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Using Satellite Observations to Characterize the Response of Estuarine Turbidity Maxima to External Forcing

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31 theoretical result that along channel distributions of turbidity should have an
32 asymmetric distribution, and emphasizes the role of bottom topography.

33 1.0 Introduction:

34 Estuaries and the surrounding coastal environment are greatly influenced
35 by sediment distribution and transport processes. Variations in river flow, tides,
36 particle settling velocity and bathymetry help control transport, deposition, and
37 erosion patterns that shape the morphology of these regions (Meade, 1969;
38 Stevenson et al., 1985; Jay et al., 1990; Sherwood et al., 1990; Geyer et al., 2001;
39 Elias et al., 2012). Since cohesive sediment particles also include nutrients,
40 organic matter, and contaminants, the distribution of sediment also helps regulate
41 ecological activity (Simenstad et al., 1990). For example, water column turbidity
42 inhibits algae growth (e.g., May et al., 2003), such that the spatial distribution of
43 turbidity helps set the spatial distribution of algal blooms (de Swart et al., 2009).
44 Similarly, the minimum oxygen concentration is often found in the estuarine
45 turbidity maximum (ETM) zone, due to the large concentration of organic
46 material (e.g., Talke et al., 2009a).

47 Despite the importance of estuarine turbidity to system functioning,
48 measuring and interpreting spatial patterns of turbidity remains challenging (Jay
49 et al., 2015). Few estuaries have the well-developed, long-term in-situ monitoring
50 programs required to assess seasonal patterns and the spatial effects of tidal,
51 riverine, and wind forcing (Jay et al., 2015). Logistics and costs limit the spatial
52 resolution of *in-situ* sensors. Moreover, *in-situ* point data are inherently difficult to
53 interpret when large spatial variability is present, since it becomes difficult to

54 separate vertical mixing processes from horizontal advection (i.e., the usual
55 assumption that advective transport is small breaks down when large spatial
56 gradients exist). To combat this deficiency, multiple studies starting with Postma
57 & Kalle (1955) have analyzed the turbidity distribution measured from moving
58 ships (see also Kappenberg and Grabemann, 2001; Uncles et al., 2002; Talke et
59 al., 2009; de Jonge et al., 2014). While shipboard transects obtain better spatial
60 resolution than in-situ data, the measurements are often obtained over multiple
61 hours or even a day, especially in long estuaries (Talke et al., 2008). Spatial
62 measurements are therefore only quasi-synoptic and often occur at different tidal
63 phases, depth, and mixing conditions. Hence, interpretation of the spatial
64 distribution is affected or biased by ship travel time and temporal aliasing.

65 Satellite measurements offer an opportunity to address the limitations of
66 in-situ measurements by obtaining instantaneous snapshots of an estuary with
67 much higher spatial resolution. A number of studies have developed methods for
68 remotely measuring suspended sediment concentration (SCC) in estuarine
69 environments using the MODerate Imaging Spectroradiometer (MODIS) and
70 other satellite-based instruments (e.g. Ruhl et al., 2001; Doxoran et al., 2003,
71 2006, 2009; Lehner et al., 2004; Chen et al., 2006; Palacios et al., 2009). These
72 studies qualitatively link spatial patterns of turbidity to tidal, fluvial, and
73 atmospheric forcing, demonstrating the potential of satellites for monitoring ETM.
74 However, these studies have typically analyzed only a few images, possibly
75 because satellite measurements have a long return period (typically 1 day to two

76 weeks) relative to tidal forcing and are affected by atmospheric conditions, such
77 that as few as 10% of images can be used. Hence, interpretation of results has
78 been qualitative and the potential of using satellite data to statistically
79 characterize estuarine turbidity has not yet been exploited. We address these
80 issues by applying a 'big data' approach and evaluating the entire ~15year MODIS
81 record. After applying quality flags and retaining only good quality images, we
82 conditionally sample the resulting data set of >1,500 images to quantify the
83 effects of river discharge, greater diurnal tidal range, and wind forcing on the
84 turbidity field. A seasonally varying climatology is produced, and the relative
85 contributions of different tributary streams to turbidity variability are determined.

86 Beyond quantifying spatial statistics, satellite measurements may allow us
87 to investigate aspects of the along-channel turbidity distribution that have been
88 difficult to ascertain with ship-board measurements or *in-situ* sensors. For
89 example, Talke et al. (2008, 2009b) suggest that the longitudinal profile of an
90 ETM is inherently asymmetrical, due to the different forcing mechanisms (river
91 flow, estuarine circulation) which dominate transport on either side of the ETM.
92 Differences in the particle size distribution along the estuarine salinity gradient
93 may also cause asymmetry in the suspended sediment distribution between the
94 two sides of an ETM. Longitudinal profiles of the ETM have been measured, but
95 (as discussed above) such shipboard measurements are not truly synoptic.
96 Moreover, such longitudinal profiles may not adequately assess lateral variability

in SSC. Hence, previously observed features in a turbidity field (such as multiple surface ETM) could (in theory) be due to either unresolved lateral variability or temporal variations in along-channel tidal forcing. In this study we analyze instantaneous transects of surface turbidity in the Columbia River Estuary (CRE) using satellite-based estimates to test the hypothesis that ETM are asymmetrical. We also investigate the existence of multiple ETM, and examine the physical mechanisms driving the turbidity distribution by monitoring the spatial and temporal variability of the ETM.

Our investigation addresses the following questions: 1) what are the tidal, fluvial, and seasonal patterns of ETM in the Columbia River Estuary? 2) are multiple surface ETM visible from satellite data? 3) what is the shape of the ETM under different conditions? and 4) which physical variables control ETM position and shape? Since satellite data are synoptic they can reveal spatially resolved features that cannot effectively be measured by either ship-based experiments or existing moored observations. These spatial snapshots can then be compared to theoretical descriptions of the turbidity distribution to reveal which mechanisms control circulation and transport.

2.0 Setting and Methods

In the following section we detail the geophysical characteristics of the Columbia River Estuary (Section 2.1), summarize data sources (Section 2.2), and

describe how turbidity is obtained from satellite remote sensing data (Section 2.3).

2.1 Site Characterization

The CRE is a drowned river valley in which salinity intrudes from the ocean into the lower estuary via two primary channels (North and South). These channels are separated by sand flats and are bordered on their other side by either coastline or large, shallow embayments. The estuary channels and subtidal flats have a shallow aspect ratio, with depths rarely exceeding 15m. In this study the system is divided into three regions (Fig. 1): The lower-estuary between river kilometer (Rkm) 0-20, where saltwater is nearly always present and mechanical energy is dominated by tidal and wave processes; A mid-estuary region (Rkm 20-30) which typically exhibits the largest salinity gradients observed in the system; and an upper-estuary region (Rkm 30-45) where salinity is often absent but density gradients occasionally occur and affect the velocity field.

Tides are mixed (predominately semidiurnal) with an M_2 amplitude that varies by $\sim 10\%$ in the estuary, with a peak of 0.95m near Rkm 30. Most of the tidal energy is focused in the lower and mid-estuary, and begins decaying upstream due to bed roughness and river flow, despite cross-sectional convergence (Jay et al., 1990). The neap/spring cycle has a clear signal throughout the year with greater diurnal tidal range ranging from 1.6 to >3.6 m at Tongue Point. Tidal range is largest at the end of summer when the effect of

river flow on tide wave propagation and attenuation is minimal (Jay, 1987; Jay, 1991; Chawla et al., 2007).

The Columbia is the largest Northeast Pacific river system, with a watershed of $\sim 660,500 \text{ km}^2$. It is the source of 60% to 90% of freshwater input to the coastal ocean of Northern California, Oregon, and Washington (Pruter and Alverson, 1972). River flow exhibits strong seasonal variability, with the largest flows usually occurring during a May/June freshet. Much shorter winter freshets are fed primarily by coastal tributaries such as the Willamette and Cowlitz Rivers, while the smallest flows occur during late summer when precipitation and snowmelt runoff are minimal (Fig. 2). The annual regime of river flow has been dramatically altered by flow regulation, and to a lesser extent climate change over the last century. Thus, contemporary spring freshet flows have been reduced by $\sim 45\%$ and occur about a month earlier than their early 20th century counterparts (Naik and Jay, 2010; Naik and Jay, 2011). This flow regulation has led to a decrease in sediment input from the main stem Columbia River, and the CRE is currently sediment-poor and erosional (Templeton and Jay, 2013).

The residual (tidally-averaged) estuarine circulation in the CRE is highly variable in space and time (Hughes and Rattray, 1980; Jay and Smith, 1990a). Irregular bottom topography (Fig. 3) augments spatial variability of the circulation by locally enhancing estuarine circulation, producing along channel convergences and divergences in the residual flow, and intensifying or reducing vertical mixing (Jay and Smith, 1990a). The salinity field transitions from weakly

161 stratified (most of the tidal month) to strongly stratified conditions during and
162 directly following neap tides (Jay and Smith, 1990b). Flushing times and
163 adjustments in the salinity field are relatively short (~few days) and nearly in
164 phase with the spring–neap cycle and fluvial regime (Jay and Smith, 1990a),
165 therefore, scalar concentrations react quickly to changes in river flow, tides, and
166 residual circulation.

167 Suspended sediment in the CRE is primarily of fluvial origin, composed
168 mostly of fines (silt and clay) and aggregates, except during large tides and river
169 flows when boundary shear stress is sufficient to suspend sands from the bed (Jay
170 et al., 1990; Sherwood and Creager, 1990). Fines are supply limited; the capacity
171 for transport is nearly always present and fine sediments will move whenever they
172 are available. Coarse sediment (sand and gravel), in contrast, is transport capacity
173 limited and there is an abundant supply that becomes mobile only under
174 appropriate flow conditions (Naik and Jay, 2011). Suspended sediment
175 concentration (SSC) is typically less than 100gm^{-3} throughout the CRE (Fain et al.,
176 2001), a level much lower than in many other estuarine systems (Talke et al.,
177 2009; Doxoran et al., 2009). However, depth-averaged ETM sediment
178 concentration can occasionally exceed 500gm^{-3} (Gelfenbaum, 1983).

179 Net sediment transport in the CRE is driven in part by tidally averaged
180 estuarine circulation induced by horizontal density gradients (Hansen and
181 Rattray, 1965; Festa and Hansen, 1978), and internal tidal asymmetry emerging
182 from tidal modulations of vertical mixing and stratification (Jay and Smith,

1990a; Simpson et al., 1990; Jay and Musiak, 1996; Burchard and Baumert, 1998). Transport is also driven by tidal asymmetries in flow and mixing caused by frictionally-produced overtides (Geyer, 1993; Jay and Musiak, 1994, 1996; Burchard and Baumert, 1998; Chernetsky et al., 2010; Burchard et al., 2013). Horizontal convergences in the residual and tidal sediment transport serve to trap particles and form ETM, the magnitude and location of which depends on the strength of the tides, river flow, stratification, and bottom topography (Geyer, 1993; Jay and Musiak, 1994; Burchard and Baumert, 1998; Schoellhamer, 2001; Talke et al., 2008; Donker and de Swart, 2013).

ETM are found in both the North and South Channels of the CRE, in locations that are close to the upstream limits of salinity intrusion (Jay and Musiak, 1994; Fain et al., 2001), though topographic lows in each channel appear to influence ETM position, as discussed below. The SSC in the ETM is up to ten times greater than the fluvial supply (Gelfenbaum, 1983), though the residence time of sediment varies from a few days to several months (Fain et al., 2001). A third, more transient ETM has also been reported in association with the tidal intrusion fronts and a topographic low at about Rkm-5.

2.2 Data Sources and Processing

Satellite data used in this study are derived from surface reflectance measurements made by MODIS. Two versions of this instrument sample the CRE about 2 hours apart near midday aboard the AQUA (2002 to present) and TERRA

204 satellites (2000 to present). Data are available at various processing levels,
205 ranging from raw data (Level-0) to highly processed end-products (Level-3).
206 Following Doxaran et al., 2009, a Level-2 swath product (MOD09) processed by
207 the Level 1 and Atmosphere Archive and Distribution System (LAADS;
208 <https://ladsweb.nascom.nasa.gov/>) was used in this study. MOD09 is
209 atmospherically corrected and includes state and quality datasets that flag pixels
210 with clouds or cloud shadows, high or low aerosol concentrations, or that cover
211 land, and mark poor quality measurements. These flags were used to filter out
212 measurements that incorrectly represent the state of the water surface. Remote
213 turbidity estimates were calibrated using measurements from MODIS Band 1
214 (620-670nm) at a 250m resolution (Fig. 1), chosen to resolve gradients in the
215 estuary as well as possible.

216 *In situ* measurements of turbidity and salinity were derived from thirteen
217 stationary buoys (Table 1 and 2; Fig. 1) managed by the Center for Coastal
218 Margin Observation and Prediction (CMOP; <http://www.stccmop.org/datamart/>).
219 Turbidity observations used in this study were recorded intermittently from 2008
220 to 2015 (Fig. 4) at sub-minute intervals using WETLab's ECO FLNTU (measuring
221 at ~700nm) or Turner Designs' Cyclops 7 (620-715nm) fluorometers. Salinity
222 measurements are available intermittently at various locations from 2003 to 2015.
223 At each location, turbidity measurements are made within 2.5 meters of the water
224 surface. Data from the Saturn 01 profiling system in the North Channel (Table 1;

Fig.1) were conditionally sampled and averaged over the top 2m of the water column.

Surface and near bed salinity measurements at Saturn 01 in the North Channel (2008-2015) and Saturn 04 in the South Channel (2009-2015) were used to calculate salinity stratification. The along-channel salinity profile in the South Channel was parameterized using a hyperbolic tangent function, using a least squares fit of a hyperbolic tangent curve to salinity data following Warner et al. (2005) and Talke et al. (2009):

$$S(x) = 0.5 S_o \left(1 - \tanh \frac{x - x_c}{x_L} \right) \quad (\text{Equation 1})$$

where S_o is the ocean salinity, and x_L and x_c are parameters that scale the salinity gradient and mark its maximum location, respectively. A least-squares approach was applied to estimate x_c , x_L , and S_o , for near bed measurements (data from 2003 to 2010) and near surface measurements (data from 2003 to 2006) to construct a salinity profile. The along-channel coordinate with 2 psu salinity was then used as marker for salinity intrusion. Data in the North Channel were too sparse to allow estimation of salinity intrusion.

Factors that affect turbidity include river discharge, tidal range, wind speed and direction, and ocean swell. These data are obtained from the USGS (United States Geological Survey) and NOAA (National Oceanic and Atmospheric Administration) and were recorded at various locations throughout the CRE

(Table 3). River flow was processed as daily averages and all other measurements as hourly averages. Ocean swell was defined as the integrated spectral energy at frequencies <0.1 Hz (period $T \geq 10$ seconds), while wind-speed was separated into two time series according to wind direction to take into account the effect of wind direction on wave fetch, following the approach of Talke and Stacey (2003). One wind-speed time series consisted of wind blowing along the estuary axis from the west northwest (typical for summer conditions), while the other consisted of wind blowing from the east southeast, more typical of winter conditions.

The forcing variables—wind, waves, tides, and river discharge—were then regressed against MODIS-based surface turbidity estimates (described below) to determine the spatial zones of influence for each forcing variable. Other forcing variables are held approximately constant during each regression via a binning approach, to minimize the possibility that correlations are driven by another, jointly correlated process (such as seasonal changes in flow, wind, and tidal range). For example, we test the effect of wind direction and wave energy on turbidity by requiring that greater diurnal tidal range and river flow be in the range 2.5 ± 0.5 m, and $5500 \text{ m}^3 \text{ s}^{-1} \pm 1500 \text{ m}^3 \text{ s}^{-1}$, respectively. These bin sizes are a compromise between holding forcing variables constant, to first order (20-30% deviation from mean), while retaining enough samples to remain statistically significant. To account for the observation that erosion must occur over a critical stress, we further process the wind series by considering only wind-speeds above a threshold. While empirical studies at other locations find a wind threshold of ~ 5 -

6 ms⁻¹ for local erosion (e.g., de Jonge and van Beusekom, 1995; Christie et al., 1999), we find that a lower threshold of 2 m/s obtains reasonable results while retaining a statistically significant sample size. The low threshold likely results from the presence of large mid-estuary flats and extensive peripheral bays. While the mid-estuary flats are primarily sand-bedded, they can temporarily retain fines (Sherwood and Craeger, 1990).

2.3 Remote sensing of turbidity

Satellite-based instruments record *Top of the Atmosphere* (TOA) radiation, which is converted to surface reflectance (the ratio of reflected/backscattered radiance to incident solar irradiance) by adjusting raw measurements for solar and sensor zenith angles, gaseous and aerosol scattering and absorption, adjacency effects, and cloud contamination (cf. Vermote and Vermeulen, 1999). Surface reflectance is measured in discrete frequency bands that correspond to peaks in the reflectance of various optically active constituents (OACs), such as suspended sediment or chlorophyll-a. Since the optical properties (e.g. absorption and backscatter spectra) of OACs change with concentration and significantly alter the measured reflectance signature (Bricaud et al., 1981; Bricaud et al., 1995; Neukermans et al., 2012), a relationship between satellite and *in situ* data can, in principle, be established.

Calibration of remote-based measurements of OACs is carried out by empirical or semi-analytical algorithms. Semi-analytical algorithms are based on

modeling radiative transfer phenomena and rely on the deconvolution of the measured reflectance spectrum of a water body into individual OAC concentrations (Maritorena et al., 2002; IOCCG, 2006). By contrast, empirical algorithms rely on statistical correlations between remote and *in situ* data. They often involve band arithmetic formulae and have been successfully applied in various estuarine systems (Doxaran et al., 2003, 2006, 2009; Hu et al., 2004; Chen et al., 2006). Since OAC optical properties are not fully characterized in the CRE, semi-analytical methods applied to the Oregon coast typically fail in the estuary (Palacios et al., 2009). Therefore, we use an empirical model to derive satellite-based estimates of CRE surface turbidity.

The empirical model of surface turbidity was calibrated using a total of 205 concurrent measurements of turbidity and cloud-free surface reflectance during the observation period. To remove noise and small scale variability, *in situ* measurements were averaged over a 5-min period centered at the time of each satellite passing. MODIS data with a Band 2 to Band 1 ratio >0.8 were excluded in the analysis. Ratios >0.8 correspond to land and cloud pixels missed in the MOD09 processing, and proved useful for screening pixels which overlapped the shoreline and exposed sand flats.

MODIS and *in situ* data are significantly correlated ($R^2 = 0.85$, error variance = 1.25ntu; Table 4 and Fig. 5), a relationship that most likely exists due to the proximity of the MODIS Band 1 spectral range (620-670nm) and the

detection wavelength of the *in situ* turbidity sensors (~700nm). Experimentation showed that a linear combination of Band 1 and Band 2 correlated slightly better with *in situ* data. Because this difference was not statistically significant, we use the simpler model based on Band 1 only (Equation 2 and Table 4). This calibration was then applied to more than 1,500 images with good quality flags from the years 2000-2015.

$$\text{Turbidity} = 0.59 + 137 * [\text{surface reflectance}] \quad (\text{Equation 2})$$

Many sources of error can degrade the relationship between MODIS and *in situ* data, including (but not limited to) atmospheric scattering, the presence of other OACs, the disparity between the *in situ* and remote detection wavelengths, reflectance off the bed in shallow regions, and land contamination for pixels near the shoreline. Furthermore, inherent differences in instrumentation and processing may also lead to errors in the model. Buoy observations are a point measurement that may not accurately assess spatial variability, whereas MODIS integrates variability in the top portion of the water column over a 250m by 250m area. Stratification in the water column, which is prevalent in the CRE, may augment depth-related discrepancies. However, the generally low error variance suggests that the sources of error described above do not significantly influence results within the turbidity maximum zone.

The contribution to surface reflectance of other OACs such as Chlorophyll-
a can also produce errors. Because primary production in the CRE is relatively low
and the local food chain is predominantly detritus-based (Simenstad et al., 1990),
we assume here that the most significant contribution to surface reflectance is
suspended sediment. SSC obtained from spot water samples by the USGS at
Beaver Army Terminal (Rkm 86) from 2010 to 2014 shows good agreement with
in situ turbidity measurements made at the Saturn 05 station across the channel
(data not shown; $R^2 = 0.88$, $N = 21$); hence, the assumption that surface
reflectance and turbidity are dominated by suspended sediment (and not
chlorophyll) appears appropriate to first order.

While we use estimates of turbidity as a proxy for sediment concentration,
interpretation of in-situ and remote turbidity estimates may be confounded by
particle size variability in time and space. Since increased bed stresses during
large tides and periods of strong current suspend larger particles from the bed,
and since aggregation processes vary within an ETM and with tidal phase (Reed
and Donovan, 1994), turbidity/sediment calibrations which are sensitive to
changes in particle size can undermine analysis of ETM behavior. Jay et al. (1999)
found that calibrations of suspended particulate matter against acoustic
backscatter varied over the tidal cycle in the CRE in response to changing
scattering properties of different sized material. Nonetheless, Neukermans et al.
(2012) found that WET labs optical sensors are driven to first order by particle

concentrations, with a weaker, second order dependence on particle size. Although this observation gives confidence in our calibration and interpretation of results, further research is required to definitively link sediment concentrations to our satellite-based turbidity estimates.

3.0 Results and Discussion

Next we analyze seasonal patterns (Section 3.1) and along-channel distributions of MODIS-based surface turbidity (Section 3.2) in relation to external forcing variables.

3.1 Seasonal Turbidity Distribution

MODIS-based estimates of surface turbidity display a high degree of temporal and spatial variability. Four basic system states are shown in Fig. 6: low ($3,800\text{m}^3\text{s}^{-1}$) vs high ($12,000\text{m}^3\text{s}^{-1}$) river flows during moderate tidal ranges (Fig. 6a and c), and neap (1.8m) vs spring (3.5m) tides during low river flow (Fig. 6b and d). As the source of most of the suspended sediment in the system (Jay et al. 1990), river flow has a direct relationship with surface concentrations throughout the estuary. Dependence on the neap/spring cycle is also apparent; greater concentrations are observed during spring tides when the larger tidal velocities mix sediment from the bed to the surface.

Monthly averaged turbidity fields in the estuary (Fig. 7) exhibit considerable seasonal variability during the study period; hence, the seasonally

changing supply of sediment from the rivers is likely important. Maximum system wide concentrations are observed in January and December during the winter rainy season, when considerable flow is input from tributaries west of the Cascade mountain range (daily average of $\sim 2,000 \text{ m}^3 \text{ s}^{-1}$; Fig. 2); i.e., from the Coastal Sub-basin (Fig 1.; Naik and Jay, 2011). Elevated turbidity levels persist throughout the spring freshet in May and June, with river flow measured at The Dalles averaging $\sim 8,000 \text{ m}^3 \text{ s}^{-1}$ and fed primarily by snow-melt in the Interior Sub-Basin. Both interior and coastal river flows subside during the summer dry season, reaching a minimum in September. The satellite-based climatology of turbidity (Fig. 7) is consistent with studies that show *in situ* sediment concentrations vary by a factor of 3 to 4 seasonally and up to ten spatially (Gelfenbaum, 1983; Sherwood and Creager, 1990; Fain et al., 2001). Fig. 7 also suggests that the spatial distribution of monthly averaged turbidity is relatively constant and less variable than seasonal changes in concentration. Large gradients in the turbidity field are always present in the lower/mid-estuary, while gradients in the upper-estuary are more muted.

Coastal turbidity patterns also exhibit seasonal variability. During winter months a large coastal plume deflects northwards, consistent with in-situ studies (Horner-Devine et al., 2009; Hickey et al., 2010) and the prevailing wind direction (see also Fig. 6). In upwelling conditions marked by northwesterly winds, the average turbidity is spread more symmetrically around the CRE mouth

(e.g., June and July). During parts of the year (e.g., September), some estimates are below the noise-floor (the error variance of the empirical model is 1.25ntu; Table 4) and results are inconclusive.

The factors controlling the turbidity distributions in Fig. 6 and Fig. 7 are examined by correlating time series of river discharge, greater diurnal tidal range, estuarine wind speed, and offshore wave energy with corresponding estimates of surface turbidity at each pixel location ($p\text{-value} \leq 0.01$). Results suggest that turbidity is significantly correlated with along channel, summer wind-speed ($>2\text{ms}^{-1}$) in shallow regions such as the mid-estuary sand flats, Young's Bay, and Cathlamet Bay (correlation coefficient, $R \sim 0.2\text{-}0.4$; Fig 8a). Winter wind-speed correlates with surface turbidity in the mid-estuary and Young's Bay as well, but also in Baker Bay (Fig 8b; see Fig. 1 for place-names). The differences between summer and winter turbidity patterns in shallow water likely stem from seasonal differences in prevailing wind direction, fetch, and therefore local wave forcing. Winter winds primarily come from the east-southeast, and therefore have a larger influence on turbidity in Baker Bay. By contrast, summertime winds are primarily west-northwesterly, such that their influence is concentrated along southern regions in the CRE. Note that although white-capping of wind-driven waves may influence the surface reflectance signal, evaluation of in situ data at Saturn 07 in Baker Bay confirm that turbidity increases during large wind events ($R \sim 0.5$, $N > 2,000$).

413 Wave energy from ocean swell (period ≥ 10 seconds) correlates with
414 surface turbidity estimates at the mouth of the CRE near the jetties ($R \sim 0.4$; Fig.
415 9), although we cannot evaluate whether the signal is due to white-capping from
416 wave breaking, turbidity, or both. Nonetheless, we observe that the elevated
417 MODIS reflectance at the jetties is consistent with observations that suggest that
418 wave-induced erosion is an important factor in jetty evolution (e.g. Hickson and
419 Rodolf, 1950; Elias et al., 2012). Unlike other estuaries (e.g., Talke and Stacey,
420 2008), the effect of ocean swell on turbidity cannot be statistically shown beyond
421 the entrance, though some low-frequency wave energy likely propagates inland
422 and to Baker Bay.

423 Greater diurnal tidal range, defined here as the difference between the
424 daily maximum and minimum tidal elevations, is significantly correlated with
425 turbidity throughout the system (median $R \sim 0.36$; Fig. 10). Since tidal currents
426 increase during periods with large tidal range (spring tides), vertical eddy
427 viscosity increases, stratification is reduced, and more sediment is mixed to the
428 surface (Allen et al., 1980; Jay and Smith, 1990a). The correlation between
429 greater diurnal tidal range and turbidity is maximal in mid-estuary, where tidal-
430 monthly variations in stratification and ETM concentration are greatest
431 (Gelfenbaum, 1983; Jay and Smith, 1990a). The greater diurnal tidal range also
432 shows a positive relationship with turbidity in the coastal ocean, consistent with
433 previous studies that suggest more sediment is exported to the coast during spring
434 tides when vertical mixing is amplified (Jay et al., 2007; Fain et al., 2001).

River flow is the most statistically significant process influencing variability in surface turbidity (Fig. 11). The total flow measured at the Beaver Army Terminal (Rkm 86), which measures an average of 97% of the total freshwater flow at the mouth (Orem, 1968) and includes both the main stem Columbia and most major coastal tributaries, shows a median R of 0.51 over the estuary (Fig. 11). We find, however, that different tributaries exhibit statistically distinct influences on the CRE turbidity. Coastal tributaries such as the Willamette (lagged by 1 day) and the Cowlitz (lagged by 1 day) exhibit a dominant influence within the estuary, with median R of 0.71 and 0.63 respectively. The main stem Columbia River (measured with a 2 day lag from The Dalles) exhibits a much lower correlation within the estuary (median R of 0.29), despite carrying an average of 75% of the total annual discharge.

Despite its weaker effect on surface turbidity levels, main stem Columbia River discharge is evidently large enough in May/June to make a significant impact on overall turbidity levels (Fig. 2; Fig. 7); during this period, the discharge from coastal tributaries is relatively small. The relatively weak correlation between surface turbidity and main stem discharge is likely augmented by a combination of the main stem sediment supply (see below) and stratification during the Columbia River freshet that damps vertical mixing and therefore the observed surface signal. Moreover, the larger river discharge in May/June exports significant quantities of sediment to the shelf (Spahn et al., 2009), by increasing seaward advection, reducing the salinity containing volume of the estuary that

can trap sediment, lowering the trapping efficiency (which is inverse to supply concentration), and decreasing suspended sediment residence time (Fain et al., 2001). Jay et al. 2007 provides a more complete discussion.

The influence of coastal tributaries, which are primarily driven by winter storm events (rain and sometimes rain-on-snow), helps explain why the largest turbidities are observed during winter (Fig. 7). Analysis of US Geological Survey water quality data and sediment concentration data indicate that the Willamette River sediment load typically has a higher percentage of fine sediment (with a strong influence on surface turbidity) than the load from the main stem (Naik and Jay, 2011). Wind may also play a role in enhancing winter-time turbidity (Fig. 8).

The observed influence of different tributaries on CRE turbidity is consistent with local geology but is strongly influenced by the reservoir system (Naik and Jay, 2010, 2011). As discussed by Gates (1994), the Lower Columbia River is thought to have been historically sediment poor, possibly because late Pleistocene floods scoured much of the Columbia River Basin. By contrast, coastal tributaries such as the Cowlitz and Willamette carry runoff from regions with active volcanoes and fine alluvium and aeolian deposits (Thesien, 1958; Balster and Parsons, 1968). The Cowlitz, which drains the Mt St. Helens watershed, has historically discharged large amounts of sediment to the CRE (Sherwood et al., 1990; Sherwood and Creager, 1990) and USGS measurements suggest elevated suspended sediment concentrations related to the eruption in May 1980 persist to this day (data not shown). Local development and land use practices may also

affect coastal tributaries like the Willamette (Lyons and Beschta, 1983; Naiman et al., 1992; Benner and Sedell, 1997; Norman et al., 1998). Since 1937 and 1941, dams in both the Columbia and Willamette River Basins have increasingly curtailed sediment export (Naik and Jay, 2011; Templeton and Jay, 2012; Wentz et al., 1998). However, the high correlation of turbidity with the Willamette hydrograph suggests that significant amounts of turbidity enter the system during coastal storm events. Because the Willamette and Cowlitz may also act as a proxy for smaller, unregulated coastal streams, we conclude that coastal tributaries generally are important sources of suspended sediment exported to the estuary and the coastal ocean.

3.2 Turbidity Transects

We next examine the longitudinal distribution of turbidity along the North and South Channels as defined in Fig. 1. Example transects are shown during low/high flows ($3,000\text{m}^3\text{s}^{-1}$ and $11,000\text{m}^3\text{s}^{-1}$ with $500\text{m}^3\text{s}^{-1}$ window; Fig. 12a,b) and neap/spring tides (2.0m and 3.5m with 0.15m window; Fig. 12c,d). Note that transects exhibit variance around the median (illustrated by the 33rd and 67th quantiles as shaded regions in Fig. 12), which is a result of inherent variance of environmental conditions within the bin window and possibly uncorrected atmospheric artifacts in remote measurements. Tidal excursion of the ETM ($\sim 20\text{km}$; Gelfenbaum, 1983) and time varying mixing also contribute to the variance because individual transects are observed during different tidal phases.

Since a large number of data are concatenated in each bin (>50), we argue that the median curves in Fig. 12 approximate the tidally averaged turbidity distribution for each greater diurnal tidal range/river flow combination.

3.2.1 Asymmetry in Turbidity Transects

The MODIS-based turbidity transects show that turbidity distributions are typically asymmetrical, with larger gradients observed downstream of the maximum. For example, gradients of $\sim 0.3 \text{ ntu km}^{-1}$ are observed downstream of the ETM vs 0.15 ntu km^{-1} upstream of the ETM during large river discharge and large tides in the South Channel (Fig. 12). These observations qualitatively agree with theoretical studies of ETM behavior, which suggest that asymmetry can develop because different physical processes dominate sediment transport on either side of the maximum (Talke et al., 2008, 2009b). Landward of an ETM, upstream sediment fluxes are small and the turbidity distribution is set by factors such as river flow, dispersion, tidal pumping, and channel convergence. By contrast, the turbidity distribution seaward of the ETM is controlled by the same factors that produce the salinity field (Talke et al., 2008), resulting in similar longitudinal length scales of variation (e.g., as encapsulated in Equation 1). The relationship between salinity and turbidity scales is observed in MODIS-based transects in the South Channel (Fig. 12d); during the transition from spring to neap, both the near-bed salinity field and the ETM move landward (increasing x_c in Equation 1) and become stretched over a larger distance (increasing x_l in

Equation 1). High flow conditions (Fig. 12b) exhibit increased gradients in both the salinity and turbidity fields downstream of the ETM.

Asymmetry in surface turbidity during high flow conditions can be quantified by fitting a hyperbolic tangent curve to either side of the ETM (similar to Equation 1 for the salinity field). Results confirm that asymmetry is especially prominent during high river flows as the ETM region approaches Rkm 15 in the North Channel and Rkm 20 in the South Channel (Fig. 12a,b); turbidity gradient lengthscales (x_L) are 4 to 5 times greater upstream of the ETM than downstream (Table 5). Because these locations have local depth maxima (Fig. 3) and landward sediment fluxes are enhanced with increasing channel depth (Talke et al., 2009b; de Jonge et al., 2014), topographic features may contribute to observed asymmetry.

3.2.2 Response of ETM to River flow and tidal forcing

The response of ETM magnitude and position to forcing data is examined with bin-median transects in the river flow ($1,000\text{m}^3\text{s}^{-1}$ window) and greater diurnal tidal range (0.25m window) parameter space (Figure 13). In both the North and South Channels, the magnitude of the ETM (C_{max}) is a function of greater diurnal tidal range and river flow. Surface C_{max} increases by a factor of 2-3 when flow rate increases from $2,500\text{m}^3\text{s}^{-1}$ to $10,000\text{m}^3\text{s}^{-1}$ (Fig. 12c and d), with the largest increase occurring during neap tide conditions (Fig. 13). For the largest river discharge bin with extensive data available ($10,000\text{m}^3\text{s}^{-1}$), C_{max} is

approximately constant regardless of greater diurnal tidal range. By contrast, turbidity values approximately double when tide range increases from 2 to 3.5m during low flow conditions ($< 5,000\text{m}^3\text{s}^{-1}$). These observations are consistent with the hypothesis that local resuspension and mixing processes dominate sediment concentration magnitudes during low river discharge, but that external sediment input from the river dominates estuarine sediment concentrations during large discharge events. We note that the background river turbidity of about 2 NTU during low discharge conditions (e.g., measured at Saturn 05 at Rkm 86) is significant relative to estuary conditions, and may contribute to the observed C_{max} , particularly upstream of Rkm 40, where channel width converges and salinity is generally absent. The dilution of river water downstream of Rkm 40 during low discharge conditions (as measured by increasing salinity) may therefore contribute to the overall longitudinal profile observed in Fig. 12b, in which no clear maximum is observed and turbidity decreases all the way to the ocean boundary.

The location of the ETM (X_{max}) in the North Channel is fairly stable, between Rkm 20-25, for a majority ($\sim 55\%$) of river discharge and tidal conditions (Fig. 13a). The possible exception is non-neap tides during low river discharge, in which X_{max} switches to an upstream locus between Rkm 25-30 (Fig. 12a and 13a). However, as shown in Fig. 12a, the statistical significance of the upstream movement is somewhat ambiguous. Since the $\sim 10\text{km}$ range over which

X_{max} varies and turbidity is elevated is much smaller than the tidal excursion we conclude that the tidally averaged position of the North Channel turbid zone or ETM is relatively fixed for all conditions, at least to within the accuracy of our estimate.

X_{max} in the South Channel is also found near Rkm 20-25 for a large proportion (49%) of the river discharge and tidal range parameter space. Interestingly however, X_{max} exhibits a different dependency on tidal forcing in the South Channel than in the North Channel, and is found further upstream (Rkm 30-35) as tidal range increases. During neap tides, reduced vertical mixing and higher stratification result in increased salinity intrusion and landward sediment transport (Gelfenbaum, 1983; Jay, 1990a; Talke et al., 2009); similarly, reduced river flow results in the upstream movement of salinity and turbidity (Fig. 12). The South Channel X_{max} location is especially sensitive whenever tidal ranges are close to 2.7m; X_{max} decreases by 5km as tidal range exceeds 2.7m for nearly all river discharges (Fig. 13b).

The location of the surface ETM is influenced by the stratification and mixing within the system and responds in concert with the salinity field to forcing data (Fig. 12). During spring tides, tidally averaged South Channel salinity intrusion (marked by the 2 PSU isohaline X_2) differs by <3km at the surface and bed (Fig. 12d; compare positions of pink square and circle). Under these conditions X_{max} is downstream of the near bed X_2 and occurs near the 8psu

contour. By contrast, surface and bottom X_2 are separated by $O(10\text{km})$ during neap tide conditions (Fig. 12d), indicating a strained salinity field and increased stratification. X_{max} is shifted landward of X_2 in this case and occurs near the 1psu contour. For intermediate greater diurnal tidal ranges, South Channel X_{max} exhibits a behavior in-between the spring and neap extremes for both small and large discharge (Fig. 12b).

3.2.3 Comparison of North and South Channels; Topographic trapping

The North and South Channels exhibit different reactions to increased river discharge. While the South Channel salinity field and zone of stratification moves downstream 10-15km as river discharge increases (Fig. 12b and 14c,d), the North Channel remains stratified at Rkm 16 for all but the largest discharge conditions ($>15,000\text{m}^3\text{s}^{-1}$; Fig. 13a,b). This behavior occurs both because the North Channel receives less river discharge (Jay and Smith, 1990) and is thus less sensitive to changing flow conditions, but also because of the topographic holes at Rkm 16 and Rkm 25 (Fig. 3) that trap saline water. Vertical mixing at Rkm 16 in the North Channel is thus nearly always inhibited to some degree, resulting in decreased surface turbidity compared to locations upstream. Strong stratification at Rkm 16 also implies vigorous landward bottom currents due to gravitational circulation (MacCready, 2007) and internal asymmetry (Jay, 2010), both of which scale as H^2 and strongly affect the ETM position (Jay, 2010; Talke et al., 2008;

Hudson, 2014). The decrease in depth from $\sim 16\text{m}$ to $\sim 10\text{m}$ on either side of Rkm 16 implies a two-fold alteration in upstream circulation over a small distance (not accounting for altered eddy diffusivity), and is a barrier to downstream sediment transport. The South Channel salinity and turbidity fields are much more sensitive to altered river discharge than the North Channel, but also exhibit evidence of persistent turbidity gradients downstream of the topographic holes at Rkm 20 during most flow and tidal conditions (Fig. 12b,d).

The scaling of estuarine circulation with depth supports the notion that trapping of turbidity occurs due to circulation patterns related to topography – i.e., due to “topographic trapping”. To demonstrate this idea, we note that two features of the turbidity field are associated with the holes at Rkm 16 and 25 in the North Channel. First, a large gradient of turbidity occurs downstream of the hole at Rkm 16 in the North Channel, consistent with a strong gradient in salinity and landward transport. Observations of Jay et al. 2007 confirm that landward transport and trapping efficiency in the CRE ($E = C_{max}/C_R$; where C_R = river concentration of particles) are indeed greatest in the North Channel near Rkm 15. Second, sediment trapping appears to be fixed between the topographic holes at Rkm 16 and Rkm 25 for much of the greater diurnal tidal range/river flow parameter space (Fig. 12 and 13). Although the surface ETM appears to persist in this region, surface and bottom turbidity maximum may not be coincident (as with the salinity field in the South Channel) and further research is necessary.

3.2.4 Trapping Efficiency

Scaling and analysis of data from 15 systems (Jay et al., 2007) suggests that trapping efficiency, E decreases as the supply number S_R increases:

$$S_R = \frac{U_R}{U_T} \frac{W_s}{\kappa U_*} \quad (\text{Equation 3})$$

where W_s is particle settling velocity, κ is von Karmann's constant, U_* the shear velocity, U_R is river flow velocity, and U_T the tidal flow velocity.

The Jay et al. (2007) scaling suggests that the following observations drive ETM behavior in the CRE:

- For constant river flow, supply number S_R increases and trapping efficiency E decreases during small (neap) tides, since weak tidal mixing during these times fails to move sediment from the bed. Fluvial sources of sediment may then represent the maximum concentrations in the system. Our results, in which relatively large concentrations are observed in the river relative to the ETM during neap conditions, are consistent with this interpretation (e.g., Fig. 12d).
- Conversely, large (spring) tides enhance vertical mixing in the estuary, local resuspension increases, and salinity intrusion decreases due to a reduction in estuarine circulation. These factors increase E , decrease S_R ,

and move the ETM downstream from the fluvial source (for the same river flow). Our results largely agree with this scaling (Fig. 12d).

4.0 Conclusions

In this study MODIS-based surface reflectance measurements and *in situ* data were used to gain further insight into surface turbidity distributions and transport processes in the CRE. Because MODIS data are synoptic and span a considerable time frame (~15 years), a “big data” approach yields spatial and temporal insights not easily available from either *in situ* measurements or field cruises. More than 1,500 images were used to develop a climatology of turbidity in the estuary and to quantitatively explore the processes that drive ETM behavior. Analyses of these images highlight the following conclusions:

- 1) Surface turbidity variations in the estuary are most strongly correlated with river flow; thus, monthly averaged concentrations are greatest during the winter months when river flows from turbid, coastal tributaries are maximum. Elevated turbidity persists, however, during the spring freshet in the mainstem Columbia, and annual minimum turbidity levels are observed during late summer. Greater diurnal tidal range controls surface turbidity concentrations to second order. As a result SSC is maximal and the ETM most prominent during spring tides when tidal currents and vertical mixing are elevated.

- 2) The synoptic nature of satellite data reveals two turbidity maxima in the CRE—one in the North Channel and one in the South Channel—with different responses to external forcing. Both ETM increase in magnitude with increasing river flow and tidal range. However, while the North Channel ETM location is relatively stable over the observed range of river flow and tidal range, the South Channel ETM migrates sea/landward about 20km with increasing/decreasing river flow and tidal range.
- 3) Although turbidity levels fluctuate during the year, the spatial distribution is more consistent, and large gradients are often observed in the mid/lower–estuary. MODIS–based turbidity transects also validate the theoretical expectation that ETM are inherently asymmetric. Because the distribution of sediment downstream of the ETM location (X_{max}) scales with salinity intrusion, and upstream of X_{max} with river discharge, an asymmetric profile can develop. In the CRE, turbidity gradients seaward of the ETM can reach twice those upstream of the ETM during large river flows and tidal ranges.
- 4) While large river flows and greater diurnal tidal ranges prompt seaward migration of the ETM, topographic trapping in mid–estuary near local depth maxima limits further downstream movement. Because internal asymmetry and gravitational circulation increase nonlinearly with water

683 depth, they provide a mechanism for limiting ETM movement at this
684 location.

685 These results highlight a fundamental characteristic of the CRE—since ETM
686 are not observed seaward of topographic lows in the mid/lower-estuary, and
687 since asymmetry in the turbidity distribution increases as X_{max} approaches this
688 region from upstream, bottom topography is likely an essential component
689 contributing to sediment transport and ETM formation. Indeed, maximal values of
690 trapping efficiency (E) occur in the North Channel at Rkm 15, near a local
691 depression in the bed (Jay et al., 2007). Topographic trapping of ETM has also
692 been observed in other estuaries (Roberts and Pierce, 1976; Schoellhammer,
693 2001; Ralston et al., 2012) and is therefore fundamental to the sediment
694 dynamics in many estuarine systems—especially those with strong baroclinic
695 gradients that drive depth-sensitive circulation. Despite considerable stratification
696 in the mid/lower-estuary (Fig. 12; Fig. 14), our measurements indicate that
697 surface signatures of topographic trapping can be observed with satellite data.

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References

Allen, G.P., J. C. Salomon, P. Bassoullet, Y. Du Phenhoat, C. De Grandpré. 1980. Effects of tides on mixing and suspended sediment transport in macrotidal estuaries. *Sedimentary Geology* 26: 69-90.

Balster, C. A. and R. B. Parsons. 1968. Geomorphology and soils Willamette Valley, Oregon. Soil Conservation Service, Special Report 265.

Benner, Patricia A. and James R. Sedell. 1997. Upper Willamette River landscape: A historic perspective. In *River quality: dynamics and resotation*, ed. Antonius Laenen and David A. Dunnette, 23-47. New York: CRC Press.

Bricaud, Annick, André Morel, Louis Prieur. 1981. Absorption by dissolved organic matter of the sea (yellow substance) in the UV and visible domains. *Limnol. Oceanogr.* 26: 43-53.

Bricaud, Annick, Marcel Babin, André Morel, Herve Claustre. 1995. Variability in the chlorophyll-specific absorption coefficients of natural phytoplankton: Analysis and parameterization. *Journal of Geophysical Research*, 100(C7): 13,321-13,332.

Burchard, H. and H. Baumert. 1998. The formation of estuarine turbidity maxima due to density effects in the salt wedge. A hydrodynamic process study. *Journal of Physical Oceanography* 28(2): 309-321.

726

727 Burchard, H., H. Schuttelaars, W. Geyer. 2013. Residual sediment fluxes in
728 weakly-to-periodically stratified estuaries and tidal inlets. *Journal of Physical*
729 *Oceanography* 43(9): 1841-1861.

730

731 Chawla, Arun, David A. Jay, Antonio M. Baptista, Michael Wilkin. Charles Seaton.
732 2007. Seasonal variability and estuary–shelf interactions in circulation dynamics
733 of a river–dominated estuary. *Estuaries and Coasts* 31(2): 269-288. doi:
734 10.1007/s12237-007-9022-7.

735

736 Chen, Zhiqiang, Chuamin Hu, Frank Muller–Karger. 2006. Monitoring turbidity in
737 Tampa Bay using MODIS/Aqua 250–m imagery. *Remote Sensing of Environment*
738 109: 207-220.

739

740 Chernetsky, A.S., H.M. Schuttelaars, S.A. Talke, 2010. The effect of tidal
741 asymmetry and temporal settling lag on sediment trapping in tidal estuaries.
742 *Ocean Dynamics* 60: 1219-1241. doi: 10.1007/s10236-010-0329-8

743

744 Christie, M.C., K.R. Dyer, P. Turner. 1999. Sediment flux and bed level
745 measurements from a macro tidal mudflat. *Estuarine, Coastal and Shelf Science*
746 49: 667–688.

747

748 Donker, Jasper J. A. and Huib de Swart. 2013. Effects of bottom slope,
749 flocculation and hindered settling on the coupled dynamics of currents and
750 suspended sediment in highly turbid estuaries, a simple model. *Ocean Dynamics*
751 63: 311-327.

752

753 Doxaran, David, Jean–Marie Friedefond, Patrice Castaing. 2003. Remote–sensing
754 reflectance of turbid sediment–dominated waters. *Reduction of sediment type*

variations and changing illumination conditions effects by the use of reflectance ratios. *Applied Optics* 42: 2623-2634.

Doxaran, David, Patrice Castaing, S.J. Lavender. 2006. Monitoring the maximum turbidity zone and detecting fine-scale turbidity features in the Gironde estuary using high spatial resolution satellite sensor (SPOT HRV, Landsat ETM+) data. *International Journal of Remote Sensing* 27: 2303-2321.

Doxaran, David, Jean-Marie Friedefond, Patric Castaing, Marcel Babin. 2009. Dynamics of the turbidity maximum zone in a macrotidal estuary (the Gironde, France): Observations from field and MODIS satellite data. *Estuarine, Coastal and Shelf Science* 81: 321-332. doi: 10.1016/j.ecss.2008.11.013.

Elias, E. P. L., G. Gelfenbaum, A. J. Van der Westhuysen. 2012. Validation of a coupled wave-flow model in a high-energy setting: The mouth of the Columbia River. *Journal of Geophysical Research* 117: C09011. doi: 10.1029/2012JC008105.

Fain, Annika M. V., David A. Jay, Doug J. Wilson, Phil M. Orton, Antonio M. Baptista. 2001. Seasonal and tidal monthly patterns of particulate matter dynamics in the Columbia River Estuary. *Estuaries* 24: 770-786.

Festa, John F. and Donald V. Hansen. 1978. Turbidity maxima in partially mixed estuaries: A two-dimensional numerical model. *Estuarine and Coastal Marine Science* 7: 347-359

Gates, Edward Breed. 1994. The holocene sedimentary framework of the lower Columbia River basin. Masters thesis, Portland State University.

784 Gelfenbaum, Guy. 1983. Suspended-sediment response to semidiurnal and
 785 fortnightly tidal variations in a mesotidal estuary: Columbia River, USA. *Marine*
 786 *Geology* 52: 39-57.
 787

788 Geyer, W. Rockwell. 1993. The importance of suppression of turbulence by
 789 stratification on the estuarine turbidity maximum. *Estuaries* 16: 113-125.
 790

791 Geyer, W. Rockwell, Jonathan D. Woodruff, Peter Traykovski. 2001. Sediment
 792 transport and trapping in the the Hudson River Estuary. *Estuaries* 24(5): 670-679.
 793

794 Hansen, D. V. and M. Rattray, Jr. 1965. Gravitational circulation in straits and
 795 estuaries. *J. Mar. Res.* 23: 104-122.
 796

797 Hickey, B. M., R.M. Kudela, J.D. Nash, K. W. Bruland, W. T. Peterson, P.
 798 MacCready, E. J. Lessard, D. A. Jay, N. S. Banas, A. M. Baptista, E. P. Dever, P. M.
 799 Kosro, L. K. Kilcher, A. R. Horner-Devine, E. D. Zaron, R. M. McCabe, J. O.
 800 Peterson, P. M. Orton, J. Pan, M. C. Lohan. 2010. River Influences on Shelf
 801 Ecosystems: Introduction and synthesis. *Journal of Geophysical Research* 115:
 802 C00B17. doi: 10.1029/2009JC5452
 803

804 Hickson, R. E. and F. W. Rodolf. 1950. History of Columbia River jetties. In
 805 *Proceedings of the first conference on coastal engineering*, ed. J. W. Johnson, 283-
 806 298.
 807

808 Horner-Devine, Alexander R., David A. Jay, Phillip M. Orton, Emily Y. Spahn.
 809 2009. A conceptual model of the strongly tidal Columbia River plume. *Journal of*
 810 *Marine Systems* 78: 460-475. doi: 10.1016/j.jmarsys.2008.11.025
 811

812 Hu, Chuanmin., Zhiqiang Chen, Tonya D. Clayton, Peter Swarzenski, John C.
 813 Brock, Frank E. Juller Karger. 2004. Assessment of estuarine water-quality
 814 indicators using MODIS medium-resolution bands: Initial results from Tampa
 815 Bay, FL. Remote Sensing of Environment 93: 423-441. doi:
 816 10.1016/j.res.2004.08.007.
 817
 818 Hughes, F. W. and M. Rattray Jr. 1980. Salt flux and mixing in the Columbia
 819 River Estuary. Estuarine and Coastal Marine Science 10: 479-493.
 820
 821 IOCCG, 2006, Remote Sensing of Inherent Optical Properties: Fundamentals,
 822 Tests of Algorithms, and Applications. Lee, Z.-P. (ed.), Reports of the
 823 International Ocean-Colour Coordinating Group, No. 5, IOCCG, Dartmouth,
 824 Canada.
 825
 826 Jay, David A. 1991. Green's Law Revisited: Tidal long-wave propagation in
 827 channels with strong topography. Journal of Geophysical Research 96(C11):
 828 20,585-20,598.
 829
 830 Jay, David A., Benjamin S. Giese, Christopher R. Sherwood. 1990. Energetics and
 831 sedimentary processes in the Columbia River Estuary. Progress in Oceanography
 832 25: 157-174.
 833
 834 Jay, David A. and J. Dungan Smith. 1990a. Circulation, density distribution and
 835 neap-spring transitions in the Columbia River Estuary. Progress in Oceanography
 836 25: 81-112.
 837
 838 Jay, David A. and J. Dungan Smith. 1990b. Residual circulation in shallow
 839 estuaries 1. highly stratified, narrow estuaries. Journal of Geophysical Research
 840 95(C1): 711-731.

841

842 Jay, David A., Jeffery D. Musiak. 1994. Particle Trapping in estuarine tidal flows.

843 Journal of Geophysical Research 99(C10): 20,445-20,461.

844

845 Jay, David A., Jeffery D. Musiak. 1996. Internal tidal asymmetry in channel flows:

846 Origins and consequences. Coastal and Estuarine Studies 50: 211-249.

847

849 Jay, David A., Phillip M. Orton, David J. Kay, Annika Fain, Antonio M. Baptista.

850 1999. Acoustic determination of sediment concentrations, settling velocities,

851 horizontal transports and vertical fluxes in estuaries. Proceedings of the IEEE

852 Sixth Working Conference: 258-263.

853

854 Jay, David A., Phillip M. Orton, Thomas Chisholm, Douglas J. Wilson, Annika M.

855 V. Fain. 2007. Particle trapping in stratified estuaries: application to observations.

856 Estuaries and Coasts 30(6): 1106-1125.

857

858 Jay, D. A., S. A. Talke, A. Hudson, M. Twardowski. 2015. Estuarine turbidity

859 maxima revisited: Instrumental approaches, remote sensing, modeling studies,

860 and new directions. In Developments in Sedimentology: Fluvial-Tidal

861 Sedimentology, eds. Philip J. Ashworth, James L. Best, and Daniel R. Parsons, 49-

862 109: Elsevier.

863

864 de Jonge, V.N., J.E.E. van Beusekom. 1995. Wind and tide induced resuspension

865 of sediment and microphytobenthos from tidal flats in the Ems estuary. Limnology

866 and Oceanography 40: 766–778.

867

868 de Jonge, V.N., Henk M. Schuttelaars, Justus E.E. van Beusekom, Stefan A. Talke,

869 Huib E. de Swart. 2014. The influence of channel deepening on estuarine

870 turbidity levels and dynamics, as exemplified by the Ems estuary. *Estuarine,*
871 *Coastal and Shelf Science* 139: 46-59.

872

873 Kappenberg, Jens and Iris Grabemann. 2001. Variability of the mixing zones and
874 estuarine turbidity maxima in the Elbe and Weser Estuaries. *Estuaries* 5: 699-706.

875

876 Lehner, Susanne, Ivonne Anders, Gerhard Gayer. 2004. High resolution maps of
877 suspended particulate matter concentration in the Geman Bight. *EARSel*
878 *eProceedings* 3: 118-126.

879

880 Lyons, Joseph K. and Robert L. Beschta. 1983. Land use, floods, and channel
881 changes: Upper Middle Fork Willamette River, Oregon (1936-1980). *Water*
882 *Resources Research* 19(2): 463-471.

883

884 Maritorena, Stéphane, David A. Siegel, Alan R. Peterson. 2002. Optimization of
885 semianalytical ocean color model for global-scale applications. *Applied Optics*
886 41(15): 2705-2714.

887

888 May, C. L., J. R. Koseff, L. V. Lucas, J. E. Cloern, D. H. Schoellhamer. 2003. Effects
889 of spatial and temporal variability of turbidity on phytoplankton blooms. *Marine*
890 *Ecology Progress Series* 254: 111-128.

891

892 Meade, Robert H. 1969. Landward transport of bottom sediments in estuaries of
893 the Atlantic coastal plain. *Journal of Sedimentary Research* 39(1): 222-234

894

896 Naik, Pradeep K. and David A. Jay. 2010. Human and climate impacts on
897 Columbia River hydrology and salmonids. *River Research and Applications* 27:
898 1270-1276. doi: 10.1002/rra.1422.

899

901 Naik, Pradeep K. and David A. Jay. 2011. Distinguishing human and climate
902 influences on the Columbia River: Changes in mean flow and sediment transport.
903 *Journal of Hydrology* 404: 259-277. doi: 10.1016/j.jhydrol.2001.04.035.

904

905 Naiman, Robert J., Timothy J. Beechie, Lee E. Benda, Dean R. Berg, Peter A.
906 Bison, Lee H. MacDonald, Matthew D. O'Connor, Patricia L. Olson, E. Ashley
907 Steel. 1992. Fundamental elements of ecologically healthy watersheds in the
908 pacific northwest coastal ecoregion. In *Watershed Management: Balancing*
909 *Sustainability with Environment Change*, ed. R.J. Naiman, 127-188. New York:
910 Springer-Verlag.

911

912 Neukermans, G., H. Liosel, X. Meriaux, R. Astoreca, D. McKee. 2012. In situ
913 variability of mass-specific beam attenuation and backscattering of marine
914 particles with respect to particle size, density, and composition. *Limnology and*
915 *Oceanography* 57(1): 124-144.

916

917 Norman, David K., C. Jeff Cederholm, William S. Lingley Jr. 1998. Flood plains,
918 salmon habitat, and sand and gravel mining. *Washington Geology* 26(2): 1-28.

919

920 Orem, Hollis M. 1968. Discharge in the lower Columbia River basin, 1928-65.
921 U.S. Geological Survey Circular 550.

922

923 Palacios, L. Sherry., Tawnya D. Peterson, Raphael M. Kudela. 2009. Development
924 of synthetic salinity from remote sensing for the Columbia River Plume. *Journal of*
925 *Geophysical Research* 114: C00B05. doi: 10.1029/2008JC004895.

926

927 Postma, H. and K. Kalle. 1955. Die Entstehung von Trübungszonen im Unterlauf
 928 der Flüsse, speziell im Hinblick auf die Verhältnisse in der Unterelbe. Deutsche
 929 Hydrografische Zeitschrift 8(4): 134-144.

930

931 Pruter, A.T. and D.L. Alverson. 1972. The Columbia River Estuary and adjacent
 932 ocean waters. Seattle: University of Washington Press.

933

934 Ralston, David K., W. Rockwell Geyer, John C. Warner. 2012. Bathymetric
 935 controls on sediment transport in the Hudson River estuary: Lateral asymmetry
 936 and frontal trapping. Journal of Geophysical Research 117: C10013. doi:
 937 10.1029/2012JC008124

938

939 Reed, D. J. and J. Donovan. 1994. The character and composition of the Columbia
 940 River estuarine turbidity maximum. In Changes in fluxes in estuaries: Implications
 941 from science to management, eds. K. R. Dyer and Robert J. Orth, 445-450.
 942 Fredensborg, Denmark: Olsen and Olsen.

943

944 Roberts, W. P., J. W. Pierce. 1976. Deposition in Upper Patuxent Estuary,
 945 Maryland, 1968-1969. Estuarine and Coastal Marine Science 4: 267-280.

946

947 Ruhl C. A., D. H. Schoellhamer, R. P. Stumpf, C. L. Lindsay. 2001. Combined use
 948 of remote sensing and continuous monitoring to analyse the variability of
 949 suspended-sediment concentrations in San Fransisco bay, California. Estuarine,
 950 Coastal, and Shelf Science 53: 801-812.

951

952 Schoellhamer, D. H. 2001. Influence of salinity, bottom topography, and tides on
 953 locations of estuarine turbidity maxima in northern San Fransisco Bay. In Coastal
 954 and Estuarine Fine Sediment Transport Processes, ed. McAnally, W.H., and
 955 Mehta, A.J., 343-357. Elsevier Science B.V.

956

957 Sherwood, Christopher R., David A. Jay, R. Bradford Harvey, Peter Hamilton,
 958 Charles A. Simenstad. 1990. Historical changes in the Columbia River Estuary.
 959 Progress in Oceanography 25: 299-352.

960

961 Sherwood, Christopher R. and Joe S. Craeger. 1990. Sedimentary geology of the
 962 Columbia River Estuary. Progress in Oceanography 25: 15-79.

963

964 Simenstad, Charles A., Lawrence F. Small, C. David McIntire. 1990. Consumption
 965 processes and food web structure in the Columbia River Estuary. Progress in
 966 Oceanography 25: 271-297.

967

968 Simenstad, Charles A., Lawrence F. Small, C. David McIntire, David A. Jay,
 969 Christopher Sherwood. 1990. Columbia River Estuary studies: An introduction to
 970 the estuary, a brief history, and prior studies. Progress in Oceanography 25: 1-13.

971

972 Simpson, J. H., J. Brown, J. Matthes, G. Allen. 1990. Tidal straining, density
 973 currents, and stirring in the control of estuarine stratification. Estuaries: 13(2)
 974 125-132.

975

976 Spahn, Emily Y., Alexander R. Horner-Devine, Jonathan D. Nash, David A. Jay,
 977 Levi Kilcher. 2009. Particle resuspension in the Columbia River plume near field.
 978 Journal of Geophysical Research: Oceans 114(C2).

979

980 Stevenson, J. Court, Michael S. Kearney, Edward C. Pendleton. 1985.
 981 Sedimentation and erosion in a Chesapeake Bay brackish marsh system. Marine
 982 Geology 67(3): 213-235.

983

984 de Swart, H. E., H. M. Schuttelaars, S. A. Talke. 2009. Initial growth of
 985 phytoplankton in turbid estuaries: A simple model. *Continental Shelf Research*
 986 29(1): 136-147.
 987
 988 Talke, S.A., H.E. de Swart, and H.M. Schuttelaars, 2008. An analytical model of
 989 the equilibrium distribution of suspended sediment in an estuary. In *Dohmen-*
 990 *Janssen and Hulscher, eds: River, Coastal and Estuarine Morphodynamics 2008*, p.
 991 403-411, London, Taylor and Francis.
 992
 993 Talke, S.A., H.E. de Swart, V.N. de Jonge. 2009a. An idealized model and
 994 systematic process study of oxygen depletion in highly turbid estuaries. *Estuaries*
 995 and Coasts 32(4): 602-620.
 996
 997 Talke, S.A, H.E. De Swart and H.M.Schuttelaars. 2009b. Feedback between
 998 residual circulation and sediment distribution in highly turbid estuaries: an
 999 analytical model. *Continental Shelf Research* 29(1): 119-135. doi:
 1000 10.1016/j.csr.2007.09.002
 1001
 1002 Talke, S.A. and M.T. Stacey. 2003. The influence of oceanic swell on flows over
 1003 an estuarine intertidal mudflat in San Francisco Bay. *Estuarine, Coastal, and Shelf*
 1004 *Science* 58: 541-554.
 1005
 1006 Talke, S.A. and M.T. Stacey. 2008. Suspended Sediment Fluxes at an intertidal
 1007 flat: Shifting Influence of wave, wind, tidal, and freshwater forcing. *Continental*
 1008 *Shelf Research* 28(6): 710-725. doi: 10.1016/j.csr.2007.12.003
 1009

Templeton, William J. and David A. Jay. 2013. Lower Columbia River sand supply and removal: Estimates of two sand budget components. *Journal of Waterway, Port, Coastal, and Ocean Engineering* 139: 383-392.

Theisen, Arthur Albert. 1958. Distribution and characteristics of loess-like soil parent material in Northwestern Oregon. Masters Thesis, Oregon State College.

Uncles, R. J., J. A. Stephens, R. E. Smith. 2002. The dependence of estuarine turbidity on the tidal intrusion length, tidal range and residence time. *Continental Shelf Research* 22: 1835-1856.

Vermote E. F. and A. Vermeulen. 1999. MODIS algorithm technical background document, atmospheric correction algorithm: spectral reflectances (MOD09). NASA.

Warner, John C., W. Rockwell Geyer, James A. Lerczak. 2005. Numerical modeling of an estuary: A comprehensive skill assessment. *Journal of Geophysical Research* 110: C05001. doi: 10.1029/2004JC002691

Wentz, Denis A., Bernadine A. Bonn, Kurt D. Carpenter, Stephen R. Hinkle, Mary L. Janet, Frank A. Rinella, Mark A. Uhrich, Ian R. Waite, Antonius Laenen, Kenneth E. Bencala. 1998. Water quality in the Willamette Basin, Oregon, 1991-95. U.S. Geological Survey Circular 1161.

Figure Captions

Fig. 1 The Columbia River, coastal tributaries (Cowlitz, Willamette, and Sandy), and Columbia River Estuary (inset). Columbia River flows are measured at The Dalles and Beaver Army Terminal. The lower-estuary extends from the mouth (Rkm 0) to Youngs Bay (~Rkm 15), mid-estuary continues up to Tongue Point (~Rkm 30), and the upper-

estuary to the landward limits of salinity intrusion (~Rkm 50). Transects denote the two main channels (North and South) in the system. Stations measuring turbidity are denoted by square icons, those measuring salinity by circles. Grid near mouth of Columbia River represents a 10x10 MODIS Grid at 250m resolution.

Fig. 2 Annual hydrograph measured at Beaver Army Terminal, (the sum of all river flow entering the estuary; solid line), The Dalles (dashed line), and coastal tributaries (Willamette, and Cowlitz rivers; dash-dotted line), monthly averaged during the study period in daily intervals. Maximum river flows in the Columbia River (Dalles) are observed May-June during the spring freshet; minimum flows occur during late summer. Coastal rivers peak during winter.

Fig. 3 Approximate depth profiles along the North (a) and South (b) Channel of the CRE, relative to MLLW. The South Channel is maintained at a minimum depth of 13m and both channels exhibit several topographic lows.

Fig. 4 Measurement record for turbidity sensors used in this study

Fig. 5 Estimates of MODIS-based turbidity vs. in situ measurements

Fig. 6 Example turbidity distributions in the CRE derived from MODIS-based surface reflectance. The turbidity is a function of both river flow and tidal range. The left panels are measured at a time of moderate tidal ranges (~2.6m) during low (a) and high (c) flow rates. Right panels illustrate neap (1.7m tidal range) and spring (3.5m tidal range; b and d, respectively) conditions at a time of low river flow. The two plots at bottom show prominent ETM

Fig. 7 Monthly averaged turbidity distributions. Elevated concentrations begin during the winter months and persist throughout the spring while high river flows are maintained. Minimum concentrations occur in late summer at periods of low flow. Each plot is the average of at least 65 MODIS images

1070

1071 **Fig. 8** Correlation map between MODIS derived turbidity and wind-speed during
1072 summer (a) and winter (b). The effect of wind speed is confined to shallow locations and
1073 is greatest in the lower estuary bays. Summer time winds affect the southern bays,
1074 whereas wintertime winds mostly affect Baker Bay in the northern estuary near the
1075 mouth.

1076

1077 **Fig. 9** Correlation map between MODIS derived turbidity and ocean swell wave energy.
1078 Waves only affect turbidity near the mouth of the CRE

1079

1080 **Fig. 10** Correlation map between MODIS derived turbidity and tidal range. Tidal range is
1081 positively correlated with surface turbidity throughout the system

1082

1083 **Fig. 11** Correlation map between MODIS derived turbidity and river flow measured at
1084 Beaver Army Terminal, The Dalles, Willamette River, and Cowlitz River

1085

1086 **Fig. 12** Longitudinal transects of turbidity (solid lines) in the North (left) and South
1087 (right) Channels. Transects represent median values, binned according to greater diurnal
1088 tidal ranges (bottom; 2.5m and 3m with 0.15m window) and river flow (top; $3000\text{m}^3\text{s}^{-1}$
1089 and $11000\text{m}^3\text{s}^{-1}$ with $500\text{m}^3\text{s}^{-1}$ window). Shaded regions denote 33rd and 67th quantiles of
1090 binned transects. Near surface and near bed salinity intrusion (dotted and dashed
1091 lines, respectively) and 2 psu contours (squares and circles, respectively) have paired
1092 response with turbidity distribution to forcing data.

1093

1094 **Fig. 13** The location (X_{max}) and magnitude (C_{max}) of the ETM in the North (left) and
1095 South (right) Channel plotted in the river flow and tidal range parameter space.
1096 Increasing river flow and tidal range promote seaward advection and greater intensity of
1097 the ETM

1098

1099 **Fig. 14** Daily averaged stratification in the North (top) and South (bottom) Channel. At
1100 Rkm 16 stratification in the North Channel decreases with tidal range (right) and river

1101 flow (left). At Rkm 28 stratification in the South Channel decreases with freshwater
1102 discharge but depends little on tidal range
1103

1104 Tables

1105 **Table 1. Locations of in situ turbidity measurements used in OLS**

Buoy	Measurement Depth (meters)	Latitude (degrees)	Longitude (degrees)
Saturn 01	^a	46.235	-123.872
Saturn 02	0-1	46.173	-124.127
Saturn 05	2.5	46.184	-123.188
Saturn 07	1	46.287	-124.016
Saturn 09	0.5	46.177	-123.869

1106 ^aMeasurements taken as an average over top 2 meters of water column

1107 **Table 2. Locations of in situ measurements used to derive salinity intrusion and**
 1108 **stratification.**

Buoy	Measurement Depth (meters)	Latitude (degrees)	Longitude (degrees)	Distance from mouth (km)
SANDS	7.9	46.256	-123.982	7
DESD	7.3	46.226	-123.955	11
RED26	3.3	46.208	-123.951	12
TANSY	8.4	46.189	-123.919	15
Saturn 01	7.4	46.235	-123.872	16
MBS	2.6	46.196	-123.852	20
COAOF	2.1 or 3.2	46.205	-123.774	29
Saturn 04	8.6	46.204	-123.759	30
CBNC	6.5	46.210	-123.714	34

1109
 1110 **Table 3. Measurement details of forcing parameters used in the analysis.**

Forcing	Measurement Location
Wind Speed/Direction (ms ⁻¹ /degrees)	NCDC (NOAA) Station 94224. Astoria Airport
Wave Energy (m ² Hz ⁻¹)	NBDC (NOAA) Station 46029. Columbia River Bar
Tidal Range/Elevation (m)	NOAA Station 9439040. Astoria, OR
River Flow (m ³ s ⁻¹)	USGS 14246900. Columbia River. Beaver Army Terminal USGS 14243000. Cowlitz River. Castle Rock, WA USGS 14211720. Willamette River. Portland, OR USGS 14105700. Columbia River. The Dalles, OR

1111
 1112 **Table 4. Model statistics for OLS between Band1 and in situ turbidity and coefficient**
 1113 **estimates. Confidence intervals ($\alpha = 0.05$) for coefficients are provided in brackets.**

MODEL STATISTICS			
Sample size	R ²	P _{value}	Error variance
205	0.85	0	1.25 ntu
COEFFICIENT ESTIMATES			
Intercept		Slope	
0.59 [0.38; 0.80]		137 [129; 145]	

1114
 1115 **Table 5. Turbidity gradient length scales downstream and upstream of the ETM in**
 1116 **the North and South Channels during high river flows (Figure 12a and b). ¹ Length**
 1117 **scales are estimated by fitting hyperbolic tangent curve to turbidity upstream and**
 1118 **downstream of the ETM.**

	Downstream of ETM	Upstream of ETM
North Channel	11.6km	42.3km
South Channel	11.8km	55.7km

1119

1120 ¹. Length scales are estimated by fitting hyperbolic tangent curve to turbidity upstream and
1121 downstream of the ETM.