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Long Term Changes to the Lower Columbia River Estuary (LCRE) Hydrodynamics and Salinity Patterns

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Long Term Changes to the Lower Columbia River Estuary (LCRE) Hydrodynamics and
Salinity Patterns

by

Aqeel Al-bahadily

A dissertation submitted in partial fulfillment of the
requirements for the degree of

Doctor of Philosophy
in
Civil and Environmental Engineering

Dissertation Committee:
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2020

Abstract

Changing the morphological and hydrological conditions of an estuary can affect the estuarine hydrodynamics. The hydrograph of the Lower Columbia River Estuary (LCRE) and its bathymetry have been altered significantly over the past 150 years, such that the spring-freshet has decreased by 40-50% while winter flow has increased by 50%. In addition, the inlet width has been narrowed from 9.7 to 3.2km by the construction of jetties, and the controlling depth of the navigation channel has been deepened from 6 to 13m by continuous dredging. Also, ~70% of the shallow water habitat has been lost due to diking and wetland reclamation. Finally, the main shipping channel was changed from the north to the south side of the estuary. These system alterations lead to the following question: how have changes since the mid-1800s altered the salinity intrusion, freshwater distribution, transport processes, and water levels in the Lower Columbia River estuary?

In this study, I use a 3D hydrodynamic model to gain insights into changes in circulation and salinity intrusion from the mid-1800s to the present. Two models were constructed: one based on bathymetric measurements made between 1869 and 1900, and a second based on bathymetric surveys made from 2008 to 2010. The horizontal resolution is typically 50m inside the estuary and 2000m offshore, with 36 sigma layers of vertical resolution. The models have been calibrated and verified using field measurements distributed over the lower 40km of the estuary. The average skill assessment of the calibration is 0.98; the root mean square error is between 0.11m and 0.15m for water level, 2-4psu for salinity, and 0.25m/s for velocity.

Analysis of flow bifurcation (water distribution between channels at a junction) showed that morphology, tidal forcing, and water surface slope have played major roles in freshwater distribution in the multi-channel estuary. However, the asymmetry in the water surface slope between the North and the South Channels is what controls the distribution of flow between the two channels. In the historical model, the two channels surface slopes are equal. Therefore, the hydraulic radius and the cross-sectional area control the flow distribution, but the percentage of freshwater in each channel is still a function of river discharge. For the modern model, the percentage of freshwater in the South Channel decreases as the river discharge increases. During spring tide, 88% of freshwater passes through the South Channel during low river discharge conditions ($3,000\text{m}^3/\text{s}$), while 48% passes through during high river discharge ($15,000\text{m}^3/\text{s}$). Conversely, in the historical model, the percentage of freshwater in the South Channel increases as the river discharge increases. This implies that the morphological changes in the estuary causes the South Channel to export more freshwater today, as compared to 150 years ago. Finally, model results suggest that M_2 amplitude has increased by 10% in the city of Astoria due exclusively to channel deepening, and about 12.5% ($\sim 0.11\text{m}$) when both channel deepening and river discharge alteration are included.

The LCRE system alteration is modeled to change the average location and the seasonal cycle of salinity intrusion. The simulation results found that average winter-time salinity intrusion has decreased due to a 50% increase in river discharge. During the annual spring freshet, salinity intrusion has increased because of decreased river flow. Due to changes in river discharge, the seasonality of maximum salinity intrusion has shifted from

wintertime to late summer/early fall. Channel deepening exerts a strong control on salinity intrusion, and has caused greater landward salinity intrusion landward due to increased stratification. Channel deepening also amplified spring-neap variations in intrusion and altered the sensitivity of intrusion to river discharge. Altogether, the change in freshwater distribution, salinity intrusion, and stratification produced a change in the LCRE system classification. A parameter space classification based on Geyer & MacCready, (2014), suggests that morphological changes and river discharge alteration have shifted the LCRE from partially mixed 41% of the time to a salt wedge estuary 34% of the time.

Finally, the effects of future sea-level rise and a subduction zone earthquake on tidal amplitudes and salinity intrusion are investigated. A future sea-level rise of 0.21m will have an insignificant effect on salinity intrusion and tidal variation inside the estuary. On the other hand, sudden land subsidence resulting from a potential M9 earthquake and/or a future 1.5m SLR (extreme cases) could produce a substantial effect on tidal amplitudes and circulation. Modeling analyses suggest that either scenario could elevate the major tidal constituent (M_2) by 0.05m. Moreover, the maximum value could be moved upstream by 4km. Combined together, the combined effect of 1.5m SLR and sudden land subsidence might raise M_2 by 0.11m and advance the location of maximum M_2 by 10km landward. Similarly, salinity intrusion could increase 4km as a result of either the SLR or land subsidence cases, and might increase by 10km with both scenarios combined. Moreover, both cases have altered the habitat water depth and inundation by increasing the mean depth.

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Chapter 1 Overview of Dissertation**1.1 Introduction**

This chapter includes general information about the definition of estuaries, their importance, and their classification. In this chapter, I will review salinity intrusion, and then give a brief history of the Lower Columbia River Estuary (LCRE), and the morphological and hydrological changes since the mid-1800s. Finally, this chapter describes how these changes motivated the research questions about how hydrodynamic and salinity transport have changed.

1.2 General

Estuaries are typically defined as the transition region in which freshwater from river flow mixes with salt water from the ocean, producing a horizontal density gradient and (often) vertical stratification (Wilson, 1977; Hansen & Rattray, 1966; Pritchard, 1967). Estuaries are valuable natural systems both ecologically and economically (Lenanton & Potter, 1987). From the environmental point of view, estuaries provide habitat for fishes, plants, and birds. Estuaries can filter out sediment and pollutants coming from the river and going to the ocean. Economically, estuaries are useful for fishing, transportation, and recreation; also, they act as protective buffers against floods and help dissipate storm surge.

The morphological and hydrological conditions can control tides and circulation in estuaries. Further, estuaries are affected by river flow, tidal forcing, and wind stresses.

These factors can control vertical/horizontal mixing, and stratification, and hence, estuarine circulation. Moreover, these factors can be altered by human interference and climate change (Miranda et. al., 2017). Estuaries can be classified based on water balance, geomorphology, vertical salinity structure, and hydrodynamic characteristics (Figure 1-1). A positive estuary means there is a net-inflow of freshwater, while an inverse estuary means that a net loss of water due to evaporation occurs. Based on geomorphology, estuaries can be Coastal Plain (formed from sea-level rise), Fjord (marked by a deep channel with a sill), bar-built (characterized by a sand bar formed between ocean & estuary), or tectonic (formed from vertical land movement). The vertical salinity structure defines whether an estuary is classified as a salt-wedge, strongly stratified, weakly stratified, or well mixed (Figure 1-2a, b). A salt wedge is a wedge shaped boundary that separates the bottom high-density layer from the upper freshwater layer, with minimal mixing, because of large river discharge and weak tidal forcing (Figure 1-2a). A strongly stratified estuary has a large salinity difference between layers (usually associated with large river flow). Weakly stratified (partially mixed, Figure 1-2c) occurs where the river flow is weak to moderate and tidal mixing is moderate to strong, such that the tidal forcing enhances mixing of the two layers. A vertically mixed system (well mixed, Figure 1-2d) is characterized by strong mixing between the freshwater layer and ocean water layer. This mixing is a result of strong tidal forcing and weak river flow (i.e., no stratification). For these reasons, hydrodynamic classification depends on circulation (the ratio of the surface

flow to column-averaged flow) and stratification (the ratio of salinity difference between top and bottom compared the depth-averaged salinity) (Parker, 1991).

Figure 1-1. Estuarine classifications

Figure 1-2. Estuarine classification based on water mixing after Valle-Levinson, (2010)

The estuarine classification shows the importance of the morphological parameters such as channel depth and the amount of freshwater provided by river discharge in changing the system from one state to another. Hence, altering the system can modify estuarine hydrodynamics and circulation. Both river flow and morphology of the Columbia River Estuary (CRE) have changed significantly over the past 150 years, which suggests a change in the hydrodynamics of the estuary.

At present, the classification of CRE varies from a salt wedge to well mixed estuary. During high river flow, the system is highly stratified and the estuary is a salt wedge for all tidal conditions. During low flow conditions the tidal forcing becomes critical to keeping the system partially mixed (Jay & Smith, 1990). The long-term changes to the

estuary and their effect on salinity intrusion are important since they can affect freshwater availability and can affect the type of habitats and species within estuaries.

1.3 Salinity intrusion

When freshwater from the river meets saline water from the ocean, the freshwater tends to dilute the mixture, while the tidal motion tends to drive the salt upstream by advection. This is because of the density difference between seawater and river discharge. The result is intrusion of ocean salt water into the estuary. Because of the temporal variation of tidal forcing and river flow, salinity intrusion can greatly vary over time (MacCready & Geyer, 2010). The salinity intrusion length is typically defined as the point at which the bottom salinity equals 2-psu (Monismith et al., 2002; MacWilliams et al., 2015), see (Figure 1-3).

Figure 1-3. Salinity intrusion a-concept, and b- example of modeled salinity intrusion where the white curve represents the 2psu isohaline

1.4 Columbia River Estuary (CRE)

The Columbia River (CR) is the fourth largest river in North America that drains to the ocean. Its length is about 1900km and its basin drainage area is about 660,480km² (Simenstad & Sherwood, 1990). The CR passes through seven states inside the United States and a Canadian province (see Figure 1-4). The river provides hydropower with 14 dams on the main river and 450 dams on its tributaries (Osborn, 2012). The CR estuary has two channels (North and South); most of the river flow goes through the South Channel, while most of the tidal prism fills and empties through the North Channel (Chawla et al., 2008). Before 1850, the estuary was primarily used for fishing by natives who inhabited the area. At present, the estuary is used for industry, navigation, agricultural, fishing, and tourism purposes.

Figure 1-4. Columbia River watershed

1.5 Bathymetry

The major changes in the morphology of the CR are due to dredging activities and jetty construction. Before the construction of jetties, the Columbia River mouth lay between two sand spits: Point Adams, on the south, and Cape Disappointment, on the north. Cape Disappointment is about 6 miles north-northwest from Point Adams (Figure 1-5). Also, there were two areas with shoals: Peacock Spit to the south-west of Cape Disappointment, and Clatsop Spit to the north-west of Point Adams. These two spits were connected by an ocean bar across the river entrance and the navigation channel was over

the southwest quadrant of the bar. The shoals and the bar at the entrance were continuously changing position due to waves and wind (Hickson & Rodolf, 1951), (see Figure 1-5).

The construction of the south jetty began in 1885, but the channel deepening did not start until 1889, such that the channel depth changed from 17ft to 31ft by 1895 (Sherwood et al., 1990). After the completion of the north jetty (1913-1917), the depth of the entrance had increased to 37ft. However, the depth gradually increased to reach 46-48ft by 1925. By 1928, Clatsop Spit had advanced into the river, and the navigation channel was forced to the north (Figure 1-5). Furthermore, the construction of jetties had advanced the bar into the ocean by about 1 mile in 1900 and 5 miles by 1950 (Hickson & Rodolf, 1951). Upon the completion of the jetty construction in 1939, the entrance width between north and south jetties had narrowed from 6 to 2 miles (USGS, 1981), (see Figure 1-6).

Besides the construction of jetties, maintaining the navigation channel through dredging and destroying the habitat through filling have had a continued impact on the estuary morphology (Simenstad et al., 1990). The largest change in morphology was between 1867-1958, caused by navigational improvements such as channel dredging, jetties, and pile dikes. Diking and filling were responsible for the loss of 77% (8100ha) of tidal swamps, 62% (4050ha) of tidal marshes, and 7% (1195ha) of tidal flats (Sherwood et al., 1990). Dredging activities began in 1873 (U.S. Army, 2012), but major dredging did not begin until 1909 (Sherwood et al., 1990). Some portion of the removed materials were disposed of offshore, while others were dumped in wetlands. Because of the dredging activities, the estuary bathymetry changed dramatically. The controlling depth of the South

Channel is 13m in the estuary, and 18m at the mouth, while it was about 6m in the 1868-1901 period. The development of the depth and width of the navigation channel and the mouth can be found in Table 1-1.

Figure 1-5. CR mouth before jetty construction (1839), (Hickson & Rodolf, 1951)

Figure 1-6. CR mouth after jetty construction (1950), (Hickson & Rodolf, 1951)

Table 1-1. Columbia River mouth & navigation channel development by U.S. Army Corp of Engineers (USACE)

Mouth of the Columbia River		Depth	Width
1882	Public Acts of the 47 th Congress	30'	
1905	River and Harbor Act	40'	½ mile
1954	River and Harbor Act	48'	½ mile
Columbia River Channel			
1878	River and Harbor Act	20'	
1899	River and Harbor Appropriation Act	25'	
1912	River and Harbor Act	30'	300'
1930	River and Harbor Act	35'	500'
1962	River and Harbor Act	40'	600'
1999	Water Resources Development Act	43'	

1.6 River flow

The Columbia River hydrograph has changed dramatically as a result of energy demand, irrigation demand, flood control, temperature, and precipitation changes (Naik & Jay, 2011), see (Figure 1-7). At present, the maximum daily flow during spring freshet (Apr-Jun) often reaches more than $12,000\text{m}^3/\text{s}$, while minimum daily flow goes below $2000\text{m}^3/\text{s}$ during the dry season (July-Oct). Monthly average flow, however, ranges between $3,000\text{--}8,000\text{m}^3/\text{s}$. Historically (before 1900), the monthly average range was between $2,500\text{--}15,000\text{m}^3/\text{s}$, (Figure 1-7), while the daily range was $1000 - 35,000\text{m}^3/\text{s}$ (Chawla et al., 2008).

The biggest change in the seasonal cycle started around 1960 due to flow regulation (Sherwood et al., 1990). Winter season river flow has since increased and spring-freshet flow has decreased (Figure 1-7). Naik & Jay, (2011) reported a 17% reduction in the annual mean flow, a 46.5% reduction in spring freshet, and a 44% decrease in the maximum daily flow. The Spring-freshet timing has changed as well, such that, due to climate change and flow regulation, it occurs about two weeks earlier than it did before 1960.

The river hydrograph will likely continue to change in the future due to human and natural intervention (Knowles et al., 2006). For example, Cohen et al. (2000) have predicted an earlier peak and possible reduction in lower minimum flow and total annual flow. Bürger et al. (2011) have also predicted an earlier peak flow from July to June, with a decrease in summer flow and no increase in total flow.

Figure 1-7. Monthly mean flow of Columbia River at The Dalles/Oregon

1.7 Motivation and research questions

Salinity intrusion can threaten the freshwater supply in an estuary, especially in dry seasons, although it varies, specially and temporally, depending on freshwater supply and tidal forcing (Gong & Shen, 2011). Furthermore, salinity has a large effect on what species can exist in aquatic environments, since some species tolerate only specific ranges of salinity. Salinity has a profound effect on the distribution and abundance of aquatic organisms. It is one of the determinants of the abundance of organisms in estuarine habitats (Bell et al., 2003). Change in salinity affects organism distribution, behavior, and limits reproduction. Stratification, mixing, and exchange flow have a significant impact on

salinity intrusion, and both affect the suspended sediment and dissolved substances in estuaries (Geyer & MacCready, 2014; Dijkstra et al., 2017).

Changes to the tidal forcing and river discharge have a direct impact on salinity in estuaries and can change the vertical structure and intrusion length. For example, partially-mixed estuaries have a maximum salinity intrusion during neap tides (Monismith et al., 2002; MacCready, 2007; Ralston et al., 2008) while a well-mixed and salt-wedge estuaries expect a higher intrusion during spring tides (Ralston et al., 2010). Further, salinity intrusion length has a power relation with river discharge (inversely correlated). The power exponent varies in different estuaries (Monismith et al., 2002; Bowen and Geyer, 2003).

Most investigations of salinity intrusion focus on estuaries with one primary channel such that the whole river discharge and tidal forcing pass through that channel (Monismith et al., 2002; Ralston et al., 2008). This approach cannot be used on estuaries with more than one channel, since the tidally averaged flows are distributed between the channels as a variable function of river discharge and tidal forcing (Buschman et al., 2010; Sassi et al., 2011; Zhang et al., 2017). Hence, the salinity intrusion and river flow relation can change (different power exponent). Overall, river flow bifurcation along with the tidal asymmetry in flood-ebb for the multiple channels (Robinson, 1960) will produce different salinity intrusion behavior than a single channel estuary.

The Lower Columbia River Estuary (LCRE) is a multi-channel estuary with two primary channels (North & South). The majority of tidal exchange occurs at the North Channel (Lutz et al., 1975; Chawla et al., 2008; Jay & Smith, 1990) while the majority of

river flow has been diverted into the South Channel through dredging and the construction of islands and other flow training structures. The Columbia River has been significantly altered in terms of geometry and river hydrograph during the last 150 years (Naik & Jay, 2005). The major change has been the deepening of the navigation channel, which has increased the tidal range and reduced the river slope in the CRE (Helaire et al., 2019) such that the tidal range near Astoria has increased by 5.6% (~0.11m) since 1855 (Talke et al., 2020a). The river forcing alteration and the tidal change could have altered the freshwater distribution in the estuary. Accordingly, the transport of sediment and salt could have changed as well.

The long-term change to river hydrograph and the morphological change in the estuary (bathymetry, the construction of jetties) have changed the estuarine geometry and likely the freshwater distribution in the estuary (the South and North Channels). These changes have motivated my first research questions: *How is the river flow distributed between the North and South Channels? How do the river flow and bathymetric changes affect the river flow bifurcation in the Lower Columbia River Estuary?* These questions are addressed in Chapter 3.

Changes in river discharge, bathymetry, and possibly flow bifurcation raise the possibility that salinity intrusion has changed since the 1800s. This observation leads to the following question: *How have long-term modifications to the bathymetry and fluvial boundary conditions of the system altered the hydrodynamics and transport processes within the estuary?* This question is addressed in Chapter 4.

Finally, changes in bathymetry and river discharge may continue in the future, due to sea-level rise and possibly a Cascadia subduction zone earthquake. These observations have motivated the third research question: *What is the potential impact of future sea-level rise, altered river discharge, and land subsidence due to an earthquake on system tides and salinity intrusion?* This question is addressed in Chapter 5.

Chapter 2 Theory and Literature Review**2.1 Introduction**

In this chapter, I will review the importance of channel depth and river discharge on estuarine circulation, vertical salinity profile, and salinity intrusion based on available scientific literature. Further, I will review how salinity intrusion varies with river discharge in different estuaries. I will review the possible future effects of climate change on the hydrological cycle and sea-level rise. Finally, I also provide an overview of tidal asymmetry, exchange flow, flow partitioning, and estuarine classification.

2.2 Theory**2.2.1 Effect of channel depth on estuarine circulation and salinity profile**

Hansen & Rattray (1965) proposed a model to describe circulation and salt flux processes in estuaries. The model equations describing circulation and salinity distribution are the equations of motion, mass conservation, salt conservation, and the equation of state. The model assumes lateral homogeneity, i.e., the model is reduced to longitudinal & vertical dimensions, where the x-axis and z-axis are positive, seaward and downward, respectively). Further, the estuary has a constant width and depth. The analysis neglects Coriolis forces, the viscous frictional terms, and applies the approximation of Boussinesq (i.e., it neglects density variation in all but buoyancy terms). It also assumes that the pressure variations are applied to water only (atmospheric pressure is assumed constant).

The momentum equation (before time averaging) in the longitudinal direction can then be written as

$$\underbrace{\frac{\partial u}{\partial t}}_{\text{Local acceleration}} + \underbrace{\frac{\partial(uu)}{\partial x} + \frac{\partial(wu)}{\partial z}}_{\text{convective flux}} = - \underbrace{\frac{1}{\rho} \frac{\partial p}{\partial x}}_{\text{pressure force}}. \quad (2.1)$$

The momentum equation in z-direction assuming the hydrostatic conditions (inertial and convective acceleration are neglected) is

$$0 = -\frac{1}{\rho} \frac{\partial p}{\partial z} + g \quad (2.2)$$

The conservation of salt, neglecting molecular diffusion is

$$\frac{\partial s}{\partial t} + \frac{\partial(us)}{\partial x} + \frac{\partial(ws)}{\partial z} = 0, \quad (2.3)$$

and the two-dimensional mass conservation equation (incompressible flow) is

$$\frac{\partial(u)}{\partial x} + \frac{\partial(w)}{\partial z} = 0, \quad (2.4)$$

where t is time, x & z are rectangular space coordinates with origins in the mean sea surface (positive seaward and downward), u & w are horizontal and vertical velocity components; p is pressure, ρ & ρ_0 are densities of estuarine water and freshwater, g is gravitational acceleration, and s is salinity.

The analysis next applies Reynolds averaging (i.e., splitting u, w & s into depth-averaged and depth varying parts)

$$u = \bar{u}(x) + u'(x, z), s = \bar{s}(x) + s'(x, z), w = \bar{w}(x) + w'(x, z), \quad (2.5)$$

and introduces mean eddy coefficients

$$-\frac{\partial \langle \overline{u'w'} \rangle}{\partial z} = \frac{\partial}{\partial z} \left(K_m \frac{\partial u}{\partial z} \right), \quad -\frac{\partial \langle \overline{u's'} \rangle}{\partial x} = \frac{\partial}{\partial x} \left(K_h \frac{\partial s}{\partial x} \right), \quad \text{and} \quad -\frac{\partial \langle \overline{w's'} \rangle}{\partial z} = \frac{\partial}{\partial z} \left(K_v \frac{\partial s}{\partial z} \right), \quad (2.6)$$

where K_m is tidally averaged vertical eddy viscosity, K_h & K_v are horizontal and vertical eddy diffusivities. ‘ $\overline{\quad}$ ’ is turbulent average, $\langle \quad \rangle$ is average over the tidal period, $u', w',$ and s' are the turbulent components, and u & s are averages over a tidal period of horizontal velocity and salinity. With the additional assumption of vertically uniform K_m , the horizontal momentum equation becomes:

$$\frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + w \frac{\partial u}{\partial z} = -\frac{1}{\rho} \frac{\partial p}{\partial x} + K_m \frac{\partial^2 u}{\partial z^2} \quad (2.7)$$

In the analysis of the James River data, Pritchard (1954, 1956) argues that the dominant terms in the longitudinal equation of motion are the pressure gradient and the vertical eddy flux of momentum; all other terms are of second order. The set of the model equation becomes

$$0 = -\frac{1}{\rho} \frac{\partial p}{\partial x} + K_m \frac{\partial^2 u}{\partial z^2}, \quad (2.8)$$

$$0 = -\frac{1}{\rho} \frac{\partial p}{\partial z} + g, \quad (2.9)$$

$$\frac{\partial s}{\partial t} + \frac{1}{A} \frac{\partial (usA)}{\partial x} + \frac{\partial (ws)}{\partial z} = \frac{1}{A} \frac{\partial}{\partial x} \left(AK_h \frac{\partial s}{\partial x} \right) + \frac{\partial}{\partial z} \left(K_v \frac{\partial s}{\partial z} \right), \quad (2.10)$$

$$\frac{\partial (u)}{\partial x} + \frac{\partial (w)}{\partial z} = 0. \quad (2.11)$$

From equation (2.9), $\frac{\partial p}{\partial z} = g\rho \rightarrow p(z) = \int_z^\eta \rho g dz = \rho g (\eta - z)$,

and the pressure gradient becomes (assuming vertically uniform $\frac{\partial \rho}{\partial x}$):

$$\frac{\partial p}{\partial x} = \rho g \frac{\partial \eta}{\partial x} + (\eta - z)g \frac{\partial \rho}{\partial x} = \rho g \frac{\partial \eta}{\partial x} - g \frac{\partial \rho}{\partial x} z, \quad (2.12)$$

where η is neglected using the rigid lid assumption (i.e., η is insignificant compared to z).

The relationship between salinity and density is given by the approximate equation of state

$\rho = \rho_0 (1 + \beta s - \alpha T)$, where the temperature (T) effect can be disregarded. Hence

$$\rho = \rho_0 (1 + \beta s) \quad (2.13)$$

where β is the coefficient of saline contraction $= 7.7 \times 10^{-4} [\text{psu}^{-1}]$.

Hence, $\frac{\partial \rho}{\partial x} = \frac{\partial \rho_0}{\partial x} + \rho_0 \beta \frac{\partial s}{\partial x} = \rho_0 \beta \frac{\partial s}{\partial x}$, since $\frac{\partial \rho_0}{\partial x} = 0$.

Thus, equation (2.12) becomes $\frac{\partial p}{\partial x} = \rho g \frac{\partial \eta}{\partial x} - g \rho_0 \beta \frac{\partial s}{\partial x} z$, and the x-axis momentum equation becomes

$$0 = -g \frac{\partial \eta}{\partial x} + g \beta \frac{\partial s}{\partial x} z + K_m \frac{\partial^2 u}{\partial z^2}. \quad (2.14)$$

Taking $\frac{\partial}{\partial z}$ of eq (2.14) results in

$$\frac{\partial^3 u}{\partial z^3} = -\frac{g\beta}{K_m} \frac{\partial s}{\partial x}. \quad (2.15)$$

Equation (2.15) is integrated with the following boundary conditions:

$$u(z = -H) = 0,$$

$$\frac{\partial u}{\partial z}(z = 0) = 0, \text{ and}$$

$$\frac{1}{H} \int_{-H}^0 u dz = \bar{u} = \frac{Q_r}{A}, \quad (2.16)$$

where Q_r is the river flow, and A is the cross-sectional area. After integration and applying boundary conditions, the solution reads:

$$u = \bar{u} \left(\frac{1}{2} - \frac{3}{2} \zeta^2 \right) + u_E (1 - 9\zeta^2 - 8\zeta^3), \quad (2.17)$$

where $\zeta = z/H$. The scale of the exchange flow is given by

$$u_E = \frac{g}{48} \frac{H^3 \beta}{K_m} \frac{\partial s}{\partial x}, \quad (2.18)$$

and indicates the dependency of the exchange flow on the salinity gradient $\left(\frac{\partial s}{\partial x}\right)$.

Internal tidal asymmetry (Jay & Musiak, 1994) is another important cause of estuarine circulation. Internal tidal asymmetry is caused by tidally variable vertical and horizontal density gradients, which cause a difference in stratification between flood and ebb tides. Stratification suppresses the mixing in an estuary, producing an asymmetry in flood mixing compared to ebb mixing. This produces an asymmetry in the vertical shear in the along channel momentum that produces the estuarine circulation (u_E). Jay and Musiak (1996) suggest that about half of the two-layer estuarine circulation in the Columbia River under moderate to high flow conditions is caused by internal asymmetry, with the other half being caused by gravitational circulation.

The dependency of the salinity profile on depth is explored by taking the depth average for the salt conservation equation (2.10), with $w=0$, $K_v \frac{\partial s}{\partial z} = 0$ at top and bottom, resulting in

$$\frac{\partial \bar{s}}{\partial t} + \frac{1}{A} \frac{\partial (\bar{u} \bar{s} A)}{\partial x} + \frac{1}{A} \frac{\partial (\overline{u' s' A})}{\partial x} = \frac{1}{A} \frac{\partial}{\partial x} \left(A K_h \frac{\partial \bar{s}}{\partial x} \right) \quad (2.19)$$

Subtracting eq (2.19) from (2.10), yields:

$$\frac{\partial s'}{\partial t} + \frac{1}{A} \frac{\partial [us - \bar{u}\bar{s} - \overline{u' s'}]A}{\partial x} + \frac{\partial (ws)}{\partial z} = \frac{\partial}{\partial z} \left(K_v \frac{\partial s'}{\partial z} \right), \quad (2.20)$$

where $\frac{\partial s'}{\partial x}$ is assumed to be much less than $\frac{\partial \bar{s}}{\partial x}$, allowing cancelation of the K_h term. For partially mixed and well-mixed systems, the dominant steady balance for s' is between the creation of stratification by vertical shear and its destruction by vertical mixing (Pritchard 1954). Taking K_v as a constant in the z direction (MacCready, 2004) yields.

$$u' \frac{\partial \bar{s}}{\partial x} \cong K_v \frac{\partial^2 s'}{\partial z^2} \quad (2.21)$$

where $u' \frac{\partial \bar{s}}{\partial x}$ arises from $\frac{\partial us}{\partial x}$ in eq (2.20), (MacCready, 2004)

This may be integrated directly as:

$$s' = \frac{H^2}{K_v} \frac{\partial \bar{s}}{\partial x} \left[\bar{u} \left(-\frac{7}{120} + \frac{1}{4} \zeta^2 - \frac{1}{8} \zeta^4 \right) + u_E \left(-\frac{1}{12} + \frac{1}{2} \zeta^2 - \frac{3}{4} \zeta^4 - \frac{2}{5} \zeta^5 \right) \right]. \quad (2.22)$$

Equation 2.16 suggests a strong dependence of estuarine circulation on depth ($u_E \propto H^3$). Furthermore, changing depth could also modify the longitudinal salinity gradient and eddy diffusivity (Chant et al., 2018). In addition, the salt content and the associated salinity

gradient are vary with river discharge, exchange flow, and tidally induced salt transport processes, as suggested by equation 2.20. Chant et al. (2018) reported that an increase of 15% depth in Newark Bay, New Jersey depth resulted in doubling the exchange flow, a slight increase in salinity and longitudinal salinity gradient, a decrease in tidal current amplitude, and a spatially variable change in the tidal range.

2.2.2 Tidal changes

Tides can be described by the Reynolds averaged shallow water equations in one dimension that assume a nonrotating flow, and uniform density. The continuity equation (mass conservation) can be written as

$$\underbrace{\frac{dh}{dt}}_{\text{local change of surface height}} + \underbrace{u \frac{dh}{dx}}_{\text{advection of height}} + \underbrace{h \frac{du}{dx}}_{\text{volume convergence}} = 0 \quad (2.23)$$

This shows that the rate of local change in the water surface is controlled by the advection of height and the volume convergence, where (h) is the local water depth, (u) is the longitudinal velocity (x-direction), and (t) is the time (Jay, 1991; Friedrichs & Aubrey, 1994).

Similarly, using the balance of momentum with the hydrostatic pressure assumption yields the equation of motion. This states that the rate of local change of velocity is controlled by the advection of momentum, the pressure gradient, and bottom stress.

$$\underbrace{\frac{\partial u}{\partial t}}_{\text{local change of velocity}} + \underbrace{u \frac{\partial u}{\partial x}}_{\text{advection of momentum}} + \underbrace{g \frac{\partial h}{\partial x}}_{\text{pressure gradient}} + \underbrace{\frac{\tau_b}{\rho h}}_{\text{bottom stress}} = 0 \quad (2.24)$$

where τ_b is bottom stress, and ρ is water density (Jay, 1991; Friedrichs & Aubrey, 1994)

. Both equations of continuity and momentum show that tidal amplitude and velocity can be affected by the increase of the local water surface height (h) in response to sea-level rise.

2.2.3 Effect of river flow and bathymetry on salinity intrusion

The river flow has a direct effect on salinity intrusion; high flow is associated with reduced salinity intrusion, and vice versa (Hansen & Rattray, 1965; 1966; MacCready, 1999; MacCready, 2007; Monismith et al., 2002; Geyer & MacCready, 2014). A theoretical analysis of salinity intrusion, presented by Monismith et al. (2002) based on the tidally averaged salt conservation equation and the work of Hansen & Rattray (1965) follows. The analysis starts with the salt conservation equation that is averaged tidally and cross-sectionally:

$$A(x) \frac{\partial S}{\partial t} + \frac{\partial}{\partial x} (QS) = \frac{\partial}{\partial x} \left[K_x(x) A(x) \frac{\partial S}{\partial x} \right], \quad (2.25)$$

where: A =local cross-section area, x =the distance landward from the mouth of the estuary, S =the salinity, Q =the river flow, and K_x =the longitudinal dispersion coefficient. The vertical shear of the flow is given by:

$$U_g \sim \frac{\beta g \left(\frac{dS}{dx}\right) H^3}{\nu_t} \quad (\text{Hansen \& Rattray, 1965}) \quad (2.26)$$

$$\text{also, } K_x \sim \frac{(\beta g)^2 \left(\frac{dS}{dx}\right) H^8}{\nu_t^3} \quad (\text{Hansen \& Rattray, 1965}) \quad (2.27)$$

at steady state, the net flux is zero, so that

$$\frac{-Q}{A} S = K_x \frac{dS}{dx} = \frac{\alpha (\beta g)^2 H^8}{\nu_t^3} \left(\frac{dS}{dx}\right)^3, \quad (2.28)$$

where $\alpha = 5.4 \times 10^{-5}$ (Monismith et al., 2002).

$$\text{Re-arranging: } \left(\frac{dS}{dx}\right)^3 = \frac{-Q \nu_t^3}{W \alpha (\beta g)^2 H^9} S, \quad (2.29)$$

which in terms of salinity intrusion X_s gives

$$\frac{S_o}{X_s} \sim \frac{Q^{1/3} \nu_t S_o^{1/3}}{(W \alpha)^{1/3} (\beta g)^{2/3} H^3} \quad \text{or} \quad X_s = \frac{(W \alpha)^{1/3} (\beta g S_o)^{2/3} H^3}{Q^{1/3} \nu_t} \quad (2.30)$$

where X_s is the salinity intrusion length from the mouth of the estuary toward the upstream (where the bottom salinity is 2-psu (Monismith et al., 2002) and (MacWilliams et al., 2015)). Here, W is the width, α is the constant of proportionality (5.4×10^{-5}), β is the saline expansivity ($7.7 \times 10^{-4} \text{ psu}^{-1}$), S_o is ocean salinity, H is the local depth, Q is the river flow, and ν_t is the eddy diffusion coefficient (ν_t can be scaled to $u_t H$, and u_t is the tidal velocity).

Salt is transported landward due to tidal dispersion, while seaward transportation is driven by river discharge (eq. 2.30). The river flow freshens the estuary, which decreases

the salinity intrusion. On the other hand, increasing channel depth leads to an increase in stratification, reduces mixing, and eventually increases salinity intrusion landward at the bottom (eq. 2.30).

From the foregoing analysis, it appears that salinity intrusion varies with the second power of channel depth ($X_s \sim H^2$) and inversely with the cubic root of the river discharge ($X_s \sim Q^{-1/3}$). This means that, theoretically, the salinity intrusion is much more sensitive to changes in bathymetry than to river discharge (see Figure 2-1). The percentages in Figure 2-1 represent the effect of doubling the two parameters H and Q in Equation 2.30. As shown, the second power of H produces a much larger effect on salinity intrusion than the cubic root of Q . The processes controlling salinity intrusion is more complex than shown in Figure 2-1, because of tidal asymmetry, bottom drag nonlinearity, and other non-linear interactions that are functions of H and Q . Due to the complex 3D bathymetry and the presence of two estuary channels in the Columbia River Estuary a numerical model is needed to explore the response of the estuary to changed bathymetry and river discharge.

Figure 2-1. Percentage of contribution to salinity intrusion by river flow (Q) and channel depth (H) variations. The percentage approximates the theoretical scaling $X_s \sim H^2/Q^{-1/3}$. The percentage represents the effect of the 2nd power of H and the cubic root of Q , the same percentage of doubling the two parameters (H & Q).

2.2.4 Bifurcation theory

According to Buschman et al. (2010), the subtidal discharge division at a bifurcation can be quantified as

$$\Psi = \frac{\langle Q_1 \rangle - \langle Q_2 \rangle}{\langle Q_1 \rangle + \langle Q_2 \rangle}, \quad (2.31)$$

where $\langle \rangle$ indicates tidally average, Q_1 is the first channel discharge, Q_2 is the second channel discharge, and Ψ is the discharge asymmetry index. When the index is zero, both channels have an equal share of flow. If the index is positive, then the first channel discharge is larger than the second channel. If it is one, all the flow goes through channel one. The subtidal flow due to river discharge and tides can be decomposed into

$$\langle Q \rangle = Q_r + \langle Q_t \rangle + \langle Q_{rt} \rangle, \quad (2.32)$$

where Q_r is river flow, Q_t is the tidal flow, and Q_{rt} is the tide-river interaction flow.

Similarly, the asymmetry index can be decomposed into three components

$$\Psi = \Psi_r + \Psi_t + \Psi_{rt} , \quad (2.33)$$

where Ψ_r the asymmetry in freshwater flow, Ψ_t is the tidal flow asymmetry, and Ψ_{rt} is the river-tide flow interaction asymmetry (Sassi et al., 2011).

However, the vertically integrated tidal flow can be decomposed into

$$\langle Q_t \rangle = \underbrace{W \langle U' \eta' \rangle}_{Q_S \text{ (Stokes transport)}} + \underbrace{W h \langle U \rangle}_{Q_R \text{ (return discharge)}} + \underbrace{W \langle U \rangle \langle \eta \rangle}_{\text{residual}} , \quad (2.34)$$

where U is the cross-section average velocity and equal to $\langle U \rangle + U'$ (prime denotes the tidal variation), W is the channel width, h is the water depth, and η is water level (total water depth, $d=h+\eta$). The residual term is negligible since it is small when $\langle \eta \rangle$ is near zero. For a single channel with a constant width, Q_S (landward) and Q_R (seaward), the water storage balance ($Q_S=Q_R$). Hence, Q_t is small or zero. In network channels, they (Stokes and return flows) do not necessarily balance, and Q_t has a non-zero value (Buschman et al., 2010).

2.2.5 Estuarine classification theory (parameter-space)

Geyer & MacCready, (2014) used two parameters to classify estuaries: the freshwater Froude number that represents river flow effects,

$$Fr_f = U_R/(\beta g s_{ocean} H)^{1/2}, \quad (2.35)$$

and a mixing number that represents the tidal forcing,

$$M^2 = \frac{C_D U_T^2}{\omega N_0 H^2}. \quad (2.36)$$

Here, U_R is the river flow velocity, β is the saline expansivity, s_{ocean} is the ocean salinity, H is the channel depth, C_D is the drag coefficient, U_T is the tidal velocity, ω is the tidal frequency and N_0 is buoyancy frequency = $(\beta g s_{ocean} H)^{1/2}$. The freshwater Froude number, Fr_f , represents the ratio of the river flow net velocity to frontal propagation speed, while M^2 represents the ratio of the tidal timescale to the vertical mixing timescale, or equivalently the effectiveness of tidal mixing and tidal straining. Note that a critical mixing condition occurs when the tidal boundary layer penetrates the entire column and reaches the surface. This occurs when

$$\left(\frac{s_{ocean}}{\delta S_i}\right)^{1/2} M^2 \approx 1, \quad (2.37)$$

where δS_i is the vertical salinity difference. The equation of stratification from (MacCready, 1999) is:

$$\frac{\delta s}{s_{ocean}} = \alpha Fr_f^{2/3}, \quad (2.38)$$

where $\alpha = 3.4$ based on (Geyer, 2010). Substituting Eq. 2.38 into Eq. 2.37 yields the following condition for vertical mixing,

$$\alpha^{-1/2} Fr_f^{-1/3} M^2 \approx 1. \quad (2.39)$$

The parameter-space plot in Figure 2-2 depends on two parameters: the freshwater Froude number (Fr_f) and the mixing parameter (M^2). The freshwater Froude number is the river velocity scaled by the maximum possible frontal propagation speed, while the mixing number is the ratio of the tidal timescale to the vertical mixing timescale.

The red solid diagonal line in Figure 2-2 divides the estuaries that should always remain stratified (to the left of the red diagonal line) from those in which boundary generated mixing reaches the surface within a tidal cycle (to the right of the red diagonal line). Each estuary occupies an area inside the diagram, because of the tidal velocity spring-neap variation (x-axis) and seasonal river flow variation (y-axis). Furthermore, the parameter-space classified estuaries into six categories; salt-wedge, strongly stratified, SIPS, partially mixed, well mixed, and time-dependent salt-wedge. An estuary is a salt-wedge when seawater advances into the estuary as a wedge-shaped bottom layer, which slightly mixes with the freshwater upper layer because of weak tidal motion and high river flow. When the tidal mixing is significant enough to erase the salt-wedge, then the estuary is partially mixed. It is well mixed if a shallow estuary has a strong tidal mixing and low river flow. SIPS or Strain-Induced periodic Stratification is the switching between mixing and stratified states over a single tidal cycle (Simpson et al., 1990); the mixing of the water column is due to tidal straining, while the interaction of a horizontal density gradient with a vertical shear produces a periodic stratification.

From the previous definitions - the mixing parameter (M^2), and freshwater Froude number equations (Fr_f) - it can be concluded that the classification in Figure 2-2 strongly

depends on the river flow, tidal velocity, and estuarine depth. The three parameters have likely changed in the Columbia River Estuary over the last 150 years because of both bathymetric and hydrological changes.

Figure 2-2. Estuarine parameter space (Geyer & MacCready, 2014).

2.3 Literature review

2.3.1 Literature on modeling hydrodynamic and transport processes in estuaries

The hydrodynamics of estuaries are described by the shallow water equations (mass momentum conservation, and the equation of state). In order to solve these equations analytically, many assumptions and simplifications are considered. Since the estuarine

transport and circulation are usually turbulent, time-dependent, and 3-dimensional (Sheng & Wang, 1988), and because the flow features complexity and a strong density gradient (Kärnä et al., 2015), 3d-hydrodynamic modeling is often necessary to better understand the spatial and temporal structure of the flow (Aristizábal & Chant, 2013; MacWilliams et al., 2015).

Numerical modeling has been used for several decades as a useful tool in estuarine investigations (Ganju et al., 2016). In order to validate results, a calibration to real data is necessary. Model accuracy depends on many factors such as; grid resolution, length of simulations, diffusion coefficients, boundary conditions, and bottom roughness. The sensitivity to grid resolution, eddy viscosity, and eddy diffusivity were presented, for example, by Ralston (2010). He used a 3D hydrodynamic model to simulate the Merrimack estuary and compared it to observed data. The comparison included water level, salinity, and velocity. The skill assessment showed that the model well represented the salinity spatial structure as well as the along and cross estuary salt flux. Similarly, Warner et al. (2005) addressed the importance of bottom friction and grid resolution by using a 3D numerical model for the Hudson River. The model compared water level, velocity, salinity (vertical and horizontal), and salt flux. The model was able to capture the stratification during spring-neap tides. Although the top-bottom salinity difference was captured well, the vertical salinity structure was not well represented. The model was found to be sensitive to the bottom drag coefficient and salinity at the open boundaries, such that decreases in the bottom roughness led to higher velocities near the bottom. This prevented the migration

of the bottom boundary layer into the upper column and reduced mixing, and consequently increased the salinity intrusion length.

For the Columbia River Estuary, Kärnä et al. (2015) constructed a numerical model based on the finite element and finite volume principle. It was tested under variable conditions that covered the spring-neap tide period as well as the conditions of high and low flow. The model was able to simulate the CRE circulation. It used a skill assessment and found that bed friction variation is the key to get a better simulation. The model successfully predicted the salinity intrusion during low and high flow conditions but it underestimated the gravitational circulations especially during spring and neap conditions. Kärnä et al. (2015) also found that the density field is smoother in the model than in measurements, due to numerical mixing. Baptista et al. (2005) recommend that model validation should extend for a long period of time (months-years) to eliminate local bias. They used an Eulerian-Lagrangian three-dimensional model of the Columbia River estuary, plume, and shelf. This model, an earlier version of the Kärnä et al. (2015) model, simulated water levels well but could not well simulate the plume dynamics and the salinity intrusion.

Using modeling techniques is an effective way to address estuarine salinity intrusion and the factors that can contribute to its variability. Tidal advection and diffusion are among the most important factors, especially during low to moderate river flow (Wei et al., 2016). The geometry of the estuary is another important factor in salinity intrusion, as found by MacWilliams et al. (2015). In their model, they tried to simulate the

hydrodynamics of the San Francisco Estuary for a 3-year period that covered a variety of flow and tidal conditions. By comparing the model to actual data and evaluating the model using multiple metric assessment skills, they found that the model was appropriate to use to predict salinity and circulation. By applying the model, they found out that the location, area, volume, and average depth of the low salinity zone (0.5-6-psu) is proportional to salinity intrusion length (2-psu daily averaged bottom salinity) and depends on the estuary geometry. MacCready (2007) applied a tidally and width-averaged numerical model to explore the subtidal salinity adaptation to tidal mixing and river flow variation. An analytical expression [$T_{ADJ} = (1/6) L/\bar{u}$] was developed for the time that it takes the salinity fields to adjust to a step-change in river flow. He found that the adjustment time (T_{ADJ}) is directly proportional to the salinity intrusion length (L) and inversely proportional to width-averaged river velocity ($\bar{u}$). He also pointed out the importance of adjustment time compared to a forcing time scale. For example, the adjustment time for the Hudson River is much faster than the seasonal river flow variation timescale, while it is nearly equivalent to the tidal variation period for spring-neap mixing (~ 7 days).

2.3.2 Flow partitioning literature

Investigating flow division at river junctions is important to understand the transport and pathways of sediments, contaminants, and nutrients. A larger share of river flow may result from a larger water surface slope of one-channel compared to the other. The slope difference can be caused by an asymmetry upstream of the junction. Asymmetries can include different channel directions with respect to the feeding river

channel and bends or curves in channels upstream of the junction. The channels' geometry and their hydraulic roughness can also control the water division. In addition, the flow division gets more complicated by enhanced friction from non-linear interaction between river flow and tides (Buschman et al., 2010).

Swart (2015) suggested that salinity intrusion and tidal propagation in a multi-channel estuary are significantly different from a single channel estuary. He used an idealized model of a two-channel estuary with different length channels that meet at a tidal junction upstream. He found that the tidal energy could transfer from the shorter channel to the longer channel at the junction. This happens because of the tidal velocity phase difference between the two channels, where the phase difference increases linearly with the channel length, so the tidal wave in the shorter channel will round the corner faster. This can affect the water level amplitudes and gradients, which increases the freshwater percentage in the shorter channel. The result of this freshwater partitioning difference could lead to more salinity intrusion in the longer channel. This finding agrees with Buschman et al. (2010), who showed that tidal motion makes the freshwater division more unequal at a tidal junction, with smaller tidal amplitude and larger subtidal water level gradient in the shorter channel.

For the CRE, Lutz et al. (1975) reported that for mean daily river flow of less than $4670\text{m}^3/\text{s}$ or greater than $5380\text{m}^3/\text{s}$, the South Channel at Astoria Reach conveys more flow during ebb than during flood, while the reverse occurs in the North Channel and shallow mid-estuary (Middle Channel). This produces a net clockwise circulation between the two

channels. For river discharge in between these values, the net circulation was counterclockwise. The percentage of tidal flow in the North Channel near Astoria Reach is about 76% of flood flow, and 69% of the ebb flow. Furthermore, their observation near Astoria indicates that the maximum surface velocities for ebb and flood were 2.4 and 1.8 m/s, respectively, while the lateral and time-averaged velocities were 0.9 and 0.6 m/s, respectively. They also pointed out that the direction of flow in the channels (North & South) runs parallel to the riverbank, but diagonally in the Middle Channel. Although the South Channel has been deepened since the 1970s (see Table 1-1), a similar pattern may still exist today. This question will be investigated using modeling.

2.3.3 Salinity intrusion literature

As discussed above, idealized scaling of salinity intrusion (Equations 2.25 thru 2.30; Hansen & Rattray, 1965), suggest that salinity intrusion is inversely proportional to the river flow to the power of one third (assuming steady-state conditions and zero net salt flux; see e.g., (Monismith et al., 2002; MacCready, 1999). Following this theoretical work, many empirical studies have interpreted salinity intrusion using a power-law relation ($X \sim Q^{-n}$). Abood (1974), found $n=1$ in the Hudson River estuary for low flow and found $n=1/3$ for high flow conditions. For the same estuary, n was found to be $1/5$ by Oey (1984). When including the boundary layer as an effective mixing length scale, Ralston et al. (2008) found a value of n equal to 0.35. When the full water depth was used, $n=0.4$ was found. Within San Francisco Bay, Monismith et al. (2002) found a value of $n=1/7$ and determined an adjustment time of about 2 weeks. The adjustment time is the time needed

for density and estuarine circulation to adapt to the change in river flow and tidal mixing (MacCready, 1999). Finally, n was found to be 0.1 for the Delaware Bay by Aristizábal & Chant (2013).

For the Columbia River Estuary (CRE), Chawla et al. (2008) used CTD (an instrument that can measure conductivity, temperature, and depth/pressure of a water column) cast data to measure the salinity intrusion along the South Channel. They recorded the salinity intrusion during two spring freshet periods. Their measurements suggested that the salt intrusion from the estuary entrance doubles when the river flow decreases by a factor of about four; i.e., that $n \approx 1/2$. Putra et al. (2015) used a multi-station calibration of a 3D flexible mesh model of the Columbia River Estuary and found that during spring tide salinity intrusion can reach 40 km upstream of the mouth of the CRE during low river flow condition. See comparisons of n in Table 2-1 for different estuaries.

Table 2-1. Salinity intrusion length (X_s), river discharge (Q) relation for different estuaries. Where K_{mo} is a mixing coefficient that uses the whole water column depth as an effective mixing length scale, while K_{mBL} uses the bottom boundary layer height.

Estuary	Discharge power	Researchers
Delaware Bay	$X_s \sim Q^{-0.1}$	(Aristizábal & Chant, 2013)
Columbia River Estuary	$X_s \sim Q^{-1/2}$	(Chawla et al., 2008a)
Hudson	$X_s \sim Q^{-0.4} \rightarrow K_{mo}$ $X_s \sim Q^{-0.35} \rightarrow K_{mBL}$	(Ralston et al., 2008)
San Francisco Bay	$X_s \sim Q^{-0.20} \rightarrow K_{mo}$ $X_s \sim Q^{-0.17} \rightarrow K_{mBL}$	(Ralston et al., 2008)
Theoretical	$X_s \sim Q^{-1/3}$	(Monismith et al., 2002)
San Francisco Bay	$X_s \sim Q^{-1/7}$	(Monismith et al., 2002)
Hudson	$X_s \sim Q^{-1/5}$	(Oey, 1984)
Hudson	$X_s \sim Q^{-1/3} \rightarrow$ low flow $X_s \sim Q^{-1} \rightarrow$ high flow	(Abood, 1974)

2.3.4 Tidal asymmetry and exchange flow literature

Tidal asymmetry is the difference in duration and current magnitudes of the flood and ebb tides. It can be either ebb or flood dominant. Whether an estuary location is ebb or flood dominant depends on the phase difference between the tidal components, while the degree of distortion depends on the ratio of constituent amplitudes (Jay & Musiak, 1996).

A main source of asymmetry is the interaction of principal lunar semidiurnal tides (M_2) with its overtides (e.g., M_4), the harmonic frequencies that are integer multiples of a principal constituent). Another source of asymmetry is the nonlinear interaction of semidiurnal tides (e.g., M_2 , N_2 , and S_2) and diurnal tides (K_1 and O_1), which produces harmonics generated from sum or differences of two frequencies (examples include MS_4 and MK_3). Both overtides and compound tides are generated from the distortion of the astronomical tides in shallow water (Nidzieko, 2010). The tidal asymmetry can be called "barotropic asymmetry" because it distorts the free surface and influences flood/ebb dominance. Meanwhile, river flow can reinforce the tidal asymmetry by increasing friction, which is the main cause of tidal distortion (Kukulka & Jay, 2003b).

The physical mechanism of the tidal asymmetry is the non-linear, shallow water processes that transfers energy from the tide to higher and lower frequencies. The nonlinearities in the equations of motions arise from the time-varying depth and width in the continuity equation, the time-varying depth and velocity in the quadratic friction in the momentum equation, and the nonlinear advection acceleration in the momentum equation

(Parker, 1991). See Figure 2-3 for an illustration of the mechanisms involved in barotropic tidal asymmetry.

Another type of tidal asymmetry known as "internal tidal asymmetry" can cause estuarine circulation related to correlations in tidally varying stratification and mixing (Jay & Musiak, 1994, 1996). Internal tidal asymmetry leads to density field distortion associated with energy transfer from the surface tide to higher and lower frequencies (Jay & Musiak, 1996). That is, internal asymmetry leads to differences in stratification between ebb and flood, which causes tidal differences in vertical mixing. This can lead to asymmetric vertical shear of the along-channel momentum, which ultimately causes two-layer residual estuarine circulation (Becherer et al., 2015). Estuarine circulation, or exchange flow, is simply the water entering the estuary near the bottom which is compensated by the outflow near the surface (tidally averaged circulation). It is caused by both gravitational circulation and internal asymmetry. The exchange flow is important to estuarine dynamics since it affects the transport of salt, nutrients and suspended sediments (Becherer et al., 2015; Wang et al., 2017).

Figure 2-3. Physical mechanism layout of tidal asymmetry

The estuarine circulation helps define the salinity structure. It can be quantified using total exchange flow (TEF). TEF is tidally averaged salt flux through an estuarine cross section, using salinity instead of spatial position in the calculation of the exchange flow. MacCready (2011) applied the TEF method to a numerical simulation of the Columbia River estuary for the period from April to September of 2005. He found that the tidal salt flux (F_T) dominated over exchange flux (F_E) in the outer part of the estuary because this region is within one tidal excursion of the mouth. By using the volume flux through the mouth section, he found that the average water inflow occurs over a narrow salinity range near oceanic values, while seaward transport of water occurs over a much larger salinity range (0-30psu). This occurs because of mixing within the estuary. Wang et

al. (2017) utilized the TEF concept and linked it to mixing in the Hudson estuary using a numerical model. They found that the total exchange flow reaches its maximum near minimum neap tide, but vertical (entrainment) salt transport reaches its maximum during the maximum spring tide.

2.3.5 Climate change literature

2.3.5.1 River hydrograph

The Columbia River discharge has changed and it continues to evolve because of climate change and flow control. Hamlet et al. (2013) studied the climate change impact on the Columbia Basin and predicted changes in spring snowpack and fundamental shifts from snow and mixed-rain-and-snow to rain-dominant behavior across most of the domain. They also anticipated decreases in summer precipitation, increases in extreme low flows in intensity for most of the river sites, and increases in flooding in mixed-rain-and-snow basins in mid-winter. Warner et al. (2015) investigated the changes in winter atmospheric rivers along the U.S. North West coast and forecasted increases in winter mean precipitation by 11%–18% by the end of this century. Lee et al. (2016) assessed the combined effects of changing natural flow due to climate change and dam operations that determine impacts to regulated flow for the Skagit River, a river north of Seattle. The results show that climate change is likely to cause substantial seasonal changes in both natural and regulated flow, with more flow in the winter and spring, and less in summer. The regulated 100-year flood is projected to increase by 23% by the 2040s and 49% by the 2080s (Lee et al., 2016).

The USBR (2016) reported potential hydrologic impacts of climate change on the Columbia River Basin. They stated that average mean-annual air temperature has increased by approximately 2°F since the late 1800s. Also, the Columbia River Basin has experienced a general decline in spring snowpack since the mid-20th century due to more precipitation occurring as rain (rather than snow) and earlier snowmelt runoff. Projected scenarios from USBR (2016) of future Columbia River flow above the Dalles are presented in Table 2-2. The data shows the simulated percent change from the 1990s (1980 to 2009) to the 2040s (2030 to 2059) and 2080s (2070 to 2099) for mean annual runoff, winter runoff (December-March), and spring runoff (April through July). Overall, a large increase in winter runoff is projected, while spring flows may either increase or decrease.

Table 2-2. Hydrologic modeling for the Columbia River above The Dalles [after Bureau of Reclamation, (2016)]

Change from 1990s period	2040s	2080s
Mean Annual Runoff (%)	-5 to +10	-4 to +15
Mean Dec-Mar Runoff (%)	+13 to +44	+26 to +91
Mean Apr-Jul Runoff (%)	-8 to +8	-17 to +10

2.3.5.2 Sea-level rise

Sea-levels are rising globally as a result of climate change. As air and water temperature increase, the Greenland and Antarctic ice sheets melt and cause more rapid rise in sea-level. Warmer water temperatures also causes thermal expansion and a rise in mean water levels. Sea-level rise can have a significant impact on salinity intrusion (Xiao et al., 2014). Rice et al. (2012) investigated the salinity intrusion change of Chesapeake Bay because of sea-level rise using a three-dimensional model. Their results showed that

the James River's mean salinity increases (2-4 ppt) when the sea-level rises (0.5-1m), and the effect is more severe during dry years. Furthermore, the isohalines move upstream as the sea-level rises.

Regarding global mean sea-level rise (GMSL), Church & White (2011) estimated historical global sea-level rise from coastal and island sea-level measurements and satellite data. This study concluded that the global sea-level has risen about 210mm between 1880 and 2009, with a linear trend of 1.7 ± 0.2 mm/year. For the satellite data era since 1991, the trend was 3.2 ± 0.4 mm/year. Hay et al. (2015) combined tide gauge records with modeling of the various contributing signals and indicated that GMSL rose at a rate of 1.2 ± 0.2 mm/year between 1901 and 1990. Dangendorf et al. (2017) used an area-weighting technique for averaging tide gauge records and estimated a sea-level rise of 1.1 ± 0.3 mm/year before 1990 and 3.1 ± 1.4 mm/year from 1993 to 2012. The methods and calculations of various researchers are shown in Figure 2-4.

Figure 2-4. Sea-level curves calculated by different research groups with various methods. The curves show the sea-level relative to the satellite era (since 1992), <http://www.realclimate.org/index.php/archives/2015/01/a-new-sea-level-curve/>.

Relative sea-level rise (RSL) is the sea-level rise relative to a local reference land level. RSL rise is more important than global MSL rise for most management purposes. In addition to MSL, it can be affected by many factors including changes in the gravitational attraction of ice, ocean-atmosphere interaction (El-Nino Southern Oscillation, or ENSO), and Pacific Decadal Oscillation (PDO) patterns which affect wind and temperature, terrestrial water storage, groundwater withdrawal, glacial isostatic adjustment, density changes, ocean circulation, and uplift and subsidence (NRC, 2012). RSL can be affected by the gravitational pull of the large mass of glacier ice sheets. When they melt, the gravitational pull decreases and produces a spatial pattern of regional sea-level. Regarding

Oregon, the RSL increases during the warm phase (El-Nino, up to 10-30cm for several months) and can drop during the cold phase (La-Nina) by a similar amount. Melting from Alaska and Greenland causes a fall in RSL, while melting from Antarctica causes a rise in RSL near Oregon. Additionally, the land is sinking at about 0.37mm/year in Astoria and 1.3mm/year in Fort Stevens (according to glacial isostatic adjustment models, GIA), which causes an increase in sea-level in the region (Burgette et al., 2009; Talke et al., 2020). Hence, because of the subduction zone effect on the area, the land at the coast is pushed upward. Recently, sea-level rise rates in the Eastern Pacific were suppressed by a change in wind stress pattern caused by the Pacific decadal oscillation (PDO), which led to the suppression of relative sea-level rise (Merrifield et al., 2012). Once this pattern changes again, the relative sea-level rise could match the global projections of sea-level or even surpass them (Bromirski et al., 2011).

The projection of sea-level rise in Oregon (including global and local effects) suggested by NRC (2012) suggests that the RSL could increase by 1.43m by 2100; see Table 2-3. Additionally, a potential earthquake ≥ 9 in magnitude could rise the regional sea-level (RSL) an additional 1-2m over the projected values, by causing subsidence of 2m at the coast, zero at the upstream boundary, and 1.5m at Astoria (Kalmbacher & Hill, 2015), see Figure 2-5 & Figure 2-6. NRC (2012), Pfeffer et al. (2008), and Vermeer & Rahmstorf (2009) predicted that the global sea-level will rise about 1-2m along the Oregon shore, which will increase flooding and may affect salinity intrusion.

Table 2-3. projection of RSL near Oregon coast as suggested by NRC (2012)

Projection (yrs)	2030	2050	2100
RSL (cm)	-4 to + 23	-3 to 48	10 to 143

Figure 2-5. Estimated land subsidence from a magnitude 9 earthquake on Astoria and Oregon coast (OSSPAC, 2013).

Figure 2-6. Change in the bathymetry of CRE and Oregon coast from magnitude 9-modeled earthquake after Kalmbacher and Hill (2015). Note: negative values represent the upward displacement

Chapter 3 Flow Bifurcation in the Lower Columbia River Estuary**3.1 Introduction**

River discharge and tides exert a dominant influence on barotropic estuarine processes and affect stratification, mixing, circulation, and hence salinity intrusion and other transport processes (Hansen & Rattray, 1965; Hansen & Rattray, 1966; Jay & Musiak, 1996; MacCready, 1999; Monismith et al., 2002; MacCready & Geyer, 2010). Because many estuaries have only one primary channel, theory has traditionally evaluated estuarine dynamics within a single channel, assuming that river discharge enters far upstream of estuarine circulation (Monismith et al., 2002; Ralston et al., 2008). Although this approach has yielded many insights into tidal dynamics (e.g., Godin, 1985, 1999; Jay, 1991), a simple theory is inadequate when multiple channels are found in an estuary, as is frequently the case. Examples that received early attention include a) the Thames Estuary in the Southeast of Great Britain with flood-ebb asymmetric tides on each channel (Robinson, 1960); b) the Humber Estuary on the east coast of Northern England, where channel width determines whether channels are flood or ebb dominant (Robinson, 1960); and c) the Wadden Sea in the Netherlands where scalars are dispersed chaotically (Zimmerman, 1986). More recently, it has become clear that the amount of tidal flow and river flow in the channels of multi-channel systems is not simply a function of channel cross-section. Thus, the distribution of tidally-averaged flow may vary, for example, as a function of depth, channel curvature, river discharge and/or tidal forcing (Buschman et al., 2010; Sassi et al., 2011; and Zhang et al., 2017).

Here, I use the multi-channel Lower Columbia River Estuary (LCRE) to examine numerically how river flow is distributed between channels, how this division has been altered by navigational development, and finally, how LCRE tidal processes, in general, have been altered by navigational development.

The Lower Columbia River Estuary (LCRE) is a multi-channel estuary with two primary channels and multiple smaller ones. It is characterized by strong tidal currents as high as 2-4 m/s (Gonzales, 1984; Jay & Smith, 1990) and large river discharge that varies seasonally from 2000 to 17,000 m³/s (Naik & Jay, 2011). The seasonal variation in the tidal range and river discharge along with the non-uniformity of the estuary geometry can lead to laterally variable water levels (Hudson et al., 2017), which can then lead to lateral variations in the along channel water surface slope and an uneven distribution of flow among channels.

The Columbia River has been strongly altered in terms of geometry and river hydrograph over the last 150 years (Sherwood et al., 1990; Simenstad et al., 1992; Naik & Jay, 2005, 2011). Deepening the navigation channel has increased the tidal range and reduced the river slope in the CRE (Helaire et al., 2019) such that the tidal range has increased by 5.6% (~0.11m) near Astoria since 1855 (Talke et al., 2020). Changes in river flow and tidal range have likely altered the distribution of freshwater fluxes in the estuary along with the transport of sediment and salt. The changes in salt transport are investigated in chapter 4. Previous analyses have suggested that until the beginning of the 20th century, the North Channel was the primary conduit for river inflow and tidal transport of water

(Hamilton, 1990). Over the past century, however, the majority of river flow has been diverted into the South Channel, through dredging and the construction of islands and other flow training structures (Sherwood et al., 1990), although the majority of tidal exchange still occurs at the North Channel (Lutz et al., 1975; Chawla et al., 2008; Jay & Smith, 1990).

In this chapter, I analyze via numerical modeling how the river discharge distribution between the North and South Channels of LCRE has changed over time, and I explore the factors that have affected the distribution. Also, I compare the modern hydrodynamics of the estuary to the historical one (150 years ago) to quantify the tidal change inside the estuary due to human alteration of the system (channel deepening and river discharge management).

3.2 Theory and methods

According to Manning's equation (Manning, 1891), steady, uniform flow (Q) in open channels depends on bed friction (n), channel cross-sectional area (A), hydraulic radius (R_h), and channel slope (S), where the bed slope is the same as water surface slope under the uniform condition assumption. This leads to:

$$Q = \frac{1}{n} A R_h^{2/3} S^{1/2} \quad [SI], \quad (\text{Manning, 1891}) \quad (3.1)$$

Using Manning's arguments, the ratio of river flow entering two channels at a bifurcation is a function of the ratios of Manning's equation in the two channels (Ramamurthy et al., 1990):

$$\frac{Q_1}{Q_2} = \frac{\frac{1}{n_1} A_1 R_{h1}^{2/3} S_1^{1/2}}{\frac{1}{n_2} A_2 R_{h2}^{2/3} S_2^{1/2}} \quad (3.2)$$

If I define the following terms,

$$R_Q = \frac{Q_1}{Q_2}, R_n = \frac{n_2}{n_1}, R_A = \frac{A_1}{A_2}, R_{R_h} = \frac{R_{h1}^{2/3}}{R_{h2}^{2/3}}, \text{ and } R_S = \frac{S_1^{1/2}}{S_2^{1/2}} \quad (3.3)$$

Then,

$$R_Q = R_n \cdot R_A \cdot R_{R_h} \cdot R_S, \quad (3.4)$$

Thus, the ratio of flow of two channels braided from a channel is a function of the ratio of their cross-sectional areas (R_A), relative roughness (R_n), hydraulic radii (R_{R_h}), and slopes (R_S). From this simple scaling, it is possible to explore which factors dominate the bifurcation of flows, and how changes from the historical to modern bathymetry alter the balance. However, as shown in Buschman et al. (2010), tidal currents affect the tidally-averaged friction, thereby influencing bifurcation and making it a function of the spring-neap cycle. Moreover, the above analysis assumes barotropic conditions and would require modification if salinity stratification were present. The possible effects of tidal forcing and stratification are investigated later using model results.

The residual flow in the estuary contains brackish water. To distinguish the freshwater percentage, I used the fraction of freshwater (F) equation (Neal, 1966):

$$F = \frac{S_o - S_x}{S_o}, \quad (3.5)$$

where S_o is the ocean salinity, and S_x is the salinity inside the estuary. A fraction equal to one indicates 100% freshwater, while $F = \text{zero}$ means there is no fresh water at all. The fraction is calculated during spring and neap tides (tidally averaged) and a range of river discharges to explore the tide and river flow effect on freshwater bifurcation.

3.2.1 Salt fluxes

In a steady-state condition, the tidally averaged salt transport from river flow is balanced by the estuarine circulation. The salt transported seaward by the river discharge is equal to the input of salt due to estuarine circulation, or:

$$U_r S_o = a_0 U_e \Delta S, \quad (3.6)$$

where U_r is the river velocity and is defined by the river discharge over a cross-sectional area (Q_r/A), S_o is tidally averaged salinity, a_0 is a constant ~ 0.5 , U_e is the estuarine circulation, and ΔS is the stratification (the difference between the top and bottom salinities). The scaling of estuarine circulation (U_e) and stratification (ΔS) gives:

$$U_e = a_1 \frac{\beta g \frac{\partial s}{\partial x} h^2}{C_d U_t}, \quad (3.7)$$

$$\Delta S = a_1 a_2 \frac{\beta g \left(\frac{\partial s}{\partial x}\right)^2 h^3}{(C_d U_t)^2}, \quad (3.8)$$

where a_1 & a_2 are constant and depends on the shape of the salinity and velocity profile (~ 0.3 and ~ 50 , respectively, for the Hudson River), β is saline expansion, g is gravitational acceleration, $\frac{\partial s}{\partial x}$ is the salinity gradient, h is the depth, C_d is the drag coefficient, and U_t is

the tidal velocity). The combination of the above three equation (salt balance, estuarine circulation, and stratification gives:

$$\frac{\partial s}{\partial x} = \frac{C_d}{a} \frac{U_t U_r^{1/3} S_0^{1/3}}{(\beta g)^{2/3} h^{5/3}}, \text{ and } a = (a_0^2 a_1 a_2)^{1/3}. \quad (3.9)$$

This steady-state equation indicates that salt flux is sensitive to the 1/3 power of river flow, the first power of tidal velocity, and 5/3 power of depth. However, the salt flux is less sensitive when vertical mixing is included (Geyer, 2010). The possible change in river flow bifurcation and bathymetric changes are likely impacted the salt flux distribution in the LCRE. For the purpose of this research, the salt flux was calculated as follows:

$$q_s = u_{(x,z,t)} s_{(x,z,t)}, \quad (3.10)$$

$$q_s = (\bar{u} + \acute{u}) (\bar{s} + \acute{s}), \quad (3.11)$$

$$q_s = \bar{u}\bar{s} + \overline{\acute{u}\acute{s}}, \quad (3.12)$$

$$Q_s = \iiint q_s \, dzdydt, \quad (3.13)$$

where q_s is the salt flux in each cell, u & s are the instantaneous velocity and salt such that they are variable in time (t), and space (longitudinal (x), and vertical (z) coordinates). $\bar{u}$ & $\bar{s}$ are depth average velocity and salinity while $\acute{u}$ & $\acute{s}$ are depth varying. Q_s is the total tidally averaged salt flux through a cross-section.

3.2.2 Hydrodynamics models

I have built two 3D hydrodynamic models using Delft3D Flexible Mesh (D-flow FM) Suite 2017 HMWQ software by Deltares, a model that allows the use of unstructured grids (Roelvink and Van Banning, 1995). The modern and historical models represent, respectively, modern bathymetry (MB, Figure 3-1-a, b) and pre-1900 historical bathymetry (HB). The D-flow FM is a three-dimensional (3D) hydrodynamic simulation program that can solve the shallow water equations (continuity, momentum, and transport). The model grids cover the LCRE from the Beaver Army Terminal (Rkm 86, where hourly river discharge is available by USGS) to about 30km offshore (E-W) and cover about 60km of the coastal area (N-S), as shown in Figure 3-1-c. Each model contains over 58,000 cells (ranging from 100m inside the estuary to 1000m in the farthest point in the ocean, while the vertical resolution is 36 sigma layers (the vertical spacing and the height of a layer adjusts to the local water column height). This high resolution in the vertical is needed to simulate vertical density variation. Bottom friction is represented using the Chezy formulation, such that the Chezy coefficient (C) for the MB is 65 for the area between the ocean boundary and Skamokawa station (Rkm55), then decreases gradually to 40 near Wauna (Rkm66); it is constant thereafter (Figure 3-1a). The HB Chezy coefficient is depth-dependent (65 for deep channels and 25 for shallow areas) because the historical floodplain had more vegetation (and was much larger) than it is at present (Helaire et al., 2019).

The horizontal coordinate system is referenced to NAD83/Washington South, while the vertical datum is referenced to NAVD88. The modern digital elevation map

Chapter 3 Flow Bifurcation in the Lower Columbia River Estuary

(DEM) is based on data collected between 2008-2010 as multi-beam surveys (in water) and LiDAR surveys (subaerial, USACE, 2010). The historical model bathymetry was collected between 1868-1901 by the US Coastal Survey and the US Geodetic Survey. They were digitized and geo-rectified by Burke (2010).

3.2.3 Study area

The Columbia River (CR) is one of the biggest rivers in the US. Its length is about 1900km, and it has a drainage area of 660,480km² (Simenstad & Sherwood, 1990). The CR passes through seven states inside the United States and a Canadian province. It provides hydropower and has 14 dams on the main river and 450 dams on its tributaries (Osborn, 2012). Reservoir management has reduced peak spring flows by about 45% and enhanced low flow during fall and winter. Irrigation diversion and climate change have reduced peak flows by about 15-17%. The present mean annual river discharge to the ocean is about 7,300 m³/s (Naik and Jay, 2005, 2010; Jay and Naik, 2011; Talke et al., 2020).

The Lower Columbia River Estuary (LCRE) is characterized by mixed diurnal and semidiurnal tides with a tidal range at Astoria Tongue Point of 1.7-3.7 m (Kukulka & Jay, 2003 and Jay et al., 2016). As discussed above, the estuary has two primary channels: South and North, and several braided channels that connect between them (Figure 3-1a). I refer to these collectively as the “Middle Channel.”

The area of study covers the estuary from the mouth to the Beaver Army Terminal (BAT), which is about 87 km upstream of the mouth to the Pacific Ocean boundary (see

Figure 3-1a). The modern and historical models represent the four estuarine zones of eight LCRE zones, as defined by Jay et al. (2016). The four zones modeled are: The Entrance, and the Lower, Middle, and Upper Estuary Zones. Salinity intrusion is usually absent from the Upper Estuary. The zones not represented make up the tidal river part of the system, using the tidal definition of the boundary between the estuary and tidal river suggested by Hoitink & Jay (2016). More specifically, BAT is the most seaward point in the system where the lowest low waters occur on neap tides, not spring tides. Low waters are lower in a tidal river on the neaps, because the neap-spring difference in river stage is larger than the neap-spring difference in tidal range.

The estuary has seen dramatic changes in morphology over the past 150 years (Sherwood et al., 1990, Simenstad et al., 1994). Historical uses like fishing and log transport and storage have largely been replaced by a variety of industrial, agricultural, fishing, and tourism purposes. Navigational use has been ongoing since the pre-historical period, and the system has been heavily modified for navigation since the 1870s (Helaire et al., 2019). Major changes include jetty construction, channel deepening, and straightening. Blockage of side channels and installation of pile dikes have been used to concentrate flow in the navigation channel. Extensive wetland loss has occurred due to wetland reclamation for industry and agriculture (Simenstad et al., 1990). Marcoe & Pilson (2017) reported a loss of 68-70% of the tidal wetlands over the past 140 years. Dredging activities began in 1873 (U.S. Army, 2012), but major dredging began about 1909 (Sherwood et al., 1990). Some portion of the removed materials were moved offshore,

while 30-40% were likely placed in wetlands (Templeton & Jay, 2013). Especially relevant, the maintained controlling depth of the South (navigation) Channel is 13m today (and 14.6m at the entrance), while the natural controlling depth was about 6m before 1900 (Figure 3-1a & b). These depths are relative to mean lower low water (MLLW). Also, dredging and jetty construction at the mouth have created a narrower entrance (3.2 instead of 9.7 km wide) (Helaire et al., 2019).

Extensive hydrological change is primarily due to energy and irrigation demand, flood control, and temperature and precipitation changes (Naik & Jay, 2011). The biggest changes in the seasonal flow cycle started around 1960 due to water management (Sherwood et al., 1990), such that the flow in winter is now higher than 150 years ago by about 50% and lower by ~45-50% during Spring (Naik & Jay, 2011), see Figure 3-2. Higher flows are needed in winter for electric power generation, while flood control and irrigation withdrawal have reduced spring freshet flows and augmented low fall flows.

Figure 3-1. CR Bathymetrical change over the past 150-year: a) modern bathymetry (USACE, 2010); also shown are the grid boundaries, the calibration point locations (Cape Disappointment, Astoria Downtown, Tongue Point, Skamokawa, and Wauna), the North & South Channels layout, and the cross-sections along the estuary. b-historical bathymetry (Burke, 2010). Note: coordinates are in Washington South-State Plane Coordinate System (WA-SPCS).

Figure 3-2. Changes in the CR discharge at The Dalles (30-day moving average). The solid black and the dashed black curves represent, respectively, the historical (1879-1929) and the modern (1967-2016) daily mean river discharge (50th percentile). The shadow curves for each period extend from the 25th to 75th percentiles of river discharge.

3.2.3 The boundary conditions

I apply river discharge, tidal forcing, and meteorological forcing at the model boundaries because these are the major external forces that affect the circulation in the CRE (Baptista et al., 2004).

3.2.3.1 Meteorological forcing

Wind is the most significant factor among the meteorological factors, and it varies spatially and temporally. Over the coastal ocean, a median wind speed of 6m/s with a range of 0.1-24m/s was measured 1984-2016 at the National Data Buoy Center (NDBC) station 46029 at the Columbia River bar. Wind conditions can be categorized into three periods, according to Chawla et al. (2008a): winter, spring, and summer-fall period. The winter period occurs from October to March and is characterized by a strong but variable wind that comes from the south and south-west, which leads to downwelling near the coast. The spring freshet season occurs from April to June and is considered a transition period from downwelling to upwelling wind. The last period is from July to November (summer-fall), where wind from the Northwest is dominant and leads to upwelling conditions. For the modern bathymetry model (MB), a climatological average of wind speed and direction from NDBC station 46029 were applied to the offshore area of the model, while data from the National Oceanic and Atmospheric Administration (NOAA) at Astoria, OR-station 9439040 was used inside the estuary. The same wind forcing was used for the historical bathymetry model (HB) because there are no accurate historical records and so the models could be compared under the same conditions.

3.2.3.2 River boundary

Hourly river flow records were used to force the riverside boundary. The modern hourly flow records (50 percentile of the 1967-2016 river flow) were obtained from USGS 14246900 site on the Columbia River at Port Westward near Quincy, OR, at the Beaver

Army Terminal (BAT). The flow at BAT includes the Columbia River and its major tributaries, such as the Willamette, Lewis, Cowlitz, and Clatskanie rivers (USGS, 2016). This part of the river is also tidal so that the hourly data reflect tidal variability. The historical flow (1879-1929) is available as a daily record for the Columbia River (CR) at The Dalles (Naik & Jay, 2005), and the Willamette River (WR) at Albany (Naik & Jay, 2011). These daily records were used as rivers boundary conditions in a 2D hydrodynamic model (from Helaire et al., 2019). The Helaire et al. (2019) model extended from Bonneville Dam at CR and Oregon City at WR to 30 km offshore. I used the modeled hourly discharge from Helaire et al. (2019) at BAT as a boundary condition for my historical model. See Section 3.2.4 below for details of the historical simulations.

3.2.3.3 Ocean boundary

Eight major constituents (Q_1 , P_1 , O_1 , K_1 , N_2 , M_2 , S_2 , and K_2) were used to force the astronomical water level from the ocean side for both bathymetry models. The amplitude and phase for each constituent were found using a tidal prediction software called the Tide Model Driver (TMD) (Egbert & Erofeeva, 2002). At the north and south side boundaries, I used Neumann boundary conditions (water level gradient). The assumed mean sea-level is +1.15m (NAVD88, <https://vdatum.noaa.gov/>) and is kept constant for all simulations. The salinity and temperature boundary conditions for the model are described in section 4.2.4.

3.2.4 Model simulations:

Two hydrodynamic model grids were used to represent the LCRE modern and historical bathymetric models (MB and HB). For calibration, the 2013 river discharge hydrograph was used with the MB; this run is known as MM0. HB was calibrated using the 1876 river hydrograph (HH0). The daily river flow during the simulation period was not available, so an estimation from a similar hydrological year (1880) for high flow period and a 10-year hydrological average (1880-1890) for low flow periods were used. Three other model runs were used to investigate the long-term changes in tidal forcing; these were run with the 50th percentile river shown in Figure 2 (modern & historical). The first run uses MB with a modern hydrograph (MM). The second run uses the HB and historical river hydrograph (HH). A hypothetical case was also run using MB and the historical river discharge (MH); this run was used to distinguish the effects of changes in river discharge variation vs. bathymetric changes. Each run covered a one-year period with tidally variable river discharge applied at BAT.

Furthermore, to investigate the effect of river discharge on channel bifurcations, 13 additional simulations were needed, denoted MB1-6 and HB1-7, adapted for constant river discharge. Six simulations were used to represent a range of constant river discharge (3000-15000m³/s in increments of 2000m³/s,) for MB and modern river discharge. Seven simulations were used to cover the larger historical river flow range (2000-18000m³/s in increments of 2000m³/s) with HB. The bifurcation simulations were run for 1.5 months for each river discharge, see Table 3-1. The range of river discharge was selected to represent

the area between the 25th and 75th percentiles of the modern and historical river discharge for each model.

Finally, eighteen cross-sections (2.5km apart) were defined along the estuary transection (Figure 3-1-a) for the bifurcation investigation. These many cross-sections were selected to check the spatial variability of freshwater distribution among the channels. At each cross-section, the boundaries of the channels were then selected according to the cross-section bed topography, with three channels defined at each location, South Channel (SO), North Channel (NO), and Middle Channel (MI). The SO and NO channels are connected through the MI and many other braided channels, which allows water exchange between them. Therefore, an estimate of flow at a specific cross-section is unreasonable. To overcome this problem, a spatially averaged percentage was used for the area that covers both channels (cross-sections 10-14, see Figure 3-1a). The flow percentage was found through each channel at each cross-section, after which the mean of all cross-sections was considered (Table 3-2). A Godin filter (Godin, 1972) was used to estimate the mean flow from hourly simulation data at each cross-section. The Godin filter is a low-pass filter that applies a moving average over 24, 24, and 25 hours to remove the tidal signal.

Table 3-1. The Hydrodynamic models

Model	Bathymetry	River discharge, Q [m ³ /s]	Simulation Period	Purpose
MM0	Modern	Modern (2013 hydrograph)	1 year	Calibration
HH0	Historical	Historical (1876 hydrograph)	10 months	
MM	Modern	Modern (50th percentile)	1 year	Tidal change investigation
HH	Historical	Historical (50th percentile)	1 year	
MH	Modern	Historical (50th percentile)	1 year	
MB	Modern	3000, 5000, 7000, 9000, 12000, and 15000	1.5 months for each Q	Flow bifurcation investigation
HB	Historical	2000, 3000, 5000, 7000, 9000, 12000, 15000, and 18000	1.5 months for each Q	

Table 3-2. Example of total flow (river + tide) percentage at each cross-section (CR) of MB during low river discharge and spring-flood tide. SO is the South Channel, NO is the North Channel, and MI is the Middle Channel.

Cross-section	SO%	MI%	NO%
CR10	31	15	54
CR11	34	4	62
CR12	34	6	61
CR13	37	7	56
CR14	36	13	51
Mean	34	9	57

3.3 Model calibration

The modern model (MM0) was calibrated to water level measurements at Cape Disappointment (CD, NOAA station ID: 9440581), Astoria Downtown (ASD, a radar gauge with data spanning from October 2015 to the present (Talke et al., 2020)), Tongue Point (TP, NOAA station ID: 9439040), Skamokawa (SK, NOAA Station ID: 9440569), and Wauna (WA, Station ID: 9439099); see Figure 3-1a for locations. An example of the water level time series at TP showed a good agreement between the actual water levels and the modeled water levels during low river discharge (Figure 3-4a), with mean root square error (RMSE) of 0.11m and 0.98 of the variance accounted for adjusted R^2 . On the other

hand, the high river discharge simulation (Figure 3-4b) produced 0.15m RMSE and 0.97 adjusted R^2 . Additionally, the major tidal constituents (M_2 , S_2 , K_1 , and O_1) of the MM were calibrated at variable stations along the estuary. The percentage difference shows that the major constituent M_2 has the best calibration values (0.5-3%) among the other tidal components, while the worst are O_1 and K_1 with max percentage differences (34 & 25%, respectively) at Wauna (Table 3-3).

The historical model (HH0) was calibrated (for 1876) using historical water level data that are available from Astoria (ASD) from 1853-1876 (Talke et al., 2020). Due to the unavailability of discharge measurements for 1876, I approximated the 1876 hydrograph by using measured 1880 discharge. The 1880 hydrograph is known to be quite similar to the 1876 hydrograph, with a peak that was only 0.9m lower in Portland and occurred ~1 week later. More recently, Talke et al. (2020) have produced discharge estimates for 1876, which I compare to our boundary condition in Figure 3-3. Because the method used by Talke et al. (2020) is effectively a 30d average, the peak is likely somewhat attenuated and seems to occur somewhat earlier than the measured peak in water level. Still, the estimated flow for 1876 from July-September is close to the 1880 hydrograph. Therefore, for our model calibration, I compared statistics from this time period.

Figure 3-3. CR 1876-hydrograph at the Beaver Army Terminal (BAT). The FM hydrograph used in modeling is derived from the 1880 hydrograph. The green curve is the estimate of river discharge in 1876 at BAT by Talke et al. (2020).

The calibration at Astoria downtown (Rkm20) produced 0.13m of RMSE and $R^2 = 0.97$ during low flow conditions (Figure 3-4c) and 0.15m RMSE and $R^2 = 0.96$ during high river discharge (Figure 3-4d). Additionally, the model was calibrated at Cathlamet (RKM59) using records from the fall of 1877, a low flow period (0.13m, RMSE, and $R^2 = 0.97\%$). These records are described in Helaire et al. (2019).

All simulations in both models were tested using a skill assessment (SK) that measures agreement between models and observed data (Willmott, 1981), and the Nash-Sutcliffe (NS) number, an indicator of how well the modeled vs simulated data fits a 1:1 line (Nash & Sutcliffe, 1970). The minimum values of SK and NS were found to be 0.97 and 0.99 for the modern model and 0.94 and 0.98 for the historical model, respectively. Note that a value of one for these coefficients means a perfect agreement between the model and the actual data, while zero indicates no agreement at all.

Figure 3-4. Modern Bathymetrical Model (MM0) calibration at Tongue Point (TP) during low river discharge (a) and high river discharge (b). Historical Bathymetrical Model (HH0) at Astoria downtown (ASD) during low river discharge (c) and high discharge (d); the blue plot is the error for each case (modeled-observed).

Table 3-3. Major Constituent comparison of the MM0 model in 2013 at different locations: Cape Disappointment (CD), Astoria Downtown (ASD), Tongue Point (TP), and Wauna (W).

Location	Constituents	O₁	K₁	S₂	M₂
CD	Simulated [m]	0.260	0.465	0.243	0.865
	Actual [m]	0.263	0.413	0.238	0.859
	% difference	1	12	2	0.7
ASD	Simulated [m]	0.262	0.456	0.248	0.918
	Actual [m]	0.243	0.394	0.237	0.942
	% difference	7	15	4	2.5
TP	Simulated [m]	0.264	0.458	0.253	0.939
	Actual [m]	0.244	0.400	0.240	0.970
	% difference	8	14	5	3
SK	Simulated [m]	0.237	0.399	0.206	0.834
	Actual [m]	0.183	0.321	0.200	0.838
	% difference	25	22	3	0.5
WA	Simulated [m]	0.229	0.381	0.196	0.793
	Actual [m]	0.163	0.294	0.182	0.800
	% difference	34	26	7	0.9

The source of errors in Figure 3-4 for both models probably include: a) using a uniform wind distribution over the estuary and the coastal area, b) not including seasonal variations in elevation in the ocean boundary condition, and c) using an approximate flow for the historical model. To understand the importance of these factors, I compared the monthly averaged water level for both models (Figure 3-5). The modern model monthly averaged water level at Tongue Point (Figure 3-5a) shows a good agreement between the simulated and the observed water level with an RMSE of 0.04m. The comparison of the simulated water level to in-situ data (green dashed line in Figure 3-5a) show that the MM model captured the coastal upwelling/downwelling well in all months except in September. The mismatch in September produced the most significant error in the monthly averaged data, about 0.12m error between modeled and observed (Figure 3-5b). These errors could

be the result of using uniform wind speed and direction over the entire model area, or because of neglect of offshore wind effects on the ocean boundary condition.

The monthly averaged water level in the historical model produced a 0.08 m RMSE compared to ASD measurements and showed a good agreement during high river flow (Figure 3-5c). The largest percentage of error occurred during low flow conditions. This likely occurs because: a) the flow data are from a different year than the tidal data, b) the wind forcing used in the historical model did not represent the upwelling/downwelling accurately enough, c) the effects of El Nino-Southern Oscillations conditions are not accounted for in the ocean boundary, and/or d) ocean tides may have changed over the last century (Jay, 2009).

Figure 3-5. Monthly averaged WL for the MM0 calibration (2013, a & b), and HH0 (1876, c & d)

3.4 Results

Measurements in 1968 found that 76% and 69% of the total flow passes through the North Channel during flood and ebb, respectively (black dashed lines in Figure 3-6, Lutz et al., 1975). Our results suggest that the bifurcation of river discharge and tidal flow between the North and South Channels has since remained relatively stable, despite the deepening of the estuary entrance ca 1984 (from 14.6m in 1956 to 16.8m in 1984, according to Corps of Engineers, 1989); no more recent measurements are available. Specifically, my analyses show that 66% of total flow goes into the North Channel during flood-spring and 71% during flood-neap tides (filled black dots, Figure 3-6); the corresponding values are 64% during ebb-spring and 70% during ebb-neap (hollow black dots, Figure 3-6). The dashed lines in Figure 3-6 represent the average percentage of total flow passing through the North Channel (dashed-black) and the South Channel (dashed-red). The bars in Figure 3-6 are the percentage of the total flow in the channels at different river discharges from Lutz et al., 1975. The solid and hollow black circles represent the percentage of flow in the North Channel during spring and neap tides, respectively, while the red ones are for the South Channel.

The small differences between the 1975 result and mine could be due to several reasons, such as morphological and hydrological changes in the last 45 years, changes in the tides over the previous 50 years, or differences in the calculation methodology and experimental error. Also, I am using a constant river discharge, which does not happen in practice. Considering these differences, the agreement is quite good.

Figure 3-6. Total flow percentage in the North and South Channels. The Lutz et al. (1975) data is represented in black bars to show the portion of the flow in the North Channel at each flow rate and the red bars for the South Channel percentage (dashed lines are the average percentage at each channel). The circles represent data from our analysis using the modern bathymetry model (MB).

3.4.1 Tide changes

Tidal constituents have changed in magnitude and phase due to system alterations such as channel deepening, mouth narrowing, and river discharge modification (Figure 3-7a & b). The modern model (MM) is compared to the historical model (HH). Simulation MM includes modern bathymetry (2008-2010) and the mean of modern river discharge (1967-2016). The HH simulation uses historical bathymetry (1868-1901) and the mean of historical river discharge (1879-1929).

Results show an increase in the M_2 tidal constituent of ~12.5% at Astoria Downtown (ASD). Thus, the historical (HH) M_2 amplitude is modeled to be 0.88 m while the modern (MM) one is ~ 0.99m, an increase of 0.11 m of M_2 at ASD (Figure 3-7c). This increase of tidal amplitude is a result of both bathymetric and river discharge changes in the past 150 years. Note that the difference is less in the actual data. Talke et al. (2020) show that the change is from 0.88m to 0.95m; hence, the modern model appears to be over-predicted. The reason could be using an average of river discharge for the period of 1967-2016, which is lower than the actual flow during the calibration year (2013).

When historical river flow is applied to modern bathymetry (case MH, the M_2 change is reduced to 0.1m. Hence, approximately 0.01m of the modeled change in M_2 is due to altered river flow, or roughly 10%. Using empirical data and tidally-based estimates of river flow, Talke et al. (2020) also estimated that approximately 10% of the changed constituent values were due to river discharge. The effect of channel deepening is also observed in the phase speed of the tide, which arrives earlier at Tongue Point and upstream locations today, compared to historical conditions (Figure 3-7d). Another difference is a shift in the location of maximum M_2 inside the estuary. In the HH model, the maximum M_2 was near Rkm 20, while in the MM simulation, it is 8 km landward, near Rkm 28.

Figure 3-7. M_2 Maps: a-modern bathymetry model, b-historical bathymetry model, c- M_2 amplitude in the South Channel for MM (modern bathymetry/modern river flow), HH (historical bathymetry/historical river flow, and MH (modern bathymetry/historical river flow) models, the actual values represented by squares at Tongue Point (TP) at 2013 and Downtown Astoria (ASD) in 2015 and 1876, and d- M_2 phases at South Channel.

3.4.2 Modern model bifurcation

The bifurcation analyses show that the flow percentage going through the North and South Channels in the modern model (MM) is spatially variable along the longitudinal axis of the estuary. This occurs due to exchange between the North and South Channels via the small braided channels that connect them. The tidally averaged river discharge percentages (Figure 3-8) were carried out for low (LO: 3,000m³/s) and high (HI: 15,000m³/s) river discharge. The river discharge at each case was constant during a 1.5-

month simulation period. The percentage was then averaged at each channel location for two tidal days (for stability) using a Godin filter around the highest water level (spring) and lowest water level (neap) during a month of simulation.

The river discharge percentage is represented in dotted colors in Figure 3-8. The red dots represent the percentage of river discharge in the South Channel, the blue dots show the percentage in the North Channels, magenta coloring shows the percentage of river discharge circulated around tidal marshes, and green dots depict the water circulated between the South and North Channel. The dot size represents the percentage of river discharge at each location relative to the total discharge leaving the mouth (100%).

Figure 3-8. Modern-model river discharge distribution for low (LO)/ high (HI) river discharge and spring/neap scenarios (two days mean around the highest water level (spring) and lowest water level (neap)). The red dots represent the percentage of river discharge along the South Channel (SO), blue represent the North Channel (NO), and green the Middle channels (MI).

In order to quantify the tidally averaged river discharge in each channel, a spatially averaged percentage that covers the area between cross-sections 10-15 is used. These cross-sections cover the domain of the three channels, SO, MI, & NO, between the junction near the mouth and the upstream boundary of the North Channel (Figure 3-1a). This analysis shows that during spring tides the South Channel takes a larger percentage of the river discharge than the North Channel, except when river discharge is greater than 12,000 m³/s (Table 3-4).

The dependence of partition with the spring-neap cycle likely occurs because most of the tidal exchange occurs in the North Channel (~56%, NO vs. ~34% SO, and the rest in the MI). Therefore, high tidal energy in spring tides might hinder the river flow in the North Channel and push it toward the South Channel. Increasing the tidal energy during spring tides results in a larger effective friction. A similar result was found in Buschman et al. (2010) on a river delta in Indonesia. During the neap tides, the North Channel takes a larger percentage of the river discharge (Table 3-4).

Table 3-4. River discharge percentage of the modern model for different river discharges averaged over two days during high (spring) and low (neap) tides periods, where SO is South Channel, NO is the North Channel, and MI is the Middle Channel.

Q [m ³ /s]		SO [%]	NO [%]	MI [%]	Ψ		SO [%]	NO [%]	MI [%]	Ψ
3,000	Spring	80	21	-1	0.58	Neap	44	48	9	-0.03
5,000		62	36	3	0.26		40	54	7	-0.14
7,000		56	40	4	0.17		38	56	6	-0.18
9,000		53	42	5	0.11		38	57	6	-0.20
12,000		49	46	5	0.03		36	59	6	-0.24
15,000		46	49	6	-0.03		36	59	5	-0.23

The river discharge percentage in each channel varies nonlinearly with river discharge.

The percentage in the South Channel decreases as river flow increases, with more going

to the North Channel (Table 3-4). I found the same pattern when using the asymmetry index (Ψ , Equation 2.31) proposed by Buschman et al. (2010), see Figure 3-9. This figure shows that the larger percentage of river discharge is being routed through the South Channel during spring tides (positive Ψ). In contrast, during the neap tide, the large percentage is passing through the North Channel (negative Ψ).

Figure 3-9. Variation of discharge asymmetry index with river discharge

The pattern of river discharge under modern bathymetric conditions can be explained in terms of water surface slope in the channels (Figure 3-10). Because water levels in the Tongue-Point to Millar Sands Reach are a function of river flow, the slope

between North and South Channel increases as discharge increases from 3000 to 15000m³/s (Figure 3-10). The water surface slope of the North Channel through the Middle Channels (4.1×10^{-6} to 1.5×10^{-5} m/m) is larger than the slope in the South Channel (3.3×10^{-6} to 1.1×10^{-5} m/m). This large slope produces tidally-averaged flow towards the North Channel. Therefore, a significant amount of river discharge is diverted towards the North Channel, despite the shallower and more frictional mid-channel geometry.

The mismatch between North and South Channel water levels is driven by several factors. First, the distance between point 1 and points 2-5 is much longer (~ 750m, 1950m, 3250m, and 3250m, respectively) along the South Channel than the North Channel. Since the total water level drop between the point where they split (point 1) and the point that they meet up again (point 5) is the same, this leads to an inherent asymmetry in slopes. Secondly, the lower flow within the North Channel between point 1 and point 5, compared to the South Channel, requires a relatively small slope to drive flow downstream. Since this slope is small, the slope between points 2-5 must be large. The overall result is that a larger percentage of river flow is routed through the Middle Channel during periods of high discharge vs. low flow conditions (Table 3-4).

Figure 3-10. Tidally averaged surface slope (m/m) of each point on each junction and point 1 (where the two channels meet). The color plots represent the spatial water level map *e*- mean water level comparison between the South and North Channels (dashed lines are the water level in the North Channel), max difference in water levels between North and South Channel between the North Channel boundaries (South-North) are a-4.7, b-1.7, c-5.3, and d-2.1 cm.

3.4.3 Historical bifurcation

Results from the HH Model show that the river discharge partition was reversed compared to the modern situation. The spatially averaged percentage for the cross-sections that cover the three channels (South, North, and Middle) show that the North Channel always takes a larger percentage of river discharge than the South Channel (Table 3-5). One reason for this result is that the Middle Channels in the historical model conveys more discharge toward the north than in the modern configuration, due to their larger cross-sectional area. Also, the South Channel is deeper in the modern model (MM) due to continuous dredging. Another reason is the use of pile dikes and construction of dredged

material islands such as Rice Island and Miller Sands (Rkm40, Figure 3-1-a), which help to direct more river discharge to the South Channel.

Table 3-5. River discharge percentage of the historical model for different river discharges averaged over two days during high (spring) and low (neap) tides periods, were SO is South Channel, NO is the North Channel, and MI is the Middle Channel.

Q [m ³ /s]		SO [%]	NO [%]	MI [%]		SO [%]	NO [%]	MI [%]
2,000	Spring	22	55	22	Neap	17	55	28
3,000		24	58	28		20	58	22
6,000		25	61	14		24	60	16
9,000		25	62	13		25	60	15
12,000		25	61	13		25	61	14
15,000		25	62	13		24	62	14
18,000		25	62	14		24	62	14

The river discharge distribution is illustrated in Figure 3-11 for the historical model. Upstream of Tongue Point, the South Channel feature is not a straight, dredged channel, as at present; instead, there are several braided channels. Downstream of Tongue Point, there are two main channels: The South and the Middle Channel. The Middle Channel conveys more water than the South Channel, probably because it has a larger cross-section area than the South Channel (Figure 3-11).

Figure 3-11. Historical-model river discharge distribution. The red dots represent the percentage of river discharge in the South Channel, and the blue is the percentage in the North Channels, the green is the percentage in the Middle Channel, the magenta is the percentage in the braided channels in Cathlamet Bay area that eventually meet in one location close to about 25 km from the mouth (just upstream Tongue Point, Point2 in Figure 3-12)

I investigated why the Middle channel takes a larger percentage of river discharge than the South Channel by estimating the ratios of the river flow in the two channels and the other parameters of Manning's equation (friction, water surface slope, hydraulic radius, and cross-section area). The surface slope was found to be almost the same for all channels between point 2 at Tongue Point and point 1, where the South and North Channels meet near the mouth. This is because the distance between these two points along the channels between them is the same (Figure 3-12). Since the slopes are equal, there must be other factors that drive the higher percentage of flow through the Middle channel compared to the South Channel downstream of Tongue Point. Therefore, using the simulation data, I

compared the ratio of river discharge of the two main channels (R_Q) to the ratio of the other parameters' ratios in Manning's equation (cf. (3.1) to (3.4) above). I found that n , S , and P have little or no effect on river discharge distribution between the South and Middle Channels (their ratios are equal to ~ 1 , Figure 3-13e, f, and g, respectively). On the other hand, R_h , A , and H have the most effect on freshwater distribution such that the hydraulic radius and the cross-sectional area of the Middle channel are both about five times bigger than the South Channel (Figure 3-13b & c). In contrast, the Middle Channel water depth is double than the South Channel water depth (Figure 3-13d). Therefore, the Middle Channel conveys more water than the South Channel due to its larger cross-section.

Figure 3-12. tidally averaged slopes of the South, North and the Middle Channels in the historical model

Figure 3-13. Effect of Manning equation parameters on flow distribution between the Middle channel (MI) and the South Channel (SO) near Tongue Point. R^Q is the ratio of river discharge of Middle to the South Channel, (Q is river discharge, Rh is the hydraulic radius, A is the cross-sectional area, H is the mean water depth, n is bed friction coefficient, S is the surface slope, and P is the wetted parameter).

The flow bifurcation analyses show that the river discharge-tidal forcing interaction and the morphological differences between the modern and historical cases alter the flow distribution. In the modern model (MB), the percentage of freshwater discharge is found to have a negative relation to the river discharge. Hence, the percentage of freshwater flow in the South Channel decreases as the river discharge increases (Figure 3-14a). This occurs because the slope in the MB is a function of river discharge (positively correlated) and the slope in the North Channel is higher than the South Channel, see Figure 3-14c (i.e., the percentage of the South Channel decreases with higher river discharge, Table 3-4).

However, the neap tide can reduce the availability of freshwater to the South Channel, especially in low river discharge (Figure 1-a), because the water surface slope is lower in neap than in spring tide (Figure 3-14c). In the historical model (HB), since the slope of the south, middle, and North Channel are nearly equal, the most effective parameter in the discharge distribution between the channels is the cross-sectional area. Therefore, the percentage of freshwater in the South Channel is positively correlated to river discharge (Figure 3-14b) with the tendency of neap tides to reduce the South Channel share--the same as in the MB.

Figure 3-14. The river flows vs. South Channel percentage in MB (a) and HB (b), (c): represents the relation between river discharge and slopes for North & South Channels during neap & spring tides.

3.4.4 Salt fluxes

Salt flux analysis (Figure 3-15) shows that most of the salt transport is confined within the South and North Channels (Hamilton, 1990 concluded the same result). In the MM model, about 44% of mean salt is transported through the North Channel, 51% in the South Channel, and 5% in the Middle Channel. However, 75% of fluxes pass through the

South Channel during low flow (2,000 m³/s) vs. 40% during high river discharge (15,000 m³/s). During low river flow, the South Channel is highly stratified due to the high percentage of freshwater passing through the Channel (Figure 3-14a), which enables more salt flux to be exchanged through the South Channel (i.e., salt transport increases during high stratification). The North Channel salt exchange trend is opposite that of the South Channel transport (Figure 3-15a).

In the historical model, the percentage of freshwater in the North Channel is always higher than the South Channel percentage (Figure 3-14b). Consequently, about 60% of salt exchange happens through the North Channel, 24% in the South Channel, and 16% in the Middle Channel. This percentage is averaged over the spring-neap tidal cycle and then averaged along the river flow range (Figure 3-14b).

Figure 3-15. Subtidal salt fluxes percentage variation with river discharge (averaged over the spring-neap cycle).

In the MM model, The North Channel conveys more tidal flux during the flood (61%) than the ebb (55%). The same pattern also found in the Middle Channel (12% flood vs. 8% ebb). In contrast, the South Channel conveys less flux during the flood (27%) than the ebb (37%), which indicates a clockwise circulation. This means that salinity flux is imported in the North and the Middle Channels and exported in the South Channel (Figure 3-16a). Similarly, in the HH model, the salt flux rotates clockwise, but the time it returns through the middle and South channels (Figure 3-16b). The flood percentage in the North Channel is 76% in flood and 60% in ebb, the South channel flood percentage is 22 % vs. 25% in ebb, and finally, the Middle Channel flood flux is 2% compared to 15% during ebb.

Figure 3-16. Salt flux circulations

3.5 Conclusions

Flow bifurcation in the LCRE is controlled by the interaction of tidal forcing, river discharge, and channel geometry. In the modern bathymetry model (MB), a larger percentage of river discharge is found to be routed toward the North Channel during periods of high discharge, vs. low flow conditions, because of the asymmetry in water surface slopes between the North and South Channels. This asymmetry is driven by two main factors: the difference in length of the water path along each channel and the non-linear tidal forcing-river discharge interaction (water level along the estuary is a function of river flow and tides in a non-linear way). These factors produce a mismatch between the two channels in terms of the surface slope.

In the historical bathymetry model (HB), the geometry of the South Channel is less clear because of the presence of the braided channels upstream of Tongue Point. Downstream of that point, the estuary has three channels (North, Middle, and South) that convey river discharge toward the mouth. The tidally averaged surface slope between those

channels was found to be equal, which gives the advantage to the channel geometry to control the water distribution between the channels.

For both models, it was found that the percentage of total non-tidal transport for the channels is proportional to the river discharge and follows a power rule. However, the proportion has a negative trend in the modern model (MB), as it is controlled by the asymmetry in water surface slopes, and is positive in the historical model (HB) since it is proportional to channel geometry. Moreover, the state of the tide controls the channel bifurcation through altering the water surface slopes between the channels, such that the percentage of freshwater in the South Channel is higher in spring tides than neap tide because of the higher water surface slope. The freshwater distribution between the channels has impacted the salt fluxes. Historically, most of the salt was transported through the North Channel since it is more stratified than the South Channel during all river flow. Nowadays, during low river flow, a higher percentage of salt flux is found on the South Channel. Conversely, a higher rate of salt flux goes through the North Channel when the river flow is high because of the stratification. The salt flux lateral circulation has changed as well. The salt flux used to be imported through the North Channel and exported through both the Middle and South Channel. In the modern model, the circulation is the same, but the difference is that the flux is imported in the Middle Channel as well.

Finally, the system alteration has impacted the tidal forcing inside the estuary. The tidal constituent M_2 amplitude has increased by 12.5% at Tongue Point (~0.11m) due to channel deepening, mouth squeezing, and river discharge alteration while it has increased

by 10% due to channel deepening only. This means the river alteration has elevated M_2 by 2.5% only. The deepening of the channel reduces the velocity, which decreases the effect of friction and hence, increases the tidal amplitude.

Chapter 4 Changes in the Lower Columbia River Seasonal Cycle of Salinity Intrusion Related to Change in the River Flow Regime and Bathymetry**4.1 Introduction**

Estuaries are often defined as the transition region in which freshwater from river flow mixes with saltwater from the ocean, producing a horizontal density gradient and vertical stratification (Wilson, 1977; Hansen & Rattray, 1966; Pritchard, 1967). Previous literature suggests that salinity intrusion is sensitive to alterations in river flow and bathymetric changes. For example, on the Columbia River Estuary, Chawla et al. (2008) found that the salinity intrusion length doubles when the river flow decreases by 75%, though this result is somewhat sensitive to the zero point chosen. Chant et al. (2018) reported that 15% increase in the depth of Newark Bay, New Jersey resulted in a doubling of the exchange flow, but caused only a small increase in salinity and longitudinal salinity gradient, possibly because of the limited extent of deep water in the system. In the Hudson River Estuary, salinity intrusion has increased by 30% over the past century due to channel deepening and decreased river discharge (Ralston & Geyer, 2019). Also, Mulamba et al. (2019) reported an increase in the lower St. Johns River salinity in Florida in response to sea-level rise.

Many studies have interpreted salinity intrusion using a power-law relation ($X \sim Q^{-n}$). For San Francisco Bay, Monismith et al. (2002) found $n = 1/7$, while n was found to be 0.1 for Delaware Bay (Aristizábal & Chant, 2013). Ralston et al. (2008) found

a value of $n = 0.35$ when accounting for stratification, but 0.4 when using full water depth in the vertical mixing coefficient for the Hudson estuary. However, aside from Hamilton (1981) and Ralston & Geyer (2019), few studies have analyzed the interaction of historical changes to depth and river flow in altering salinity intrusion and stratification.

In this chapter, I investigate the changes to the Lower Columbia River Estuary (LCRE) salinity intrusion since the 19th century due to the changes in seasonal river hydrograph (flow regulation and diversion) and navigational development (primarily channel deepening and jetty construction). Two questions are addressed:

- How have long-term modifications to the channel depth and river discharge altered the tidal and subtidal transport and distribution of salt within the estuary?
- How have the changes in bathymetry and river hydrograph changed mixing processes, density stratification, and the position of the system on the Geyer-MacCready (2014) classification diagram?

For this study, I use records of LCRE river inflow (modern flow, 1966-2016, and historical flow, 1878-1928) for boundary forcing of 3D Delft3D-FM of modern and historical (pre-1900) numerical model grids.

4.1.1 Columbia River Estuary (CRE)

The Columbia River is the largest in the northwest (Jay et al., 2007 and Naik & Jay, 2011) and the second largest in the US (Simenstad et al., 1990). It has a large mean river discharge ($7,300 \text{ m}^3/\text{s}$, Bottom et al., 2005) with a modern seasonal range of about 2000 to

17,000 m³s⁻¹. There are also strong tidal currents caused by mixed diurnal and semidiurnal tides with a typical annual range of 1.6-3.7m (Jay et al., 2016). The highly variable tidal energy and fluvial buoyancy input produce large temporal variations and a persistently strong horizontal salinity gradient along the estuary (Jay, 1984; Jay & Smith, 1990b). Depending on flow and tidal conditions, the estuary typically fluctuates between partially mixed to highly stratified (Jay et al., 2007). The estuary has two main channels (South & North) and many other braided channels divided by shoals (Sherwood & Jay, 1990) but most of the flow is confined within the main channels (Jay, 1984), especially in the modern system, which has been highly developed for navigation. At present, most of the river discharge goes through the South Channel while most of the tidal exchange occurs through the North Channel (Jay & Smith, 1990b).

The LCRE has been highly modified in terms of flow regulation (hydrological modification) and channel development (topographical modification). The topographic modification has included deepening the navigation channel, constructing three jetties at the entrance (total length, 18 km), and diking and filling intertidal areas (Sherwood et al., 1990; Simestad et al., 1994). The controlling channel depth in the estuary has doubled (from about 6 to 13m relative to MLLW, Sherwood et al., 1990) to accommodate cargo ships, while the entrance is now 18m deep. The construction of jetties has increased the inlet depth and reduced the mouth width by a factor of three (9.7 to 3.2km); see Figure 4-1. Also, diking and filling have reduced the tidal prism by 15% and caused about a 70% loss in tidal wetland over the last 140 years (Marcoe & Pilson, 2017).

Water management, including irrigation diversion, flood control, and hydropower, started in the 1930s, but integrated system management did not begin until the early 1970s. Flood control and power generation have altered the river hydrograph such that it caused a reduction in spring freshet (by 45-50%) and an increase in winter, fall, and late summer river discharge. Irrigation and climate change have reduced the mean in annual flow by 17% (Naik & Jay, 2005, Figure 4-2). Before modification of the hydrograph, river discharge was ranged from 1,000 to 35,000 m³/s in the river above the head of the estuary (Naik & Jay, 2011), the present range is about 2000-17,000 m³/s (Naik & Jay, 2011; Yeh et al., 2012), with rare floods that may reach 26,000 m³/s (Helaire et al., 2019). Although climate change has caused (along with deforestation) earlier spring floods (Sherwood et al., 1990) and a reduction in annual mean discharge of about 7%, its effect remains less significant than direct engineered modifications (Naik & Jay, 2011). Hydrologic modifications have altered the system in obvious ways, such as increasing salinity intrusion (the subject of this chapter; see also Naik & Jay, 2011) and reducing higher elevation wetland habitats in the tidal river (Jay et al., 2016), but also in less obvious ways such as increasing the tidal range (Jay et al., 2016; Talke et al., 2020). Increasing the tidal, changing depth, and any subsequent changes of the tidal velocity range is then a secondary cause of changes to the salinity intrusion, because of the role of wave (tidal) salt fluxes in setting salinity intrusion.

Figure 4-1. Columbia River Estuary (CRE) topographical changes: a-modern bathymetric topography (2008-2010), observation point's locations (OC2, SAT3 & Tongue Point), North & South channels layout, and river boundary (BAT), b-historical bathymetric topography (1870-1900), and c-South Channel bathymetric changes. The axes coordinate system referenced to Washington South State Plane Coordinate System (SPCS), E is for easting, and N is for northern.

Figure 4-2. Modern vs. historical flow hydrograph (30 days moving average) at The Dalles. The solid lines are the 50th percentile of 50 years while the shaded areas are for the 25th and 75th percentiles.

4.2 Materials and methods

4.2.1 Salinity intrusion theory

Salt is transported landward due to tidal advection (wave fluxes) and two-layer estuarine circulation (including the gravitational and internal asymmetry modes; (Jay & Musiak, 1994, 1996), while seaward transport is driven by river discharge. The gravitational circulation (exchange flow) results from the horizontal density gradient between the ocean and the river. Internal asymmetry also produces landward salt fluxes and results from the tidal wave deformation (difference in flood and ebb velocities and mixing). The river flow has a direct effect on salinity intrusion, and high flow is associated with reduced salinity intrusion (Hansen & Rattray, 1965; 1966; MacCready, 1999;

MacCready, 2007; Monismith et al., 2002; Geyer & MacCready, 2014). A theoretical salinity intrusion was presented by Monismith et al. (2002) based on the tidally averaged salt conservation equation, where the net salt flux is zero and the effect of the gravitational circulation is excluded (Hansen & Rattray, 1965):

$$X_s = \frac{(W\alpha)^{1/3}(\beta g S_o)^{2/3} H^3}{Q^{1/3} \nu_t}, \quad (4.1)$$

where X_s is the salinity intrusion length from the mouth of the estuary to the location where the bottom salinity is 2-psu (Monismith et al., 2002; MacWilliams et al., 2015). The channel width is W , α is the constant of proportionality (5.4×10^{-5}), β is the saline expansivity ($7.7 \times 10^{-4} \text{ psu}^{-1}$), S_o is the ocean salinity, H is the local depth, Q is the river discharge, and ν_t is the eddy diffusion coefficient. This scaling assumes a steady-state, constant width and depth, and that the vertical eddy diffusivity is equal to the eddy viscosity coefficients. It also assumes a uniform eddy viscosity over the water column.

From the scaling in eq 4.1, salinity intrusion varies directly with the third power of channel depth ($X_s \sim H^3$) and inversely with the cubic root of the river discharge ($X_s \sim Q^{-1/3}$). However, estuarine circulation and stratification both can increase salinity intrusion (MacCready & Geyer, 2010). From the scaling of the tidally averaged along channel momentum and salt equations, the estuarine circulation and stratification can be found as:

$$U_E = \frac{\beta g H^3}{48 \nu_t} \frac{\partial s}{\partial x}, \text{ and} \quad (4.2)$$

$$\Delta S = U_E \frac{H^2}{K_S} \frac{\partial s}{\partial x}, \quad (4.3)$$

by relating the eddy viscosity and eddy diffusivity (K_S) to tidal velocity and local depth, $v_t \sim u_t H$ and $K_S \sim v_t / S_c \sim u_t H / S_c$; S_c is Schmidt number (Geyer & MacCready, 2014), the scaling of the above equations show that $U_E \propto H^2$, $\Delta S \propto H^3$, and $X_s \propto H^2$. This means that, theoretically, the salinity intrusion is much more sensitive to changes in bathymetry than to river discharge. A further complication is that depth changes can alter the velocity magnitude. The depth-averaged tidal velocity at the mouth can be found, according to Friedrichs, (2010), from the continuity equation as:

$$u_t = \frac{L\omega A}{H}, \quad (4.4)$$

where L is estuary length, ω is the tidal frequency, and A is the tidal amplitude. Thus, the change in tidal amplitude can affect salinity intrusion (Jay & Naik, 2011; Geyer, 2009). Because of the complexity of many contributors to salinity intrusion, the presence of two estuary channels in the Columbia River Estuary, and a 3D bathymetry, a numerical model is needed to evaluate the changes in salinity intrusion that have occurred since the mid-1800s.

4.2.2 Previous analyses of LCRE salinity intrusion

Using salinity data observations, Neal (1966) measured the salinity intrusion length in the LCRE and found that during low river flow ($\sim 4,000 \text{ m}^3/\text{s}$), the intrusion from lower low tide to higher high water was between 28 and 37km, respectively. Moreover, the range

was 9-24km during high river discharge ($\sim 17,000\text{m}^3/\text{s}$). Hamilton (1990) predicted a maximum salinity intrusion length of 50 km via numerical modeling. The same salinity intrusion length was reported by Jay & Smith (1990) while a minimum of 20 km was observed, depending on the tidal stage and river discharge. More recent modeling (e.g., Baptista et al. (2015)) predicted a maximum salinity intrusion of up to 45 km in the South Channel. However, salinity intrusion in the North Channel is generally greater than in the South Channel, since a large portion of river flow is diverted to the South Channel during high river discharge. In contrast, during low river discharge, the difference in salinity intrusion between channels is small (Jay, 1984).

During low river discharge in the LCRE, salinity intrusion is greater during neap than spring tides. This occurs because the vertical mixing is weaker during neap tides. In contrast, when the river discharge is very high (historical flow), a greater salinity intrusion is expected during spring tides, because of a greater tidal excursion (Jay & Smith, 1990). However, there is a small variability for spring-neap salinity intrusion on moderate to high river discharge (Jay, 1984). The reason is that the tidal mixing during such flow is unable to overcome the moderate/high stratification (Jay & Smith, 1990)

4.2.3 Parameter space theory

Geyer & MacCready (2014) proposed a parameter space for estuarine classification based on an analytical solution for the relationship between stratification and the freshwater Froude number for idealized estuaries. In this classification, the effect of river flow is represented by the freshwater Froude number:

$$Fr_f = U_R/(\beta g s_{ocn} H)^{1/2}, \quad (4.2)$$

and the relative strength of tidal mixing vs. stratification effects is represented by the mixing number M :

$$M^2 = C_D U_T^2 / \omega N_0 H^2, \quad (4.3)$$

where U_R is river flow velocity, s_{ocn} is the ocean salinity, H is the depth, C_D is the drag coefficient, U_T is tidal velocity, ω is the tidal frequency, and N_0 is the buoyancy frequency.

$$N_0 = (\beta g s_{ocn} H)^{1/2}. \quad (4.4)$$

Further, Geyer & MacCready (2014) propose that complete vertical mixing occurs when

$$\alpha^{-1/2} Fr_f^{-1/3} M^2 \approx 1, \quad (4.5)$$

where $\alpha = 3.4$ (Geyer, 2010). Geyer & MacCready (2014) suggest that each estuary occupies an area inside the diagram given its specific bathymetry and typical tidal/fluvial forcing. From the definitions of the mixing number and freshwater Froude number, it can be concluded that the classification strongly depends on the river flow, tidal velocity, and estuarine depth. Since all three parameters have changed in the LCRE over the last 150 years, I hypothesize that the parameter space occupied by the Columbia River has shifted as well.

4.2.4 Methods

I have built two hydrodynamic models using Delft3D Flexible Mesh; the first one uses the modern bathymetry (2008-2010, Figure 4-1a) and the second uses the historical bathymetry (1868-1901, Figure 4-1-b). From the ocean side, I forced both models with the major tidal constituents (Q_1 , P_1 , O_1 , K_1 , N_2 , M_2 , S_2 , and K_2) obtained from the Tide Model Driver (TMD), a tidal prediction software by Oregon State University (Egbert & Erofeeva, 2002). Oceanside salinity and temperature were obtained from the Hybrid Coordinate Ocean Model (HYCOM) and the global Navy Coastal Ocean Model (NCOM) (after Kärnä et al., 2015; Baptista et al., 2004). The HYCOM model (26 layers) was interpolated to match the LCRE model (36 layers).

The models were forced from the river side using modern (1967-2016) and historical river discharge records (1879-1929). The modern records at the Beaver Army Terminal (BAT) were obtained from the USGS, and the historical records were obtained from the flow estimates of Naik & Jay (2005, 2011). For each era, the 25th, 50th, and 75th percentiles of river flow were extracted. Accordingly, three simulations were run for each river discharge percentile. For example, for the 50th percentile river flow, the simulations represented: the modern bathymetry and the modern river discharge (MM), the historical bathymetry and the historical river discharge (HH), and a hypothetical case where modern bathymetry combined with the historical river discharge (MH). The same was used for the other river flow percentiles such that a total of nine cases that ran for a full year of the simulation were used to represent various conditions of tidal forcing and river discharge.

The salinity was taken as zero at all times at the landward boundary. The mean daily water temperatures for the Columbia River were obtained from combined sources: USGS station 453845121562000 at Bonneville Dam (<https://www.streamnet.org/>) for the period 1938-2018. Wind data were obtained from the National Data Buoy Center (NDBC) station 46029 (Columbia River Bar) to represent the offshore area for the period 1984-2016, while wind data from the National Oceanic and Atmospheric Administration (NOAA) at Astoria Tongue Point (TP), OR-station 9439040 was used to cover the inside region of the estuary (2005 – 2016). For both water temperature and wind data, I used the mean of the available period, and the same date used for both models (historical and modern). Because of the steep topography along the LCRE, actual winds are highly variable and likely differ in many locations from the values used.

Averaging over a long period enables a climatological approach, such that the mean seasonal response of the system is simulated. I therefore implicitly assumed that wind forcing and water temperature have not changed over time. By holding these forcing factors constant, the factors producing long-term changes are reduced to just bathymetry and river flow alterations, which is a more tractable problem. However, it is known that water temperature is increasing in the Columbia River (Bottom et al., 2005), and it is possible that wind/wave forcing is also increasing (Ruggiero et al., 2011). Further, there are well-known variations in wind/temperature associated with El Nino and the Pacific Decadal oscillation (Newman et al., 2016). A detailed investigation of these factors, and how they have changed over time, is beyond the scope of this modeling effort; for now, I

note that some of the subtidal variations in water level that occur in data, but not the model, may be related to the use of climatological averages (see calibration, results, and discussion thereof).

4.3 Calibration of historical and modern models

Both models (historical and modern bathymetry) were calibrated to available water levels. In addition, the modern model was calibrated/compared to measured salinities and velocities. The model was also compared to the model results -- salinities and velocities - - of Kärnä et al. (2015). The calibration/validation of water levels can be found in (3.3). Here I show calibration results for salinities and velocities of the modern model only as there is no historical data. The good agreement with modern salinity measurements (see below) improves our confidence in the historical hindcast. Four measures of model quality were employed: The coefficient of determination (R^2), the mean root square error (RMSE), the Nash-Sutcliffe coefficient (NS), and the skill coefficient (SK). For NS and SK, one means a perfect match between the model and the observation, while zero is no match at all.

Time series and profiles of velocity and salinity were collected in the North Channel (OC2 on Figure 1) during May and October 2012 and used to calibrate the SELF numerical model (Kärnä et al., 2015). The observation data and the modeled data from (Kärnä et al., 2015) are available online at (<http://dx.doi.org/10.5281/zenodo.13782>); see Sanford et al. (2015). Using the salinity time series data for high flow (spring) and low flow (fall) at the OC2 location, I found a close agreement in terms of R^2 and NS. The

modern model calibration results show that R^2 and NS are about 0.9 between the modeled and observed salinity data during high flow conditions and 0.93 during low river flow (Figure 4-3a & Figure 4-3b). These scenarios are compared to the Kärnä et al. (2015) model in Figure 4-3c & Figure 4-3d. The statistical comparison shows good agreement between the model and the observations for both salinity and velocity. Compared with the two measured conditions, the salinity results show minimum values of $R^2=0.91$, NS =0.9, SK=0.97, and maximum RMSE= 3.4psu. These results compare favorably with the values of $R^2=0.85$, NS =0.83, SK=0.96, and maximum RMSE= 4.2psu reported by Kärnä et al. (2015)(Table 4-1). Results (Figure 4-3, Table 4-1) therefore show that the Delft3D-FM model is better than the Kärnä et al. (2015) model at reproducing measured velocity and salinity patterns. Model data in the FM model are clustered more closely around the 1:1 line than in the Kärnä et al. (2015) model, with fewer outliers (compare Figure 4-3b to Figure 4-3d, for example). Particularly at high flow, salinity (and therefore salinity intrusion) shows less variance and outliers. The good fit to data in the modern model increases our confidence that the historical model is reasonably modeling historical intrusion patterns.

Figure 4-3. Observed vs. modeled salinity comparison, FM-models (top panels: a & b). Kärnä et al. 2015 model (bottom panels: c & d). Left panels for low flow (fall: a & c) while right panels for high flow (spring: b & d).

Table 4-1. Model Calibration Statistics: R^2 is the coefficient of determination, RMSE is the mean root square error, NS is the Nash-Sutcliffe coefficient, and SK is the skill coefficient (one indicates there is the perfect match between the model and the observation, while zero is no match at all)

	River Flow	FM-model				Kärnä et al., (2015)-model			
		R^2	RMSE	NS	SK	R^2	RMSE	NS	SK
Salinity	High	0.91	3.38	0.90	0.97	0.85	4.23	0.83	0.96
	Low	0.93	2.79	0.93	0.98	0.89	3.79	0.89	0.97
Velocity	Low	0.94	0.25	0.92	0.98	0.95	0.23	0.92	0.98

*velocities of high flow are not available on the in-situ data

The modern model was also calibrated to selected salinity measurements made by the Center for Coastal Margin Observation and Prediction (CMOP) stations that are

distributed along the estuary (see locations marked Jetty A, SAT3, SAT1, and Grays in Figure 4-1). Depending on the availability of the data, the salinity extracted from our model was calibrated at the top, middle, and bottom of the water column (Figure 4-4). This figure shows calibrations during high river discharge conditions in Figure 4-4 (a1 to a3) and low discharge in Figure 4-4 (b1 to b3) at SAT3; other stations appear in Figure 4-4 (c1-c3). The statistical results for Figure 4-4 range from 0.7-0.8 for R^2 & NS and 2-4.5psu for the RMSE.

Figure 4-4. Calibration of the CMOP stations

4.4 Results

4.4.1 Salinity intrusion

Comparison of the historical and modern models (**MM**, **MH** and **HH**) shows that hourly estimates of the X2 (2psu) isohaline follow the seasonal pattern of the river hydrograph for each case, but are modulated by the tidal signal on daily and monthly tidal time scales (Figure 4-5). Comparison of Figure 4-5a (**MM**) and Figure 4-5b (**MH**) illustrates the effect of the changing hydrograph on salinity since both the **MM** and **MH**

models are based on modern bathymetry. A significant seasonal change in salinity intrusion can be found during winter and spring. A maximum decrease (~ 10 km) in the intrusion length observed in hourly model data occurs during winter because modern discharge is significantly larger (50 %) than historical discharge (Figure 4-5). By contrast, salinity intrusion during the spring-freshet is greater than it used to be by up to 4km because the modern freshet is only 44% as large as it was historically. Considering all tidal phases, the salinity intrusion range is 8-40 km for the modern 50th percentile run (**MM** case) while it was 5-45 km for the historical flow (**MH** case; see Table 4-2). The larger variability in the tidally varying X2 intrusion during historical discharge conditions stems from the overall greater variability in discharge over a year. The average historical hydrograph varied from 2,500 to 14,500 m³/s, while the modern hydrograph varies from 3,000 to 8,500 m³/s (see white lines in Figure 4-5a and Figure 4-5b). Despite the 40% greater range, however, the overall difference in X2 range is only 11%; the relative insensitivity of salinity intrusion to discharge changes, as shown later, is driven by the approximately $Q^{-1/3}$ relationship between salinity intrusion and flow (see also Equation 4.1).

The effect of the bathymetric change is observed by comparing cases (b) **MH** and (c) **HH** in Figure 4-5. The two cases show the same seasonal pattern of salinity intrusion because hydrological forcing is the same; however, much less salinity intrusion is observed in the **HH** case. The salinity intrusion for the 50th percentile run was shifted landward from 3-30 km in the **HH** case to 5-45 km due to changes in the bathymetry since the 1800s, (Table 4-2). As shown below, the deepening of the channel made the system more stratified

and pushed intrusion further landward, for the same historical hydrograph. The combined effect of both river flow alteration and channel deepening is demonstrated by comparing a-**MM** and c-**HH** in Figure 4-5. As shown, the salinity intrusion has shifted landward from 3-30 km to 8-40 km for the mean flow. Therefore, the decreased dynamic range in river discharge (which tends to decrease the range of salinity intrusion, both by increasing low-flow and decreasing high flow) somewhat counteracts the tendency of channel deepening and other bathymetric changes to move salt further landward. Overall, however, the results still suggest an upstream movement of the maximum salinity intrusion.

Figure 4-5. Instantaneous (hourly) Salinity intrusion seasonal variation of the South Channel. The 2-psu isohaline is the limit between blue color and the other colors. a) MM, b) MH, and c) HH. The white curve is the river hydrograph.

Higher salinity intrusion is correlated with lower river discharge. The salinity intrusion variation based on the 25th percentile of discharge (i.e., a relatively dry year) suggests that the tidally varying intrusion (hourly time scale) could reach up to 55km for the modern case (**MM** model) compared to a maximum of 35km in the historical scenario (**HH** model) under the 25th percentile discharge (Table 4-2). Hence, during low-flow

conditions (25th percentile discharge), as much as 10 km of landward intrusion has occurred as a result of flow management and bathymetric change. Conversely, on a wet year (75th percentile or river discharge), only a 4 km increase in salinity intrusion has occurred (Table 4-2).

Table 4-2. Maximum and minimum hourly salt intrusion in km from the mouth of the CRE (using 2-psu isohaline) for the three cases (MM, MH, and HH), using mean flow, 25th, and 75th percentiles of flow

Salinity intrusion (km)	Modern bathymetry & Modern flow (MM)			Modern bathymetry & Historical flow (MH)			Historical bathymetry & Historical flow (HH)		
	25 th	50 th	75 th	25 th	50 th	75 th	25 th	50 th	75 th
Min	11	8	6	7	5	2	5	3	2
Max	45	40	34	48	45	44	35	30	29

To better understand the net effect of flow and depth variation on the salinity intrusion, a tidally averaged salinity intrusion for each scenario was considered (Figure 4-6). The range of tidally averaged intrusion was pushed landward from 6-27km (**HH**) to 12-43km (**MH**) due only to bathymetric change. The combined effect of flow management and bathymetric alteration has increased X2 by 14-36km (**MM vs. HH**), see (Figure 4-6). In the deeper channel cases (**MM & MH**), as shown below, the salinity intrusion pushed further landward than for the shallower channel (**HH**) because of stratification. The timing of maximum salinity intrusion has also shifted, due to river discharge hydrograph alteration (Figure 4-2). The historical maximum intrusion occurred in early January but occurs in September in the modern model. For the 25th and 75th percentile discharge, the same time-reversal was noted. The salinity intrusion results shown are for typical historical and modern flow patterns, but the actual salinity intrusion depends on year-to-year variation.

Different years have very different flow patterns, with the extremes of mean annual flow varying by a factor of about 2.7 between 1878 and the present (Naik and Jay, 2011).

Furthermore, a change in the spring-neap intrusion was noticed due to the bathymetric and flow variation. The spring-neap transition has changed such that a mean of ~7km variation between successive spring-neap cycles occurs in the **MM** model compared to ~5km in the **HH** model (Figure 4-6). Changes in the spring-neap transition between the **MM** and **HH** model are caused by both discharge changes and bathymetry changes. The effect of bathymetry can be isolated by comparing the **MM** and **MH** runs. Here, I see that under historical river flow, the modern spring-neap transition would be slightly enhanced, with an average of 10km. Therefore, I conclude that bathymetric change is the primary driver of weekly and monthly variability in intrusion (while discharge dominates the seasonal variability). The increased variability of salinity intrusion over a two-week and monthly time scale occurs due to greater salinity stratification caused by channel deepening, see section 4.4.2. However, it remains true that LCRE shows maximum salinity intrusion on neap tides for most flow conditions. Only for the highest flows in the **HH** case does the maximum salinity intrusion shift to spring tides, a situation suggested by Jay & Smith (1990).

Figure 4-6. Tidally averaged salt intrusion based on the mean flow, gray lines mark the neap tide

4.4.2 Salinity intrusion sensitivity to the river flow.

Theory suggests that the salinity intrusion (X_s) is inversely proportional to the river flow (Q) as $X_s \sim Q^{-n}$, where $n = 1/3$. Using the river discharge portion that passes through the South Channel only (see Chapter 3), the value of the exponent n was found to be ~ 0.42 for both river flow scenarios on modern bathymetry (models **MM** and **MH**). As might be expected, changing the seasonal flow magnitude and distribution does not affect this relation. However, changing the bathymetry does alter the salinity intrusion-river discharge relation. For simulations run on historical bathymetry (case **HH**), the exponent n is found to be 0.56, rather than 0.42 (Figure 4-7a). Therefore, changing channel bathymetry has not

only pushed the salinity intrusion further landward but may have reduced its sensitivity to river flow as well. These results are different when assuming the whole river discharge is passing through the South Channel. Using BAT river discharge, rather than accounting for flow bifurcation between the north and south channels (see Chapter 3), produces $n=0.33$ for both MM & MH models and $n=0.46$ for the HH model (Figure 4-7b). Therefore, the change in the flow sensitivity has been caused in part by the change in the redistribution of flow between the North and South Channels.

A possible reason for the difference in exponent n between time periods is methodological, rather than physical. Since the exponent is fit with nonlinear regression, it is quite sensitive to the (somewhat arbitrary) position of the coordinate system. The modern 'River mile zero' is defined by the Army Corps of Engineers to be at the intersection of the shipping channel and a line that bisects the seaward limit of the North and South Jetty. When the origin (X_0) is moved further seaward in our nonlinear regression, the exponent n increases, and vice versa (Figure 4-7c). Since the historical entrance to the estuary is landward of the present-day X_0 , it is therefore likely that our estimated coefficient value n is at least somewhat of an artifact of how one fits a line. When the origin of Clatsop spit is used, the historical n -value is 0.42. These results are consistent with the boundary of the estuary having moved further seaward since the 1800s. However, there is significant uncertainty about the correct location to put the estuary boundary, and therefore the correct functional dependence of the exponent n . For example, if I move the

origin of the modern model landward by 8km, one could obtain the $n = 1/3$ theoretical relationship.

Figure 4-7. (a) River discharge tidally averaged intrusion relation at the South Channel, where $X=0$ km (at the mouth) is obtained from the corps of engineers (see Figure 4-1-a). Q_H refers to the historical hydrograph while Q_M is the modern one. (b) river discharge power (n) sensitivity to zero km (X_0) selection.

Salinity intrusion is also sensitive to stratification, such that the vertical mixing is restricted in the presence of high stratification, which leads to greater intrusion (Monismith, et al., 2002). Model results confirm that greater salinity intrusion is associated with larger

stratification during neap tides (Figure 4-8, region B). However, larger river flow can increase the salinity gradient which pushes the salinity intrusion towards the ocean (Geyer & MacCready, 2014). This results in a strongly stratified system with a short intrusion length that is close to the river mouth (Figure 4-8, region A). The reverse is also true, such that during low river flow the salinity flux can reach further upstream (Figure 4-8, region C).

Figure 4-8. South Channel stratification -salinity intrusion relation for the modern model (MM). The blue curve is tidally averaged salinity intrusion (X_2) time series. The orange is a tidally averaged stratification index (Φ). Region A represents a high flow period, region C represents a low flow, and region B shows the correlation between X_2 & Φ

4.4.3 Stratification

Stratification changes were analyzed using the stratification index ($\emptyset$) which is the difference in the bottom to top salinity divided by the bottom salinity (zero indicates well mixed, while one means strongly stratified). I define the system as ‘stratified’ when the index is greater than 0.5 stratification (i.e $\emptyset > 0.5$). For the South Channel and the mean flow scenario, it is found that the **HH** model is stratified about 59% throughout the whole year of the simulation period (hourly estimates), while the **MH** model is stratified 77% of the time (Figure 4-9b and Figure 4-9c). This increase in stratification is due to bathymetric change only since both models have the same river discharge but different bathymetry. Likewise, the **MM** model is found to be stratified about 92% of the time during the simulation year (Figure 4-9a). In other words, deepening the channel and altering the river hydrograph have increased the likelihood of exceeding the $\emptyset = 0.5$ threshold from 59% to 92% (18% due to bathymetric change and 15% due to river flow change). The system alteration has made the system stratified most of the year except during very low flow periods (Figure 4-9a), though it was historically strongly stratified mostly during above average flows (Figure 4-9c). The change in stratification is primarily because of the seasonal redistribution of the flow during the modern periods for hydropower generation and flood control, as more water is released during winter for power generation (Naik & Jay, 2011). The increased discharge during low-flow periods leads to more stratification. Secondary causes of altered river flow (and hence salinity stratification) include deforestation and climate change (more precipitation falls as rain in winter and the spring snowmelt is earlier;

Matheussen et al., 2000; Hamlet et al., 2013), and irrigation withdrawal, which reduces mean flow, most strongly at the end of the spring freshet (Naik & Jay, 2011).

Figure 4-9. South Channel stratification. (a) MM, (b) MH and (c) HH.

4.4.4 Parameter space evaluation

To show the aggregate change of altered tides, depth, and river forcing on the salinity dynamics of the Columbia River estuary, I employ the non-dimensional parameter space plot developed by Geyer & MacCready (2014). Plotting model results onto the Geyer & MacCready parameter space is challenging in an estuary with irregular geometry since it is unclear what representative depth, width, and tidal velocity to use. The choice of a specific depth, width, and tidal velocity could move the location of an estuary from one

regime to another. To overcome this problem, I apply both Eulerian and Lagrangian approaches and qualitatively compare the resultant parameter space to actually modeled results. In this way, I ensure that a forcing condition that is placed, for example, in the "highly stratified" parameter space, is actually highly stratified over a modeled tidal period.

The Eulerian method is applied at three locations along the South Channel of the estuary: at the mouth (rkm0), in the middle of the estuary (SAT3, rkm13), and Tongue point (rkm29), see Figure 4-1-a for locations. At each location, the mixing time scale (T_m) was found from the relation $T_m = H^2/K_m$ and compared to a half-tidal cycle time scale (T). Note that the bottom boundary layer would likely reach the surface if $T_m < T$. The local depth is H , and K_m is the eddy viscosity coefficient as averaged from model results over the water column at each station. The mixing time scale (T_m) was found for each flood and ebb period during high flow (9,000CMS) and low flow (3,000CMS) periods. The local depth and eddy viscosity coefficient were averaged over a half-tidal cycle, see Figure 4-10. The estuary is considered completely mixed if mixing time is less than the half-tidal time for all three stations. In the Geyer & MacCready (2014) parameter space plot (Figure 2-2 and also below), the mixed condition is the parameter space located to the right of the diagonal red line.

The Lagrangian method uses the 2-psu isohaline to define a time-varying region over which to evaluate the freshwater Froude number (Equation 4.2) and the Mixing number (Equation 4.3). The water depth (H), channel width (W), and tidal velocity are spatially averaged over the region in between the mouth and the 2-psu isohaline in 2.5km

intervals (Figure 4-11). Then (H) and (W) are averaged over the duration of both the flood and ebb tide, and the tidal velocity is estimated as the root-mean-square of hourly-spaced model output.

A comparison of the Lagrangian and Eulerian methods of estimating tidally averaged Froude and mixing numbers shows that they both agree with the vertical salinity structure profile. An example of this analysis is the vertical salinity structure plots of the modern model (MM) during high river discharge (9,000 CMS, mean; Figure 4-12). Each plot represents either the end of a flood period (F) or the end of an ebb (E) while the numbers in Figure 4-12 represent the day of that month (June). For both flood and ebb, the estuary is found to be mostly stratified during the whole month except for the five tidal periods that are enclosed by a green rectangle in Figure 4-12. The analyses of high (9,000CMS) and low (3,000CMS) river discharge simulation data showed that the Eulerian method matched the Lagrangian approach in 93% of cases (i.e., it is in the same bin in the parameter space). The occasional mismatch occurs when the Eulerian approach slightly overestimates the mixing time scale. This may result from averaging the local depth and eddy viscosity coefficient over a half-tidal cycle. For the historical model (HH) analysis, I found full consistency between the Eulerian and the Lagrangian approaches for the high flow, and 88% matches between Lagrangian and Eulerian for the low flow condition.

Figure 4-10. Eulerian Approach, (1) is the RKM0 location at the mouth, (2) is the SAT3 location (11.5km from the mouth), and (3) is the Tongue point location (25Km from the mouth). The red dots are for the end of the flood, while the green ones are for the end of the ebb period. The black solid line is the mean flood time and the dashed line is the mean ebb time. Finally, the top row is for the high flow conditions and the lower panel is for the low flow period.

Figure 4-11. Eulerian vs. Lagrangian approaches. The three Eulerian locations are RKM0, SAT3, and Tongue point (TP). The Lagrangian approach is from Rkm0 to the white line (2psu isohaline moving boundary).

Figure 4-12. Vertical salinity structure of the modern model (MM) for the South Channel during the high flow period. F & E denote the spatial structure at the end of flood and ebb periods; numbers represent the number of floods/ebbs since the beginning of the month. Green rectangles enclose the situation where the total vertical mixing is likely to happen.

After applying the Eulerian & Lagrangian approaches for quality control, the model data for both modern model (**MM** case) and historical model (**HH** case) were plotted on the Geyer & MacCready (2014) parameter space using the Lagrangian approach (as shown in Figure 4-13). Each point in Figure 4-13 represents the end of each flood and each ebb for the whole simulation year. In the modern model (**MM** case), it is found that the estuary occupies the strongly stratified regime 50% of the time, the salt wedge regime 34% of the time, the partially mixed region 10% of the time, and the time-independent salt wedge 6% of the time. By comparison, in the historical model (**HH** case), the percentages are 50%, 4%, 41%, and 4%, respectively. Thus, changes in the river discharge and bathymetry have

caused the modern system (MM case) to spend more time in the salt wedge regime but less time in the partially mixed zone than the historical system (HH case).

Although the total river discharge (Q_T) range is higher in the HH model, the percentage of freshwater going through the South Channel (Q_S) is lower in the MM model. Because of the higher discharge and the greater cross-sectional area (A), the river velocity (U_R) range in the MM model is larger than in the HH model (Table 4-3). Hence, the MM model occupies the salt wedge zone (determined by $Fr_f \sim U_R/H^{1/2}$, from equation 4.2) more frequently despite the bathymetric changes. Likewise, even though the tidal velocity (U_T) range in the MM model is larger than in the HH model (Table 4-3), the range of mixing number is still larger in the HH model because of its shallower depth ($M \sim U_T/H^{3/4}$, from equation 4.3). Hence, the MM model spends a lower percentage of time in the partially mixed regime.

Figure 4-13. Parameter space for the modern and historical models, where (a) & (b) represent the Lagrangian approach for flood and ebb of the 41 tidal cycles during June (high flow period), and (c) is the presentation of modeling results of modern and historical models

Table 4-3. The effective variables and ranges in the parameter space: Q_T is the river discharge at the BAT, Q_S is the freshwater in the South Channel, H is the water depth range, A is the cross-section area, U_T is the tidal velocity range, and U_R is the river velocity range in the South Channel.

	MM model	HH-model
Q_T [m ³ /s]	2,000-10,000	1,400-15,000
Q_S [m ³ /s]	1,100-5,200	160-3,750
H [m]	11-15	6.5-11
A [m ²]	$(2.4-4.3) \times 10^4$	$(1.7-3.9) \times 10^4$
U_T [m/s]	0.33-1.3	0.23-1.05
U_R [m/s]	0.04-0.17	0.008-0.12

4.5 Conclusions

Both river flow and bathymetrical variation can affect salinity intrusion in estuaries. The analysis of river flow effects on LCRE salinity intrusion shows that the time series profiles of salinity intrusion follow the same pattern of the river flow hydrograph. However, a significant seasonal change in salinity intrusion is found during winter and spring. Less salinity intrusion occurs during winter while greater intrusion takes place in spring because modern discharge is significantly larger in winter and much smaller in magnitude during spring-freshet than historical river flow. Moreover, the change in the bathymetry of the estuary caused by deepening the South Channel made the system more stratified, so that the intrusion pushed further landward.

The combined effect of both river flow alteration (50th percentile) and bathymetric change suggests an increase of about 10km in maximum hourly salinity intrusion into the system. Additionally, the time of occurrence of maximum intrusion has shifted from early January to the end of September, as it corresponds to the minimum river flow. Furthermore, a change in the spring-neap intrusion was noticed due to the bathymetric variation, such that there is more salinity intrusion variation between spring and neap in the modern model than in the historical one, though there was greater seasonal variability in neap-spring transitions historically. The deeper channel's salinity intrusion pushed further landward than the shallower channel because of the neap-spring transition in stratification. The bathymetric changes to the CRE have not only affected the timing and salinity intrusion

position, but also the intrusion-river flow sensitivity, on account of the change in water distribution between the South and North Channels.

Finally, the location and the range of the LCRE has changed on the Geyer and MacCready (2104) diagram. Given that the scales are logarithmic, small visual changes in Figure 4-13 are quite significant in process terms – on the whole, the system is now more stratified than it was historically. For both models, historical and modern, the estuary occupies four regimes in the parameter space: Salt wedge, time-dependent salt wedge, strongly stratified, and partially mixed. However, the estuary has changed in two major regimes such that it used to spend more time in the partially mixed and less time in the salt wedge regimes. Though the river flow alteration has contributed slightly to the regime changes as the change in the channel geometry has balanced the flow alteration, these regime changes are due to mainly to bathymetric changes.

In this research, I have shown the use of historical and modern bathymetry and flow with corresponding numerical model grids to evaluate changes in salinity intrusion patterns in the LCRE over the past ~150 years. By studying salinity intrusion over a year, I have expanded on previous short-term scale models of the LCRE and provided a broader overview of the system hydrodynamics due to variable tide and fluvial forcing. Finally, I have demonstrated the importance of the variable selection (depth, width, and tidal velocity) on the estuarine classification by utilizing different (Eulerian and Lagrangian) approaches.

Chapter 5 After the Earthquake and Climate Change: How Sudden Relative Sea-Level Changes May Alter Salinity Intrusion in the Columbia River Estuary?**5.1 Introduction**

The estuarine depth of the Lower Columbia River Estuary (LCRE) is increasing over time. This depth alteration can be man-made (dredging) or natural (sea-level rise or sudden land subsidence). In this chapter, I will explore the effect of possible future depth changes on the LCRE estuarine hydrodynamics. The Cascadia Subduction Zone (CSZ) is a 1000 km long fault along a subduction zone between Northern California and British Columbia, Canada (Clague, 1997). It is thought that mega- earthquakes (greater than M8.5) recur about every 400-600 years (Atwater et al., 2005). The last known mega-earthquake in the northwest occurred over 300 years ago (January 1700). The effects of the tsunami waves associated with this earthquake have been found in sediments up to Rkm 56 up the Columbia River (Atwater et al., 2005). In addition to a tsunami wave, average land subsided by up to 0.5m along the Oregon coast (Hawkes, et al., 2011; Mori et al, 2011). The maximum wave height resulting from the M8.7-9.2 earthquake could range from 4 m to 25 m along the shoreline of Bandon, Oregon as predicted via simulation by Witter et al. (2013). A simulation of a hypothetical tsunami at the Columbia River mouth by Yeh et al (2012) used a maximum tsunami amplitude of 5.6 m at the mouth. The wave height was modeled to decrease to ~1.5 m by Astoria Tongue Point and to about 0.35 m near Beaver Army Terminal (Rkm 86). Further, the subsidence in the LCRE resulting from the M9

earthquake could lower the land by 2m at the coast and 1.5m at Astoria with small or negligible change further upstream (Kalmbacher & Hill, 2015; OSSPAC, 2013).

There are three tectonic plates on the Pacific Northwest that are in continuous horizontal movement and could cause a major earthquake: The Pacific Plate, the North American plate, and the Juan De Fuca plate (Dipietro, 2013). The result of these movements causes a spatially variable uplift of the Oregon coastal zone. Talke et al. (2020) reported uplift activity of ~1.5 mm/yr near the coast of the Columbia River and uplift of 0.35-0.4 mm/yr in the Tongue Point; a similar pattern was reported by Burgette et al. (2009). Variable coastal uplift is confirmed through an analysis of sea-level data, leveling surveys, and Global Navigation Satellite System (GNSS) data (Burgette et al., 2009; Talke et al., 2020).

The projection of sea-level rise in Oregon (including causes both global and local) suggested by NRC (2012) suggests that sea-level rise could reach up to 1.43 m by 2100. Additionally, a potential earthquake ≥ 8 in magnitude could raise the regional relative sea-level (RSL) another 1-2m over these projected values, through land subsidence (Pfeffer et al., 2008). Based off of my previous simulations which shows that depth increases cause increased salinity intrusion (Chapter 4), I hypothesize that future increased depth would increase salinity intrusion. Further, sea-level rise could impact estuarine habitats by alter salinity, water levels, and inundation of shallow water habitat. Tidal marshes, for example, could convert to tidal flats or lagoons because of changes in their elevation range (Horton et al., 2018).

The potential for sudden land subsidence, projected coastal sea-level rise, and predicted changes to river hydrology (more winter flow, less spring and summer flow) have motivated this study. In this chapter, I test both the effect of a constant sea-level rise and a linearly varying bathymetry on salinity intrusion and tides, as possible scenarios, and compare them to a validated simulation of the present-day situation. Both these scenarios are plausible given current projections, and in fact may occur together, making the effects worse. These scenarios constrain the order of magnitude changes that might occur. However, any morphodynamic changes that result from the tsunami and its aftermath are not modeled, and could merit further consideration in the future.

5.2 Theory

A theoretical model of salinity intrusion was presented by Monismith et. al. (2002), based on the cross-sectionally and tidally averaged salt conservation equation (Hansen & Rattray, 1965) as:

$$X_2 = \frac{(W\alpha)^{1/3}(\beta g S_0)^{2/3} H^3}{Q^{1/3} \nu_t}, \quad (5.1)$$

Equation 5.1 assumes steady-state, constant depth, vertical eddy diffusivity is equal to the eddy viscosity, and assumes that tidally averaged circulation is the primary mechanism of salinity intrusion. The variable X_2 is the salinity intrusion length from the mouth of the estuary toward the upstream point at which the bottom salinity is 2-psu (Monismith et al., 2002; MacWilliams et al., 2015). W is the width, α is the constant of proportionality

(5.4×10^{-5}) , β is the saline expansivity (7.7×10^{-4} psu⁻¹), S_o is ocean salinity, H is the local depth, Q is the river flow, and ν_t is the eddy diffusion coefficient. Assuming that ν_t scales as $u_t H$, where u_t is the tidal velocity, then the salinity intrusion scale become:

$$X_2 = \frac{(W\alpha)^{1/3} (\beta g S_o)^{2/3} H^2}{Q^{1/3} U_t}. \quad (5.2)$$

The scaling in equation 5.1 shows that the salinity intrusion scales as water column depth squared (H^2), holding the other parameters constant (especially river discharge). In this theory, estuarine circulation (U_E) and stratification (ΔS) scale as follows (MacCready & Geyer, 2010):

$$U_E = \frac{g H^3 \beta}{48 \nu_t} \frac{\partial s}{\partial x}, \quad (5.3)$$

$$\Delta S = U_E \frac{H^2}{K_S} \frac{\partial s}{\partial x}, \quad (5.4)$$

where K_S is the eddy diffusivity. Using $\nu_t \sim u_t H$ and $K_S \sim \nu_t / S_c \sim u_t (H / S_c)$ (S_c is Schmidt number), the estuarine circulation and stratification scaling equation can be written as:

$$U_E = \frac{g H^2 \beta}{48 u_t} \frac{\partial s}{\partial x}, \quad (5.5)$$

$$\Delta S = U_E \frac{H S_c}{u_t} \frac{\partial s}{\partial x}, \text{ or } \Delta S = \frac{\beta g S_c}{48 u_t^2} \left(\frac{\partial s}{\partial x} \right)^2 H^3, \quad (5.6)$$

The scaling above suggests that the estuarine circulation and stratification are proportional to H^2 and H^3 , respectively. Monismith et al. (2002) & Ralston et al. (2008) approximate $\frac{\partial s}{\partial x}$ to be $\frac{S_o}{X_2}$ for the steady-state salt balance, so that estuarine circulation and stratification equations can be written as:

$$U_E = \frac{g H^2 \beta}{48 u_t} \frac{S_o}{X_2}, \quad \text{or} \quad U_E = \frac{1}{48} \left(\frac{g \beta S_o Q}{W \alpha} \right)^{1/3} \quad (5.7)$$

$$\Delta S = \frac{\beta g S_c}{48 u_t^2} \left(\frac{S_0}{X_2} \right)^2 H^3, \quad \text{or} \quad \Delta S = \frac{S_c (S_0 Q)^{2/3}}{48 H (\beta g W \alpha)^{1/3}}, \quad (5.8)$$

The last two equations suggest that estuarine circulation is independent of depth during steady-state conditions, and stratification is proportional to H^{-1} .

For the Hudson River Estuary, the results of Ralston & Geyer, (2019) agreed with the scaling in equation 5.7, in that the estuarine circulation is independent of channel depth (i.e., when the channel deepened, the salinity intrusion increased but there was no change in estuarine circulation). However, they found that stratification increased with increasing depth, which contradicted the steady-state balance in equation 5.8. This area of estuarine circulation theory remains unresolved.

5.3 Methods

The effects of relative sea-level rise and possible earthquake-induced land subsidence on tidal properties and salinity intrusion were simulated using a 3D hydrodynamic model. The model was calibrated earlier in Sections 3.3 and 4.3. Six possible scenarios were considered. The base case (B-2019) represents the current bathymetry and no rise in sea-level, case B-2100 simulates a 0.21m sea-level rise, and B-2300 simulates 1.5m sea-level rise. For each of the sea-level rise cases, I made an additional simulation which represented the effect of spatially variable land subsidence caused by a subduction-zone earthquake (Table 5-1). All cases were simulated using a river discharge boundary condition that applied two constant flow conditions (3,000 m³/s, labeled “LO”

and 15,000 m³/s, labeled “HI”). Each discharge was applied for two weeks to equilibrate the model, and then the model was run for a month to cover two full neap-spring cycles.

The projections of relative sea-level rise in 2100 and 2300 were selected based on the projections of Kopp et al. (2017) with a high greenhouse emission scenario (Figure 5-1). They projected a SLR with high (RCP8.5), moderate (RCP45), and low (RCP2.6) greenhouse emission scenarios. The relative sea-level rise utilized is the 50th percentile projection of the worst-case scenario (i.e., the model with high greenhouse emissions, Figure 5-1a). The Earthquake-induced land subsidence scenario is based on a hypothetical change in depth that is 2m at the coast, zero at the upstream boundary (Kalmbacher & Hill, 2015), and 1.5m at Astoria for a magnitude 9 earthquake (OSSPAC, 2013), see Figure 5-2

Table 5-1. Simulation scenarios for the constant river discharge.

Case	Sea-level Rise [m]	Subsidence
B-2019	0	No
B-2100	0.21	No
B-2300	1.50	No
T-2019	0	Yes
T-2100	0.21	Yes
T-2300	1.50	Yes

Figure 5-1. (a) Projected relative sea-level rise scenarios near CRE with 5 and 95% confidence interval, RCP8.5 is a sub-model with a high greenhouse emissions scenario, RCP4.5 is a moderate greenhouse emissions scenario, and RCP2.6 is a low greenhouse emissions scenario. The green plot is from NRC, 2012 predictions. (b) Sea-level rise stations along the west coast (Kopp et al., 2017).

Figure 5-2. Projected subsidence scenario in the LCRE a result of a Cascadia zone subduction earthquake; the scenario is 2m subsidence at the coast, zero at the upstream model boundary (Rkm 86), and 1.5m at Astoria, after Kalmbacher & Hill (2015).

I also assessed the impact of hydrological changes on the sessional variation of salinity intrusion. First, projected future river discharge conditions were extracted from USBR (2016) which predicts the potential hydrologic impacts of climate change in the Columbia River Basin. Their projected scenarios for the Columbia River above The Dalles are presented in Table 5-2. The tabulated data shows the percentage of changes in river hydrograph from the 1990s to the 2040s and 2080s. From this projection, I have chosen the minimum values of the 2080s to represent the future change in the hydrograph, where

the Winter discharge increases by 26% and spring freshet reduces by 17% (Figure 5-3). Note that this does not consider any changes in the management of the system. To account for changed river flow, four more simulations were conducted to represent the combined effect of SLR and river hydrograph on salinity intrusion (Table 5-3).

Table 5-2. Hydrologic modeling for the Columbia River above The Dalles [after the Bureau of Reclamation, (2016)]

Change from 1990s (1980-2009)	2040s (2030-2059)	2080s (2070-2099)
Mean Annual Runoff (%)	-5 to +10	-4 to +15
Mean Dec-Mar Runoff (%)	+13 to +44	+26 to +91
Mean Apr-Jul Runoff (%)	-8 to +8	-17 to +10

Figure 5-3. Future hydrograph changes (Winter increases by 26% and spring freshet reduced by 17%)

Table 5-3. Simulation scenarios for a variable river discharge

	River hydrograph	SLR [m]	Description
P-2019	Present	0	Recent river hydrograph w/o any SLR
F-2019	Future	0	Future river hydrograph w/o any SLR
F-2100	Future	0.21	Future river hydrograph w/ Future SLR
F-2300	Future	1.50	Future river hydrograph w/ Future SLR

5.4 Results

5.4.1 Tidal changes

Increases in depth are simulated to cause an increase in tidal amplitudes everywhere within the estuary, consistent with past measurements and models (Talke et al., 2020; Helaire, et al., 2019). The effect of depth change applied equally everywhere is illustrated in the comparison of Figure 5-4a to Figure 5-4b (low river discharge) and Figure 5-4c to Figure 5-4-d (high river discharge). Consistent with past studies (Jay et al., 2011; Moftakhari et al., 2013; Talke & Jay, 2020) higher river discharge tends to dampen the tidal amplitude (Figure 5-4a & Figure 5-4c or Figure 5-4b & Figure 5-4d). The comparison among all cases including sea-level rise and/or Earthquake-induced land subsidence can be illustrated using Figure 5-4e and Figure 5-4f for low and high river discharges, respectively. In addition to increased tidal range (M_2 amplitude), the location of the maximum tide has shifted into the upstream for all cases in response to the increased depth.

Results show that 0.21m RSL has a slight effect on M_2 , but a 1.5m RSL rise causes an increase in M_2 of 5-6%. The M_2 amplitude maximum is pushed ~4.1km and 4.2km further upstream for low and high river discharge, respectively. The earthquake-induced

land subsidence case has a similar effect as 1.5 m RSL on M_2 as 1.5m RSL (5-6% increase and 2.6-4.2km propagation), see Table 5-4. The combined effect of 1.5m RSL rise and land subsidence is simulated to increase the peak of M_2 amplitude by 11% (0.11m) with landward propagation of the M_2 maximum by about 10km during the low-flow simulation.

Figure 5-4. Changes in tides for scenarios indicated; squares represent the location of maximum amplitude. For South Channel location, see Figure 5-2.

Table 5-4. Sea-level rise and earthquake-induced land subsidence effects on M_2 amplitude and location of Maximum M_2 inside the South Channel of the estuary. Mean water depth is during low flow, averaged over neap-spring, and averaged spatially along the South Channel.

Cases	Location from mouth [km]	Max M_2 location shift [km]	M_2 Amplitude [m]	Increase in M_2 [cm]	M_2 Increment percentage	Mean depth, H [m]	ΔH [m]
Low Q							
B-2019	29.7	-	1.03	-	-	18	-
B-2100	29.7	0	1.04	1	1%	18.2	0.2
B-2300	33.8	4.1	1.08	5	5%	19.5	1.5
T-2019	32.3	2.6	1.08	5	5%	19.3	1.3
T-2100	33.8	4.1	1.09	6	6%	19.6	1.6
T-2300	39.7	10	1.14	11	11%	20.8	2.8
High Q							
B-2019	25.3	-	0.98	-	-	18.3	-
B-2100	25.3	0	0.99	1	1%	18.5	0.2
B-2300	29.5	4.2	1.04	6	6%	19.7	1.4
T-2019	29.3	4	1.04	6	6%	19.6	1.3
T-2100	29.3	4	1.05	7	7%	19.8	1.5
T-2300	32.5	7	1.08	10	10%	21	2.7

5.4.2 Salinity Intrusion

Salinity intrusion is greatly impacted by channel depth as suggested by salinity intrusion scaling in equation (5.1). The channel deepening effect is observed by the comparison of the baseline case and the subsidence case (without any sea-level rise) in Figure 5-5. Results suggest that the salinity intrusion would advance by 3.2km landward during spring (compare Figure 5-5a with Figure 5-5c) and 4.6km during neap tides (compare Figure 5-5b with Figure 5-5d). The advance of salinity intrusion is in response to the land subsidence, which in turn increased the local depth inside the estuary. The larger depth has reduced the bed stress by about 0.06 N/m^2 and 0.08 N/m^2 for spring and neap (averaged along the South Channel), respectively, compared to the base case (current bathymetry). The reduction in bed stress enhances the stratification (see section 4.4.2) and

reduces the vertical mixing, hence increasing salinity intrusion. The discrepancies between spring and neap intrusion are due to the higher spring tidal energy which enhances mixing and reduces the intrusion compared to neap tides (see Figure 4-6).

Figure 5-5. Salinity intrusion maps, white contours represent 31-psu & 2-psu isohalines.

A similar increase in salinity intrusion is found in other sea-level rise scenarios, or when earthquake-induced subsidence and SLR occur together. The worst-case scenario (Figure 5-6b) is associated with the higher sea-level rise case (1.5m) combined with the land subsidence case during neap tides (T-2300). In this scenario, the X2 (2psu) isohaline intrudes by 9.3 km landward during neap/low river flow conditions of 3000 m³/s, from 34.4 km to 43.6 km from the mouth.

Figure 5-6. Tidally averaged bottom salinity profile along the South Channel

A statistical comparison (Table 5-5) shows that there is little change in salinity intrusion for 0.21m SLR (0.6km increase in X2) but a substantial increase could happen in the case of 1.5m SLR or sudden land subsidence resulting from a major earthquake (9-magnitude) or both of them combined. For 1.5m sea-level rise, the spring salinity intrusion could increase by about 3.8km and 4.6km for neap tides. This is nearly the same change in the intrusion as when land subsidence occurs without any SLR (3.2km/spring & 4.6km/neap). In other words, the change in salinity intrusion that might occur in the next 300-years could happen nearly instantaneously, if a subsidence zone earthquake were to occur and cause the simulated subsidence.

Results show that other salinity isohalines have also propagated upstream. This can be seen in the 31-psu isohaline in Table 5-5, such that for 1.5m SLR, X31 advanced about 1.1km during spring tides but only 0.2km during neap tides. The land subsidence case pushed the 31-psu isohaline upstream by 3.2km and 0.4km during spring and neap tides, respectively.

Simulations, therefore, suggest that the subsidence case has slightly more impact on the horizontal salinity gradient than the sea-level rise case. Both the B2300 case and the land subsidence case are modeled to have a 1.5m increase in water level near Astoria, which is closer to the position of the X2 isohaline. Given similar depths, tides, and river flow, salinity intrusion behaves similarly. However, near the mouth, the 0.5m greater depth in the earthquake subsidence scenario leads to a greater intrusion of the X31 isohaline. As a result, the gradient of salinity is slightly larger in the subsidence case, in the mid-estuary.

Table 5-5. Tidally averaged Salinity intrusion in the South Channel during spring (SP) and neap (NE) tides and low river discharge (Q).

2-psu Salinity Intrusion (X2)									
Base	X [Km]				Tsunami	X [Km]			
	SP	Diff.	NE	Diff.		SP	Diff.	NE	Diff.
B-2019	30.0	0	34.3	0	T-2019	33.2	3.2	38.9	4.6
B-2100	30.6	0.6	34.9	0.6	T-2100	34.3	4.3	39.3	5.0
B-2300	33.8	3.8	38.9	4.6	T-2300	38.5	8.5	43.6	9.3
31-psu salinity Intrusion (X31)									
B-2019	4.2	0	11.4	0	T-2019	7.4	3.2	11.8	0.4
B-2100	4.2	0	11.4	0	T-2100	7.4	3.2	11.8	0.4
B-2300	5.3	1.1	11.6	0.2	T-2300	7.6	3.4	12.3	0.9

The net salinity intrusion (X2) changes, excluding the neap/tide variation, show that the intrusion could increase by 0.7km for 0.21m SLR and 4.6km for 1.5m SLR. The salinity intrusion could propagate about 4.6km upstream for a land subsidence case and about 9km for the combined effect of land subsidence and SLR (Figure 5-7).

Figure 5-7. South Channel average salinity transaction (averaged over the spring-neap cycle)

5.4.3 SLR & future hydrograph

Simulation results suggest that a change in future river discharge may also affect salinity intrusion. Further, any fluvial effects also interact with the change caused by increased depth (see section 4.4.1), by altering the freshwater distribution and increasing local depth. As shown in section 4.4.1, the temporal pattern of salinity intrusion in the South Channel follows the same pattern of the river discharge. The maximum salinity intrusion length (**X2mx**) occurs around October (low river flow period) while minimum intrusion length (**X2mn**) is in June (high Q) corresponding to the annual spring freshet (Figure 5-8).

The alteration of river discharge by increasing the winter flow by 17% and decreasing the spring freshet up to 26% produces a slight impact on salinity intrusion. This alteration pushed the minimum tidally averaged salinity intrusion of 0.6km (landward) and reduced **X2_{mx}** by 0.7 km only. The combined effect of both future hydrograph and 0.21m SLR also has a small effect, and increases **X2_{mn}** by 1.2km and **X2_{mx}** by 0.4km. The largest effect is for 1.5m SLR and the future hydrograph. For this condition, salinity intrusion could be 9.1km & 5km for **X2_{mn}** and **X2_{mx}**, respectively (Table 5-6).

Figure 5-8. Tidally averaged salinity intrusion (X2) along the South Channel

Table 5-6. Maximum and minimum limits of X2 for river discharge and SLR interaction (+ for percentage increase, and – for decrease)

Case	Description	Min. X2 [km]	%	Max. X2 [km]	%
P-2019	0m SLR + Present Q	15.0	--	37.0	--
F-2019	0m SLR + Future Q	15.6	+4	36.7	-1
F-2100	0.21m SLR + Future Q	16.2	+8	37.4	+1
F-2300	1.50m SLR + Future Q	24.1	+60	42.0	+13

5.4.4 Impact of SLR on habitats

Sea-level rise and/or land subsidence can alter the shallow water habitat by increasing salinity intrusion length (X2), water depth, tidal range, and inundation. The water depth comparison of recent bathymetry (B-2019) with 1.5m sea-level rise showed that the mean water depth over four islands within the LCRE (Figure 5-2) would increase as a result of SLR. Moreover, the habitat near the mouth has a larger depth change than those in the upstream. The mean water depth over Desdemona Sands could increase by 1.48m, while simulations suggest only a 0.65m increase on Marsh Island during low river discharge (Figure 5-9 & Table 5-7).

The projected river discharge changes also help in raising the water depth over those habitats. The river influence is also function of river mile, but in the opposite sense of the SLR effect. The maximum effect of river discharge starts at the river boundary upstream and then decreases in the ocean direction; the depth increases by 0.12m near the river boundary due to elevated river and decreases toward the ocean. Near the mouth, the river discharge does not affect the habitat's water depth (Figure 5-9). The change in water depth could alter the habitat classification and the plants and animals that live on/in it (Horton et al., 2018).

Figure 5-9. Change in water depth in shallow water habitats. Shading represents the range between low (3,000 CMS) and high river discharge (15,000 CMS).

Table 5-7. Water depth (averaged during spring-neap cycle) increment due to SLR and river discharge variation.

	River Q	Case	Desdemona Sand [m]	McGreder Island [m]	Karlson Island [m]	Marsh Island [m]
Water Depth [m]	High	B2019	1.43	0.53	0.27	0.11
		B2300	2.90	1.75	1.26	0.88
		ΔH	1.47	1.22	0.99	0.77
	Low	B2019	1.25	0.41	0.17	0.05
		B2300	2.73	1.45	1.06	0.7
		ΔH	1.48	1.04	0.89	0.65

5.5 Conclusion

Numerical simulation of the Lower Columbia River shows that climate change represented by river hydrograph alteration and/or water depth change from SLR or land subsidence has important implications for tidal properties and salinity intrusion.

A hypothetical SLR of 0.21m has no significant impact on salinity intrusion and tidal hydrodynamics inside the LCRE. On the other hand, a future 1.5m SLR has a

substantial impact on the estuarine hydrodynamics as measured by the M_2 tidal constituent. M_2 increased by 0.01m for 0.21m SLR without changing the location of maximum M_2 . However, M_2 increased by 0.05m for 1.5m SLR and the location of maximum M_2 advanced by about 4km landward. Sudden land subsidence that might occur from a potential major earthquake (M9) along the Oregon coast was simulated to produce a similar dynamic response in salinity intrusion and tidal amplitude to the 1.5m sea-level rise case.

The tidal analysis of both cases showed an increase in M_2 amplitude by 5% (0.05m) in the South Channel. Moreover, the location of maximum M_2 inside the estuary has moved upstream by ~4km for the SLR case and 3.5km for the subsidence case. The combined effect of both scenarios can double the effect of each case such that it can increase M_2 by 11% (0.11m) and push its maximum up to 10km (34%) landward for low flow conditions.

Salinity intrusion analysis showed the same trend for both SLR and subsidence scenarios (B-2300 & T-2019). Simulations suggest that X2 may advance by ~4km during spring and neap tides for both cases. However, the discrepancy between the two cases is in the horizontal salinity gradient. In comparison, the subsidence case caused a higher salinity gradient than the SLR case since it can push the 31psu isohaline (X31), because of the deeper water column at the mouth of the estuary. The deeper depth reduces the bed stress and lessens vertical mixing.

Alteration of the river hydrograph by +17% in winter and -26% during spring freshet produced a small effect on salinity intrusion, such that winter X2 advanced by 0.6km while spring freshet X2 reduced by 0.7km. However, 1.5m of SLR plus the future

hydrograph change produced a much greater effect. Together, these joint alterations increased X2 by 9km and 5km during low and high river discharge seasons, respectively.

Finally, future changes have a strong impact on shallow-water habitats, not only through the increase of salinity intrusion but also because of water depth changes. The water depth variation of the habitats (ΔH) has a maximum value near the mouth and decreases toward the upstream.

Chapter 6 Summary and Conclusion**6.1 Summary**

This study analyzed the effect of long-term changes in the Lower Columbia River Estuary (LCRE) that have occurred in the past 150 years on its estuarine hydrodynamics and transport. The bathymetry of the estuary has been changing since the middle of the 19th century as a result of navigation channel improvement and maintenance to accommodate increasingly large cargo ships. Additionally, large scale changes have included the construction of jetties, islands, and other navigational structures, and the reclamation and diking of shallow water habitat along the estuary. The other major change is hydrological--water withdrawal for agriculture and river flow management. Together, these have reduced the spring freshet for (flood control and irrigation) and increased the winter release of water for power generation.

In this thesis, I have investigated via numerical modeling the variation in freshwater distribution between the two main channels (South & North) inside the estuary, and also the tidal changes. The importance of these variables (buoyancy and tidal forcing) comes from their influence on estuarine hydrodynamics, especially salinity intrusion. The second question addressed was, '*How has salinity intrusion changed over the past century due to the system alteration (hydrological and morphological)?*'. The natural and man-made alteration will likely continue as a result of land development and climate change. The final research chapter, therefore, addressed the following question: *What will happen in the future to the estuarine hydrodynamics as a result of future river management, future sea-*

level rise, and a sudden land drop from a potential mega-earthquake along the Pacific Northwest?

6.2 Conclusion

Considering the forgoing research questions, I have used 3D-hydrodynamic models to simulate the historical bathymetry, historical river hydrograph, and modern bathymetry, modern river hydrograph. These models were calibrated intensively under barotropic and baroclinic conditions.

The results from the first research question (Ch.3) suggest that the river discharge distribution between the North and South channels is a function of tidal forcing, river discharge, and channel geometry. In the modern model conditions, the slope plays a major role in water distribution between the two channels. This asymmetry is driven by the length difference of water paths along each channel and by the non-linear interaction between the tidal forcing and river discharge. On the other hand, in the historical model, Manning's roughness coefficient and surface slope of both channels were found to be equal, hence the water bifurcation was controlled by the channel's geometry. The effect of tidal range on bifurcation was also noticed in the spring-neap variation, such that the percentage of freshwater in the South channel is higher on spring tides than on neap tides since spring tides tend to elevate the water surface slope. A further effect of system alteration was found on the tidal constituents such that modeled M_2 is elevated by 0.11m near Astoria Tongue Point mainly due to channel deepening.

The second research question (Ch.4) investigation revealed a significant seasonal change in salinity intrusion. Relative to the historical model, a larger intrusion was found in the modern model during summer due to spring-freshet reduction while the intrusion reduced during winter in response to larger modern winter flows. However, channel deepening has increased the stratification and helped propagate salinity intrusion further landward. The combined effect of the 50th percentile of river flow alteration and channel deepening suggests an increase of about 10km in maximum tidally varying salinity intrusion into the system, which occurs during low river flows and neap tides. Additionally, the time of occurrence of maximum intrusion has shifted from early January to the end of September, consistent with changes in the seasonality of minimum river flow.

The results also suggest a major change in the power law relating the X_2 salinity intrusion distance to river flow ($X_2 \sim Q^{-n}$). This change is mainly due to a change in river discharge distribution between the North and South Channel in the historical and modern estuary (discussed in Ch.3). Finally, the parameter-space examination showed that the estuary used to be more often partially mixed and act less salt wedge regime due to the bathymetrical change (on the whole the system is now more stratified than it was historically). The river flow change has a minor effect on the classification change since it has been balanced by the change in channel geometry.

An analysis the future changes to the system and in the event of rapid tectonic subsidence (Ch.5) showed that a 1.5m SLR and a linear bathymetrical change profile has a similar effect on salinity intrusion and tidal changes. In each case, the major tidal

component (M_2) is predicted to increase by around 0.05m and the maximum M_2 is predicted to advance ~4km landward. The combined effect of both scenarios can approximately double the effect of the individual changes such that it can increase M_2 amplitude by 0.11m and push its point of maximum amplitude about 10km landward for low flow conditions.

Similarly, both cases showed an advance of about 4km of salinity intrusion (X2), though the salinity gradient became larger in the land subsidence case than the SLR case. In other words, the change in salinity intrusion that might occur in the next 300-years due to SLR could occur much sooner if an earthquake occurs. The worst-case scenario is the combination of the two extreme cases (1.5m SLR and land subsidence) such that the effect on M_2 and X2 is more than double compared to each individual case. Moreover, a future 1.5m SLR and a future decrease in river flow can push X2 up to 9km landward during low river discharge seasons. The future changes would probably impact the shallow-water habitat as well. Simulation results showed an increase in water depth (1.5-0.8m) along the estuarine habitat (near the mouth and near the river boundary in the upstream, respectively).

6.3 Future steps and recommendations

Future studies could enhance my modeling results by expanding the hydrodynamic grid further into the ocean (200-300km), and overall investigate the possible influence of changed oceanic conditions (such as any changes to upwelling) on the oceanic salinity and

water temperature condition. Further investigation is also needed to address the relationship between salinity intrusion and channel depth (more simulations with different depths). The effect of the mouth modification (depth and width) on the hydrodynamics needs to be examined closely as well. More effort is needed to study the effect of salinity intrusion and climate change on water temperature variation, since water temperature is an important factor (like salinity) in the estuarine environment. Finally, future studies might study the effect of river discharge change, bathymetric change, and jetty constructions on plume dynamics.

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