. The difference between the two models is largest in winter, peaking in April.

![Monthly Flushing Times for the Peel-Harvey Estuary](image)

**Figure 4.3.4 Monthly flushing times of the Peel-Harvey estuary based on the external and internal models.**

### 4.3.3 Tidal Influence on exchange flow

The tide is diurnal, the tidal amplitude is 0.4m and the length of the tidal excursion is 5 km (Ranasinghe and Pattiaratchi, 2000). No monthly variation is taken into account, the tidally forced exchange flow rate is considered constant. $Q_t$ equals 4.97 m$^3$s$^{-1}$.

Figure 4.3.5 shows the variation in the tidal influence parameter $E_t$ with each month in the Peel-Harvey estuary for the year 2001. For each month the value of $E_t$ is much greater than 5, indicating that the system is dominated by baroclinic forcing. The month of April is an anomaly, the tidal influence on exchange flow is greatest, although $E_t$ is still greater than 5.
Figure 4.3.5 Monthly Variation of $E_t$ in the Peel-Harvey Estuary.

Figure 4.3.6 illustrates the monthly variation of the parameters $\text{GrtA}^2$ and $E_t$. The values of $\text{GrtA}^2$ are within the transitional region, with the exception of April, which is diffusively limited. The values of $E_t$ are always greater than 5, influence of buoyant forcing on the exchange flow is much greater than that of tidal forcing.

Figure 4.3.6 The hydrodynamic classification of the Peel Harvey Estuary, calculated on a monthly basis.
4.4.0 Shark Bay

4.4.1 Exchange flow estimates from internal and external model

Due to the presence of the Faure Sill most of the exchange flow occurs through a channel with dimensions; height (h) 6m, width (b) 2000m and length (L) 20 000m (Burling et al, 1999). The surface area of Hamelin Pool is (BL) $10^9$ m$^2$ and the volume of Hamelin Pool is $4 \times 10^9$ m$^3$.

The exchange flow based on external parameters for Hamelin Pool is calculated on an annual basis, as there is little seasonal variation in the salinity range of the entire bay area (Burling et al, 1999). Herald Loop is the channel between Hamelin Pool and Hopeless Reach, it is vertically well mixed, with a linear gradient in salinity. The density difference through the channel is $\Delta \rho = 8$ kgm$^{-3}$ (Burling et al, 1999). Due to this density difference the reduced gravitational force equals $g^\prime = 0.075$.

Based on an annual evaporation and precipitation rate the buoyancy flux out of the system can be calculated. The average precipitation, $P = 200$mm/year, the average evaporation, $E = 2000$ mm/year (Burling et al, 1999). This is equal to an evaporation rate, $E = 5.7 \times 10^{-8}$ ms$^{-1}$.

Hamelin Pool experiences strong wind and tidal mixing, following Burling et al (1999) estimates of the mean turbulent eddy viscosity; $K_v = 1 \times 10^{-3}$.

Based on an annual evaporation and precipitation rate the buoyancy flux out of the system can be calculated. The average precipitation, $P = 200$mm/year, the average evaporation, $E = 2000$ mm/year (Burling et al, 1999).

The total evaporation over a yearly period $E = 2 - 0.2 = 1.8$ m/year.

This is equal to an evaporation rate, $E = 5.7 \times 10^{-8}$ ms$^{-1}$.

<table>
<thead>
<tr>
<th>Flow Parameters</th>
<th>External Model</th>
<th>Internal Model</th>
</tr>
</thead>
<tbody>
<tr>
<td>$G_e A^2$</td>
<td>-</td>
<td>5.4</td>
</tr>
<tr>
<td>$\delta/h$</td>
<td>-</td>
<td>2.2</td>
</tr>
<tr>
<td>$Q$ (m$^3$s$^{-1}$)</td>
<td>1525</td>
<td>1860</td>
</tr>
</tbody>
</table>
4.4.2 Tidal Influence on exchange flow

The maximum tidal current \( u \) is equal to the tidal amplitude divided by the tidal period, \( u = 1 \text{ ms}^{-1} \). \( L \) is the length of the tidal excursion, \( L=5000\text{m} \), and \( b \) is the width of the channel, \( b=2000\text{m} \) (Burling et al, 1999). \( E_t \) is \( 2\times10^{-4} \).

This value of \( E_t \) indicates that Shark Bay is dominated by tidal forcing rather than buoyancy forcing. Therefore considering only the effects of tidal forcing and buoyancy forcing the exchange flow rate can be approximated by the expression for barotropic flux.

4.5.0 Summary

Figure 4.5.1 illustrates the dominant hydrodynamic forcing for Wilson Inlet, the Swan River, the Peel-Harvey estuary and Hamelin Pool. It is evident from this diagram that the dynamic regime of the systems alter on a seasonal basis and the hydrodynamics of each basin are significantly different. \( G_{rt}A^2 \) is plotted on a logarithmic scale as the range of values is large. The Swan River and Peel-Harvey Estuary show similarity in the exchange flow dynamics, the relative influence of the buoyant to diffusive forcing is within the same range through out the year. The strength of the tidal forcing is greater in the Swan River during the winter months. Wilson Inlet is represented by the two extreme cases, the winter conditions is close to the hydraulic limit, there is little internal mixing and the summer has strong internal mixing, the tidal forcing is weak compared to the buoyant forcing for both conditions. Hamelin Pool shows the greatest variation between the sites, it has very strong internal mixing, strong tidal forcing and the system is diffusively limited.
5.0 Discussion

Using hydraulic estimates in the external model provides an upper bound estimate of the exchange flow rate. As the intensity of the vertical mixing increases, the magnitude of the horizontal flux decreases (Winters and Seim, 2000). So assuming that the system is hydraulically limited will give an overestimate of the exchange flow rates and a smaller residence time than will actually occur in the estuary. Therefore it can be expected that the external model calculations will provide an upper bound on the exchange flow rate as the model assumes the system is hydraulically limited with no mixing between the two layers in the channel. The internal model takes into account the viscous diffusive forcing and the extent of internal mixing; therefore this exchange flow rate is expected to be a more realistic estimation. However, in the applications of the models the internal model does not automatically give a significantly larger estimate of residence times than the internal model, despite strong mixing in the channel. Implying that even though some mixing is occurring, most systems are in practice, much closer to the hydraulic limit than the diffusive limit.

Another factor contributing to the difference between the exchanges flow rates predicted by the models are the different theories behind the calculations. The internal model uses the actual density difference between the ocean and the inlet entrance as the forcing of the exchange flow rate. Compare this to the external model which links the buoyancy fluxes in the system to the exchange flow rate. The ocean is considered an infinite basin and therefore the density of the system is equal to ocean. Evaporation, precipitation, river inflow, freezing and melting of ice can all cause a buoyancy flux in the inlet. The buoyancy flux is considered to act equally over the entire surface of the inlet, the mechanisms causing density differences. The internal mechanisms spreading the buoyancy deficit/gain and thus altering the density of the inlet are not considered. The model simply assumes this buoyancy forcing is mixed in the inlet and causes a density difference that forces the exchange flow with the ocean.

The internal data model uses an estimate of vertical diffusivity, which determines the vertical thickness of the intermediate layer. This is a source of some speculation, as it is not an easily estimated parameter. The iterative process estimates the vertical diffusivity as per Fisher et al (1979), for Wilson Inlet, the Swan River and the Peel-Harvey estuary, and the variation in the vertical diffusivity can lead to a large difference in the mixed layer depth, non-dimensional
exchange flow rate and the Grashof parameter. This is evident from the case study on Shark Bay, where the magnitude of the vertical diffusivity is greater.

5.1.0 Wilson Inlet

5.1.1 Comparison of the internal and external model results

Comparison of the exchange flow rates for the internal and external model calculations for Wilson Inlet can only be made for the months of July, August and December. These months represent the two extremes of conditions in the inlet. In July the inlet has just opened, the channel dimensions are greatest and the inlet volume is at its maximum due to winter stream flow and the closure of the bar. July and August are representative of winter conditions. December is illustrative of summer conditions, where the inlet has been opened for the maximum time possible and the channel dimensions are greatly diminished, the density difference between the ocean and the inlet is decreased due to the exchange flow throughout the year and high evaporation. There is very little difference between the exchange flow rates predicted by the internal and external model for the three months. This is interesting to note, as in the month of July the system has not been at steady state at all, there has been no stabilizing exchange flow between the ocean and the inlet, and therefore one of the underlying assumptions the external model is based on is not strictly valid. The month of December is also important, as the channel experiences strong internal mixing; therefore the hydraulically limited assumption of the external model is not valid. Despite these assumptions, the external model still predicts exchange flow rates very similar to that of the internal model. The flushing times for July and August are similar and there is little difference between the two models. The flushing time for December is greatly increased, the reduced channel dimensions hinder the exchange flow. The difference between the model’s flushing time predictions is very large; this is a result of the small magnitude of exchange flow rate. A difference in exchange flow rates of 1 m$^3$s$^{-1}$ corresponds to a difference in flushing time of 600 days. The flushing times predicted by both models are realistic, within the range calculated by Ranasinghe (1995) from 35 days to greater than 1 year.

5.1.2 The hydrodynamic classification

The classification of Wilson Inlet based on the seasonal values of $G_tA^2$ and $E_t$ show a varying range of hydraulic conditions in the estuary. The Grashof number times the aspect ratio squared (referred to for simplicity as the Grashof parameter) shows a wide range of forcing
between the two seasons. The summer conditions show strong internal mixing is occurring and the exchange flow is dominated by the diffusive forcing rather than the buoyant forcing. For the winter conditions the exchange flow is close to the hydraulic limit, there is only weak internal mixing and the buoyant forcing is much greater than the diffusive forcing. The strength of the tidal forcing is greater in winter than summer, throughout the entire year the system the dominant forcing is buoyant rather than tidal, although during the winter condition the value of $E_t$ is only just outside of the transitional region. The strength of tidal influence has a strong correlation with the magnitude of the channel dimensions.

There is large seasonal variation in the hydrodynamic forcing of Wilson Inlet. The winter exchange flow can be considered buoyancy dominated, the summer exchange flow is dominated by diffusive forcing and characterized by strong internal mixing in the channel.

5.1.3 Hypothetical dredging scenarios

When considering the internal model in a predictive sense it is important to note the results are only useful in a comparative sense. The exchange flow rates predicted are not accurate as the density difference between the inlet and the ocean cannot be predicted. The values used are representative of current conditions, and therefore not applicable in an inlet with a permanent opening and increased exchange flow. The internal model is useful in indicating the trends in the residence time for increasing channel width and length as compared to current residence times. The channel length is incorporated in the model through the aspect ratio, $a=h/L$, the width is incorporated directly into the exchange flow rate equation. The height of the channel is not altered for these dredging scenarios, as the maximum change is approximately 1m. The exchange flow rate shows an increase with increasing channel width and length. The exchange flow increases with increasing channel length, however when the channel length is dramatically increased the exchange flow actually decreases. The length is equal to 2000m, the model does not take into account that the channel is then connected to the main basin, which would increase the exchange flow rate. The width of the channel has a greater effect on the exchange flow rate than the length. The maximum exchange flow rate occurs for the maximum width possible, 500m, where the entire channel is dredged.

The external model is of greater use in a predictive sense, as the buoyancy flux forcing the exchange flow is an external parameter that is unchanged within the basin. The external model shows no variation in residence time with changing channel length although the residence time decreases with increasing channel width. The maximum channel width, 500m,
corresponds to the maximum exchange flow rate. While the external model does not account for the dampening effect of internal mixing it still adequately predicts the exchange flow rate for Wilson Inlet, for conditions of strong and weak internal mixing. Therefore the flushing times can be considered representative of the exchange flow rates.

5.2 The Swan River

5.2.1 Comparison of the internal and external model results

The buoyancy flux for the Swan River is relatively constant over the summer months, which a period of high evaporation. A large decrease in buoyancy fluxes for the months of May and June is seen due to the low evaporation, low precipitation and little river inflow into the Swan River. A maximum stream flow event in August causes the peak buoyancy flux for the year.

The monthly exchange flow rates estimated by the external and internal model appear to follow the same general trend. The external model exchange flow rates are generally higher than that of the internal model. The exchange flow rate differences between the two models are quite high, ranging from a difference of 5 m$^3$s$^{-1}$ to 80 m$^3$s$^{-1}$. This considerable difference for each month is related to the strength of the internal mixing in the channel, the mixed layer depth ranges between 47% of the channel depth to being entirely mixed. The months where the internal model is very similar to or greater than the external model the mixed layer has minimum values. Therefore the influence of the internal mixing and subsequent dampening of the exchange flow is minimized. The external exchange flow model over predicts the exchange flow to a lesser extent when the internal mixing is weaker and the models predictions then show greater similarity. The greatest difference between the two models occurs for the month of April, when the Swan River is diffusively limited, the water column is entirely mixed. Therefore the exchange flow rate for the external model can be assumed to be over predicting by a large amount, the difference being a flow rate of 80 m$^3$s$^{-1}$. This difference between the internal and external model is much less than that of Shark Bay, 335 m$^3$s$^{-1}$ and much greater than that of Wilson Inlet 1.35 m$^3$s$^{-1}$, the difference is relative to the magnitude of the exchange flow estimates for both models. The difference in the flushing time is evidence of the comparability of two models, as the magnitudes of the systems are accounted for.

The monthly flushing times for the Swan River are comparable for the internal and external model for August through to March. The differences between the residence times for each
model range between 8 days to 1055 days. The months of April and July show similar estimates in flushing times, 1055 and 961 days, this is despite the exchange flow rate difference being 40 m$^3$s$^{-1}$. This indicates that the exchange flow rate differences between the models become larger with increasing mixed layer depth, yet this does not directly translate into an increase in the flushing times. The flushing time increases are much more dramatic for smaller exchange flow rates. Therefore an equal difference in the exchange flow rates between the models for a high flow rate and low flow rate does not translate into an equal flushing time difference.

5.2.2 The hydrodynamic classification

The Grashof parameter shows high seasonal variation, it ranges within the transitional hydrodynamic region defined by Hogg et al (2001). The channel has significant internal mixing throughout the year, the winter months are close to the VAD limit with strong internal mixing, the summer condition has greater values of $G_ntA^2$. The relative strengths of buoyant and diffusive forcing vary throughout the year, as does the strength of the tidal forcing. During the winter period the system is in a transitional region, where the forcing is a combination of tidal and buoyant forcing. The summer months are dominated by buoyancy rather than tidally forced exchange flow.

5.3 Peel-Harvey Estuary

5.3.1 Comparison of the internal and external models

The monthly buoyancy flux shows considerable variation on a seasonal basis, decreasing from the summer period maximum to a minimum value in June and then gradually increasing again and directly related to the evaporation rate. The mixed layer is significant throughout the entire year, ranging from 65% of the water column to being entirely internally mixed.

The magnitude of the exchange flow through the Mandurah channel is such that it is of negligible effect when considering the flushing time estimates from the models. The maximum flow rate is 20 m$^3$s$^{-1}$, compared to the maximum 260 m$^3$s$^{-1}$ through the Dawesville channel. The flow through the Mandurah channel has little impact on the residence time in the estuary and inlet, therefore for simplicity the exchange flow is only considered through the Dawesville channel.
The internal and external model exchange flow rates shown significant differences for this system. The high salinity of the Peel-Harvey Inlet originates over long shallow flats adjacent to the entrance of the channel and intrudes into the bottom of the canal as gravity current (Humpries et al, 1983). This isn’t reflected in the buoyancy forcing for most months, contributing to the external model calculating a much lower exchange flow rate, with differences up to 175 m$^3$s$^{-1}$. The external model overestimates the exchange flow as it does not take into account the dampening effect of internal mixing in the channel. The effect of internal mixing in the channel is particularly prominent for the months of September through to November and April through to June, interestingly these months show the greatest resemblance between the models. The greatest difference between the two models is experienced in the winter period, where the buoyancy forcing is at a minimum and the strength of internal mixing is at a minimum. It is interesting to note that the predictions of the two models are very similar for these months where the internal mixing in the channel is particularly prominent. The external model shows much less seasonal variation than the internal model. The internal model shows great monthly variation and no particular seasonal trend.

The flushing times of the inlet and estuary are usually calculated separately, the Peel Inlet has a cited residence time of between 10 and 30 days, the Harvey Estuary has a flushing time of between 17 and 50 days (Hodgkin, 1998). The flushing times predicted by the internal and external model adequately represent the flushing times in the inlet and estuary. The predicted exchange flow is between 10 and 80 days for the inlet and estuary. The models show highly comparable results for the summer months, there is greater variation in the winter months, with flushing time differences between 5 to 10 days. However considering the use of this model as a rough tool these predictions are reasonable.

5.3.2 The hydrodynamic classification

The hydrodynamic classification shows some seasonal variation. The parameter $G_t A^2$ indicates that the dominant forcing is within the transitional region for the majority of the year, therefore the exchange flow forcing is a combination of buoyant and diffusive forcing. The month of April is an anomaly, whereby the system is diffusively limited and has higher tidal influence. The strength of the tidal influence is weak compared to the buoyant forcing throughout the year. The hydrodynamic classification of the exchange flow forcing in the Peel-Harvey estuary can be described as a combination of buoyant and diffusive forcing, the relative magnitudes of each showing monthly variation.
5.4 Shark Bay- Hamelin Pool

5.4.1 Comparison of the internal and external model

The exchange flow rates between Hamelin Pool and Hopeless Reach are large and there is also a large difference between the internal and external modelling. The high exchange flow this translates into a small flushing time difference between the two models. The flushing time differs by 7 days and the external flushing time is greater. The external model assumes that the system is hydraulically limited, this is clearly a poor approximation as the system lies well within the diffusive limit. The entire water column is internally mixed, despite this dampening effect on the exchange flow, the external model still predicts a greater flushing time than the internal model.

5.4.2 The hydrodynamic classification

The water column in the channel is entirely internally mixed, therefore the system is within the VAD limit. The diffusive forcing dominates rather than the buoyant forcing. Comparing Hamelin Pool’s tidal influence to buoyancy forcing shows that the system is entirely dominated by the barotropic exchange flow. Therefore the hydrodynamics of the system can be described as a combination of diffusive and tidally driven exchange flow.

5.5 Summary

The magnitude of the exchange flow flushing time estimates in both the internal and external model. Small differences in exchange flow rates estimated by the models correspond to large differences in flushing times when the exchange flow rates are small. However larger differences between the exchanges flow rates estimated by the internal and external model correspond to smaller differences in the flushing time difference when the magnitude of the flow rates is larger. Therefore caution is necessary when using the external model flushing time estimates for systems with low exchange flow rates. Despite the over-prediction of the exchange flow rate by the external model this does not always correspond to the internal model having a higher residence time than the external model.
The strong seasonal variation within the hydrodynamic classification of the estuaries and between the estuaries, corresponds to variance in the dominance of the buoyant, diffusive and tidal forcing of the exchange flow. The influence of the tide on exchange flow appears to be linked closely to the magnitude of the width of the channel. This is evident from the Wilson Inlet case study, where the value of $E_t$ varies greatly depending on the channel dimensions. Also for the study of Hamelin Pool, this is the only system where the exchange flow is tidally dominated- it has a much greater channel width than any of the other applications. The Swan River and the Peel-Harvey estuary have channel dimensions similar in magnitude, and there is a corresponding similarity in the hydrodynamic range, however the Swan River has greater tidal influence during winter. The dimensions of the topographic constriction play an important role in the prediction of the exchange flow rates and the dominant hydrodynamic forcing. In the tidal transitional flow regime the observed flux is greater than predicted by steady state hydraulic theory as there is two mechanisms of exchange, tidal forcing and buoyancy induced flow (Phu, 2001).

The internal and external models give a good indication of the residence times and exchange flow rates through out the year in the systems they have been applied too. Despite the simplifying assumption of the hydraulically limited system, the external model does not greatly overestimate exchange flow rates in practice. The internal and external models correlate reasonably well and are useful in considering the dynamics of the systems. The flushing time estimates improve as the exchange flow rates increase in magnitude. Together with the classification system an overall picture of the system is depicted, from which the dominant exchange mechanisms can be determined as well as perceiving the focus for more complex modelling using numerical circulation models.
6.0 Conclusion

The internal and external models calculate the exchange flows and residence times based on fundamentally different underlying assumptions. The internal model uses field data and incorporates the strength of internal mixing in the channel and the dampening effect this has on the exchange flow. The external model assumes the system is at steady state and no internal mixing occurs, the forcing is calculated from external parameters. The application of these models was to four estuarine systems of differing channel magnitudes, buoyancy forcing, densities, location and volume. The models show varying degrees of discrepancy in their results between each system and within each system seasonally. The residence times predicted by both models in Wilson Inlet and Shark Bay show superior similarity compared to the Peel-Harvey estuary and the Swan River. Considering the use of these models to give basic understanding of the processes and timescales in the estuarine system it can be concluded that the models predict the exchange flow and residence time in the basins adequately.

The hydrodynamic classification took into account the relative dominance of buoyant, diffusive and tidal forcing in the estuary. The relative magnitudes of the buoyant and diffusive forcing are dependant on the strength of internal mixing in the channel, which is characterised by the value $G_{n}A^2$. The tidal forcing relative to the buoyant forcing in the system is characterised by the non-dimensional parameter $E_t$. The application of this classification system to Wilson Inlet, the Swan River estuary, the Peel-Harvey estuary and Hamelin Pool showed hydrodynamic variation between each of the systems and within each of the systems on a seasonal basis. The hydrodynamic classification can be considered useful in terms of understanding of the dominant exchange mechanisms of the system and determining the focus for more complex numerical circulation models.
7.0 References


Garret, C., Bormans, M. & Thompson, K. 1990. *Is the exchange through the Straits of Gibraltar maximal or sub maximal?* In the Physical Oceanography of Sea Straits 271-294. NATO ASI Series, Kluwer.


8.0 Appendix

Table 8.1.1 The exchange flow rate and flushing time for Wilson Inlet using the internal model.

<table>
<thead>
<tr>
<th></th>
<th>Q[m$^3$s$^{-1}$]</th>
<th>T [days]</th>
</tr>
</thead>
<tbody>
<tr>
<td>July</td>
<td>23.53</td>
<td>77.4</td>
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<tr>
<td>August</td>
<td>24.97</td>
<td>50.7</td>
</tr>
<tr>
<td>December</td>
<td>1.16</td>
<td>848.3</td>
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</tbody>
</table>

Table 8.1.2 contains the evaporation rate (E), buoyancy flux due to evaporation and precipitation (B$_o$). The river inflow (R) and the river inflow distributed over the surface area of the inlet (U), the buoyancy flux due to the river inflow and the total buoyancy flux.

Table 8.1.2 Wilson Inlet buoyancy flux.

<table>
<thead>
<tr>
<th></th>
<th>E [ms$^{-1}$]</th>
<th>B$_o$ m$^2$s$^{-3}$</th>
<th>R [m$^3$s$^{-1}$]</th>
<th>U [ms$^{-1}$]</th>
<th>B$_o$ [m$^2$s$^{-3}$]</th>
<th>Tot B$_o$</th>
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</thead>
<tbody>
<tr>
<td>Jan-95</td>
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<td>1.17×10$^{-8}$</td>
<td>0.035</td>
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</tr>
<tr>
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<td>9.3×10$^{-9}$</td>
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<tr>
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<td>2.82×10$^{-8}$</td>
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<td>1.83×10$^{-9}$</td>
<td>4.4×10$^{-10}$</td>
<td>1.3×10$^{-8}$</td>
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</tbody>
</table>


** R- Waters and Rivers Commission, 1994/95
Table 8.1.3 External model predictions of flushing time and exchange flow for Wilson Inlet.

<table>
<thead>
<tr>
<th></th>
<th>Q [m$^3$s$^{-1}$]</th>
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<tr>
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Table 8.2.1 Exchange flow and flushing times for the Swan River estuary using internal model.

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Table 8.2.2 Buoyancy forcing for the Swan River Estuary

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<th>$B_o$ m$^2$s$^{-3}$</th>
<th>$R$ m$^3$s$^{-1}$</th>
<th>$U$ [ms$^{-1}$]</th>
<th>$B_o$ [m$^3$s$^{-3}$]</th>
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<td>1.2×10$^{-7}$</td>
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Table 8.2.3 The exchange rate and flushing time of the Swan River Estuary calculated using external model.

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<th>Tₑ [days]</th>
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<td>June</td>
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<td>November</td>
<td>97.4</td>
<td>89.9</td>
</tr>
<tr>
<td>December</td>
<td>101.5</td>
<td>86.3</td>
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Table 8.3.1 The exchange flow rate and flushing time for the Peel-Harvey estuary using the internal model.

<table>
<thead>
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<th>2001</th>
<th>Qₜ [m³s⁻¹]</th>
<th>Tₑ [days]</th>
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<tr>
<td>April</td>
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<td>80.4</td>
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<tr>
<td>May</td>
<td>109.7</td>
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<td>June</td>
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<td>12.7</td>
</tr>
<tr>
<td>July</td>
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<td>6.7</td>
</tr>
<tr>
<td>Month</td>
<td>E [m$^3$s$^{-1}$]</td>
<td>B$_o$ m$^2$s$^{-3}$</td>
</tr>
<tr>
<td>------------</td>
<td>------------------</td>
<td>---------------------</td>
</tr>
<tr>
<td>January-95</td>
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<td>1.0×10$^{-7}$</td>
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<tr>
<td>Feb-95</td>
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<tr>
<td>Dec-94</td>
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<td>6.5×10$^{-8}$</td>
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</table>

* *R- Waters and Rivers Commission, 2001

### Table 8.3.3 The exchange flow rate and flushing time for the Peel-Harvey Estuary based on the external model.

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<th>Year</th>
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<td>Value2</td>
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<td>---------</td>
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