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2 ESTUARINE CIRCULATION

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12 Definition

13 An estuary is a semi-enclosed coastal body of water which
14 has a free connection with the open sea and within which
15 salt water is measurably diluted with fresh water derived
16 from land drainage.

17 Introduction

18 Estuaries are unique aquatic environments that share
19 terrestrial and marine contributions, creating a set of
20 interesting physical characteristics including bimodal
21 circulation patterns (fresh water flowing along the surface
22 and saline water entering along the bottom) and mixing
23 processes induced by tides. To a lesser extent, continental
24 shelf processes such as storm surges and wind-driven
25 circulation can also influence circulation in estuaries.
26 At first glance, most estuaries are governed by tidal
27 oscillation and influenced by river discharges, so
28 mixing processes account for most of the hydrodynamic
29 characterization of estuaries. The resulting velocity
30 and salinity fields have strong time variability, and
31 the non-steady-state current patterns may be altered by
32 sediment erosion, transport, and sedimentation, which
33 contribute to bathymetric and margin changes.

Formation and geological age

34

35 Estuaries are transitional environments that were formed
36 in the narrow coastal boundary zone between the ocean
37 and the continents during the Holocene rise in sea level
38 which began about 15,000 years ago, when the sea was
39 about 125 m below its present level. They were formed
40 due to the Flandrian Transgression, the secular increase
41 of global sea level that ended about 7,000 years ago. Local
42 changes in sea level as well as tectonic activity may have
43 also contributed to the formation of estuaries. At the end
44 of that period, the continental coastal plains and the river
45 valleys were slowly flooded forming the bays, coastal
46 lagoons, inlets, and estuaries.

47 The geomorphology of estuaries depends on coastal
48 topography, geomorphology, continental shelf character-
49 istics, and river discharges, which change due to natural
50 processes of erosion and sediment deposition. These pro-
51 cesses have often been modified due to the anthropogenic
52 activities in estuaries and in their drainage basins.

53 Recent studies conclude that during the late Holocene
54 sea-level rise amounted to 5–7 m. This was corroborated
55 by the earlier work of Fairbridge (1961). Villwock
56 (1972), Suguio and Martin (1978), and Kowsmann
57 et al. (1977) also confirmed this phenomenon in the south
58 and southeastern coastal plain of Brazil.

59 Transitional environments associated with estuaries,
60 such as salt marshes and mangroves, are colonized by
61 the salt-tolerant species and distributed in regions of tem-
62 perate and tropical climates. They are environments ulti-
63 mately dependent on the dynamics of estuarine systems,
64 but have been strongly affected by human activities
65 around the world Herz (1992). As such, their areas have
66 declined significantly. Some have been protected by
67 reserve status since the nineteenth century. Human
68 impacts and their interference on estuaries and coasts are
69 described in French (1997).

Herz (1992) used digital mapping, based on processing of multispectral images, to describe the mangrove colonization area along the 8,000 km of the Brazilian coast. The mangrove area decreased from 85.0 % in the northern coast of Brazil to 9.9 % and 5.1 % in the eastern and southern coastal regions.

Pioneer investigations

In 1875, the Swedish researcher F. L. Ekman performed experiments in the Götaelf River Estuary (Sweden) to investigate estuarine processes (quoted in Defant (1961), 539). From the analysis of the salinity distribution on August 1875, Ekman observed that a strong outflowing current could not be compensated only by river discharges, but instead by a compensating upward motion (upwelling) of saline water from the bottom.

Upwelling is a vertical advective motion that breaks the continuity of the bidirectional motion field observed by Ekman. The vertical motion was calculated theoretically under steady-state conditions for the first time by Pritchard (1954), solving the salt balance equation using extensive experimental data from the James River Estuary (Virginia, USA) during the summer of 1950.

Ekman investigated the occurrence of currents up-estuary in the lower layers of the estuary in a paper published in 1899, quoted in Defant (1961, 539). Although this paper was only descriptive, Ekman was aware that the river discharge input and the salinity stratification were the main forcing mechanisms of the observational results.

It was only in 1952 – almost 60 years later – that D. V. Pritchard became the first researcher to link estuarine circulation to the forcing by the horizontal density gradient. He used observations from the James River estuary to demonstrate this mechanism, quoted in Geyer (2010, 13).

Definitions and variability

Ketchum (1951) focused on the exchanges of fresh and salt water in tidal estuaries and defined an estuary as “a body of water in which the river water mixes and measurably dilutes sea water.” The most classical definition is that stated by Pritchard (1952) and Cameron and Pritchard (1963), “An estuary is a semi-enclosed coastal body of water which has a free connection with the open sea and within which salt water is measurably diluted with fresh water derived from land drainage.” Several other definitions are found in the literature. Perillo (1995) defined an estuary as “a semi-enclosed coastal body of water that extends to the effective limit of tidal influence, within which sea water entering from one or more free connections with the open sea, or any other saline coastal body of water, is significantly diluted with fresh water derived from land drainage, and can sustain euryhaline biological species from either part or the whole of their life cycle.” These definitions along with others by Dionne (1963), Kjerfve (1987), and Dalrymple et al. (1992) have the

ability to include all basic marine disciplines, which are necessary for studying this transitional water body as an ecosystem. There are interchanges between three estuarine domains: the tidal river (TR), the mixing zone (MZ), and the estuarine plume (EP), which must be investigated together as a unique transitional system Okubo (1970).

In these definitions, salt-water dilution is due to mixing processes, which involve advection (large-scale) and diffusion (small-scale). These occur internally in the estuary and tend to produce uniformities in the concentration of dissolved properties (Bowden, 1963; Okubo, 1970). The small-scale component of mixing is generated by surface (due to the wind stress) and bottom shears and the vertical internal shears.

The estuarine drainage basin and the fresh water discharge (Q_f) into the estuary may change as a result of natural processes and human activity, such as inadequately planned land use and pollutants from homes, farms, and factories. According to Ji (2008), every surface water system is unique and many face similar environmental problems such as eutrophication, pathogen contamination, toxic chemicals, loss of habitat, and declines in fish and wildlife. Estuaries also have problems that cause declines in water quality, living resources, and overall ecosystem health. Rivers, lakes, and estuaries contain a very small fraction of the total earth's total water budget driven by the hydrologic cycle.

Circulation variability in estuaries is classified as intertidal when it occurs at semi-diurnal or diurnal tidal frequencies (>1 cycle/day), or subtidal at lower frequencies (<1 cycle/day). The beat period of the main semi-diurnal tidal lunar and solar components (M_2 and S_2) results in fortnightly subtidal frequency (modulation between successive spring tides, ≈ 15 days).

Classification of estuaries

Advection and mixing processes can be used to compare and classify different estuaries. The major classification schemes are based on salinity stratification, circulation, and mixing. Figure 1 shows a simple steady-state dynamic balance analysis of the interaction of these processes based on a longitudinal section presented by Geyer (2010).

The first estuary classification was suggested by Stommel (1951) taking into account the main forces as tides, fresh water discharge, and wind. Raritan, Pamlico Sound, and Mississippi river estuaries of New Jersey, North Carolina, and Louisiana (USA) have as primary forcings the tide, wind, and river, respectively. The important characteristic of these estuaries is the vertical salinity stratification. The Raritan River estuary, which is the shallowest one, is nearly vertically homogeneous. The Mississippi River estuary is the deepest, being forced by micro-tides.

Pritchard (1952) introduced a classification scheme based on the main geomorphologic estuarine features. Four types of estuaries were differentiated: (1) coastal

181 plain estuaries or drowned river valleys; (2) bar built estu-
182 aries; (3) fjords, typically in higher latitude regions; and
183 (4) tectonic estuaries.

184 Taking into account vertical salinity stratification,
185 Stommel (1953) suggested the following classification:
186 (1) salt wedge estuaries, river discharge dominates with
187 vertical mixing absent; (2) fjords, deep estuaries (several
188 hundred meters) characterized by a highly stratified upper
189 layer; and (3) moderately and highly stratified estuaries,
190 dependent on the intensity of the vertical mixing and
191 establishment of a steady-state bidirectional circulation.
192 This classification was described by Pritchard (1955),
193 Cameron and Pritchard (1963), and Schubel and Pritchard
194 (1972), taking into account improvements related
195 to dynamic conditions due to estuarine circulation and
196 salinity stratification.

197 During the 1960s, estuarine classifications were based
198 on quantitative criteria. Ippen and Harleman (1961)
199 introduced the stratification number, based on laboratory
200 experimental results, and Hansen and Rattray (1966)
201 introduced the stratification and circulation parameters.

202 Hansen and Rattray (1965) were the first investigators
203 to link estuarine classification with its physical properties.
204 They used salinity stratification and circulation in
205 a theoretical development based on the analytical solu-
206 tions of a bi-dimensional system of equations (motion,
207 mass and salt conservation, and a linear state equation)
208 under steady-state conditions, applied to a laterally homo-
209 geneous estuary. The results were vertical profiles of the
210 longitudinal velocity and salinity used for a theoretical
211 deduction of the stratification-circulation diagram
212 (Hansen and Rattray, 1966). The coordinates (axis) of this
213 diagram are the stratification (p_e) and the circulation (p_c)
214 parameters defined by $p_e = \frac{S_r - S_s}{\bar{S}} = \frac{\delta S}{\bar{S}}$ and $p_c = \frac{u_s}{u_f}$, respec-
215 tively, where S_f and S_s are the time mean salinities on the
216 bottom and surface, respectively, and $\bar{S}$ its mean-depth
217 value. The velocity u_s is the time mean value on the
218 estuary surface and u_f is the one generated by the river
219 discharge (Q_f) defined by $(u_f = \frac{Q_f}{A})$, where A is the
220 cross-sectional area.

221 The stratification (p_e) and circulation (p_c) parameters
222 are linked to a third parameter v defined by $v = \frac{\phi_D}{\phi_D + \phi_A}$,
223 where ϕ_D and ϕ_A are the salt flux into the estuary due to
224 turbulent diffusion and the advection, respectively. When
225 $\phi_D \gg \phi_A$, it follows that $v \rightarrow 1$ and the salt flux are
226 dominated by the tidal mixing; for $\phi_D \ll \phi_A$, $v \rightarrow 0$ and
227 the salt flux is dominated by advection (gravitational
228 circulation). Then, this parameter varies from 0 to 1, and
229 $v = v(p_c, p_e)$ is expressed theoretically by Miranda
230 et al. (2012):

$$(p_e)^{-1} [210 + 252(p_c - 1.5)] v^2 + [32 - (p_e)^{-1} (210 + 252(p_c - 1.5)) + 76(p_c - 1.5) + \frac{152}{3} (p_c - 1.5)^2] v = 0, \quad (1)$$

For $v = 0$ (1) has no physical meaning, and for $v = 1$ the
salt transport is due to the turbulent diffusion only.
The equation is thus reduced to:

$$32 + 76(p_c - 1.5) + \frac{152}{3} (p_c - 1.5)^2 = 0 \quad (2)$$

This equation with the unknown ($p_c - 1.5$) has no solution
in the field, unless the constant 32 is disregarded. If so, it
has two solutions: $p_c = 0$ (with no physical meaning)
and $p_c = 1.5$, which indicates that when the salt flux is
due to the turbulent diffusion ($v = 1$), its solution is inde-
pendent of the stratification parameter (p_e). Under this
simplification, it is possible (1) to define a set of isolines
in a Cartesian Coordinate system ($p_e \times p_c$) with $v = \text{cte}$
and interval $0 < v \leq 1$.

Hansen and Rattray (1966) confirmed that theory by
analyzing an experimental data set from several estuaries,
from which four previously classified estuarine types
emerged: (1) Type 1, well-mixed estuary (unidirectional
circulation); (2) Type 2, partially mixed estuary
(circulation reverses at depth); (3) Type 3, fjords; and
(4) Type 4, salt-wedge estuary. Types 3 and 4 were classi-
fied because their experimental data fits very well in the
stratification-circulation diagram. Subdivisions a and
b for Types 1, 2, and 3 are low and high stratification when
 $p_e < 0.1$ and $p_e > 0.1$, respectively.

Figure 2 shows application of this diagram, with obser-
vational data for the tropical Caravelas River Estuary,
located in the SE Brazilian coast (lat. $17^\circ 45' 14.0''$; long.
 $039^\circ 13' 53.0''$). The classification changes from well
mixed ($v \approx 1.00$ – all salt transport is due to diffusion in
the spring tide) to partially mixed with low stratification
($v = 0.80$, meaning that 80 % and 20 % are the
up-estuary salt transport due to diffusion and advection,
respectively, at neap tide), due to the fortnightly tidal
modulation.

These theoretical results were revisited and confirmed
with the introduction of alternative parameters by Fisher
(1972), Prandle (1985), Jay and Smith (1988), and Scott
(1993). In Prandle's paper, the nondimensional p_c axis
was replaced by the ratio of the residual accelerations
associated with the horizontal density gradient and bed
friction, yielding a more direct assessment of the classifi-
cation based on more readily available parameters. The
demarcation line, which separates estuaries of types
1 and 2, can then be explained by the occurrence of flow
reversal. Prandle's diagram was applied by Miranda
et al. (2012) to classify the estuarine Bertioga Channel
(São Paulo).

Two classification diagrams were recently developed in estuarine physics. They introduced parameters based on salinity stratification and estuary circulation: the vertical Ekman (E_K) and the Kelvin (K_e) numbers (Valle-Levinson, 2008) and the nondimensional tidal (U_T) and freshwater (U_R) velocities (Geyer, 2010).

The Amazon River is the largest river system on Earth, containing ~20 % of the global fresh water supply. The large Amazonian drainage basin, which exceeds $7 \times 10^6 \text{ km}^2$, combined with the equatorial and tropical climate results in many tributaries and tremendous discharge. Tidal ranges are as high as 6 m at the mouth of the Amazon, where intertidal and subtidal periodicities are the dominant control on river-level changes, and these tidal influences extend more than 800 km upstream to Óbidos (Sioli, 1984; Oltman, 1967; quoted in Archer, 2005, p. 18). Many aspects of this transitional system are unique and not easily characterized within the existing definitions and classifications. The extreme tidal oscillations at the mouth create ideal conditions for the development of tidal bores throughout the mouth and inner areas, as first described by Rongel (1943). Kjerfve and Ferreira (1993) made time series measurements of water level, velocity, salinity, and temperature in the presence of a tidal bore in the macrotidal Mearin River (São Marcos Bay, Maranhão) in northeastern Brazil. This hydrodynamics was complex with an ephemeral flow of $1.5\text{--}2.0 \text{ ms}^{-1}$ transient velocity surge and propagation speed as high as 7.2 ms^{-1} .

The estuarine coastal embayment of the Amazon mouth (enclosing the North and South channels) is nearly 300 km in width. These channels are not estuaries sensu stricto in terms of saltwater-freshwater mixing and dilution (Bowden, 1978; quoted in Archer, 2005). Thus, for the Amazon system the definition of a drowned river valley estuary with an inner delta can be applied until further investigations are carried out.

Equations of motion, mass and salt conservation

Oceanic tides and land runoff are typical examples of processes that control the hydrodynamics in estuarine environments. Advection and small-scale turbulent motions affect salinity and temperature mixing processes, among other physical aspects related to erosion and transport of pollutants and organisms.

The basic system of equations that drives estuarine circulation and mixing are the mass and salt conservation equations, the momentum conservation, and the equation of state. Fluid density (ρ) and its velocity ($\vec{v} = u\vec{i} + v\vec{j} + w\vec{k}$) will be assumed continuous functions of space and time in a Cartesian Coordinate System (Oxy is the horizontal plane and Oz the depth, oriented against the gravity acceleration, $\vec{g}$). With Ox oriented along the estuary axis, the velocity components u and v are named longitudinal and transversal (or secondary), respectively.

The relationship between fluid density (ρ) and its velocity ($\vec{v}$) is derived from the principle of mass conservation (continuity equation). Its analytical deduction may be made with different theoretical developments found in oceanography texts (Sverdrup et al., 1942; Lacombe, 1965; Kinsman, 1965, among others). A convenient mathematical expression is the Eulerian formulation:

$$\frac{\partial \rho}{\partial t} + \nabla \cdot \rho \cdot \vec{v} = 0, \text{ or } \frac{1}{\rho} \frac{d\rho}{dt} + \nabla \cdot \vec{v} = 0 \quad (3)$$

The symbol $\cdot$ indicates the scalar product of the gradient operator ∇ by the mass flux vector $\rho \cdot \vec{v}$ ($[\rho \cdot \vec{v} = \text{ML}^{-2}\text{T}^{-1}]$). It states that the local (total) variation of the density is in balance with the divergence of the mass flux vector. If the fluid density is constant ($\rho = \text{cte}$), or the density does not change during the motion ($\frac{d\rho}{dt} = 0$), the fluid is defined as incompressible and the continuity equation simplifies to:

$$\nabla \cdot \vec{v} = 0, \text{ or } \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0 \quad (4)$$

The continuity equation holds for a laminar single fluid flow like water. However, if we approximate the estuarine water mass as oceanic seawater, which is a binary fluid (pure water + salt) and usually in turbulent motion, we must be aware of the following approximations: (1) the velocity $\vec{v}$ is the time mean value of the turbulent velocity (the over bar indicates a time mean value); (2) the mass conservation is also valid for the turbulent velocity, $\nabla \cdot \vec{v}_{\text{turb}} = 0$, with $\vec{v}_{\text{turb}} = \vec{v} - \bar{\vec{v}}$; and (3) the net salt diffusion across a closed boundary may be disregarded, which is a good approximation as demonstrated by Csanady (1982).

When (4) is applied to an estuary and the details of its circulation are not known in the interior fluid domain, we may use the continuity equation in its integrated form, using the Gauss theorem (or divergence theorem), under the assumption that all geometric and physical properties have all the regularity conditions imposed by its hypothesis. Then, if V denotes the estuarine volume boundary, A is a closed area and $\vec{n}$ is its normal unity vector oriented positively from the interior to the exterior, then:

$$\int_V (\nabla \cdot \vec{v}) dV = \int_A (\vec{v} \cdot \vec{n}) dA = 0 \quad (5)$$

the surface integral, which is the volume transport (L^3T^{-1} , $\text{m}^3 \text{ s}^{-1}$) through its closed boundary is zero. In estuaries, the closed area A has free and bottom boundaries which may have mass contributions from external water sources or sinks (evaporation, precipitation, and subsurface water springs from aquifers), and two vertical sections enclosing the estuarine water body.

A simple integrated equation of salt conservation, taking into account only the advection process on salt mixing, may be easily obtained applying the divergence operator to the mass flux vector $S \cdot \rho \vec{v}$ $[S \cdot \rho \vec{v}] = M \cdot L^{-2}T^{-1}$. Under the hypothesis that the salinity (density) and the velocity field in the estuarine domain are in steady state represented by its time mean value during tidal cycles, the divergence of the salt mass flux is zero:

$$\nabla \cdot S \rho \vec{v} = 0 \quad (6)$$

and as in (5)

$$\int_V (\nabla \cdot S \rho \vec{v}) dV = \int_A (S \rho \vec{v} \cdot \vec{n}) dA = 0 \quad (7)$$

the salt transport (MT^{-1} , $kg s^{-1}$) over a closed area is zero. With simplified geometry, (5) and (7) may be transformed in a system with two unknowns. They were used in the Knudsen hydrographic theorem applied for the first time around 1,900, enabling mean velocities to be calculated in cross-section areas under steady-state conditions, with known mean salinities in a highly stratified estuary (salt wedge) and with the fresh water discharge (Q_f) as the main forcing mechanism. In this type of estuary, the salt transport is driven by river discharge and the vertical turbulent mixing is absent. Let A_s (u_s) and A_i (u_i) be the superior and inferior cross-section areas (mean velocities) limited by the halocline, and S_s and S_i the mean salinities, respectively.

Disregarding the mass inflow and outflow across the free surface and the bottom, the equation system of (5) and (7) may be applied, taking into account that $\vec{v} \cdot \vec{n} \neq 0$ only on the areas A_s and A_i , then:

$$\int_A (\vec{v} \cdot \vec{n}) dA = u_s A_s - u_i A_i - Q_f = 0, \quad (8a)$$

$$\int_A (S \rho \vec{v} \cdot \vec{n}) dA = S_s \rho_s v_s A_s - S_i \rho_i v_i A_i = 0 \quad (8b)$$

Disregarding the density differences in the upper and lower layers ($\rho_s \approx \rho_i$) in the (8b), this system may be resolved for the mean current velocities (u_s , u_i) and/or the volume transports (Q_s , Q_i):

$$u_s = \frac{S_i Q_f}{A_s (S_i - S_s)}, \quad \text{or} \quad Q_s = \frac{S_i Q_f}{(S_i - S_s)}, \quad (9a)$$

$$u_i = \frac{S_i Q_f}{A_i (S_i - S_s)}, \quad \text{or} \quad Q_i = \frac{S_s Q_f}{(S_i - S_s)}. \quad (9b)$$

An application of this result may be found in Miranda et al. (2012) using the following experimental data of the Fraser River, according to Geyer (1986): discharge $Q_f = 3,000 m^3 s^{-1}$; sections geometry $A_s = 3,750 m^2$

and $A_i = 4,500 m^2$, and salinities $S_s = 14.0$ psu and $S_i = 30.0$ psu. Then, the theoretical mean velocities and volume transports are $u_s = 1.5 m s^{-1}$, $u_i = 0.6 m s^{-1}$, $Q_s = 5,525 m^3 s^{-1}$, and $Q_i = -2,525 m^3 s^{-1}$, respectively. These results show that the transport volumes are in balance with the fresh water discharge.

Consider an estuary with a surface area (A) delimited by the bottom and the surface and two vertical sections A_1 and A_2 at the river zone (RZ, where $S = 0$) and mixing zone (MZ), respectively. It follows from (8a), the mean longitudinal velocity under steady-state conditions across the area A_2 , which is given by: $-u_f \cdot A_1 + u_2 \cdot A_2 = -Q_f + u_2 \cdot A_2 = 0$, and thus $u_2 = u_f = Q_f/A_2$ is the fresh water velocity driven by the river discharge (Q_f).

From the continuity equation (3), the conservation salt principle due only to the advective process is:

$$\frac{\partial(\rho S)}{\partial t} + \nabla \cdot \rho S \vec{v} = 0 \quad (10a)$$

and

$$\frac{\partial S}{\partial t} + \vec{v} \cdot \nabla S = 0 \quad (10b)$$

However, the local variation $(\frac{\partial S}{\partial t})$ also depends on the turbulent salt-diffusion flux (ϕ_s), which is simulated by Fick's law $[\phi_s = -D(\frac{\partial S}{\partial n})]$ (D is the dynamic diffusion coefficient ($[D] = M \cdot L^{-2} T^{-1}$) and $\frac{\partial S}{\partial n}$ is the directional salinity gradient). The composition of this partial salt flux with (10b) takes the expression of the salt conservation equation (Sverdrup et al., 1942; Pritchard, 1958):

$$\begin{aligned} \frac{\partial S}{\partial t} + u \frac{\partial S}{\partial x} + v \frac{\partial S}{\partial y} + w \frac{\partial S}{\partial z} &= \frac{\partial}{\partial x} \left(K_x \frac{\partial S}{\partial x} \right) \\ &+ \frac{\partial}{\partial y} \left(K_y \frac{\partial S}{\partial y} \right) + \frac{\partial}{\partial z} \left(K_z \frac{\partial S}{\partial z} \right) - S_{\text{sinks}} + S_{\text{sources}} \end{aligned} \quad (11)$$

and the local salinity variation $(\frac{\partial S}{\partial t})$ is determined by the advection and diffusion (small-scale motion) processes, and sources and sinks of salt (precipitation, evaporation, bottom springs and sinks). In this equation, the Fickian coefficients are, according to Osborne Reynolds in 1884, parameterized in terms of the small-scale velocity (u' , v' , w') and salinity fluctuations (S'):

$$K_x = -\frac{\langle u' S' \rangle}{\frac{\partial S}{\partial x}}; K_y = -\frac{\langle v' S' \rangle}{\frac{\partial S}{\partial y}}; K_z = -\frac{\langle w' S' \rangle}{\frac{\partial S}{\partial z}}; \quad (12)$$

the symbol $\langle \rangle$ indicates "time mean value" of the correlations of small-scale velocity components multiplied by the small-scale salinity variation. These values, multiplied by the density ($K_x = -\rho \langle u' S' \rangle$; $K_y = -\rho \langle v' S' \rangle$; $K_z = -\rho \langle w' S' \rangle$), are the salt fluxes $[ML^{-2}T^{-1}]$ generated by turbulent diffusion.

The equations of motion are based on Newton's second law applied to the fluids, taking into account the forces and energy dissipation. The equation for a turbulent fluid was also presented in the classic paper of O. Reynolds. However, the estuarine water body has a particular geometry and has one open surface boundary; thus, special attention will be paid to the simplifications for their analytical and numerical solutions. For a hydrostatic fluid the equations of motion are:

$$\begin{aligned} \frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} + w \frac{\partial u}{\partial z} - fv \\ = -\frac{1}{\rho} \frac{\partial p}{\partial x} + \frac{\partial}{\partial x} \left(N_x \frac{\partial u}{\partial x} \right) + \frac{\partial}{\partial y} \left(N_y \frac{\partial u}{\partial y} \right) \\ + \frac{\partial}{\partial z} \left(N_z \frac{\partial u}{\partial z} \right) \end{aligned} \quad (13)$$

$$\begin{aligned} \frac{\partial v}{\partial t} + u \frac{\partial v}{\partial x} + v \frac{\partial v}{\partial y} + w \frac{\partial v}{\partial z} + fu \\ = -\frac{1}{\rho} \frac{\partial p}{\partial y} + \frac{\partial}{\partial x} \left(N_x \frac{\partial v}{\partial x} \right) + \frac{\partial}{\partial y} \left(N_y \frac{\partial v}{\partial y} \right) \\ + \frac{\partial}{\partial z} \left(N_z \frac{\partial v}{\partial z} \right) \end{aligned} \quad (14)$$

$$\frac{1}{\rho} \frac{\partial p}{\partial z} = -g \quad (15)$$

where f is the Coriolis parameter [$f = 2 \times \Omega \sin(\theta)$, Ω is the angular velocity of the earth and θ is the latitude], and g is the acceleration of gravity. In (11) and (14) N_x (K_x), N_y (K_y), and N_z (K_z) are the eddy kinematic viscosity (diffusion) coefficients. Since the estuarine water mass is assumed to be a system composed of pure water + salt, it will also be necessary to include in the hydrodynamic framework the equation of state of seawater, and the mass and salt conservation equations (4 and 11).

In analytical solutions, it will be assumed that the velocity does not change along its lateral axis (Oy), which is a good approximation because the secondary circulation intensity (v) is usually too low in comparison with the longitudinal (u) (Figure 3). Then, all terms of the preceding equations (11, 13, and 14) must be integrated along the Oy axis, and the mean value across its width (B) is calculated; thus, the equations are reduced to (Pritchard, 1958):

$$\begin{aligned} \frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + w \frac{\partial u}{\partial z} = -\frac{1}{\rho} \frac{\partial p}{\partial x} \\ + \frac{1}{B} \left[\frac{\partial}{\partial x} \left(B N_x \frac{\partial u}{\partial x} \right) + \frac{\partial}{\partial z} \left(B N_z \frac{\partial u}{\partial z} \right) \right] \end{aligned} \quad (16)$$

$$\frac{\partial(uB)}{\partial x} + \frac{\partial(wB)}{\partial z} = 0 \quad (17)$$

$$\begin{aligned} \frac{\partial S}{\partial t} + u \frac{\partial S}{\partial x} + w \frac{\partial S}{\partial z} = \frac{1}{B} \left[\frac{\partial}{\partial x} \left(B K_x \frac{\partial S}{\partial x} \right) + \frac{\partial}{\partial z} \left(B K_z \frac{\partial S}{\partial z} \right) \right] \\ - S_{\text{sinks}} + S_{\text{sources}} \end{aligned} \quad (18)$$

In (16), (17), and (18), the quantities u , w , and S are mean values across the estuary width (B). This set of equations has the effect of decoupling the motion and mixing equations, i.e., the velocity components (u , w) obtained from the solutions of (16) and (17) are used in the (18) for the salinity profile solution. In the assumption that the width is constant ($B = \text{cte}$) and $N_z \gg N_x$, and $K_z \gg K_x$, these equations may be further simplified.

Equation (15) assumes that the hydrostatic balance and the expression of the horizontal gradient pressure force ($-\frac{1}{\rho} \frac{\partial p}{\partial x}$) may be obtained in terms of its barotropic, baroclinic, and barometric components. On the assumption that the density (ρ) is known, the only unknown is the pressure (p), which may be easily obtained by vertical integration along the water column, from a depth z up to the sea surface slope $\eta(x, y, t)$ (relative to a level surface):

$$p(x, y, z, t) = p_a(x, y, t) + g \int_z^\eta \rho dz \quad (19)$$

where p_a is the atmospheric pressure on the sea surface. Taking the differential of (19) using the Leibnitz differentiation rule it follows that:

$$\frac{\partial p}{\partial x} = \frac{\partial p_a}{\partial x} + g \left[\left(\rho_\eta \frac{\partial \eta}{\partial x} \right) + \left(\int_z^\eta \frac{\partial \rho}{\partial x} dz \right) \right] \quad (20)$$

where ρ_η is the density on the surface. From this result, with $\frac{\rho_\eta}{\rho} \approx 1$, the expression for the longitudinal component of the gradient pressure force (per mass unit) has three components:

$$-\frac{1}{\rho} \frac{\partial p}{\partial x} = -\frac{1}{\rho} \frac{\partial p_a}{\partial x} - g \frac{\partial \eta}{\partial x} - \frac{g}{\rho} \int_z^\eta \frac{\partial \rho}{\partial x} dz \quad (21)$$

namely, barometric (a), barotropic (b), and baroclinic (c), respectively:

- The barometric component is related to transient weather systems (typically 3–10 days) associated with low-pressure centers and has subtidal variability. Under steady state, the sea surface acts as an inverted barometer; for $\Delta p = \pm 1.0$ mbar the sea surface decreases/increases by 1.0 cm. However, if a storm surge was to reach an estuary, it may cause severe dangerous floods, especially during spring tide.
- Is independent of depth and varies according to the sea surface slope oscillation. In normal tidal conditions, its highest and smallest values occur during

the spring and neap tide, respectively. Its order of magnitude varies approximately in the interval $-10^{-3} - +10^{-3} \text{ (ms}^{-2}\text{)}$. Thus, it is considered to have inter or subtidal variability.

(c) This component is zero on the surface ($z = \eta$) and increases with the depth, up to a magnitude order of -10^{-4} ms^{-2} . Due to the inter and subtidal variability of the density field, its numerical value is not an easy quantity to be determined.

During the flood tides, barotropic (b) and baroclinic (c) forces act up-estuary, but during the ebb they act in the opposite direction.

In the variability analysis of the half-hourly Eulerian profiles at the spring tide (Figure 4-left), the higher barotropic tidal forcing, generating bidirectional motions up to 1.5 and -1.0 ms^{-1} (ebb and flood, respectively) preclude the baroclinic forcing, but the opposite occurs during the neap tide (Figure 4-right). The action of the less intense baroclinic pressure force is clearly seen in generating bidirectional motions and in the speed increase (in intensity) at mid-depths during the flood ($u < 0$).

In analytical solutions, the expression of the baroclinic component may be simplified on the assumption that it is independent of the depth $\left[\frac{\partial}{\partial z} \left(\frac{\partial p}{\partial x}\right) = 0\right]$, by using a depth time-mean estimated value $\left(\frac{\partial p}{\partial x} = \rho_x\right)$. Then, for a simple geometry ($B = \text{cte}$), kinematic eddy viscosity coefficient N_z independent of depth and $N_z \gg N_x$, the simplified steady-state equation of motion is:

$$-g \frac{\partial \eta}{\partial x} - \frac{g}{\rho} \rho_x (z + h) + N_z \frac{\partial^2 u}{\partial z^2} = 0, \quad (23)$$

which states a balance between the barotropic and baroclinic components of the pressure gradient force and the vertical velocity shear associated with the estuarine circulation; the influence of tides within this formulation enters only in the value of N_z (Geyer, 2010).

Analytical and numerical solutions

Mathematical models can be either analytical or numerical Ji (2008). An analytical model has an exact mathematical solution to the differential equations describing processes in estuaries Blumberg (1975). They may be applied to relatively restrictive conditions, usually for one or two dimensions, constant parameters and steady-state conditions. In spite of the severe assumptions that must be invoked, analytical models are often used to (Ji, 2008):

1. Check the accuracy of numerical models (e.g., Blumberg, 1975).
2. Provide first-order estimates of relatively simple systems.
3. Give insights into hydrodynamic and water quality processes in estuaries.

A numerical model is a discretized version of a set of mathematical equations, as presented in this chapter

(continuity, equation of motion, salt conservation), which describes processes in the estuary, and can be implemented as a computer program. By entering the input data and mode parameters into the computer model, numerical and graphical simulations of an estuary, in response to a set of forcing conditions and boundary conditions, may be obtained.

However, analytical and numerical model solutions must be calibrated or validated based on observational data. This may be done numerically with nondimensional parameters such as the Relative Mean Absolute Error (Walstra et al., 2001) and the Skill parameter (Wilmott, 1981), further applied by Warner et al. (2005). The vertical mean Skill parameter was adapted by the validation of vertical velocity and salinity profiles (Andutta et al., 2006). The Skill parameter is calculated taking into account the model solutions and the observational data and varies from 1 to 0 (zero) indicating the best fit and a complete disagreement between observation and the theoretical results, respectively.

Analytical models

The first steady-state analytical model for determining time mean longitudinal velocities in a coastal plain estuary was developed by Pritchard and Kent (1956) using the lateral and longitudinal components of the equation of motion, the tidal velocity amplitude, and the relationship between the vertical and lateral eddy stress. The method was applied to stations in the James River Estuary, studied in detail during several tidal cycles in the summer (June and July, 1950). The theoretical velocity profiles agreed well with the observational data, showing typical velocity profiles of a partially mixed estuary – seaward and up-estuary motion on the upper and lower layers, respectively, and no motion at mid-depths. The Pritchard and Kent paper is a pioneering article showing the importance of the comparison of theoretical versus experimental data.

Hansen and Rattray (1965) developed a steady-state analytical model for the circulation and mixing of partially mixed estuaries. The laterally averaged equations of motion and the mass and salt conservation equations (16, 17, and 18) were used with the simplifications: $\left(\frac{\partial u}{\partial x} \ll \frac{\partial u}{\partial z}\right)$, $\left[\frac{\partial}{\partial z} \left(\frac{\partial S}{\partial x}\right) = 0\right]$ and a linear equation of state of seawater $\rho = \rho(S)$ for the hydrodynamic equations closure. Using similarity solution techniques, the model considers the balance of the barotropic and baroclinic modes, with wind stress forcing on the surface ($u = 0$) and no-sleep condition at the bottom [$u(-h) = 0$]. The central regime solutions for $u = u(z)$ and $S = S(x, z)$ depend on subjective numerical values such as the longitudinal density gradient and the mean salinity at the mouth, as well as the eddy viscosity and diffusion coefficients. The results of this model used for the Piaçaguera channel (upper reaches of the Santos Channel – São Paulo – Brazil), validated with the mean vertical Skill parameter, are presented in Figure 5, according to Miranda et al. (2012).

A simple and direct solution to (23), conducive to the same result of Hansen and Rattray (Figure 5), was given by Officer (1976). On the assumption that the longitudinal density gradient ρ_x is a known quantity, the equation has two unknowns: $\left(\frac{\partial \eta}{\partial x}\right) = \eta_x$ and $u = u(z)$. Thus, a second equation is necessary to complete the equation system. In this solution, the equation of continuity integrated in the estuary volume, called the integral boundary condition, is used:

$$\frac{1}{h} \int_{-h}^0 u(z) dz = \frac{Q_f}{A} = u_f \quad (24)$$

where Q_f is the river discharge, and A , h , u_f are the cross-section area, the depth, and the fresh water velocity, respectively.

To achieve the solution, the upper and the lower boundary conditions are the same as in the Hansen and Rattray analytical model: $\rho N_z \frac{\partial u}{\partial z} |_{z=\eta} = -\tau_w$ and $u|_{z=-h} = 0$

For (23), which is a second order ordinary differential equation, its general solution will be dependent on two integration constants C_1 and C_2 ,

$$u(z) = \frac{1}{2} \frac{g}{\rho N_z} \eta_x z^2 + \frac{g}{\rho N_z} \rho_x \left(\frac{h}{2} z^2 + \frac{1}{6} z^3 \right) + C_1 z + C_2 \quad (25)$$

which are determined according to the upper and lower boundary conditions:

$$C_1 = \frac{\tau_w}{\rho N_z}, \text{ and } C_2 = -\frac{\tau_w h}{\rho N_z} - \frac{1}{2} \frac{g \eta_x}{N_z} h^2 - \frac{g}{3 \rho N_z} \rho_x h^3 \quad (26)$$

if these values are known, the general solution is:

$$u(z) = \frac{g}{\rho N_z} \rho_x \left(\frac{z^3}{6} + \frac{h}{2} z^2 - \frac{h^3}{3} \right) + \frac{1}{2} \times \frac{g}{N_z} \eta_x (z^2 - h^2) + \frac{\tau_w}{\rho N_z} (z - h) \quad (27)$$

or in nondimensional depth $Z = \frac{z}{h}$,

$$u(Z) = \frac{g}{6 N_z \rho} \rho_x h^3 (Z^3 + 3Z^2 - 2) + \frac{1}{2} \times \frac{g}{N_z} \eta_x h^2 (Z^2 - 1) + \frac{\tau_w h}{\rho N_z} (Z + 1) \quad (28)$$

This solution is still dependent on the second unknown (η_x) and we must apply the integral boundary condition, written in terms of the nondimensional depth Z :

$$\int_{-1}^0 u(Z) dZ = \frac{Q_f}{A} = u_f, \quad (29)$$

and the result for the surface slope η_x is:

$$\eta_x = -\frac{3 N_z}{g h^2} u_f - \frac{15}{24} \frac{\rho_x}{\rho} h + 3 \frac{\tau_w}{\rho g h} \quad (30)$$

and depends on three quantities: (1) the fresh water velocity, (2) the baroclinic component and (3) the wind stress. An order of magnitude analysis indicates that the baroclinic term, associated with the longitudinal density gradient, is the dominant equivalent to the M. Margules rule for the slope of an interface in the atmosphere, which was adapted to oceanographic use by A. Defant in 1929 (quoted in von Arx, 1962, 383).

Combining (28) and (30), the final solution for $u = u(Z)$ is Officer (1976):

$$u(Z) = \frac{g h^3}{48 N_z \rho} \rho_x (8Z^3 + 9Z^2 - 1) + \frac{3}{2} u_f (Z^2 - 1) + \frac{1}{4} \times \frac{h}{\rho N_z} \tau_w (-3Z^2 - 4Z - 1) \quad (31)$$

and its graphical results agree with those of Hansen and Rattray (1965) shown in (Figure 5b).

From (31), it is possible to simulate the seaward and unidirectional circulation of a well-mixed estuary ($u > 0$) by changing parameters values as increasing the water column height (h), decreasing the longitudinal density gradient (ρ_x) and the kinematic viscosity coefficient (N_z) (Figure 6).

Further classical and up-to-date analytical solutions of salt wedge, well-mixed, and partially mixed estuaries may be found in Prandle (2009) and Miranda et al. (2012).

Secondary circulation

The secondary estuarine circulation is normal to the along-channel currents and is an integrated component of the estuarine circulation. Its dynamics have been presented in several articles since the pioneering works by Okubo (1973). Taking into account experimental results Dyer (1977), presented diagrammatic representations of the mean along-channel currents and the associated secondary circulation for different estuarine types: salt-wedge (types A or 4), partially mixed (types B or 2), and well mixed (types C or 1). The dynamic balance of the secondary flows was established by taking into account the interaction of the following factors: bottom topography and channel geometry, lateral stratification of density due to vertical mixing, and gradient pressure forces, Coriolis acceleration and centrifugal acceleration (Nunes and Simpson, 1985).

In the bi-dimensional analytical simulation of the secondary circulation, according to Nunes and Sympton (op. cit.), the simplifying hypothesis are applied: (1) stationary conditions; (2) lateral density (salinity) stratification $\frac{\partial \rho}{\partial y} = g(y)$, and; (3) straight channel. The solution of the secondary circulation $v = v(y, Z)$ and its associated vertical component $w = (y, Z)$ obtained with this model is exemplified by Miranda et al. (2012).

The importance of the secondary flows and mixing across a channel as interrelated processes, were recently discussed in detail by Chant (2010).

Summary

Estuaries were formed in a narrow coastal boundary zone between the sea and land, during the last interglacial period. This contribution is intended to present some aspects of estuarine circulation related to: (1) transitional environments along the coastline that are ultimately dependent on estuarine dynamics; and (2) investigations of the processes that have focused on estuarine preservation, water quality, morphology, biodiversity, and fisheries which are strongly dependent on the dynamic behavior of estuaries.

Some focus is given to the pioneer research published in the last half of the twentieth century. Additionally, experimental and theoretical developments have recently been published focusing on the following issues: estuary definition, classification, mixing processes, variability, and circulation.

Transitional environments along the coastline are ultimately dependent on the dynamics of estuarine systems; however, they have been strongly affected by human activities. Some of them have been protected by reserve status since the eighteenth century. Human impacts and their effects on estuaries and coasts are important factors that need to be assessed.

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Estuarine Circulation, Figure 1 Schematic longitudinal section of an estuary showing the influence of advection ($\Rightarrow$ due to river discharge U_R and tidal currents) and vertical mixing (thin waving lines) in the local salt balance. The thick lines are isohalines. Box A (*upper layer*), horizontal advection causes a reduction of salinity, but vertical mixing compensates by replacing the low-salinity water with underlying high-salinity water. The relative roles of advection and mixing are reversed in Box B (*lower layer*). Bidirectional circulation is shown in the vertical profile of the u -velocity component (*dashed line*) with the depth of no motion, and U_E and $-U_E$ are upper and lower maxima values shown in this profile. According to Geyer (2010).

Estuarine Circulation, Figure 2 Stratification-circulation diagram with experimental data of the Caravelas River Estuary (southern Bahia State, Brazil) for the August (spring and neap tide) and for January (spring tide) experiments. The values of the parameter ν are indicated close to the symbols + and o. The circulation parameter (u_s/u_t) was approximate to u_s/u_a , where u_a is the residual velocity (time mean-depth value).

Estuarine Circulation, Figure 3 Time variations of the longitudinal (u) and the secondary (v) velocities components at the tropical Caravelas River Estuary (Bahia, Brazil) during spring tidal cycle of January, 2008.

Estuarine Circulation, Figure 4 Eulerian profiles of the u-velocity component at half-hour time intervals at spring (*left*) and neap tide (*right*) in the tropical Caravelas River Estuary (Bahia, Brazil). Note the *intratidal* and *subtidal* variabilities (From Andutta, 2011).

Estuarine Circulation, Figure 5 (a) Theoretical results for each component mode: V_g = baroclinic forcing; V_{df} = river discharge; V_v = wind stress and V_c = composite profile, in the Piaçaguera channel in the upper reaches of the Santos channel (São Paulo – Brazil, using the Hansen and Rattray (1965) analytical model. (b) Experimental versus observational of the u-velocity profile. Skill is the mean vertical parameter to validate the theoretical simulation (From Miranda et al., 2012).

Estuarine Circulation, Figure 6 Theoretical velocity profiles for a partially mixed estuary (tick line – the same as in Fig. 5b) and the unidirectional motion of a well-mixed estuary with ebbing currents ($u > 0$ – thin line).